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ADVANCES IN ORDOVICIAN GEOLOGY

Edited by
Christopher R. Barnes and S. Henry Williams

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Frontispiece. *Proximal facies of Cow Head Group, Cow Head Peninsula, western Newfoundland; contact between beds 6 and 7, the latter bearing large clasts of white biohermal (Epiphyton) limestone near base of Ordovician (see paper by B.S. Norford).*

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Cover Description

Proposed Cambrian-Ordovician boundary stratotype, Cow Head Group, Green Point, western Newfoundland; proposed base of Ordovician is at level of third person from right. Section is overturned and represents distal toe-of-slope environment on continental margin (see paper by B.S. Norford).

Camera-ready figures provided by authors

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Advances in Ordovician Geology

Christopher R. Barnes[1] and S. Henry Williams[2]

INTRODUCTION

The Ordovician Period is one of the longest and most complex in the Phanerozoic. In order to review current research and develop strategies for future studies and international collaboration, the Fifth International Symposium on the Ordovician System (VISOS) was held at Memorial University of Newfoundland in St. John's in August, 1988. The present volume contains selected papers from the symposium, and is one of a series of benchmark volumes on Ordovician geology that have been produced from the five symposia held to date.

The Ordovician Period has many distinctive characteristics, having been a time of major eustatic sea level changes, prominent anoxic conditions, widespread development of carbonate platforms on different cratons, marked faunal provincialism, and an overall greenhouse climatic state that reversed to an icehouse state at the end of the period. The Ordovician biota marks the change from the arthropod-dominated faunas of the Cambrian to more abundant and diverse faunas and floras, which developed in more complex community patterns through a wider range of habitats. The actual physical and chemical controls and thresholds that were dominant during this period are poorly understood and, unfortunately, the chronostratigraphic framework that is essential for global correlation of events is inadequately defined.

The precise recognition and definition of time in stratigraphy is a continuing priority in geological research. This has been accentuated in recent years by the focus toward short-term geological and biological events in the stratigraphic record. A principal thrust of the symposium was therefore on chronostratigraphy, with particular reference to the definition of the series within the Ordovician. The program of publication of Ordovician correlation

INTRODUCTION

L'Ordovicien est une des périodes les plus longues et les plus complexes du Phanérozoïque. Le cinquième symposium international sur le Système ordovicien [Fifth International Symposium on the Ordovician System (VISOS)] a eu lieu à l'Université Memorial à St. John's (Terre-Neuve), en août 1988; on y a passé en revue les recherches en cours et élaboré des stratégies en matière d'études futures et de collaboration internationale. Le présent volume se veut un recueil de certains rapports qui ont été présentés au cours du symposium; il fait partie d'une série de volumes sur la géologie de l'Ordovicien qui sont le résultat des cinq symposiums tenus à ce jour.

Les caractéristiques de l'Ordovicien sont nombreuses : fluctuations eustatiques majeures, conditions principalement anoxiques, construction étendue de plates-formes carbonatées sur divers cratons, provincialisme marqué de la faune et, dans l'ensemble, climat de serre qui s'est transformé en climat de glacière à la fin de la période. L'Ordovicien marque le passage des faunes cambriennes à prédominance d'arthropodes aux faunes et aux flores plus abondantes et plus variées qui ont produit des communautés plus complexes dans une vaste gamme d'habitats. Nous connaissons mal les seuils et contrôles physiques et chimiques qui ont dominé à l'Ordovicien; en outre, le cadre chronostratigraphique qui est essentiel à la corrélation globale des événements n'est pas suffisamment bien défini.

Les travaux stratigraphiques continus d'être axés sur la reconnaissance et la définition exacte du temps. En outre, depuis quelques années, les chercheurs se penchent sur des événements géologiques et biologiques de courte durée dans le profil stratigraphique. Par conséquent, le symposium a porté principalement sur la chronostratigraphie, notamment sur la définition des séries de l'Ordovicien. La publication de tableaux de corrélation de l'Ordovicien, selon le programme établi par le sous-comité de la stratigraphie de l'Ordovicien, est en grande partie terminée.

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charts initiated by the Subcommission on Ordovician Stratigraphy is largely complete and has enabled groups in many countries to redefine regional chronostratigraphic subdivisions. In this volume, authors from the United Kingdom, Canada, Australia, and the United States contribute to this process with the aim of achieving an internationally accepted internal chronostratigraphic subdivision for the Ordovician System by the mid-1990's. The development of refined geochronological methods is particularly noteworthy and illustrates the potential for achieving an integrated, high precision chronostratigraphy and geochronology for the Ordovician in the near future. This, in turn, will enable new interpretations of the dynamic systems operating during the Ordovician, such as those emerging from Mesozoic and Cenozoic research programs.

A section of this volume is devoted to recent advances in Ordovician paleontology and biostratigraphy. Emphasis is given to conodonts, graptolites and trilobites, which have been of greatest value to date in biostratigraphic studies. There is particular interest by several authors in the response of faunas to the Late Ordovician glaciation event. The effects of severe faunal provinciality during the entire Ordovocian is documented in some studies.

The interrelationships of stratigraphy, eustasy, tectonics and volcanism are explored in another section of this volume. Particular attention is paid to the Appalachian Orogen, for which there has been a rapid evolution of ideas for tectonic and event-stratigraphic models. Valuable new data are also reported from the less well known regions of the Argentine Precordillera, and from Kazakhstan and Tien Shan in the U.S.S.R.

The final section of the volume documents some of the new data and interpretations of Ordovician paleo- oceanography, paleoclimatology and paleogeography. A new paleogeographic reconstruction for each series of the Ordovician System is provided, as well as interpretations of the changing patterns of Ordovician oceanographic circulation. The application of isotope geochemistry to Ordovician problems is still in its infancy, but some current work related to the glacial event is reported and many different studies are now in progress.

Future research on the long and complex Ordovician Period must focus more on the particular physical, chemical and biological controls and thresholds that were operative. Intensive studies and new techniques are needed to resolve these problems and to construct more precise models of Ordovician paleogeography, paleoclimatology, and paleooceanography. The present volume is not intended to be summary volume for these issues, but to report primary search on various facets. We hope that this volume will assist in the advance of knowledge of Ordovician geology and provide a foundation for the scientific studies to be reported at the Sixth International Symposium on the Ordovician System, University of Sydney, Australia, in July, 1991.

Grâce à ces tableaux, des groupes dans de nombreux pays ont pu redéfinir les subdivisions chronostratigraphiques régionales. Dans le présent volume, des chercheurs provenant du Royaume-Uni, du Canada, de l'Australie et des États-Unis contribuent à ce processus en vue de l'établissement, d'ici 1995, d'une subdivision chronostratigraphique interne de l'Ordovicien qui aura un statut international. La mise au point de méthodes géochronologiques perfectionnées est digne de mention et montre la possibilité d'établir, dans un proche avenir, une chronostratigraphie et une géochronologie intégrées, hautement précises, de l'Ordovicien. Cela donnera lieu à de nouvelles interprétations des systèmes dynamiques qui ont existé à l'Ordovicien, comme ceux qui émergent des programmes de recherche sur le Mésozoïque et le Cénozoïque.

Une section du volume porte sur les réalisations en paléontologie et en biostratigraphie de l'Ordovicien. On met l'accent sur les conodontes, les graptolites et les trilobites qui, à ce jour, sont les organismes les plus utiles pour les travaux biostratigraphiques. Plusieurs chercheurs s'intéressent particulièrement à la ráection des faunes à l'événement glaciaire de l'Ordovicien supérieur. Certaines études renseignent sur les effets de la provincialité marquée de la faune tout au long de l'Ordovicien.

Une autre section du volume examine les liens qui existent entre la stratigraphie, l'eustasie, la tectonique et le volcanisme. On accorde une attention particulière à l'orogène des Appalaches, pour lequel il y a eu évolution rapide des modèles tectoniques et des modèles de la stratigraphie des événements. On y présente aussi de nouvelles données fort utiles qui proviennent des régions moins bien connues de la Précordillère en Argentine et du Kazakhstan et du T'ien-Shan en URSS.

La dernière section du volume renseigue dur certaines nouvelles données sur la paléo-océanographie, la paléoclimatologie et la paléogéographie de l'Ordovicien, et les nouvelles interprétations qui en découlent. Il présente une nouvelle reconstitution paléogéographique pour chaque série de l'Ordovicien, de même que des interprétations des changements survenus dans la circulation océanique à l'Ordovicien. Bien que l'application de la géochimie des isotopes aux problèmes ordoviciens en soit encore à ses débuts, le volume présente des comptes rendus de certains travaux en cours qui portent sur l'événement glaciaire; en outre, plusieurs autres études différentes ont été commencées.

À l'avenir, les recherches sur l'Ordovicien devront porter davantage sur les seuils et les contrôles physiques, chimiques et biologiques qui ont existé au cours de cette période longue et complexe. Il faudra entreprendre des études approfondies et élaborer de nouvelles techniques afin de résoudre ces problèmes et d'établir des modèles plus exacts de la paléogéographie, de la paléoclimatologie et de la paléo-océanographie de l'Ordovicien. Le volume ne se veut pas un simple résumé de ces questions, mais il cherche plutôt à présenter les recherches de base qui portent sur leurs divers aspects. Nous espérons qu'il contribuera à élargir nos connaissances de la géologie de l'Ordovicien et qu'il servira de fondement pour les études scientifiques qui seront présentées à l'Université de Sydney, en Australie, en juillet 1991, à l'occasion du sixième symposium international sur le Système ordovicien.

Finally, we gratefully acknowledge the sponsors of the Fifth International Symposium on the Ordovician System: the I.U.G.S. Subcommission on Ordovician Stratigraphy, the International Geological Correlation Project 216 (Global Bioevents in Earth History), the Natural Sciences and Engineering Research Council of Canada, the Geological Survey of Canada, the Newfoundland Department of Mines and Energy, and Memorial University of Newfoundland. Financial support or support-in-kind was contributed by these sponsors to the symposium, the field trips, the program and guidebooks, and to this Geological Survey of Canada Paper. We particularly thank many colleagues in the Centre for Earth Resources Research and the Department of Earth Sciences at Memorial University of Newfoundland for their organizational and technical assistance with the symposium. Copies of the program and abstracts volume and of the three separate field guidebooks may be purchased from the department.

Enfin, nous tenons à remercier les promoteurs du cinquième symposium international sur le Système ordovicien : la Subcommission on Ordovician Stratigraphy de l'Union internationale des sciences géologiques, le International Geological Correlation Project 216 (Global Events in Earth History), le Conseil de recherches en sciences naturelles et en génie du Canada, la Commission géologique du Canada, le ministère des Mines et de l'Énergie de Terre-Neuve et l'Université Memorial de Terre-Neuve. Ces organismes ont fourni un soutien financier ou autre pour le symposium, les excursions, le programme, les guides et cette étude de la Commission géologique du Canada. Nous tenons à remercier tout particulièrement nos nombreux collègues au Centre for Earth Resources Research et au Département des sciences de la Terre de l'Université Memorial de Terre-Neuve, qui ont contribué à l'organisation et à l'aspect technique du symposium. On peut se procurer des exemplaires du programme, du volume des résumés et des trois guides auprès du Département des sciences de la Terre.

Progress and problems in the selection of stratotypes for the bases of series in the Ordovician System of the historical type area in the U.K.

R.A. Fortey[1], M.G. Bassett, D.A.T. Harper, R.A. Hughes, J.K. Ingham, S.G. Molyneux, A.W. Owen, R.M. Owens, A.W.A. Rushton, and P.R. Sheldon

Fortey, R.A., Bassett, M.G., Harper, D.A.T., Hughes, R.A., Ingham, J.K., Molyneux, S.G., Owen, A.W., Owens, R.M., Rushton, A.W.A., and Sheldon, P.R., Progress and problems in the selection of stratotypes for the bases of series in the Ordovician System of the historical type area in the U.K.; in Advances in Ordovician Geology, C.R. Barnes and S.H. Williams (ed.); Geological Survey of Canada, Paper 90-9, p. 5-25, 1991.

Abstract

The series of the Ordovician System in the Anglo-Welsh area have been a standard reference in Ordovician chronostratigraphy for many years, and should remain so in the interest of a common language for international use. Their original definition was not in accord with modern stratigraphic precision, and to this end it is necessary to redefine bases for the series within continuous sequences. Such sequences have been identified for the bases of all the series. Even though there are breaks in the succession locally or in the original type areas, these shortcomings are minimized by considering these other sections in the Anglo-Welsh area as potential stratotypes. The Anglo-Welsh faunas include widespread taxa virtually throughout the Ordovician, which permit international correlation; the monographic coverage of these faunas compares with that from anywhere in the world. It is thought to be possible to modify the horizon at which series boundaries were drawn originally in the interests of international correlation, as long as nomenclature is conserved. The possible stratotypes for bases of the series of the Ordovician System in the type area are reviewed briefly, and new faunal logs are given where necessary. The base of the Tremadoc is taken at an horizon coincident with the appearance of nematophorous graptolites, the only practicable horizon in the type area. The base of the Arenig poses particular problems because of the poorly fossiliferous nature of available sections. However, newly discovered sections in the Lake District indicate that there may have been no major break in sedimentation at the top of the Tremadoc. A fossiliferous mixed graptolite-trilobite section is available for the base of the Llanvirn. Although there are candidates for the base of the Llandeilo (current usage), the comparative brevity of this series suggests that it may be better united with the Llanvirn to form a new, combined series commensurate in extent with the other series of the Ordovician. The name Llandeilo could be retained for this new concept. The Llanvirn could also be retained as the earlier stage within it, and a new stage name, for which the Dyneforian is suggested, proposed for the later part, equivalent to some or all of the traditional Llandeilo; just how much will depend on the definition of the base of the Caradoc, for which three possibilities are proposed as a basis for discussion. The base of the Ashgill has been located precisely in a section in the Murthwaite Inlier, Cautley, which is proposed as a suitable international type section.

Résumé

Les séries du Système ordovicien qui se trouvent dans la région anglo-galloise servent d'étalon pour la chronostratigraphie ordovicienne depuis de nombreuses années et, par souci d'un langage commun international, il ne faudrait pas les remplacer. Puisque leur définition d'origine n'avait pas la précision

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[1] See end of report for Authors' addresses

stratigraphique d'aujourd'hui, il devient nécessaire de redéfinir les bases des séries qui se situent dans des séquences continues. Or, les bases de toutes les séries se trouvent dans de telles séquences. Bien que la succession soit interrompue localement ou dans les régions types d'origine, on peut minimiser ces problèmes en prenant comme stratotypes éventuels les autres coupes de la région anglo-galloise. Presque tout au long de l'Ordovicien, les faunes de cette région comprennent des taxons dont la répartition géographique est vaste, ce qui permet leur corrélation mondiale; leur couverture monographique se compare à celle de n'importe quel endroit au monde. Il serait vraisemblablement possible, à des fins de corrélation mondiale, de modifier l'horizon auquel avaient été fixées à l'origine les limites des séries, dans la mesure où l'on conserve la nomenclature. Le rapport donne un aperçu des stratotypes éventuels pour les bases des séries du Système ordovicien dans la région type et, au besoin, il présente de nouveaux enregistrements fauniques. La base du Trémadocien est fixée à un horizon qui coïncide avec l'apparition de graptolites nématophores, soit le seul horizon praticable dans la région type. La base de l'Arénigien pose des problèmes particuliers en raison de la nature peu fossilifère des coupes disponibles. Or, des coupes nouvellement découvertes dans le «Lake District» indiquent qu'il pourrait n'y avoir eu aucune interruption majeure de la sédimentation au sommet du Trémadocien. Il existe une coupe fossilifère à graptolites et à trilobites pour la base du Llanvirnien. Bien qu'il y ait des possibilités pour la base du Llandeilien (usage courant), cette série couvre un intervalle de temps relativement court, et il faudrait peut-être la réunir au Llanvirnien afin de former une nouvelle série combinée dont l'étendue serait comparable à celle des autres séries de l'Ordovicien. On retiendrait le nom «Llandeilien» pour cette nouvelle série. En outre, le nom Llanvirnien désignerait l'étage plus ancien qui s'y trouve et un nouveau nom, par exemple le Dynéforien, serait donné à la partie plus récente, qui équivaut à l'ensemble ou à une partie du Llandeilien classique, selon la définition de la base du Caradocien, pour laquelle on propose trois possibilités à des fins de discussion. La base de l'Ashgillien est située avec précision dans une coupe dans la fenêtre de Murthwaite, à Cautley, que l'on propose comme coupe type mondiale.

INTRODUCTION

The series of the Ordovician System, as defined originally in England and Wales, have been used for many years as a standard reference in Ordovician chronostratigraphy. These series have provided a common language that stratigraphers and paleontologists can use to discuss international correlation. The more recent emergence of event stratigraphy has also required a common calibration for international use. Nonetheless, the scope of the series has been employed somewhat loosely outside the historical type area, and it is a proper objective of the Subcommission on Ordovician Stratigraphy to urge greater precision. Following the Fourth International Symposium on the Ordovician System in Oslo, a group of British workers (Whittington et al., 1984) attempted to summarize the historical background to the proposition of the series, and to encapsulate the problems involved in obtaining a more refined definition of their bases.

There were numerous problems, and many of these remain, but a certain measure of progress has been made over the last few years that is worth summarizing. In this paper it is assumed that the Tremadoc Series is included within the Ordovician System. The earlier literature was reviewed in Williams et al. (1972) and Whittington et al. (1984). The introductory sections of this paper were drafted by R.A. Fortey, with the approval of the Stratigraphy Committee of the Geological Society of London. The authorship of the accounts of possible stratotype sections is indicated where appropriate.

THE BRITISH AREA REMAINS APPROPRIATE FOR DEFINITION OF STANDARD SERIES

Although the British Isles has a historical claim to precedence as the type area of the Ordovician System — and this is not to be disregarded — there are other reasons why the Anglo-Welsh basin is a suitable region in which to define the type sections of the series. The limitations of the type area are rather obvious. Structural complexities are common. Exposure of long, continuous sequences is not usual. Clastic rock types predominate, which makes the recovery of conodonts in some parts of the sequence difficult. These limitations should not be allowed to obscure several positive aspects, which may not be obvious to those accustomed to working in platform carbonate successions. These aspects are:

1. The Anglo-Welsh area occupied a marginal site through the Ordovician Period; platform endemic faunas are juxtaposed closely with deeper water faunas, which include species with wide distribution and international potential for correlation. At the base of the Tremadoc, for example, several of the deep water trilobite species have a distribution beyond the confines of western Gondwana. For most parts of the column, graptolitic and shelly facies are present in nearby areas, even if the different kinds of facies faunas do not co-occur in the same section.
2. Sedimentation was generally rapid, and thick sequences are present across many of the critical intervals. It should be possible, therefore, to define stratotypes in sections free

of hiatuses, such as may be present in cratonic sequences. For example, marginal sites may escape the effects of eustatic cycles — or local tectonic effects may serve to offset their influence. The paradox is that platformal sites may afford a "clearer" boundary, precisely because new taxa appear there simultaneously, but the arbitrary nature of a boundary (drawn at the first appearance of a particular species, for example) in a thicker, fuller succession is to be seen as an advantage, not as a disadvantage.

3. The "provincial" affinities of the Anglo-Welsh shelly faunas in toto change through the Ordovician, from Gondwanan to Baltic/North American — although even in the Caradoc Series some Gondwana-type trilobites such as *Kloucekia* remained part of the fauna. Whether this is because of an independent tectonic history of the Avalon microcontinent, or a general convergence of the Gondwanan with the Baltic and North American continents, is immaterial in this context. What does matter is that the British Ordovician encapsulates some of the significant changes that occurred during Ordovician times. No area in the world has sections that allow for correlation *everywhere* throughout the Ordovician, but the type area is as representative as any.
4. Fossil faunas from the type Ordovician are relatively well known, and are still being studied actively. For example, since the publication of the previous review of Ordovician Series (Whittington et al., 1984), substantial additions to our knowledge of the ostracodes (Jones, 1986, 1987), acritarchs (Turner, 1985; Molyneux, 1987), graptolites (Hughes, 1989), the Arenig Series faunas and stratigraphy (Fortey and Owens, 1987), and conodonts both across series boundaries and within some of the series (Savage and Bassett, 1986; Bergström et al., 1987) have been published. All these have proved to be useful stratigraphically. Few other areas can boast such intensive coverage, especially now when systematic studies are not generally fashionable.

These four arguments are scientific, and separate from the undeniable historical claims of the Anglo-Welsh area as the location of the type sections. The series of the Ordovician System provide the common language for chronostratigraphy, and it is a practical common language.

CONDITIONS FOR SUITABLE STRATOTYPE SECTIONS

The historical problem is that the series were defined originally in a patchwork but by no means random fashion, taking their names from areas scattered over Wales and northern England. Boundaries were drawn at unconformities, or at least where radical faunal changes appeared (Whittington et al., 1984). Modern stratigraphic usage demands that these boundaries be moved to continuous sequences. This should present no formal problem, as it is obvious that the Anglo-Welsh area *as a whole* constitutes the type area, in which the various series names are distributed. We are justified, therefore, in selecting sections, possibly at some distance from the historical type areas, that show the best continuous sequences, and which are of service in international correlation. Partly because of the marginal and tectonically unstable setting of the type area, it is possible to discover such sections for all the series boundaries with which we are concerned. The following criteria should apply:

1. Sections should be within the Anglo-Welsh area, and south of the Solway Line. This is because the Solway Line is generally accepted as marking the Iapetus suture (e.g., Fortey et al., 1989; McKerrow and Soper, 1989) and separates (broadly speaking) the Ordovician Gondwana and Laurentia paleocontinents. There is no doubt that the Ordovician series as defined originally belonged within the former Gondwana geographic entity.
2. Only the base of a series need be defined. What lies below that base belongs in the underlying series *by definition.* This avoids apparent problems such as the so-called "unclassified intervals"; one of these was illustrated recently by Bergström et al. (1987, Fig. 18.8) as lying between the Llanvirn and Llandeilo. If the base of the Llandeilo is defined formally, then the interval below is simply part of the upper Llanvirn.
3. The horizon at which the base was originally defined need not be rigidly adhered to, especially if it is acceptable to move type sections for the bases of series from the original areas of definition. Because these bases were often described at the bottom of a lithostratigraphic unit (e.g., Llandeilo) or at an unconformity (e.g., Caradoc), in accordance with nineteenth century practice, these horizons may themselves pose considerable problems in correlation. There is no reason why such horizons may not be adjusted upward or downward *in the interests of usefulness for international correlation* (Fortey, 1988). There is already some precedent for this in the inclusion of the Pusgillian Stage within the Ashgill (Ingham and Wright, 1970). By adopting this flexible approach, we believe that the horizon eventually chosen can be one that is capable of being recognized in each of the main paleogeographic regions. If a similarly flexible approach is applied also to chronostratigraphic divisions in other regions, such as North America, there is a chance that the standard series may be understood readily in the same way internationally (Fig. 8). We therefore discuss more than one possible horizon in the accounts that follow; which one is chosen will depend on international application, following further study and discussion by the Subcommission.

POSSIBLE STRATOTYPES FOR THE BASE OF EACH SERIES

There now follows a chronological account of sections that we believe to be suitable for definition as international stratotypes for each series within the Ordovician System. The information currently available varies from one section to another; some have only been investigated recently. A preliminary account is useful at this point, however, in order to focus attention on the problems with the horizon chosen, and the limitations of the sections themselves. A brief account of the history of the series is given in Whittington et al. (1984).

We believe that each of the sections discussed below has the potential to meet the criteria recommended for the establishment of Global Standard Stratotypes as set out by Cowie et al. (1986).

TREMADOC (A.W.A.R.)

The Tremadoc Series takes its name from a small town in North Wales (now spelled "Tremadog") and the stratigraphic concept originated in the faunal characteristics of the Tremadoc Slates of the area. The history of the use of "Tremadoc" is given in Henningsmoen (1973) and in Whittington et al. (1984).

In Britain there are few sections that expose a sedimentary succession across the base of the Tremadoc Series (Stubblefield, 1956), and of these the most satisfactory in which to define a base to the series is on the forestry road at Bryn-llin-fawr, about 25 km east-southeast of Tremadog. This section was described in detail by Rushton (1982), with further information in Allen et al. (1981) and Allen and Jackson (1985). Figure 1 summarizes the rock types and the distribution of fossils, and indicates the litho-, bio-, and chronostratigraphy of the sequence.

The section is easily accessible and even though it is an artificial exposure, of which parts will require cleaning from time to time, the boundary itself is exposed permanently. The rock type is fairly uniform and although there is a regional cleavage, this is slight in the area of Bryn-llin-fawr. Shelly fossils, especially trilobites, occur at regular intervals through the section, and range from the *Acerocare* Biozone (>20 m thickness exposed) to the *R. flabelliformis* Biozone, *socialis* Subzone (estimated to be >40 m thick). Graptolites are found at and above the proposed base of the Tremadoc. Conodonts are rare and lack variety at Bryn-llin-fawr. There is a permanent confirmatory section in the stream at Dol-cyn-afon, 2 km south of Bryn-llin-fawr, but this has not been studied thoroughly. In North Wales we cannot readily recognize the equivalents of the base of the *Cordylodus proavus* conodont Biozone, which would lie in the Upper Cambrian sequence, and there is no justification for moving the base of the Tremadoc from the horizon at the appearance of *Rhabdinopora.*

Biostratigraphy and correlation

As well as the widely distributed graptoloid *Rhabdinopora* and the trilobite *Boeckaspis hirsuta*, many of the trilobites from Bryn-llin-fawr are of considerable use in correlation outside Wales. Since the description of the section in 1982 it has been possible to refer several trilobite taxa to species described from China. *Parabolinella*? sp. of the *Acerocare* Biozone can be referred to *P. contracta* Lu and Zhou (Lu et al., 1981). This species has been described under various names from several sections in China, as discussed by Rushton (1988); everywhere it occurs at a horizon near the base of the Tremadoc. Some of the *Niobella* from the *Acerocare* Biozone appear to be referable to the finely drawn species *N. preciosa* Lu and Zhou (treated as a subspecies of *N. homfrayi* in Allen and Jackson, 1985). This form occurs with *P. contracta* in Nei Monggol, China (Lu et al., 1981). Despite search, no further conodonts have been found apart from a cluster of *Phalekodus* ["*Prooneotodus*"] *tenuis* (Higgins and Austin, 1985, Pl. 2.1, figs. 1, 7).

Oslo region

The section at Naersnes, Oslo, described by Bruton et al. (1982) has some trilobite species in common with Bryn-llin-fawr. Their lowest records of *Parabolina acanthura* (Bruton et al., 1982, Pl. 1, figs. 1-3, 17, 19; 1988), all from concretion N1, are referable to *P. heres heres* (Nikolaisen and Henningsmoen, 1985, p. 4), which thus appears below *Rhabdinopora* and *Boeckaspis hirsuta*, as at Bryn-llin-fawr. Our criteria would place the base of the Tremadoc 0.5 m above the base of the Ordovician proposed by Bruton et al. (1982).

Western Newfoundland

The base of the Tremadoc at Bryn-llin-fawr is thought to correspond to the base recognized by Fortey et al. (1982), although there are no conodont or trilobite taxa to support the graptolite evidence. However, Fortey et al. (1982) showed that the *Missisquoia typicalis* Biozone underlies the base of the Tremadoc, and similarly Rushton (1982) showed that associates of *M. typicalis* at its type locality, namely *Plicatolina kindlei* and *Neoagnostus bilobus*, underlie the base of the Tremadoc in North Wales.

Platform carbonate sequences

Correlation of the base of the Tremadoc into platform carbonate sequences is indirect, and relies on sections such as those at Broom Point, Newfoundland, and Dayangcha, China, where there are "mixed" faunas. The appearance of the conodont *Cordylodus lindstromi* equates closely with, but apparently slightly precedes, the appearance of nematophorous graptolites (Landing et al., 1986). Nonetheless, this provides a workable approximation in at least the more marginal North American sequences to the base of the Tremadoc Series in its type area. Other propositions for the base of the Ordovician — for example, the base of the *Hirsutodontus simplex* Subzone of the *C. intermedius* Biozone (Miller, 1984, 1987), or the base of the *C. proavus* Biozone (Kazakhstan: Chugaeva and Apollonov, 1982; China: Zhou et al., 1984) are substantially older, and would not provide a ready correlation into North Wales.

ARENIG (R.A.F., S.G.M., R.M.O., A.W.A.R.)

The type area for the Arenig Series is at the mountain called Arennig Fawr in North Wales, where the succession has been redescribed recently by Zalasiewicz (1984; see also Beckly, 1987). Fortey and Owens (1987) recently reviewed the biostratigraphy of the Arenig Series in Britain and showed that its development is commensurate with that in other parts of the world, and that it is a major division of the Ordovician System. Three stages for the Arenig were proposed: in ascending order, Moridunian, Whitlandian, and Fennian. It was

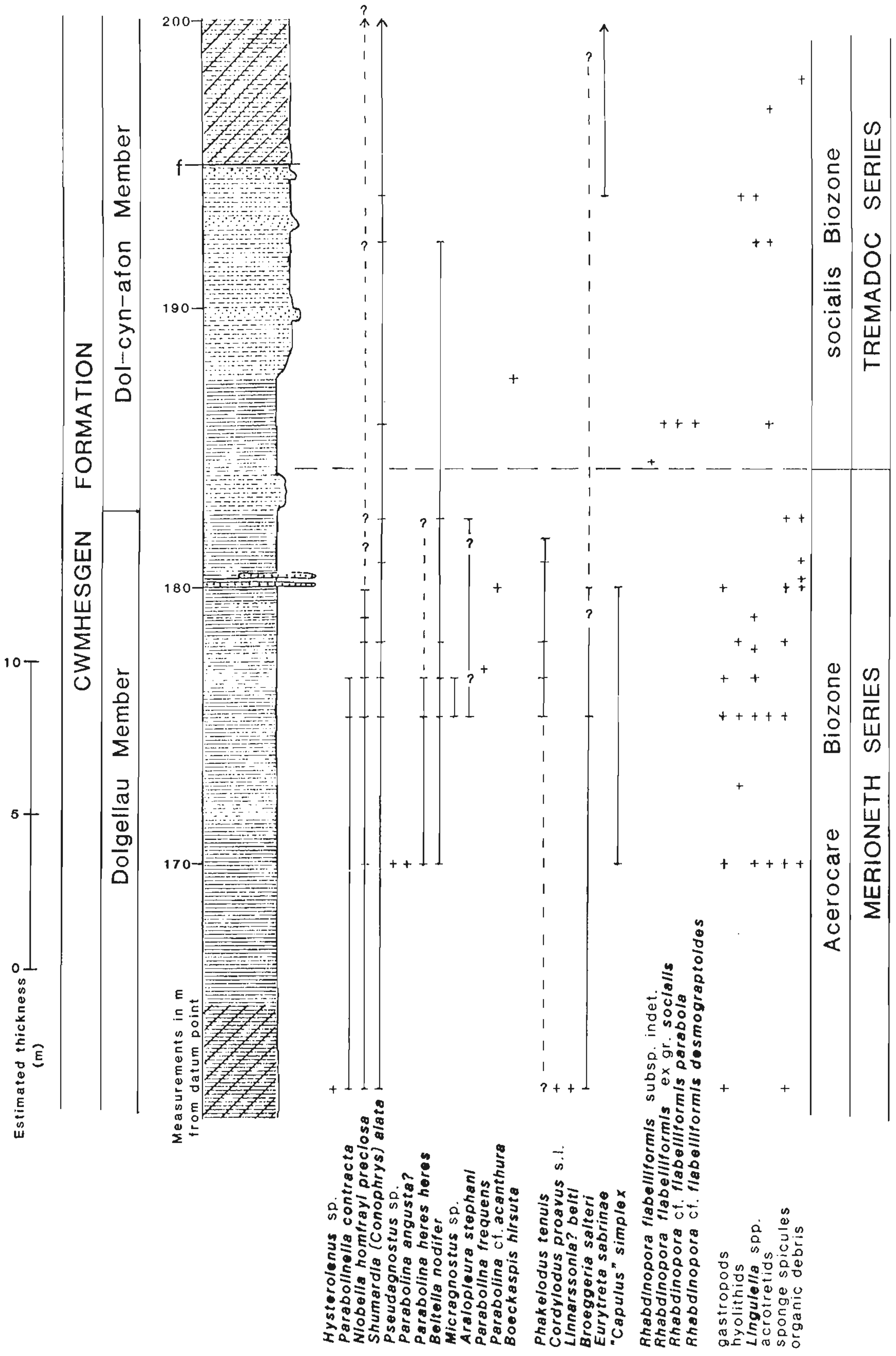

Figure 1. Proposed stratotype for the base of the Tremadoc Series, at Bryn-llin-fawr, North Wales, showing ranges of fossils; updated from Rushton (1982).

shown that the Arenig at its type section is incomplete, as much of the upper part is absent. Furthermore, at the base there is a regional unconformity, and in many places in North Wales the oldest preserved beds are unfossiliferous sandstone. No suitable section exists in North Wales that would be a candidate for a stratotype for the base of the Arenig. Of all the series boundaries, the base of the Arenig poses the most difficult problems. Because the regressive episode at the end of the Tremadoc affected almost all areas, the traditional usage of Arenig has been taken to begin with the transgressive sandstone (poor in fossils) that followed it. In Shropshire, for example, Whittard (1955-67) used as the Arenig base the Stiperstones Quartzite, which is a distinctive mappable formation, but one that has yielded only two trilobite specimens. An Arenig base taken at the base of the Stiperstones Quartzite would clearly be unsatisfactory. In South Wales the Arenig is developed most fully, but no complete section recording the passage between Tremadoc and Arenig has yet been discovered. For this reason, Fortey and Owens (1987) did not formally define a base for the earliest Arenig stage, the Moridunian, in that area. Furthermore, even when faunas appear, the shelly fossils recovered from the lower Arenig cannot be used in correlation beyond the confines of the Welsh basin. At that time, provinciality of faunas was at a maximum.

What we are seeking, therefore, is a section that is continuous between the Tremadoc and the Arenig; the horizon for the base should be taken at a level that can be correlated reasonably widely. This is not an easy matter, but some progress has been made.

Lake District

Since the publication of Whittington et al. (1984), Tremadoc faunas have been discovered in the Lake District (Molyneux and Rushton, 1985). This is particularly interesting with regard to the base of the Arenig, because the Lake District represents the most exterior (off-craton) site in the type Ordovician area, and it might be possible to find a section here that was less influenced by the regression at the end of the Tremadoc. The discovery by Rushton (1985) of the widespread upper Tremadoc graptolite *Dictyonema (Araneograptus) pulchellum* added credence to this argument, and subsequent work has shown no evidence of a break between the Tremadoc and the Arenig in the Lake District.

Molyneux and Rushton (1988) examined the graptolite fauna and acritarch flora of the Watch Hill Grits, and concluded that they may include the equivalents of the *Tetragraptus approximatus* Biozone. This is of particular interest because the base of the *approximatus* Biozone has long been regarded as a practical and widespread horizon for the base of the Arenig Series (Skevington, 1963). Hitherto there has been no good evidence of the presence of *approximatus* Biozone equivalents in England and Wales, although Fortey and Owens (1987) thought that the basal Welsh Arenig might extend this low. The Watch Hill Grits yielded a distinctive acritarch flora, which we refer to as the *trifidum* flora after *Stelliferidium trifidum*, one of its characteristic elements.

The *trifidum* flora also occurs in South Wales in beds (Login section of Cope et al., 1978, Fig. 1) immediately *underlying* the coarse clastics that had been taken as the local base of the Arenig there (Molyneux and Dorning, 1989). The flora has also been recognized in samples from the Manx Slates on the Isle of Man, and probably occurs in the Barriga Shale Formation of southwestern Spain (Molyneux and Rushton, 1988). This indicates that the *trifidum* flora is of great stratigraphic potential, both locally and perhaps more widely over Gondwana.

Only one section in the Lake District shows continuous passage from Tremadoc to Arenig. This is at Trusmadoor, 16 km east of Cockermouth, near the northern margin of the main Skiddaw inlier. A preliminary sketch of this section is shown in Figure 2. The graptolitic faunas indicate an age range from late Lancefieldian to early Arenig; equivalents of the late Lancefieldian and early Bendigonian stages are present, but it is too early to suggest the level at which the base of the Arenig can be defined. Some of the graptolite species are widespread outside Britain; for example, *D. (A.) pulchellum* and *Tetragraptus fruticosus* are recorded from Australasia and Canada, and *Temnograptus multiplex* and *Didymograptus* (s.l.) *protobalticus* are known from the early Arenig of Scandinavia. These widespread ranges are in contrast to the local distribution of the trilobites found in the type Arenig area, and the Lake District succession seems preferable as a basis for international correlation. However, fossils are sparse, and there is no shelly component in the faunas. The evidence from Trusmadoor suggests that the *trifidum* flora appears in rocks of late La2 age, and therefore predates the appearance of *T. approximatus*. However, not all the constituent species of the flora are present at Trusmadoor, and it is possible that certain taxa may appear at a higher level, coinciding approximately with the base of the Arenig, as understood internationally. This might allow eventual correlation between the Lake District, platform Gondwana, and *T. approximatus*-bearing strata elsewhere.

Shelve Inlier, Shropshire

In the spring of 1988, a section was dug below the Stiperstones Quartzite to expose the Habberley Formation. These shales would, of course, have previously been taken as Tremadoc, by definition. The Habberley Formation is younger than any of the classic Tremadoc Shineton Shales. A short distance below the Stiperstones Quartzite an asaphid trilobite referable to the genus *Asaphellus* was recovered; this differs in several characteristics from the familiar Tremadoc species *A. homfrayi* (here it is referred to as *Asaphellus* sp. cf. *A. graffi* Thoral). The same species appears to be present in the Login section, mentioned above as including the *trifidum* flora [approx. = *T. approximatus* Biozone]. Hence it is possible that *approximatus* equivalents are present in Shropshire below the Stiperstones. A possible boundary horizon is indicated in Figure 3. However, this section is very poorly fossiliferous and lacks graptolites.

In summary, two sections have been discovered that are relevant to the definition of the base of the Arenig Series. Neither is particularly fossiliferous. But there is now good reason to suppose that there is no major gap, at least in the Lake District succession of the type Anglo-Welsh area (e.g., Erdtmann, 1965), and that a horizon corresponding approximately to the base of the *T. approximatus* graptolite biozone should be recognizable in Britain, and would be useful elsewhere. The base of the Moridunian Stage will be coincident with the best of the available choices in the Anglo-Welsh area.

LLANVIRN (R.A.F. AND R.M.O.)

The base of the Llanvirn Series has been taken generally at the base of the *Didymograptus artus* Biozone [*D. "bifidus"* zone of earlier authors, and renamed by Fortey and Owens (1987, p. 90)]. The type section at the farmstead of Llanvirn on the coast of southwest Wales is not satisfactory; in particular, the contact with the Arenig below is not well exposed, nor is it in a continuous fossiliferous sequence. Good sections spanning the Arenig-Llanvirn boundary are not easy to find

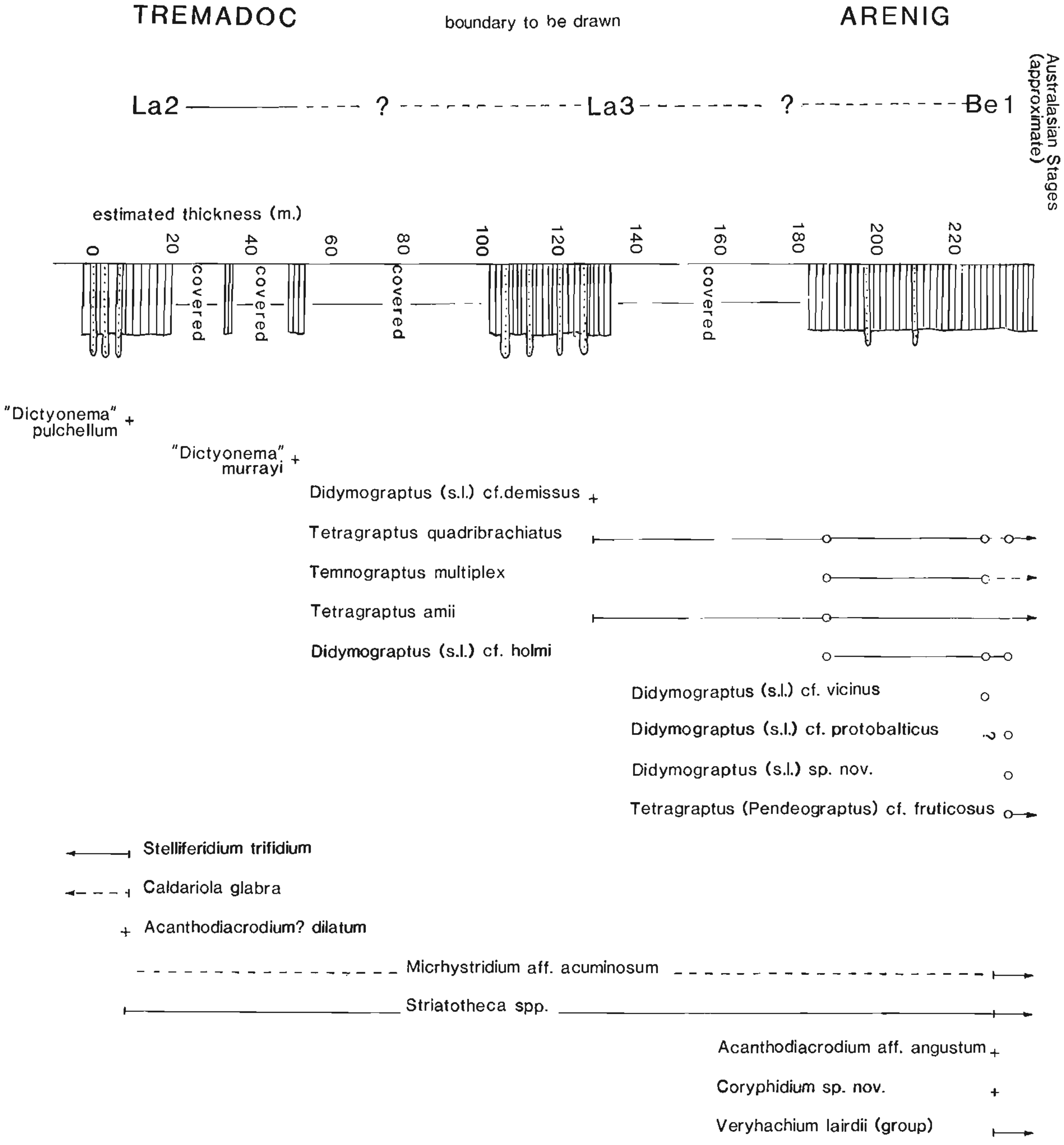

Figure 2. Newly discovered section with continuous passage between Tremadoc and Arenig series, and a possible stratotype for the base of the Arenig Series, Trusmadoor, Lake District. Fossil ranges show graptolites on left, acritarchs on right.

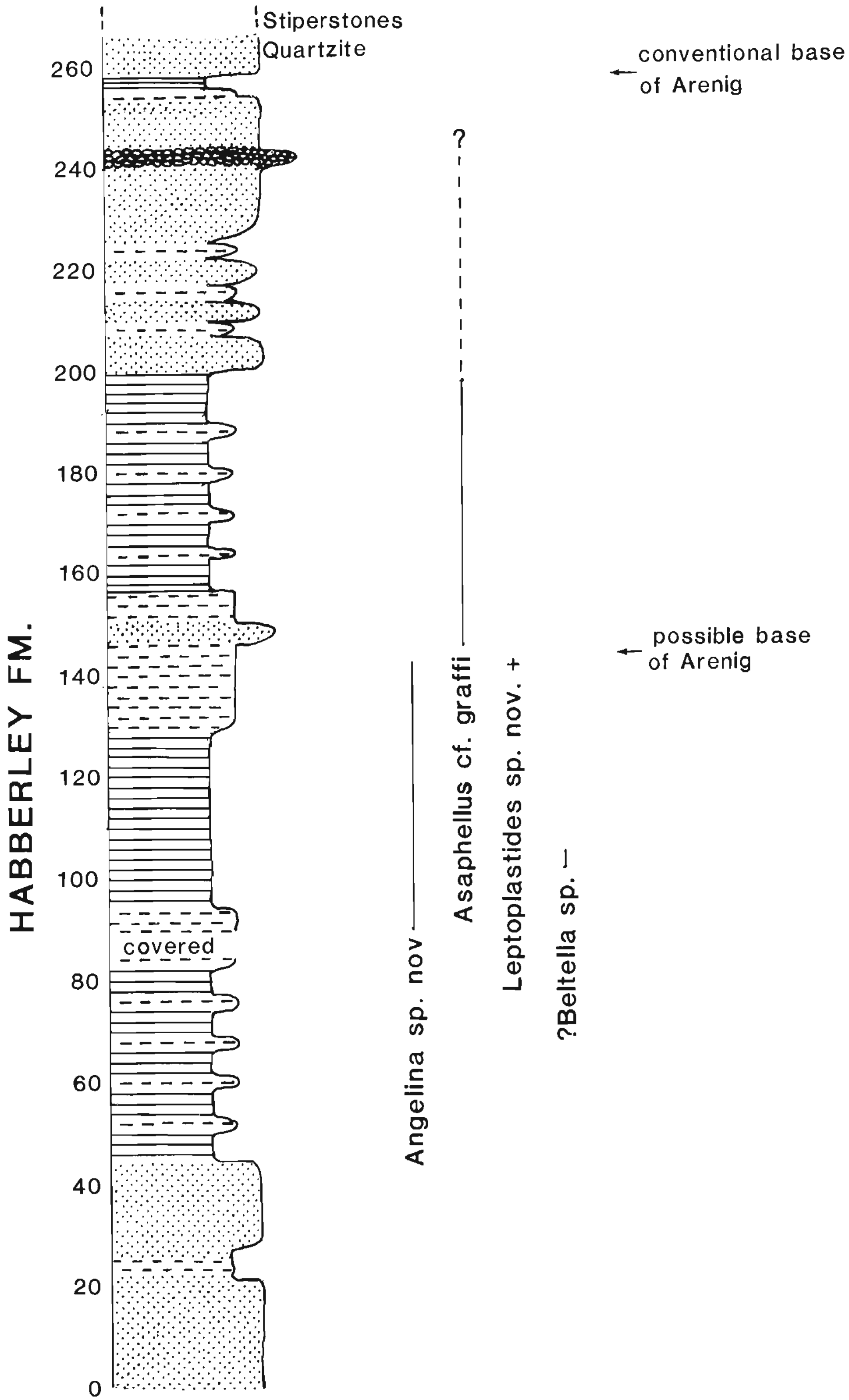

Figure 3. Newly excavated section through the Habberley Shale Formation, underlying the traditional base of the Arenig as taken at the base of the Stiperstones Quartzite Formation (top), Linley Big Wood, Shropshire, with preliminary trilobite ranges, and a possible horizon for the base of the Arenig Series.

(Fortey et al., 1990). Two have been investigated: on the river Seiont, Gwynedd (Caernarvonshire), North Wales, mentioned by Elles (1904); and along a disused railway cut at Llanfallteg, South Wales (Fortey and Owens, 1987, Textfig. 8). Only the latter is fossiliferous enough to deserve serious consideration as a stratotype (Fig. 2). This section is also not far distant from the type area on the Dyfed coast.

The Llanfallteg Formation lies at the top of a thick Arenig succession in South Wales, and consists of at least one hundred metres of soft, grey shale, which are widely exposed in an area east of Haverfordwest. Graptolites and trilobites are common, and are generally well preserved. There are no sudden changes in rock type across the Arenig-Llanvirn boundary, which is therefore placed within the Llanfallteg Formation. The definition of the boundary at the first appearance of pendent didymograptid graptolites is accordingly arbitrary. Many species have unbroken ranges extending across the boundary (Fig. 4).

Correlation

The appearance of pendent didymograptids of the *artus* group affords a convenient criterion for recognizing the base of the Llanvirn Series over a wide area. It is now well known that there is also a group of Arenig pendent didymograptid species. This was a source of much confusion in the past and resulted in the miscorrelation of North American and European so-called *bifidus* zones. True *D. bifidus* is a North American and "Pacific Province" taxon of Arenig age (Cooper and Fortey, 1982; Williams and Stevens, 1988), and is assigned to the subgenus *Didymograptus (Didymograptellus)*. The Llanvirn pendents in the subgenus *Didymograptus (Didymograptus)*, including *D. artus*, make their appearance shortly after biserial graptoloids of the "*Glyptograptus*" (now *Undulograptus) austrodentatus* group, in South Wales and elsewhere. These biserial graptoloids are also used widely in correlation. As well as throughout the Anglo-Welsh area, the Arenig-Llanvirn boundary can be recognized in major areas of Ordovician Gondwana on the basis of the appearance of pendents of the *artus* group: for example, in the Armorican Massif, Iberia, Scandinavia, Saudi Arabia, and South China (Mu et al., 1979). Certain of the trilobites appearing in the Llanfallteg Formation are also widespread in the western segment of Ordovician Gondwana, and provide confirmation of the correlation: *Pricyclopyge binodosa binodosa, Prionocheilus pulcher, Placoparia cambriensis,* and *Selenopeltis buchii* subspecies are among them. This section thus affords a satisfactory standard for the base of the Llanvirn for the Gondwana paleocontinent, over much of which it is difficult to recover conodonts from the dominantly clastic facies.

Correlation outside Ordovician Gondwana is more indirect. A comparable sequence of graptolites occurs in Baltoscandia. In this region, conodont information can be added, which permits correlation of the Arenig-Llanvirn boundary almost worldwide. Bergström (1986, Fig. 2) has summarized the ties between conodont and graptolite biozones, and suggested that the Arenig-Llanvirn boundary corresponds approximately with that between the two "assemblage subzones" into which Löfgren (1978) divided the *Eoplacognathus variabilis* Zone. This is close to, but slightly above, the base of the Kundan Stage in Baltic usage. However, a properly detailed analysis of the succession of graptolites and conodonts across the Arenig-Llanvirn boundary in Scandinavia remains to be made. The position of this boundary within the North Atlantic conodont faunal succession will influence the detail of the correlation with the Midcontinent conodont sequence.

Certain graptoloids in the upper Arenig of the type area are very widespread species, and can be used in correlations with areas as distant as Australia and Texas. These include the relatively offshelf assemblage of taxa termed the isograptid biofacies by Fortey and Cocks (1986). Isograptids evolved particularly rapidly in the later Arenig, and are capable of fine stratigraphic resolution (Cooper, 1973; Cooper and Ni, 1986). The genus *Pseudisograptus* is reported from upper Arenig, Fennian, strata in North and South Wales (Beckly, 1987). These occurrences are probably from the equivalents of the youngest Fennian biozone recognized in South Wales by Fortey and Owens (1987) (*Dionide levigena* Biozone).

The occurrence in Wales of species of the *Pseudisograptus manubriatus* complex is diagnostic of the presence there of the latest of the Arenig graptolite stages recognized in Australasia, the Yapeenian. This is consistent with the Arenig-Llanvirn boundary corresponding closely with the Yapeenian-Darriwilian boundary in Australia. The same boundary equates with the *Isograptus* (s.l.)–*P. tentaculatus* zonal boundary as used in North America. There are some differences in the ranges of graptolite species between the "Pacific" sequence and Gondwana. The most important difference is that the biserial graptoloids of the *austrodentatus* group apparently appear earlier in Wales (and South China) than in Australia, New Zealand, Texas, etc. In the Anglo-Welsh area the first biserials appear quite far down in the Fennian, whereas the "Pacific" appearances are at the base of the Darriwilian. However, the origin of the biserial graptoloids is entirely cryptic (there is no generally accepted candidate for a direct ancestor) and it is not surprising to find regional differences in their time of first appearance. For correlation purposes, the well studied and phylogenetically analyzed isograptids are considered more important (Fortey et al., 1990).

The relationship between the base of the North American Whiterock and the base of the Llanvirn requires some discussion. The Whiterock has been associated with a particular assemblage of North American trilobites and brachiopods with a "middle Ordovician" flavour. This fauna is partly a facies fauna and so its time of appearance will vary from place to place, according to local conditions (Ross and Ingham, 1970; Fortey, 1980). Now that the Whiterock is to be employed in a strictly chronostratigraphic sense, it is important to be clear about the biostratigraphy and facies components. The early Whiterock *Orthidiella* Biozone is relevant to the Arenig-Llanvirn boundary. In places where graptolites and trilobite/brachiopod faunas co-occur, there is the possibility of cross correlation with the type Arenig and Llanvirn. In the Outram Formation of Western Canada, typical Whiterock faunas coexist with *P. tentaculatus* Biozone graptolites (McKee et al., 1972). The Glenogle Shales belonging to the

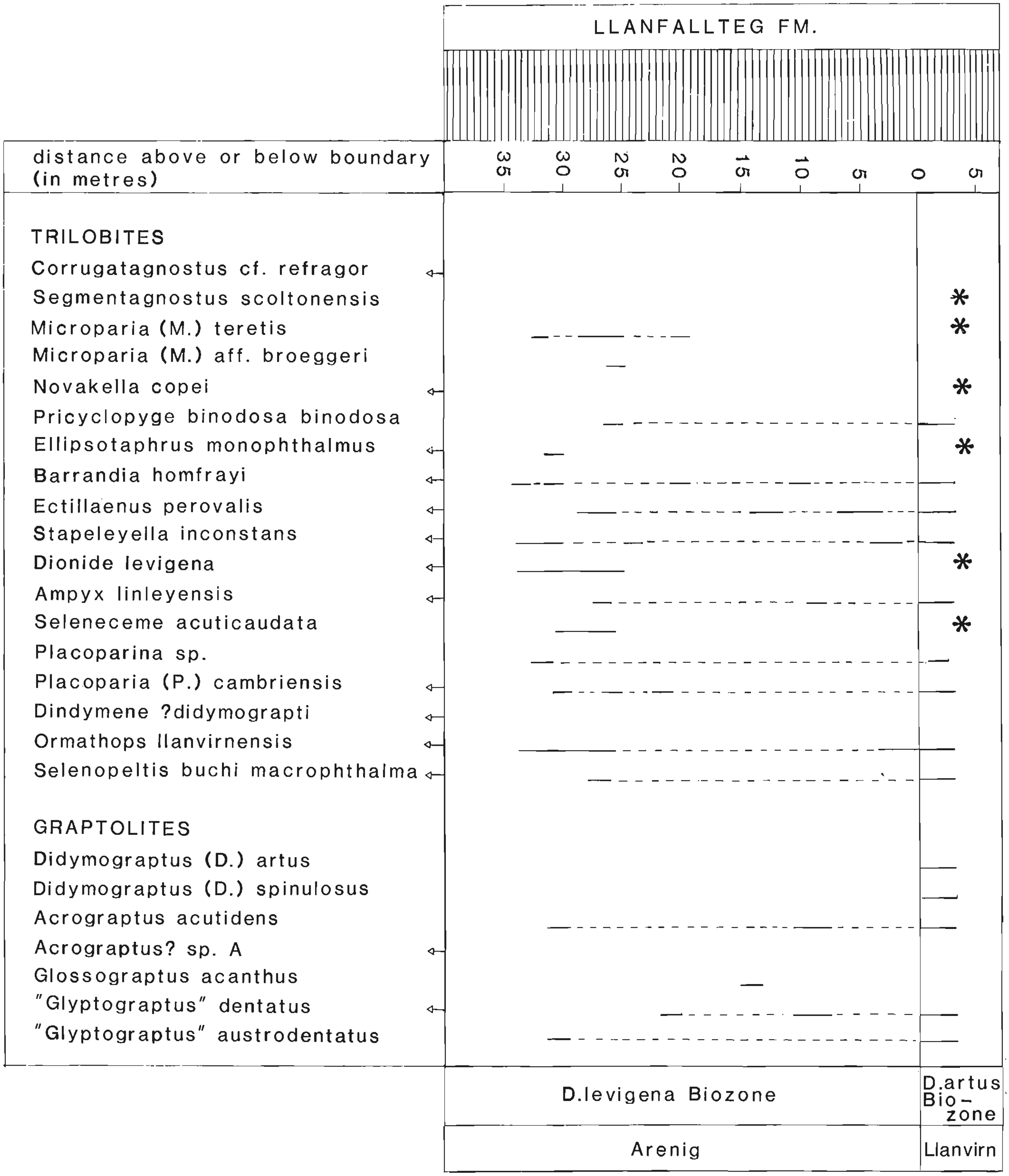

Figure 4. Proposed stratotype for the base of the Llanvirn Series within the Llanfallteg Formation, South Wales (modified from Fortey and Owens, 1987), showing ranges of trilobites and graptolites which are abundant through the section.

Isograptus Biozone beneath also include a "Whiterock fauna" in terms of generic composition, containing trilobite genera such as *Peraspis* and *Globampyx* (Norford and Ross, 1978). But none of the *species* from the Glenogle Shales is identical to those from the *Orthidiella* Biozone of Nevada. The conclusion might be drawn that the Glenogle fauna is Whiterock in facies terms — but not necessarily in chronostratigraphic terms. In Spitsbergen (Fortey, 1980, chart; Cooper and Fortey, 1982), a Whiterock fauna that *does* have some species in common with Nevada *Orthidiella* Biozone faunas occurs well above an occurrence of *Isograptus maximus*, and with *Paracardiograptus*?, suggesting that it is equivalent to Yapeenian to Da1 there. In Spitsbergen, an assemblage of "Whiterock type" genera (but with distinctly different species) occurs below this *Orthidiella* Biozone fauna — an Arenig "Whiterock fauna" in biofacies terms, which Fortey (1980) assigned to a pre-Whiterock Valhallan interval. In the Cow Head Group of western Newfoundland, boulder beds with shelf faunas are interbedded with graptolitic shales. Ross and James (1987) reported a Whiterock assemblage of brachiopods from as early as Bed 12, again basing most of their correlation arguments on genera, and conclude: "the *Orthidiella* Zone spans the range of the Australian Castlemainian, Yapeenian and Darriwilian Da1 graptolite zones". This is, to put it discreetly, a remarkably long shelly biozone; its longevity may be a function of the duration of the biofacies rather than an incapacity to refine it further. In our experience, it is possible to refine shelly zones in this part of the column to a degree not unlike the graptolite biozones in the same interval, in which case one is justified in questioning the meaning of a shelly biozone equivalent to six graptolite zones. This long *Orthidiella* Biozone spans the Arenig-Llanvirn boundary, and, indeed, includes a sizeable portion of the Arenig within it (see also Ross et al., 1982). It should be possible to refine the base of a new biozone within the current *Orthidiella* Biozone that comes close to the Arenig-Llanvirn boundary in Wales. This would increase both precision and international usefulness (Fortey, 1988).

LLANDEILO (R.A.F., M.G.B., R.A.H., P.R.S.)

Richly fossiliferous calcareous flags around the town of Llandeilo and in adjacent areas of the Tywi River valley of South Wales gave Murchison (1835, 1839) the basis to define there what is now the Llandeilo Series; his "Llandeilo flags" replaced his "Builth and Llandeilo flags" and part of his "Black Trilobite Flagstone" of earlier usage (1833, 1834). Details of the history and nomenclature of the series are given by Whittard (1960) and Whittington et al. (1984).

Modern interpretation of the Llandeilo Series in its type area stems from the work of Williams (1953), who recognized three "subformations" (lower, middle, and upper), with a succession of trinucleid trilobite biozones/subzones as a basis for correlation; the "subformations" have generally come to be regarded as divisions of chronostratigraphic stage rank (e.g., Williams et al., 1972, p. 33). Important additional faunal data for definition and correlation are available for various fossil groups, particularly brachiopods (Williams, 1949; Lockley and Williams, 1981), conodonts (Rhodes, 1954; Bergström et al., 1987), and ostracodes (Jones, 1986, 1987). The ecological relationships of faunal changes both laterally and vertically through the succession are described in detail by Wilcox and Lockley (1981).

Three sections in Wales are of immediate relevance to the definition of the base of the Llandeilo Series: the type Llandeilo area, Pant-y-Hendre quarry, and Bach-y-graig, Builth Inlier.

Type Llandeilo area

In the type Llandeilo area itself few sections expose the base of the "Lower Llandeilo Flags" in contact with beds considered to belong to the underlying Llanvirn Series. Following accustomed usage, and Williams (1953, p. 188) and Wilcox and Lockley (1981, p. 287), the "type section" is a composite running south from Dynevor Park at Llandeilo itself, through the well known Ffairfâch Railway cutting into the Cennen Valley.

Following Williams (1953), the base of the Llandeilo is marked by the appearance of the trinucleid trilobite *Lloydolithus lloydi* and by the subzone based on this taxon. In the Ffairfâch cutting, the contact with the underlying Ffairfâch Group (see Williams et al., 1981) is now unexposed, and the contact interval is poorly fossiliferous, although basal Llandeilo sands sitting on Ffairfâch rhyolite conglomerates give a hint that a break might be present at the critical level. The *Lloydolithus lloydii* level is recognized fairly well westward through Carmarthenshire into the Mydrim and Narberth areas, and also northeastward to the Shelve Inlier of the Welsh Borderland where it is present at the base of the Meadowtown Formation (e.g., Whittard, 1979, p. 43).

From information in western Carmarthenshire and in the Builth area (see Williams et al., 1972, p. 5, 33), the base of the Llandeilo Series has generally been taken to approximate closely to the base of the *G. teretiusculus* graptolite Biozone, although Bergström (1986) and Bergström et al. (1987, p. 301) interpret this base as being no older than the upper part of *teretiusculus*. Direct graptolite control is limited to the recent discovery of *G. teretiusculus* itself (identified by R.B. Rickards) in the basal Llandeilo Limestone in the old quarry at the west end of the Ffairfâch cutting above the Ffairfâch Group, but the full range of the *teretiusculus* Biozone in the Llandeilo area remains unknown. A recent assessment of the conodont data from the Ffairfâch cutting (Bergström et al., 1987) shows that the base of the Llandeilo is close to, but somewhat above, the base of the *Amorphognathus inaequalis* Subzone within the widely recognized *Pygodus anserinus* Biozone. The fact that the base of the type Llandeilo does not correspond to any recognizable level in either the graptolite or conodont zonal scale is clearly not satisfactory from a biostratigraphic point of view, especially for international use. Wilcox and Lockley (1981) have shown that brachiopod occurrences are under environmental control in the type Llandeilo, and the same presumably also applies to the trilobites. Nonetheless, a succession of trinucleid species have proved their worth for local subdivision and correlation within the Anglo-Welsh area (Williams, 1953). However, although the trinucleids of the Llandeilo have been well studied, there is no comparably detailed knowledge of the trinucleids of the underlying part of the upper Llanvirn. What is required is a detailed study of the Llanvirn-Llandeilo sequence of benthic faunas (following the brachiopod studies of Lockley and Williams, 1981) to identify the sequence of appearance of taxa. The horizon at which a base may be chosen would depend on correlative usefulness, but is likely to be somewhat below the base of the Llandeilo Flags at Llandeilo. If such could be found, for example, a horizon equating as closely as possible with the base of the *Pygodus anserinus* conodont Biozone would be suitable.

Pant-y-Hendre quarry

This quarry, near Meidrim, shows the more distal facies development of the Llandeilo Series (map in Toghill, 1970). Although overgrown by vegetation, the quarry still affords a good section from the Murchisoni Shales, through the Asaphus Ash and the local thin development of the Llandeilo Flags, into the basal part of the Hendre Shales. Trilobites, brachiopods, and graptolites in the Hendre Shales provide a good "mixed" aspect to the fauna for correlation purposes. The sequence of these faunas requires detailed study.

Bach-y-graig, Builth Inlier (R.A.H., P.R.S)

The base of the Teretiusculus Shales (Elles, 1940) is exposed in this stream section, which is the only one showing this horizon in the Builth Inlier. The lower part of the section is summarized in Figure 5. Trilobites and graptolites are found at irregular intervals throughout. There seems to be no particular faunal change associated with the disappearance of *D. murchisoni*. The interval represented by continuous exposure is also rather short. It is considered that this section, although of interest, is not at present particularly suitable as a stratotype base for the Llandeilo Series.

From this review it can be seen that the Llandeilo poses particularly acute problems for international correlation. A proposal is made below to combine the Llanvirn and Llandeilo (of current usage) within a single Series, with a base at or near the present base of the Llanvirn. This would remove the "Llandeilo problem" from immediate concern as a candidate for definition of an internationally acceptable series.

CARADOC (M.G.B., R.A.H.)

The term Caradoc was first used in stratigraphy by Murchison (1835, p. 47), who applied the name "Caradoc sandstones" as a substitute for his earlier (1834) Horderley and May Hill rocks to embrace a succession of silty and sandy strata in the neighbourhood of the Caradoc Hills of the Church Stretton district, Shropshire, west-central England. Murchison's original concept incorporated a wide range of strata now known to be of Precambrian, Cambrian, Ordovician, and Silurian ages, but subsequent studies by Salter and Aveline (1854) and by Callaway (1877) reduced the scope of the Caradoc to its now generally accepted sense. For a history of the subdivision of the Caradoc Series see Whittard (1960), Dean (1958), Hurst (1979a), and Whittington et al. (1984).

In 1929, B.B. Bancroft first erected stage divisions for the Caradoc Series and he divided the succession biostratigraphically on the basis of mainly brachiopod and trilobite faunas. Modern descriptions of the trilobites are given by Dean (1960, 1961, 1963a, b), who has also redescribed the full stratigraphy and faunal succession (1958, 1964). The upper Caradoc brachiopods have been revised by Hurst (1979c). Brenchley and Newall (1982) and Hurst (1979b, c) give valuable data on the environmental setting of the type Caradoc area, and Turner (1982, 1984) and Jenkins (1967) have summarized acritarch and chitinozoan history, respectively. Modern data on conodonts are summarized by Savage and Bassett (1986), Aldridge and Smith (1985), and Bergström (1983).

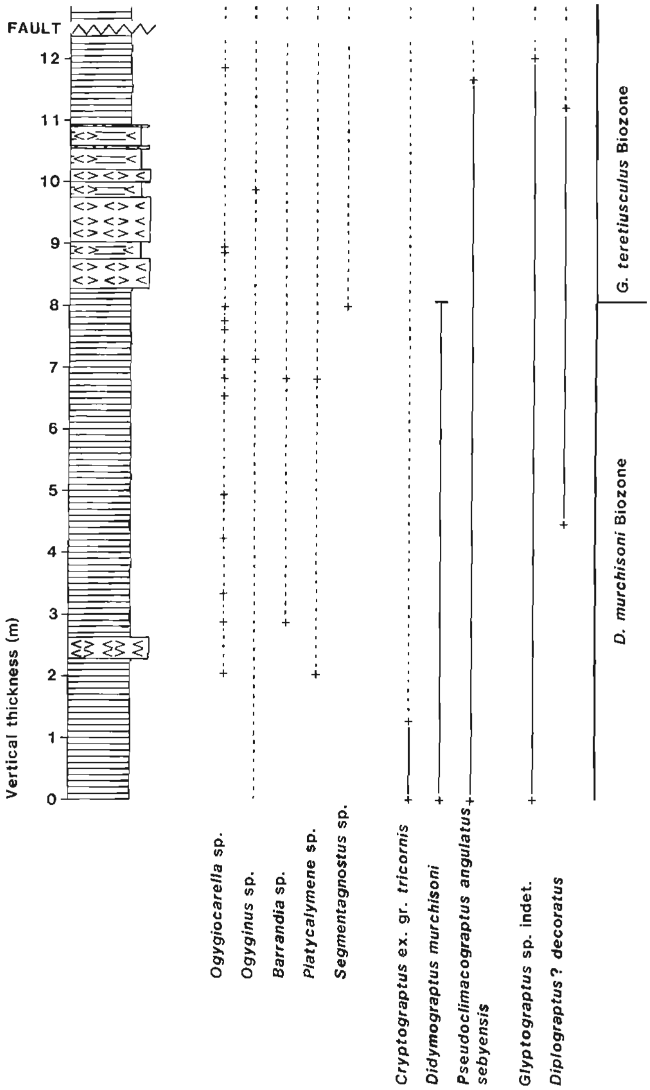

Figure 5. One of the sections relevant to the base of the Llandeilo Series, at the base of the Teretiusculus Shales, Bach-y-graig, Builth Inlier, Powys, central Wales, with graptolite and trilobite occurrences.

In the Caradoc area, the base of the Costonian Stage, which is the oldest of the local divisions of this rank, is everywhere unconformable across Precambrian to Tremadoc rocks. Basal conglomerates are mostly unfossiliferous, but they are then overlain by grit, sandstone, and limestone containing a diagnostic trilobite/brachiopod fauna that includes *Costonia ultima, Harknessella vespertilio, Dinorthis flabellulum,* and *Horderleyella plicata.* This assemblage can be traced westward through Shropshire to the Shelve district (e.g., Spy Wood Grit) and into North Wales (e.g., Derfel Limestone of the Bala District).

Beds of the Costonian Stage also include rare specimens of *Nemagraptus gracilis*, indicating a position within that biozone, with beds of the overlying Harnagian Stage containing a *Diplograptus multidens* Biozone fauna (e.g., Dean, 1958, p. 227, 228). Lower parts of the Coston Formation include conodonts of the *Amorphognathus tvaerensis* Biozone, including the eponymous species (e.g., Bergström, 1983; Aldridge and Smith, 1985; Savage and Bassett, 1986; Bergström et al., 1987). However, the levels of the base of the *gracilis* and *tvaerensis* biozones are unknown in the type Caradoc area.

The unconformity at the base of the type Caradoc Series clearly makes the Shropshire area unsuitable for definition of an international stratotype; in addition, there is no obvious internationally correlative horizon equivalent to basal Costonian. The implications are that we should consider alternative sections and various horizons within the Anglo-Welsh area which *are* capable of international recognition. The three possibilities that require consideration are: the base of the *Amorphognathus tvaerensis* Biozone, the base of the *Diplograptus multidens* Biozone, and the base of the *Nemagraptus gracilis* Biozone.

Base of *Amorphognathus tvaerensis* Biozone

The key section that yields well preserved, diagnostic conodonts is at Bryn Banc quarry near Lampeter Velfry, Dyfed (formerly Carmarthenshire), described by Addison (*in* Bassett et al., 1974) and Bergström et al. (1987, Fig. 18.7). This quarry exposes a 30 m thick sequence of limestone classified formerly within the "upper Llandeilo", and adjacent exposures allow a thicker succession to be reconstructed into overlying *Dicranograptus* shales. The conodont data of Bergström et al. (1987) suggest that the base of the *A. tvaerensis* Biozone occurs in the mid to upper part of the Bryn Banc sequence, and further recent collecting (by M.G.B. and N.M. Savage) has been carried out in order to identify this level precisely.

Slightly higher beds at the top of the quarry contain *Costonia* and other "diagnostic Caradoc" shelly fossils. Coupled with Addison's work throughout this area (*in* Williams et al., 1972; *in* Bassett et al., 1974), which shows that the *Nemagraptus gracilis* Biozone extends down at least into the upper part of the type Llandeilo succession, and probably lower, it has become increasingly clear that there is an overlap between the uppermost type Llandeilo and lowermost type Caradoc; the recognition of the *A. tvaerensis* Biozone at both Bryn Banc and in the Costonian of Shropshire now confirms this. Selection of the base of the *tvaerensis* Biozone at Bryn Banc as a revised base for the Caradoc Series would thus essentially protect the interpretation of the Caradoc in its traditional sense.

Selection of this level at Bryn Banc would also conserve the traditional interpretation of the Llandeilo Series, since slightly lower beds in the section contain the trinucleid trilobite *Marrolithus favus*, diagnostic of the uppermost Llandeilo *favus* Biozone of Williams (1953).

The *A. tvaerensis* Biozone is one of the most widespread and recognizable units in the North Atlantic conodont succession, emphasized by Bergström (1971, p. 99) as being of considerable value for intercontinental correlation. It is known from throughout Baltoscandia (in Dalby Limestone and equivalents), in Poland, the Girvan area of Scotland, and in rocks of "Porterfield to lower Barneveld" age in the Appalachians and in central and western North America. The correlation of the base of the *tvaerensis* Biozone with both the American Midcontinent conodont zonation and the graptolite succession has been documented repeatedly in papers by Bergström (e.g., 1978, 1983, 1986; Bergström et al., 1987, 1988). All these studies confirm the evidence from South Wales that the base of the *gracilis* Biozone extends for some considerable way below the base of the *tvaerensis* Biozone.

Base of *Diplograptus multidens* Biozone

In the Shelve Inlier, about 15 km west of the type Caradoc area, there is a continuous section in graptolitic facies extending throughout the *gracilis* Biozone into the *multidens* Biozone. The section in Spy Wood Brook has recently been studied in detail, although at the moment only the graptoloid succession is known species by species (Hughes, 1989). The section (Hughes, 1989, Textfigs. 6, 12a) is almost completely exposed and structurally intact. A variety of shelly fossils will provide a link with the sequences of the type Caradoc area, although this requires a detailed study of their distribution; Whittard (1955-67) and Williams (1974) have recorded a great variety of trilobites and brachiopods, respectively, through this section, and ostracodes have been added recently by Jones (1986, 1987).

From evidence of brachiopods (Williams, 1974, p. 14) and trilobites (Whittard, 1979; Dean, *in* Whittington et al., 1984, p. 26), the base of the Caradoc is placed at or below the base of the Spy Wood Sandstone Formation. This corresponds closely with the base of the *Diplograptus multidens* Biozone as recognized in Hughes' (1989) revision, which is placed just below the Spy Wood Sandstone at the top of the Rorrington Shale Formation. In the type Caradoc sequence of the Harnage area, the range of *N. gracilis* extends at least into the lower Costonian (Pocock et al., 1938), and the traditional base of the Caradoc therefore lies presumably somewhat below the base of the *multidens* Biozone within the Rorrington Shale. But the base of this Biozone would provide a horizon within a continuous sequence without seriously affecting the concept of Caradoc based on the type section. The disadvantage is that, at the moment at least, it is unclear how widely the horizon may be correlated internationally. Possibly the current active restudy of biserial graptoloids will enable its recognition quite widely.

Base of *Nemagraptus gracilis* Biozone

The base of the *N. gracilis* Biozone has been described within the Rorrington Shale Formation by Hughes (1989), as taken at the first appearance of the eponymous species. There is no doubt that the *gracilis* Biozone is one of the most widely recognized Ordovician intervals, corresponding as it does to a major eustatic transgression. As such it can be recognized on most continents outside platform limestone facies. The base of the biozone would correspond to the base of the Gisbornian of the Australian graptolitic standard.

The main problem with the selection of this horizon for a Caradoc base would be that the Caradoc would then consume a large part of the classical Llandeilo, which as shown above is now known to include some part, and possibly a large part, of the *gracilis* Biozone. This would not necessarily be a major objection if the "old" Llandeilo were to be replaced by a local stage, as we suggest below. But it does represent a radical departure from current usage. It would, in practice, represent a return to an older view (e.g., that of O.T. Jones, 1936) on the horizon of the basal Caradoc, which had been based on misconceptions about the ages of type sections.

ASHGILL (J.K.I., A.W.O., D.A.T.H.)

The Ashgill Series of Marr (1905) takes its name from Ash Gill, near Coniston in the Lake District in an area where the succession is far from complete, and where basal beds of mid-Ashgill (Cautleyan) age rest unconformably on the Borrowdale Volcanic Group (Fig. 6). The type area for the Series (Marr, 1913, p. 13, 1916, p. 89; Ingham and Wright, 1970, p. 233) is about 40 km to the east in the Cautley district of the Howgill Fells. Here, a thick and monotonous sequence, consisting largely of neritic calcareous mudstone, begins with strata of late Caradoc (Onnian) age followed conformably by a succession spanning virtually the whole of the Ashgill Series.

The Ashgill Series is divided into four stages in England and Wales: Pusgillian, Cautleyan, Rawtheyan, and Hirnantian (see Ingham and Wright, 1970). The Pusgillian was introduced by Bancroft (1945) for the highest part of the Dufton Shales at Pus Gill in the Cross Fell Inlier of the northern Pennines and was included initially in the Caradoc Series (e.g., by Dean, 1959; Ingham, 1966) in the mistaken belief that it equated entirely with the *Pleurograptus linearis* Biozone, a graptolite zone ascribed to the Caradoc Series by Elles (1922) with little reference to type sections. It has since become clear that the Pusgillian Stage is best regarded as the lowest division of the Ashgill Series. It is absent, for example, from the type Caradoc area in South Shropshire. Also it can be shown in the type Ashgill area at Cautley that the Pusgillian

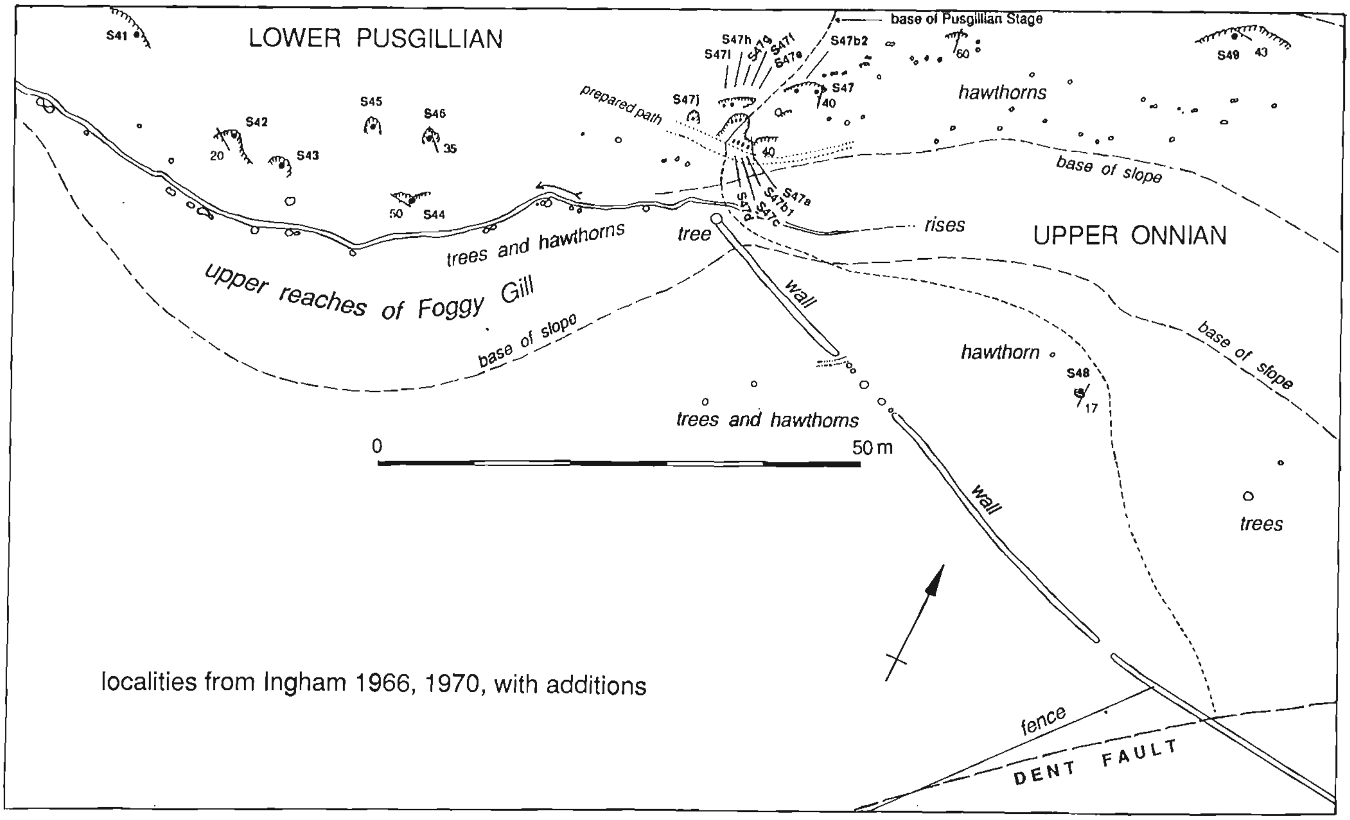

Figure 6. Proposed stratotype for the base of the Ashgill Series, Foggy Gill, Murthwaite Inlier: map shows Onnian-Pusgillian boundary and collection sites.

is partly equivalent to Marr's "*Calymene*" Beds, which King and Williams (1948) demonstrated were not of Caradoc age. There is also evidence from the Girvan area of southwest Scotland that the Pusgillian Stage includes probably all of the *D. complanatus* graptolite Biozone, widely regarded, since Elles (1922), as the lowest graptolite zone of the Ashgill Series. The base of the overlying *D. anceps* Biozone (*D. complexus* subzone) is believed to lie at or near the base of the Cautleyan Stage (Ingham, 1978).

Base of the Pusgillian Stage

The best section for the base of the Pusgillian Stage and of the Ashgill Series, which is proposed here as a stratotype, is on Foggy Gill in the Murthwaite Inlier, Cautley district. Onnian (uppermost Caradoc) strata are succeeded conformably by strata of Pusgillian age, and in 1988 we were able to establish the boundary to within a few centimetres. Our provisional faunal range chart is shown in Figure 7.

Provisional chart showing ranges of selected taxa across the Caradoc - Ashgill (Onnian - Pusgillian) boundary in its type section in the Cautley Mudstone Formation at Foggy Gill in the Murthwaite Inlier, Cautley (Cumbria). (Localities from Ingham 1966, 1970, with additions)

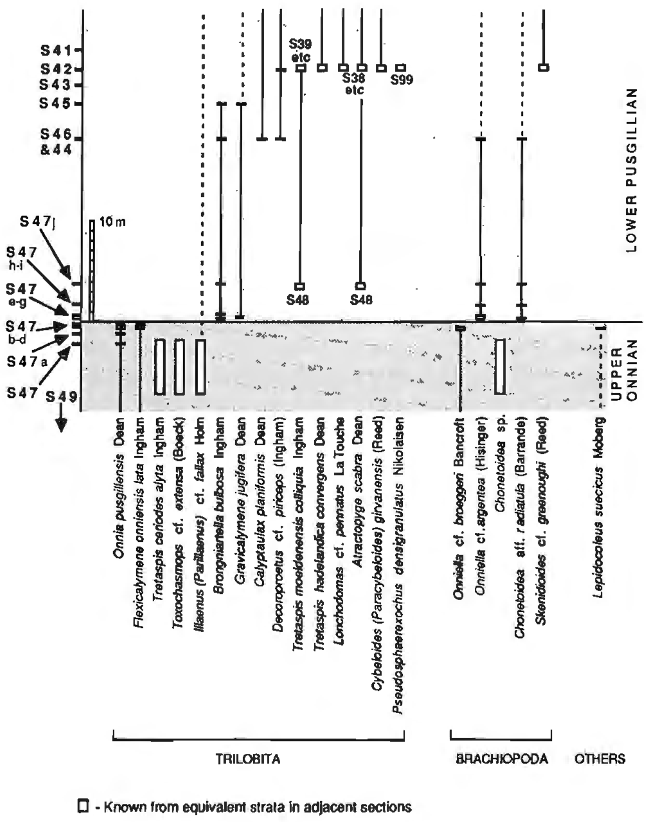

Figure 7. Proposed stratotype for the base of the Ashgill Series, base of Pusgillian, Foggy Gill, Murthwaite Inlier: faunal logs with brachiopod and trilobite ranges.

Two trilobites range to the end of the Onnian and do not continue into younger strata: *Onnia pusgillensis* and *Flexicalymene onniensis lata*. Other trilobites known from Onnian rocks in adjacent sections include *Tretaspis cerioides alyta* and *Toxochasmops* sp. cf. *T. extensa*, which are restricted to this stage, and *Illaenus (Parillaenus)* sp. cf. *I. (P.) fallax* which ranges to the top of the Cautleyan. Among the brachiopods, *Onniella* sp. cf. *O. broeggeri* and a distinctive *Chonetoidea* are restricted to the Onnian. Two trilobite species, *Brongniartella bulbosa* and *Gravicalymene jugifera* appear at the base of the Pusgillian, the former being restricted to the lower part of the stage, the latter ranging up to the lower part of the Cautleyan Stage, from whence it is known in the Cross Fell Inlier and in Wales. Adjacent sections in the lowest part of the Pusgillian have yielded *Tretaspis moeldenensis colliquia*, which is confined to it, and the longer-ranging *Atractopyge scabra*. Some twenty to twenty-five metres higher (but still in the lower part of the Pusgillian) a number of trilobites useful in correlation appear, including: *Tretaspis hadelandica convergens*, *Pseudosphaerexochus densigranulatus* and *Calyptaulax planiformis*. Among the brachiopods, *Onniella* sp. cf. *O. argentea* and *Chonetoidea* sp. aff. *C. radiatula* appear at the base of the Pusgillian and adjacent lower Pusgillian sections have yielded *Skenidioides* sp. cf. *S. greenoughi*.

Correlation

Although the faunal changes documented at the boundary no doubt partly reflect subtle environmental shifts, some of them do not, and correlative potential of some of the species is considerable. *Tretaspis ceriodes*, for example, is widespread and is confined to the upper part of the Caradoc Series (Actonian-Onnian). Its distribution extends to the Welsh Borderland, Scotland, Ireland, Norway, Sweden (see Owen, 1980; Owen et al., 1986), and as far east as China (Sheng, 1974, p. 78, Pl. 4, figs. 1, 2) — as *Nankinolithus* sp. aff. *N. nankinensis* Lu in an unnamed marly sandstone at Manju Reservoir (loc. 7129) in the Luxi district of western Yunnan.

Two *Tretaspis* stocks appear in the Pusgillian, both derived independently from *T. ceriodes*. *Tretaspis moeldenensis colliquia*, from the lower part of the Pusgillian only, is not known outside the north of England and Wales (Price, 1977), where it is of great value in recognizing early Ashgill strata. *Tretaspis hadelandica convergens* is a member of the *T. seticornis* group of Ingham (1970), all of which are confined to the Ashgill Series (see Owen, 1980 for a phylogenetic framework). This group is also distributed widely, being known from Quebec, Ireland, Scotland, England, Wales, Norway, Sweden, Estonia, Poland, and Czechoslovakia. *Tretaspis clarkei* Cooper, from the Whitehead Formation of Percé, Quebec and from the Vaureal Formation of Anticosti Island is a member of the group and may well identify with *T. hadelandica* s.l. (Owen, 1980, p. 733). The replacement of *T. ceriodes* by *T. seticornis* group in Scandinavia allows for the positioning of the Caradoc-Ashgill boundary with some precision there; for example, it falls within the Gagnum Shale member of the Lunner Formation in Hadeland, Norway (Owen, 1978, 1980, 1987; Bruton and Owen, 1979) and at the base of the Fjäcka Shale and Slandrom Limestone in central Sweden. The *Onniella-Chonetoidea/Sericoidea* brachiopod association found in the Cautley district (including the Foggy Gill section) is widespread in the deeper water clastic facies in the Upper Ordovician, particularly around the Caradoc-Ashgill boundary in the Baltic, Mediterranean, and North American provinces, thus providing potential for near global correlation.

Correlation with graptolite and conodont schemes

No graptolites are known from the proposed type section for the base of the Ashgill Series, but it is known that the boundary falls within the *P. linearis* Biozone (Ingham and Wright, 1970; Wright, *in* Whittington et al., 1984, p. 27, 28; cf. Williams and Bruton, 1983, p. 161, 162). The upper part of this biozone is recognized in the Fjäcka Shale Formation in central Sweden (Skoglund, 1963), which also yields lower Ashgill trilobites. At Girvan, the *P. linearis* fauna known from the upper part of the Lower Whitehouse Group occurs below a horizon in the Upper Whitehouse Group yielding late Caradoc *Tretaspis ceriodes* s.l. The top of the *linearis* Biozone almost certainly lies within the Pusgillian as *Dicellograptus complanatus* faunas are known from the upper part of the Upper Whitehouse Group and the lower beds of the overlying Shalloch Formation (Williams, 1987) where they are associated with lowest Ashgill trilobites — including *Tretaspis* sp. cf. *T. hadelandica convergens*. The base of the *linearis* Biozone is believed to fall within the Onnian Stage. Much has been made of the occurrence of *Onnia gracilis* in the Nod Glas Formation of the Welshpool area of mid-Wales associated with graptolites, which on balance probably indicate a level high in the *D. clingani* Biozone (Cave, 1965; Cave and Price, 1978). *Onnia gracilis* has been regarded previously as indicative of mid-Onnian but its range has been extended down to the late Actonian (Owen and Ingham, 1988, p. 850).

In Sweden, the base of the Harju Series (Jaanusson, 1963) is taken notionally at the base of the *linearis* Biozone, which was believed originally to lie at the base of the Fjäcka Shale, or, in the Siljan District, the Slandrom Limestone. For reasons already stated this is not now believed to be the case, for both these units contain lower Ashgill trilobites. In lithostratigraphic terms, for all practical purposes, the base of the Harju Series corresponds very closely with the base of the Ashgill Series (see Jaanusson, 1982, Fig. 4) and Jaanusson's Vasagaardian Stage is probably the equivalent of some lower part of the Pusgillian.

A distinctive *Dicellograptus* from the basal *D. complanatus* Biozone in the Upper Whitehouse Group at Girvan, described by Williams (1987) as *D. alector*? may be extremely valuable for intercontinental graptolite facies correlation. *Dicellograptus alector* is known otherwise from the Phi Kappa Formation of Idaho and from Nevado where its range is evidently restricted to an interval equivalent to the *D. clingani* and *P. linearis* biozones. A probable senior synonym is *D. gravis* from the latest Eastonian (Ea4) Stage of Victoria, Australia, which Williams correlates with the lower part of the British *complanatus* Biozone. It seems therefore that the Eastonian-Bolinidian boundary lies within the *complanatus* Biozone and may be closer to the top of the Pusgillian than the bottom.

With regard to conodonts, Sweet and Bergström (1971, p. 622) have demonstrated that advanced *Amorphognathus superbus* occur in the lower part of the type Maysvillian Stage of the Cincinnatian Series, which suggests that the base of the *A. ordovicicus* Biozone probably lies within that stage. Similarly, Savage and Bassett (1986, p. 683) have indicated on the basis of Welsh data that the base of the *A. ordovicicus* Biozone may also lie within the Pusgillian Stage, in spite of statements to the contrary (Orchard, 1980; Bergström and Orchard, 1985). It may be that the bases of the Pusgillian and Maysvillian stages are approximately coeval.

In summary, the base of the Pusgillian Stage is an appropriate level at which to define the base of the Ashgill Series. There are a sufficient number of species of trilobites and brachiopods to effect international correlation. If the base were to be redefined at the base of the *P. linearis* Biozone, its position within the Onnian Stage would be difficult to identify. Equally, if the boundary were to be transferred to the base of the *complanatus* Biozone, its position is unlikely to be recognizable within the Pusgillian Stage. Neither of these alternatives commends itself to the writers. The base of the Ashgill Series as currently understood has the best potential for precise definition, wide correlation, and nomenclatural stability.

A CASE FOR COMBINING LLANVIRN AND LLANDEILO WITHIN A SINGLE SERIES

In the light of our review of all available sections and their potential for definition of units of series rank, we consider that it may now be preferable to combine the Llanvirn and Llandeilo in order to form a division commensurate in scope with those others in the Ordovician System. The present series in the British type area have been criticized as being too short to be usable at this level in the chronostratigraphic hierarchy (Chlupač et al., 1981), and this objection has most substance with regard to the Llanvirn and Llandeilo. The Llanvirn (with two graptolite zones) and Llandeilo (with three trinucleid zones equivalent approximately to one graptolite zone) appear to be brief on a biostratigraphic measure, approximate though this is, and this impression is consistent with assessments of absolute age. For example, on the McKerrow et al. (1985, p. 75) timescale, the Llanvirn and Llandeilo (9 and 7 m.y. long, respectively) do appear to be short compared with the Tremadoc, Arenig, and Caradoc (21, 22 and 16 m.y. long, respectively). Combining the Llanvirn and Llandeilo would provide a series commensurate with those above and below. Neither name need be lost. We propose that the Llandeilo be retained for the whole series, and that the Llanvirnian be applied as a division of stage rank for the lower part of the Llandeilo. This would then require only the introduction of one new name for the upper stage division, approximately equivalent in scope to the present Llandeilo; for this, the Dyneforian Stage would be appropriate, named after Dynefor Park at Llandeilo itself, where rocks of this age are so well exposed (Williams, 1953).

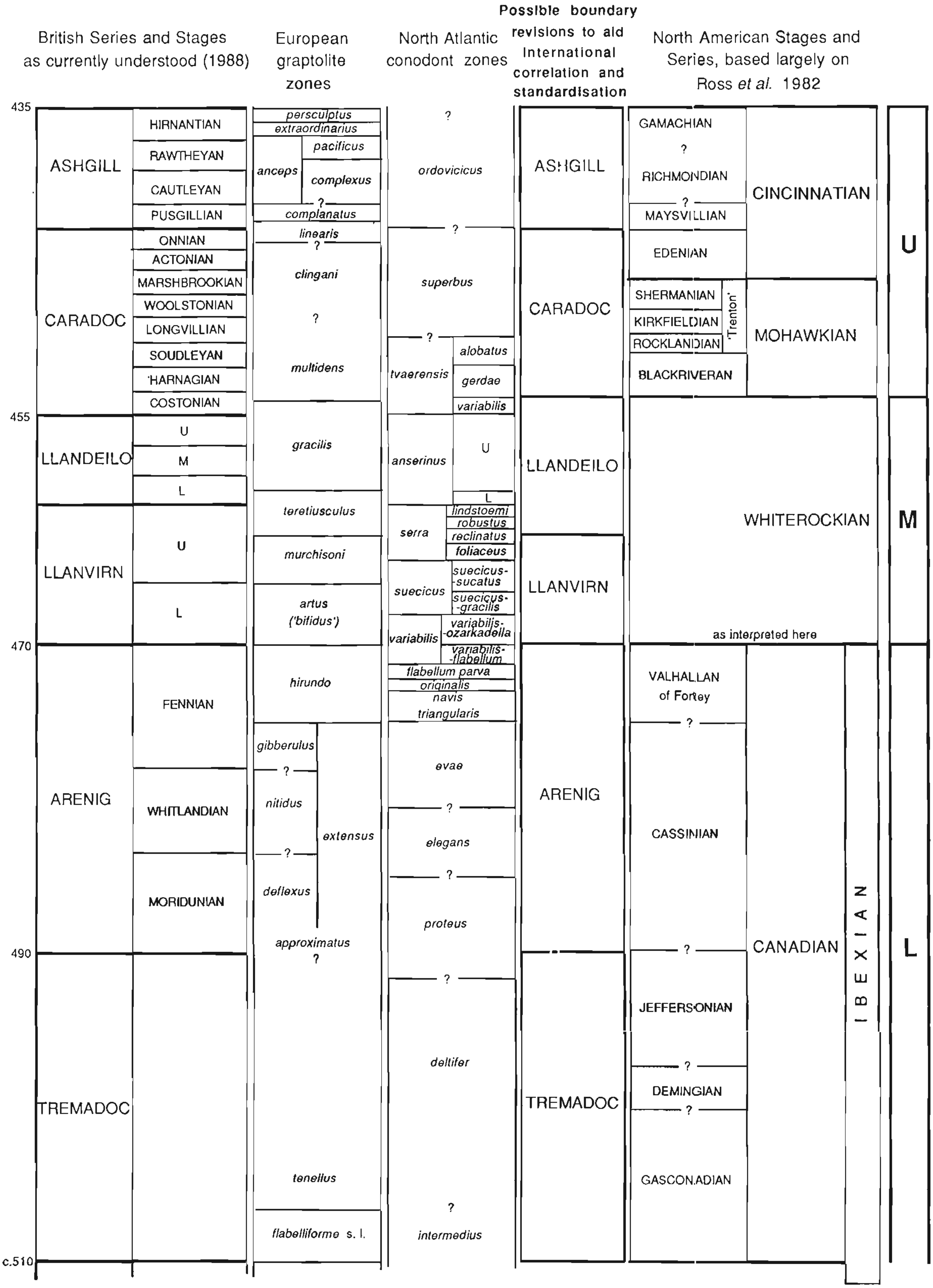

Figure 8. Correlation chart showing how some of the solutions suggested in this paper could be used for international correlation given some flexibility in definition of bases of series or stages in other countries. The North American scheme is used as an example. Note: this is a correlation chart of ***suggested**, not existing, ties.*

The proposal would, of course, remove the problem of definition of the base of the Llandeilo Series (present usage) from the immediate concern of this working group. The base of the Llanvirn is, at least potentially, capable of wide correlation. According to the possible definitions of the base of the Caradoc, as discussed above, the revised Llandeilo would either include the *Nemagraptus gracilis* Biozone, which is recognized almost worldwide, or that Biozone would be adjoined with the Caradoc. Given that the proposal would also result in a more equable distribution of time through the series of the Ordovician, without interfering with current terminology, it should find favour with those concerned with making the various primary chronostratigraphic subdivisions of the system robust and durable concepts capable of employment internationally. More generally, Figure 8 is an attempt to show how, with some common sense employed in the definition of bases of chronostratigraphic divisions of the Ordovician System internationally, a practical correlation between different continents — and between different traditional usages — could be achieved.

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The international working group on the Cambrian-Ordovician boundary: report of progress

B.S. Norford[1]

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Abstract

The Working Group has the responsibility of recommending a specific level within a suitable stratigraphic section to serve as the global stratotype for the Cambrian-Ordovician boundary. Commencing in 1974, comprehensive studies culminated in a plenary session held in Calgary in 1985 and resulted in decisions on the "golden-spike" principle of selecting the boundary at a horizon just below the first influx of planktic (nematophorous) graptolites. Conodonts are to be used as the primary guide for the selection of the specific horizon and the global stratotype section must have potential for studies using paleomagnetism, geochemistry, and other nonbiological correlation tools. The specific horizon will correspond approximately to the base of the Tremadoc Series of Wales and will be slightly higher than the base of the Canadian Series as used in western North America.

At the Calgary meeting, important sections from various parts of the world were reviewed. The Xiaoyangqiao sections at Dayangcha, Jilin Province, China and the Green Point and Broom Point sections in Newfoundland, Canada were identified as prime candidates to serve as the global stratotype. A delegation from the Working Group inspected the Dayangcha sections in 1986; the Newfoundland sections similarly were inspected in 1981. A 1988 issue of the Geological Magazine presents details of these and other globally important sections and of the biostratigraphy provided by data on conodonts, graptolites, and trilobites. In most parts of the world, the first influx of planktic graptolites has been documented as occurring within the lower part of the conodont zone of Cordylodus lindstromi Druce and Jones.

Résumé

Le groupe de travail se charge de recommander un niveau particulier dans une coupe stratigraphique convenable qui servira de stratotype mondial pour la limite cambro-ordovicienne. Des études détaillées commencées en 1974 se sont terminées par une assemblée plénière tenue à Calgary en 1985; on y a pris des décisions quant au principe de choisir la limite à un horizon qui est immédiatement au-dessous de la première apparition de graptolites planctoniques (nématophores). Des conodontes serviront de repère primaire pour le choix de l'horizon particulier; en outre, il faudra pouvoir étudier le stratotype mondial à l'aide de méthodes paléomagnétiques et géochimiques et d'autres outils de corrélation non biologiques. L'horizon particulier correspondra à peu près à la base de la série du Trémadocien du pays de Galles et sera légèrement plus haut que la base de la série du Canada selon le sens qu'on lui donne dans l'ouest de l'Amérique du Nord.

À la rencontre de Calgary, on a examiné des coupes importantes provenant de différentes parties du monde. Les coupes de Xiaoyangqiao, à Dayangcha, dans la province de Jilin, en Chine, et les coupes de la pointe Green et de la pointe Broom, à Terre-Neuve, au Canada, sont reconnues comme étant les principales possibilités pour le stratotype mondial. Une délégation du groupe de travail a inspecté les coupes de Dayangcha en 1986; les coupes à Terre-Neuve avaient été inspectées en 1981. Un numéro de la revue Geological Magazine paru en 1988 présente des détails sur ces coupes et d'autres coupes d'importance mondiale et sur la biostratigraphie établie à partir de données sur les conodontes, les graptolites et les trilobites. Presque partout au monde, les graptolites planctoniques apparaissent pour la première fois dans la partie inférieure de la zone à conodontes à Cordylodus lindstromi Druce et Jones.

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HISTORICAL PROGRESSION

The Ordovician System was originally proposed with its lower limit at the base of the lower Arenig (Lapworth, 1879, p. 14). As Henningsmoen (1973) commented, this was the base of the lower Arenig using Hick's (1875) tripartite division of the Arenig Group, which meant that the upper part of the Tremadoc Series of Wales was included within the Ordovician. Since 1879, the basal Ordovician has been correlated around the world with varying degrees of accuracy, with the base of the system interpreted for local convenience within sequences of mappable rock units. National and regional preferences have arisen and the various bases recognized for the Ordovician System have differed in age by some ten million years, ranging from near the base of the Tremadoc Series to the base of the Arenig Series (as currently defined).

Uppermost Cambrian and Lower Ordovician rocks are widely distributed in the world; many of the sequences are fossiliferous and, fortunately, three of the common and well studied groups of fossils have proved to be very effective for international correlation. Graptolites, trilobites, and conodonts have been used to establish both local and intercontinental successions of biostratigraphic zones and also, by their presence together in many sequences of rocks, to document the relationships between zonations based on these three groups. The graptolites and conodonts seem to have been cosmopolitan during this episode of geological time; faunal provinces existed for trilobites. The best known Cambrian-Ordovician sequences are in Great Britain, Scandinavia, the Soviet Union, China, Australia, western South America, the western and southern United States, and in western and easternmost Canada (Fig. 1). All these sequences can be correlated with one another with considerable precision, and the correlations show that horizons recognized regionally as the base of the Ordovician range from well below the first appearance of subspecies of the graptolite *Rhabdinopora flabelliformis* to as high as the base of the *Tetragraptus approximatus* Zone.

Following the very successful programme of the Silurian-Devonian Boundary Committee that resulted in international agreement on a horizon documented by a reference section at Klonk, Czechoslovakia (McLaren, 1977), similar studies were initiated for other systemic boundaries. During the 1974 Birmingham Symposium on the Ordovician System, the Working Group on the Cambrian-Ordovician Boundary was established (Wright, 1976, p. 3) as a constituent body of the International Commission on Stratigraphy, which itself is part of the International Union of Geological Sciences. The decisions of the Working Group are subject to ratification by the parent bodies. Titular (voting) and corresponding members were appointed for most of the countries of the world that have sequences of Upper Cambrian and Lower Ordovician rocks. The main objectives of the Working Group are as follows:

1. To stimulate intensive study of the boundary interval and of its faunas and floras.

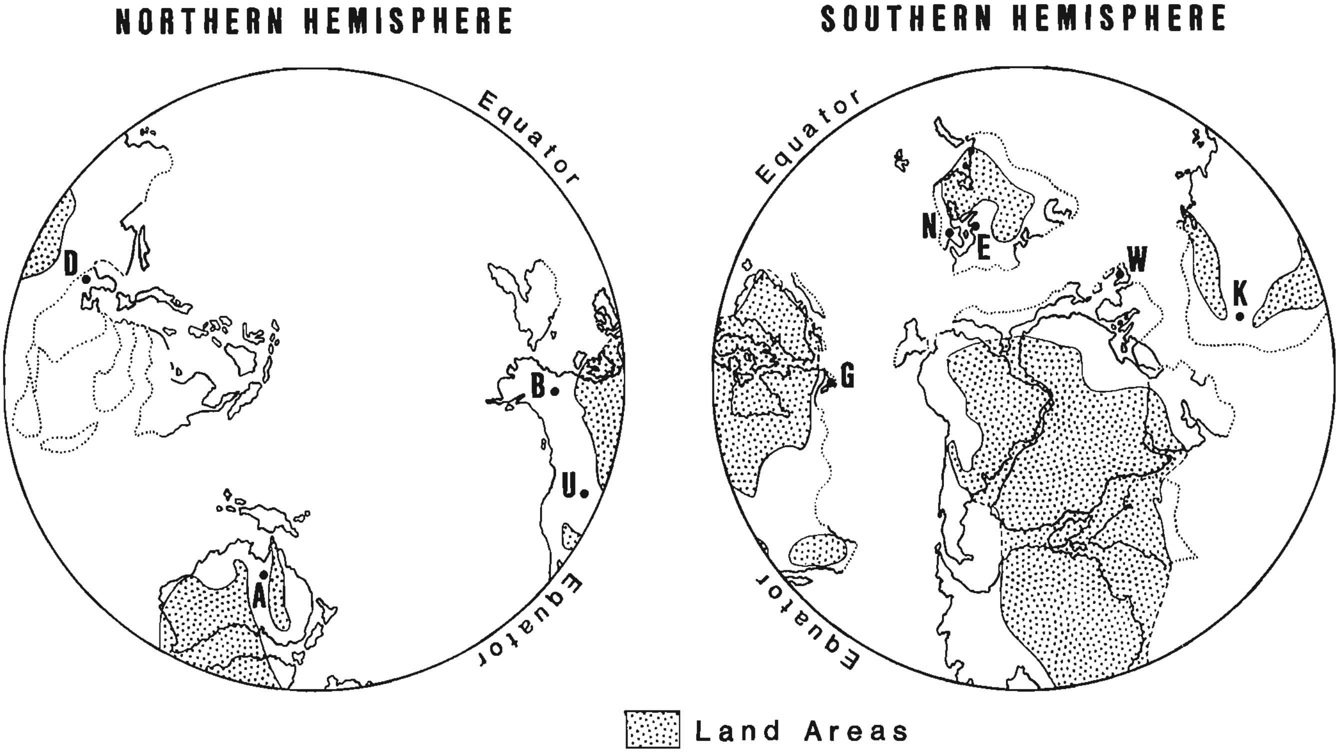


Figure 1. Distribution of continents in Cambrian-Ordovician boundary time and location of important stratigraphic sections (after Erdtmann, 1982). A, Black Mountain Section, central Australia; B, Brokenskull Section, northwestern Canada; D, Xiaoyangqiao Section, Dayangcha, China; E, Tõnismägi Section, Estonian S.S.R.; G, Green Point Section, Newfoundland, Canada; K, Batyrbay Section, Kazakh S.S.R.; N, Naersnes Section, Norway; U, Lava Dam Five Section, western U.S.A.; W, Bryn-llin-fawr Section, Wales.

2. To document the best stratigraphic sections spanning the boundary in the various parts of the world that have sequences through the boundary interval.
3. To consider recommendations for the biostratigraphic or other event to serve as the boundary and to decide on the most appropriate horizon.
4. To decide on an accessible type section within a suitable type region to serve as the permanent reference (global stratotype) for the Cambrian-Ordovician boundary.

Under the leadership of G. Henningsmoen (1974-6), W.T. Dean (1976-80), and B.S. Norford (since 1980), the members of the Working Group have conducted intensive studies of the boundary interval and have circulated within the Group summary papers of the most relevant stratigraphic sections in the various regions of the world. Many of these papers have been published in two summary volumes (Bassett and Dean, 1982; Norford and Webby, 1988).

Important stratigraphic sections considered to have potential to serve as the international stratotype have been examined and assessed by teams of the Working Group in Australia (1976), Wales, Norway, and Sweden (1980), western United States, western and eastern Canada (1981), the People's Republic of China (1983, 1986), and south-central U.S.S.R. (1984).

In 1983, by formal vote, the Working Group reached an almost unanimous consensus that the most appropriate horizon for the boundary was close to the base of the Tremadoc Series of North Wales and to stratigraphic horizons correlated with that base. In 1985, a plenary session was held in Calgary with the prime purpose of assessing a number of stratigraphic sections that had been proposed to serve as the international stratotype and if possible, to recommend a single section and a specific horizon within that section. A majority (57%) of the Group's voting members attended the Calgary meeting, together with many corresponding members. In addition to detailed presentations and discussions of stratigraphic sections in Canada (Broom Point-Green Point), China (Dayangcha), Norway (Naersnes), U.S.A. (Lava Dam), U.S.S.R. (Batyrbay and East Baltic), and Wales (Bryn-llin-fawr), comprehensive reviews were made of trilobite, conodont, and graptolite biostratigraphy, and of paleomagnetic and geochemical stratigraphic signatures. Three resolutions concerning general principles for selection of the stratotype were passed with substantial majorities, each subsequently ratified by the full voting membership (12 voted, data cited for each resolution):

1. "The group basically wishes to adopt the simple principle of driving a 'golden spike' in a stratigraphic section as the method of selection of the boundary" (12 for, 0 against, 0 abstained).
2. "We resolve that we will select conodonts as the primary guide for a horizon close to but below the first influx of nematophorous graptolites" (8 for, 3 against, 1 abstained).
3. "It is desirable that the stratotype section have potential for future study of paleomagnetism, geochemistry, and other nonbiological correlation tools" (8 for, 4 against, 0 abstained).

Detailed assessment (using the criteria shown in Table 1) and comparison of all of the stratigraphic sections that were discussed at the Calgary meeting led to focusing on two prime candidates to serve as the global stratotype. The neighbouring sections at Broom Point and Green Point in Newfoundland are well known, have been studied in detail and are well documented by many publications (see Barnes, 1988). They were assessed by a Working Group delegation in 1981. The Dayangcha Section (Xiaoyangqiao, Jilin Province) was discovered recently and since then has been intensively studied but was not available for examination when a delegation from the Working Group visited China in 1983 and assessed stratotype candidates at Wushan (Hebei Province) and at Dadoushan and Xiyangshan (Zhejiang Province). The Calgary meeting considered that both the Dayangcha and the Broom Point-Green Point sections appear to have virtually all the scientific attributes vital to a global stratotype but the documentation for Dayangcha was just beginning to be published. The meeting debated whether to recommend the Newfoundland section as the global stratotype on the grounds that the sequence has been well documented by intensive studies and is well qualified to serve as the international standard. However, the meeting decided against such a recommendation and suggested that a decision on selection of the global stratotype be delayed until more information became available on the Dayangcha Section.

Table 1. Evaluation criteria applied to the selection of international stratotypes

Historical appropriateness
Presence of good lithological succession
Continuity of sedimentation
Exposure quality and completeness
— lateral and vertical extent
— state of weathering
Structural complexity of outcrops
Favourable depositional facies
Thickness of strata
Presence of biological and nonbiological markers
Fossil abundance, diversity, continuity in sequence
— presence of key groups: condonts, graptolites, trilobites
Ease of external correlation to other regions
Thermal maturity — geochemical paleomagnetic studies
Aptitude for conversion and for sampling
Accessibility — ease, cost, political access

In 1986 a delegation from the Working Group inspected stratigraphic sections near Dayangcha (particularly the Xiaoyangqiao sections) and several days were devoted to scientific discussion at Hungjiang. Half of the voting members and several corresponding members participated. Very comprehensive documentation of the succession was compiled and published for the visit (Chen et al., 1986; Duan et al., 1986) and later summarized and updated (Chen et al., 1988).

PRESENT STATUS

Fifteen years of intense study of the boundary interval have provided extremely detailed knowledge of the rocks and faunas. Many questions have been resolved. The search for a suitable global stratotype section has focused on depositional environments of moderate depth, sited near the margins of cratonic platforms, providing a continuous record of sedimentation and including both benthic and pelagic faunas of diverse groups of fossils but especially conodonts, graptolites, trilobites, and palynomorphs. Concern for the nonbiological attributes of stratotypes requires low thermal maturity and minimal weathering so that isotope dating, ocean geochemistry, eustatic changes in sea level, paleomagnetic polarity, and other inorganic methods of correlation can be effective. Requirements for scientific reference and continuing study by future generations of specialists require lateral continuity of exposure to provide material for additional studies, together with appropriate methods of conservation to prevent waste and vandalism. Access must be monitored but must provide for scientists from all parts of the world to be able to visit a global stratotype and to conduct appropriate scientific studies.

As a result of the 1985 Calgary plenary session, the Working Group has focused its attention on the Newfoundland (Green Point) and Dayangcha (Xiaoyangqiao) sections but reconsideration of other sections could lead to their renomination as potential stratotypes.

Uncertainties that remain include: the paleomagnetic signatures of the two sections; the presence and significance of any scouring in the Newfoundland sections; replication of the Dayangcha succession by secondary sections within the region; the reliability of the *Cordylodus lindstromi* Zone as a biostratigraphic unit; and the apparently later appearance of some taxa of the cordylodan evolutionary lineage in shallow water sequences in comparison to slope and deep water environments.

Table 2. Correlation of zonal schemes: conodonts, graptolites and trilobites

Conodont Zones (North American Platform)		Conodont Zones (Slope)		Graptolite Assemblages	Trilobite Zones (North American Platform)	
Cordylodus angulatus		*Cordylodus angulatus*		Assemblage 2B	*Symphysurina*	*S. bulbosa*
				Assemblage 2A		*S. brevispicata*
Cordylodus lindstromi		*Cordylodus lindstromi*		Assemblage 1		
Cordylodus intermedius	*Cl. hintzei*					
	Hr. simplex			No Pelagic Graptolites	*Parakoldinoidia*	*P. stitti*
Cordylodus proavus	*Cl. elongatus*	*Cordylodus intermedius*				
	Fx. inornatus	*Cordylodus caboti*				
	Hr. hirsutus	*Cordylodus proavus*				*P. depressa*
					Eurekia apopsis	
Eoconodontus	*Cb. minutus*	*Eoconodontus notchpeakensis*	*C. primitivus*		*Saukia*	*Saukiella serotina*
			E. alisonae			
	E. notchpeakensis		*E. notchpeakensis*			*Saukiella junia*
Proconodontus muelleri		*Proconodontus muelleri*				*Rasettia magna*
Proconodontus posterocostatus		*Proconodontus posterocostatus*				
Proconodontus tenuiserratus					*Saratogia*	
Sources Miller 88		Barnes 1988		Erdtmann 1988	Barnes 1988	

Abbreviations: *C., Cordylodus; Cb., Cambrooistodus; Cl., Clavohamulus; E., Eoconodontus; Fx., Fryxellodontus; Hr., Hirsutodontus; P., Porakoldinoidea; S., Symphysurina*

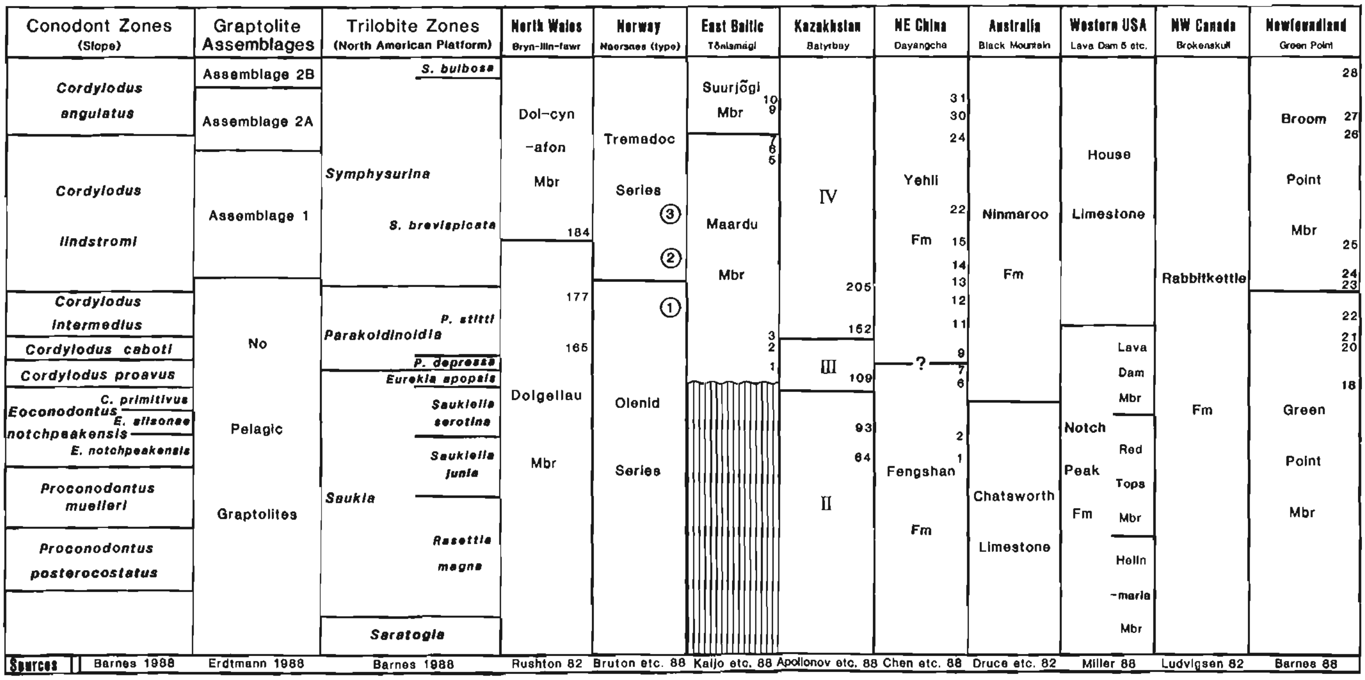
The earliest appearance of planktic (nematophorous) graptolites is a significant and readily observable event that has been recognized in shaly facies in many regions of the world. This evolutionary leap seems to have exploited a vacant ecological niche and resulted in rapid global distribution of the new forms. Where conodont faunules have been extracted from the shaly sequences and associated carbonates, this first appearance has been documented as within the *Cordylodus lindstromi* Zone. An exception is in Estonia where this event (within Bed 5 of the Maardu Member of the Kallavere Formation) is interpreted to be older and within the *C. proavus* Zone (Kaljo et al., 1988). However, there is a lack of consensus among conodont specialists on the taxonomic implications of some elements of *Cordylodus*, whether they represent *C. proavus* or *C. lindstromi*. The latter interpretation is followed in Table 3, which shows Bed 5 as high in the *C. lindstromi* Zone.

Recent studies have elucidated that the critical conodont zones *C. proavus, C. caboti, C. intermedius*, and *C. lindstromi*, are to a large extent based on evolutionary stages within the phylogeny of *Cordylodus* (Bagnoli et al., 1987; Viira et al., 1987). Each species is represented by a number of elements, whose morphologies changed at different rates. The zones are assemblage zones named for the first appearances of index species but the name bearers and other key taxa commonly continue stratigraphically higher into succeeding zones. Limits between species can be subjective and lead to overlap between the taxonomic concepts used by different specialists. Conformity of recognition of these four taxa is vital to fine biostratigraphic subdivision of the boundary interval. Such conformity has yet to be achieved but detailed taxonomic papers now being published will aid in the solution of this critical problem.

Another factor complicating the simplicity of this sequence of four conodont assemblage zones is the first appearances of some of the index species in shallow water platform environments apparently later than their first appearances in deeper water. Some of these discrepancies may be due to the different taxonomic interpretations of various elements by different specialists discussed previously. Nevertheless, in Table 2 it can be seen that such discrepancies result in the bases of the *C. intermedius* Zone and the *C. lindstromi* Zone being slightly younger on the North American Platform than in the slope facies. The onset of *C. proavus* seems to have been more instantaneous unless elements of *C. primitivus* have been interpreted as *C. proavus*, in which case a similar delay existed in the colonization of shallow environments.

Detailed correlations between nine of the most important stratigraphic sections through the boundary interval are shown in Table 3. Although no radiometric scale is attempted, the total span of the interval *C. proavus* Zone through *C. lindstromi* Zone is probably less than three million years (interpreted from Gale's, 1985, estimate of 10 million years for the Tremadoc Series). A radiometric age of 500.7 ± 7.4 million years, based on rubidium-strontium analysis of clay minerals, has been established within the *C. proavus* Zone (Table 3, sample 9) at Dayangcha (Chen et al., 1988).

Table 3. Correlation of nine important stratigraphic sections. Figure 1 shows locations. Numbers in right hand parts of columns follow published literature and have different meanings in different columns: metres within section (Bryn-llin-fawr and Batyrbay); stratigraphic position of fossiliferous nodules (Naersnes type section: 1, Acerocare concretion at 2.2 m; 2, *Boeckaspis* concretion at 3.2 m; 3, *Jujuyaspis* concretion at 4.4 m; numbering system for samples from section (Tônismägi and Dayangcha); stratigraphic units within section (Green Point).



Column	Section	Units and numbers (top to bottom)	Sources
Conodont Zones	(Slope)	*Cordylodus angulatus* *Cordylodus lindstromi* *Cordylodus intermedius* *Cordylodus caboti* *Cordylodus proavus* *Eoconodontus notchpeakensis*: *C. primitivus*, *E. elisonae*, *E. notchpeakensis* *Proconodontus muelleri* *Proconodontus posterocostatus*	Barnes 1988
Graptolite Assemblages		Assemblage 2B Assemblage 2A Assemblage 1 No Pelagic Graptolites	Erdtmann 1988
Trilobite Zones	(North American Platform)	*Symphysurina*: *S. bulbosa*, *S. brevispicata* *Parakoldinoidia*: *P. stitti*, *P. depressa* *Eurekia apopsis* *Saukia*: *Saukiella serotina*, *Saukiella junia*, *Rasettia magna* *Saratogia*	Barnes 1988
North Wales	Bryn-llin-fawr	Dol-cyn-afon Mbr 184 177 165 Dolgellau Mbr	Rushton 82
Norway	Naersnes (type)	Tremadoc Series ③ ② ① Olenid Series	Bruton etc. 88
East Baltic	Tônismägi	Suurjõgi Mbr 10, 9 Maardu Mbr 7, 6, 5, 3, 2, 1	Kaljo etc. 88
Kazakhstan	Batyrbay	IV 205, 162 III 109 II 93, 64	Apollonov etc. 88
NE China	Dayangcha	Yehli Fm 31, 30, 24, 22, 15, 14, 13, 12, 11, 9 ? Fengshan Fm 7, 6, 2, 1	Chen etc. 88
Australia	Black Mountain	Ninmaroo Fm Chatsworth Limestone	Druce etc. 82
Western USA	Lava Dam 5 etc.	House Limestone Notch Peak Fm: Lava Dam Mbr, Red Tops Mbr, Hellnmaria Mbr	Miller 88
NW Canada	Brokenskull	Rabbitkettle Fm	Ludvigsen 82
Newfoundland	Green Point	Broom Point Mbr 28, 27, 26, 25, 24, 23 Green Point Mbr 22, 21, 20, 18	Barnes 88

Paleomagnetic studies of the boundary interval are at a preliminary stage. Data have been gathered from Newfoundland (Broom Point), Texas (Lange Ranch Section), Kazakhstan (Batyrbay), and China (Dadoushan and Dayangcha) by J.L. Kirschvink and his co-workers. Reversed polarity seems to be characteristic of most of Late Cambrian and Early Ordovician times, but preliminary results from Batyrbay and Dadoushan indicate that a short episode of normal polarity may occur near the boundary interval (Ripperdan and Kirschvink, pers. comm., 1987). If these results are substantiated by the other sampled sections, the polarity signature of stratigraphic sections will provide a powerful tool for correlation of the boundary.

Substantial progress has been achieved by the Working Group during its history. Minor uncertainties remain in the details of the conodont biostratigraphic zonation but resolution of these problems is close to being achieved. This will allow decisions on the designation of a global stratotype and selection of a precise horizon within it to serve as the base of the Ordovician System.

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This paper builds on the contributions from all members of the Working Group. In particular, M.K. Apollonov, C.R. Barnes, D.L. Bruton, Chen Jun-yuan, B.-D. Erdtmann, D.L. Kaljo, J.F. Miller, A.W.A. Rushton, and J.H. Shergold have contributed data that the author has interpreted to provide the present analysis.

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Cambrian-Ordovician boundary beds in the U.S.S.R.

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Abstract

Cambrian-Ordovician boundary beds are widely distributed in many regions of the U.S.S.R. and are represented by sediments of various facies types. Correlations between facies belts are the subject of discussions in some cases. In the East European Platform, Urals, Pay-Khoy, Vaigach, South Novaya Zemlya, and in northeastern U.S.S.R., the base of the Ordovician coincides with the base of a transgressive complex. In Kazakhstan and the Siberian Platform, the change in facies is recorded close to the systemic boundary. Detailed conodont biostratigraphy of the boundary interval is established only in Kazakhstan, Estonia, and the Leningrad area. Graptolites are rare at the boundary interval; the exact level of the first appearance of nematophorous graptolites is open to dispute. The main group used for subdivision and correlation of the boundary interval is trilobites. The best horizon for the Cambrian-Ordovician boundary in the U.S.S.R. is the base of the conodont Cordylodus proavus *Zone. The widespread transgression and drastic changes in faunal composition are connected with this horizon. Geochemical anomalies and reversal of the paleomagnetic polarity are known at this horizon in Kazakhstan. The main reasons for global changes at the Cambrian-Ordovician boundary are probably the strong activation of continental plate movements and related changes in the global system of oceanic circulation.*

Résumé

Les couches limites cambriennes-ordoviciennes sont largement répandues dans de nombreuses régions de l'URSS et sont représentées par des sédiments de différents faciès. Les corrélations entre les zones de faciès font, dans certains cas, l'objet de discussions. À la plate-forme est-européenne, dans l'Oural, les collines Pay-Khoy, l'île Vaigach, la Nouvelle-Zemble du Sud et le Nord-Est de l'URSS, la base de l'Ordovicien coïncide avec la base d'un complexe transgressif. Dans le Kazakhstan et la plate-forme sibérienne, le changement de type de faciès s'observe près de la limite systémique. Des échelles détaillées de conodontes à l'intervalle limite n'ont été élaborées que pour le Kazakhstan, l'Estonie et la région de Leningrad. Les graptolites sont rares à l'intervalle limite et le niveau exact de la première apparition de graptolites nématophores est discutable. Le principal groupe utile pour une subdivision et une corrélation de l'intervalle limite est celui des trilobites. En URSS, les niveaux correspondant le mieux à la limite cambrienne-ordovicienne est la base de la zone à conodontes Cordylodus proavus. *La vaste transgression et les changements abrupts de composition faunique sont liés à ce niveau. Les anomalies géochimiques et l'inversion de la polarité paléomagnétique ont été déterminées à ce niveau dans le Kazakhstan. La principale cause des changements globaux à la limite cambrienne-ordovicienne est probablement la forte activation des mouvements de plaques continentales et les changements associés du système global de la circulation océanique.*

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GENERAL REVIEW OF THE CAMBRIAN-ORDOVICIAN BOUNDARY PROBLEM IN THE U.S.S.R.

In the U.S.S.R., the Cambrian-Ordovician boundary has been under study and discussion for several decades. In recent years it has been the subject of heightened interest, as elsewhere in the world. In many regions of the U.S.S.R., boundary beds are characterized by very different facies and environmental features. The fauna is rather endemic in many cases, resulting in many difficulties in studying the boundary interval. At the same time, some well studied successions can serve as a good basis for further study of the boundary problem (Kaljo et al., 1986; Apollonov et al., 1988). The exact correlations of boundary beds and the precise definition of the Cambrian-Ordovician boundary is important.

Until recently, the Cambrian-Ordovician boundary in the U.S.S.R. was traditionally drawn at the base of the *Dictyonema flabelliforme* graptolite Zone. The level was officially accepted by the Geological Survey of the U.S.S.R. (Sokolov, 1977). However, precise direct correlation of Cambrian-Ordovician boundary beds of most regions of the U.S.S.R. with the graptolite zonation is practically impossible. *Dictyonema flabelliforme* Zone graptolites are known in the U.S.S.R. only in the northeastern part of the East European Platform. In most parts of the U.S.S.R. (Urals, Kazakhstan, Middle Asia, Altay-Sayan region, Siberian Platform, northeastern U.S.S.R.) the Cambrian-Ordovician boundary is drawn mainly in trilobite-bearing successions.

An interesting situation arose in the U.S.S.R. when the Cambrian Commission of the Interdepartmental Stratigraphic Committee (ISC) of the U.S.S.R. accepted (conditionally) the base of the *Trisulcagnostus trisulcus* trilobite Zone in the Kyrshabakty section, South Kazakhstan as the Cambrian-Ordovician boundary (Sokolov, 1983), proceeding from the assumption that the level is correlated with the base of the Tremadocian, as was supposed by Ergaliev (1980). However, it is now well known that the level is synchronous with the middle part of the *Westergaardodina amlicava* conodont Zone and the base of the Trempealeauan. Concurrently, the Ordovician Commission of the ISC decided to draw the boundary at the base of the *Cordylodus proavus* Zone. And so, the Cambrian and Ordovician Commissions proposed quite different levels for the systemic boundary. The distance between these levels is equal to the Trempealeauan Stage of North America or the Batyrbaian Stage (series) of Malyi Karatau. The situation needs improving of course.

The boundary beds that have heterogeneous facies are distributed widely over the vast territories of the U.S.S.R. Their study has produced much data for reconstruction of the sedimentation and faunal evolution in different environments through the boundary time interval. In practically all regions,

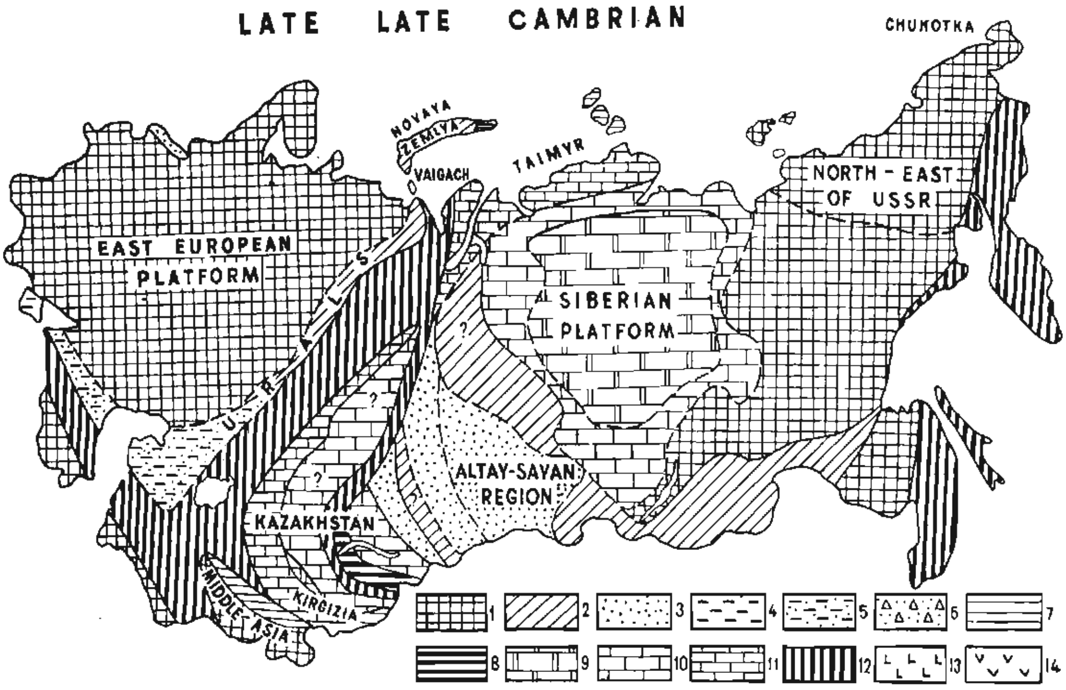


Figure 1. Schematic paleotectonic map of the U.S.S.R. territory (without palinspastic reconstruction) for the late Late Cambrian. 1) Precambrian basement of platforms and massifs; 2) folded structures of Late Precambrian and Cambrian; 3-5) shallow water basins with clastic sediments: 3) mainly sand; 4) mainly shale; 5) sand and shale; 6-8) deep water basins: 6) turbidites and olistostromes; 7) mainly shale; 8) mainly chert; 9-11) basins with mainly carbonate sedimentation: 9) shallow water carbonates of middle parts of platforms; 10) moderately shallow water carbonates of outer and middle parts of platforms and massifs; 11) deep water limestone of outer parts of platforms and massifs; 12) oceans and basins of oceanic type; 13) basalt; 14) andesites.

a major transgression is evident at the boundary interval. The transgression probably began in the latest Late Cambrian - in *Lotagnostus hedini* Zone time.

Extensive land areas existed in the late Late Cambrian in the centre of major continental plates (East European platform and northeastern U.S.S.R.) (Fig. 1). In the Early Ordovician, these were transgressed by a shallow sea (Fig. 2). In the basins of Siberia and some regions of Kazakhstan, shallow water deposits were replaced by deeper water sediments. Carbonates or alternating carbonate-shale deposits were replaced by dark shale in the basins at the edges of plates (Taymyr, North Novaya Zemlya, Bolshoy Karatau, Ulutau) and elsewhere by dark chert (Sarykum in Kazakhstan). In the oceanic basins, jasper appeared in the latest Late Cambrian (Erementau-Chu-Ili Zone in Kazakhstan). Basalts were formed locally.

Three faunal groups - graptolites, trilobites, and conodonts - are used mainly for the correlation and definition of the Cambrian-Ordovician boundary level in the U.S.S.R. The use of graptolites is limited for several reasons: boundary beds with graptolites are recorded from few areas; no zonation exists below the base of the *Dictyonema flabelliforme* Zone; the horizon of the influx (or of the first appearance) of nematophorous graptolites is not fixed exactly; other fauna are rarely found together with graptolites.

Trilobites are the most common fossil in the Cambrian-Ordovician boundary beds in many regions of the U.S.S.R. They are often endemic in shallow water environments, and correlation between facies realms and provinces is commonly difficult. Within provinces, the correlation is exact and usually adequate. Some relatively deep water trilobites provide proper interprovincial correlation. The first appearance of such widespread, well known, and easily identified trilobites such as *Hysterolenus*, *Macropyge* (*Macropyge*), *Glaphurus*, *Shumardia*, *Apatocephalus*, *Nyaya*, and others marks an important biostratigraphic level, which is correlated with the base of the conodont *Cordylodus proavus* Zone. The horizon of the first appearance of these trilobites can be used as one of the criteria for the definition of the Cambrian-Ordovician boundary.

Conodonts are the most eurytopic group and the conodont zonal scale is considered internationally to be the basis for the definition of the Cambrian-Ordovician boundary (Derby, 1986). According to the U.S.S.R. data, mainly from the Batyrbai section, South Kazakhstan, and sections in the Baltic region, the base of the *Cordylodus proavus* Zone is the best horizon for the systemic boundary (Apollonov and Chugaeva, 1982; Spizharskyi, 1986; Apollonov et al., 1988).

The base of the *C. proavus* Zone is the the only "natural" boundary at the boundary interval from the base of the *C. proavus* Zone to the base of *C. lindstromi* Zone. Eustatic events, geochemical anomalies and changes in paleomagnetic polarity are recorded at this level (Apollonov et al., 1988). Ecosystems were changed virtually completely and new biomeres were formed at that level throughout the U.S.S.R. The main reason for such catastrophic global events may be increased activation of the continental plate movements, accompanied by great transgression and changes in oceanic circulation. These paleogeographic changes at the Cambrian-Ordovician boundary time are shown in Figures 1 and 2.

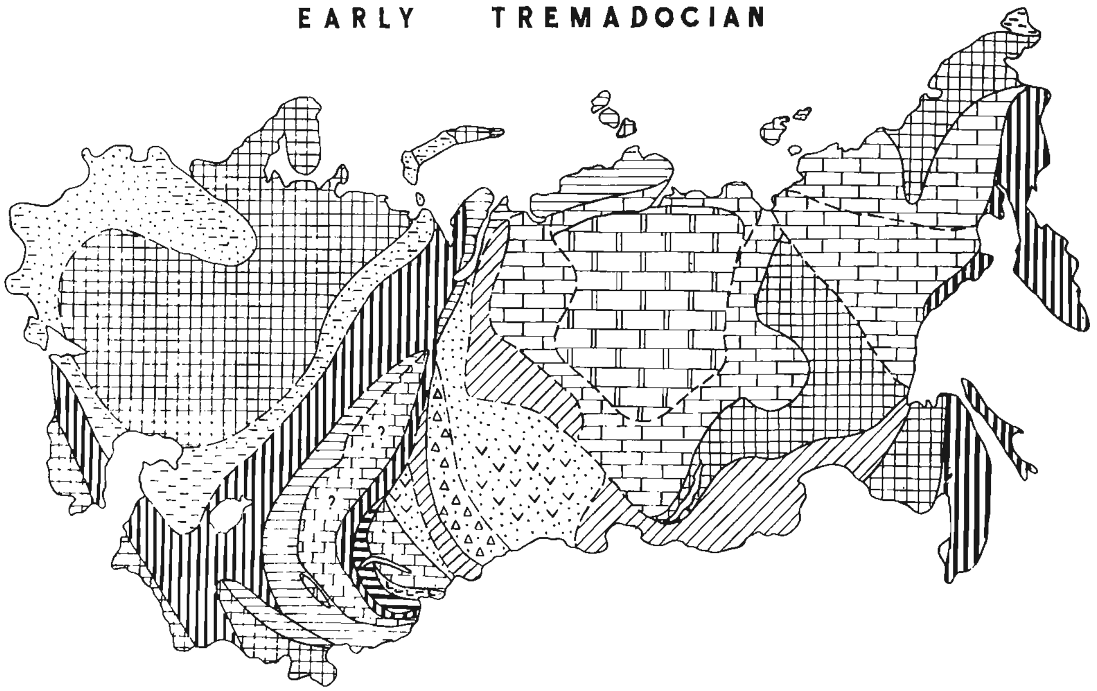


Figure 2. Schematic paleotectonic map of the U.S.S.R. territory (without palinspastic reconstruction) for the early Tremadocian. Legend as for Figure 1.

Two successions are particularly interesting for studies of the Cambrian-Ordovician transition beds: in Estonia (Kaljo et al., 1986) and in the Malyi Karatau Range, South Kazakhstan. The Batyrbay section in Malyi Karatau is proposed as a potential Cambrian-Ordovician boundary stratotype (Apollonov et al., 1988).

The analysis of various data on the numerous boundary sections in many regions of the U.S.S.R. may be useful for resolving the international Cambrian-Ordovician boundary problem.

THE EAST EUROPEAN PLATFORM

In the Soviet part of the East European (Russian) Platform, the Pakerort Horizon was erected for the lowermost part of the Ordovician (Alikhova, 1960; Männil, 1966; Grigelis, 1978). The horizon correlates with the lower Tremadoc of Wales. Rocks of the horizon were usually deposited on an eroded surface of Lower and Middle Cambrian rocks. In the Leningrad region and Estonia, an Upper Cambrian succession with conodonts of the "*Westergaardodina*" Zone was discovered recently (Borovko et al., 1980, 1984; Kaljo et al., 1986).

The Cambrian-Ordovician boundary in the East European Platform is traditionally drawn at the base of the Pakerort Horizon (Alikhova, 1960; Männil, 1966; Sokolov, 1977), although other views also exist. The history of the problem was discussed by Balashova and Balashov (1971).

After discovery of conodonts at the boundary interval in the Leningrad region (Borovko et al., 1980), the suggestion was made to draw the Cambrian-Ordovician boundary there, at the base of the *Cordylodus proavus* Zone (Borovko et al., 1984). This level was accepted as the systemic boundary for the East European (Russian) Platform in the official "Decision of the Interdepartmental Regional Stratigraphic Conference on the Cambrian of the Russian Platform (Vilnius, 1983)" (Spizharskyi, 1986).

Recently, new data on the most complete Cambrian-Ordovician boundary successions in the Baltic-Ladoga clint area (North Estonia and the Leningrad region) were analyzed (Kaljo et al., 1986). The numerous good outcrops enable detailed studies of the conodont and graptolite successions in the boundary interval (Table 1). At the level of the *Cordylodus andresi* Zone, graptolites were not found. However, the authors did not exclude the possibility that the *Dictyonema sociale* Subzone may be correlated, at least partly, with the *Cordylodus proavus* Zone. It is clear that the graptolite *Dictyonema flabelliforme* (s.l.) appears not later than the *Cordylodus intermedius* Zone. The bases of the *Cordylodus andresi* and *C. intermedius* zones were discussed as the probable Cambrian-Ordovician boundary for the East European Platform (Kaljo et al., 1986). At present, stratigraphers in Estonia and Leningrad prefer the base of the *C. proavus* Zone as the boundary level (V. Viira, pers. comm., 1988).

The sections in the East European Platform area provide very important data for interpreting the relationship of the first nematophorous graptolites with the conodont zonation at the Cambrian-Ordovician boundary interval. Shortcomings of those sections are the presence of several hiatuses, and the lack of trilobites and other fauna, except graptolites, conodonts, and inarticulate brachiopods.

THE URALS

The problem of the exact position of the Cambrian-Ordovician boundary in the Urals is not yet solved, in spite of many years of studying potential boundary sections. The problem was discussed at the conference on the Urals stratigraphy in 1977 (Breivel et al., 1980).

In the Urals, Lower Ordovician (and supposed Upper Cambrian) deposits are divided into tectonic zones - Belsk-Yeletsk, Sakmara-Lemva, Magnitogorsk-Schuchinsk, East Uralian, and Trans Uralian zones. The zones stretch from north to south over some thousands of kilometres. The boundary beds are best represented on the west slope of the Urals, in the Belsk-Yeletzk and Sakmara-Lemva zones. On the east slope in the Magnitogorsk-Schuchinsk Zone, the boundary beds are unrecorded, and in the East Uralian and Trans Uralian zones, they are only tentatively identified.

Everywhere on the west slope of the Urals, a thick complex of unmetamorphosed or slightly metamorphosed shallow water terrigenous deposits (quartz-feldspar sandstone, siltstone, mudstone, etc.) with an excellent Tremadocian fauna is known (e.g., Balashova, 1961). In the Sakmara-Lemva Zone, volcanics are recorded (Breivel et al., 1980). This transgressive complex rests on metamorphic rocks of Precambrian, and Lower and Middle Cambrian age. The presence of the Upper Cambrian is a particular problem. Ancygin et al. (1970) described the Khmelyov Horizon in the South Urals (Kidryas, Kosistek, Kayala, Abitah areas), and considered it to be of Upper Cambrian age. The horizon consists of clastic terrigenous deposits, more typical of the Ordovician. Abundant articulate brachiopods, *Billingsella akbulakensis* Andreeva, inarticulate brachiopods, *Obolus* sp., *Lingulella* sp. and others, and rare trilobites *Kujandaspis kujandensis* Ivshin, *Olentella* sp., identified by N.Ya. Acygin, are known in the horizon. But Korinevsky (1978) considered that faunal identifications need revision and that the deposits are of Ordovician age. Recently, he discovered the Tremadocian trilobites *Ceratopyge*, *Shumardia*, and *Acerocarina* in the deposits of the Khmelyov Horizon with *Billingsella akbulakensis* in the Kayala River area (Korinevsky, pers. comm., 1988).

Stratigraphic divisions of the Ordovician and Upper Cambrian, proposed by Ancygin et al. (1970) and Varganov et al. (1973) were adopted by the Geological Survey for all regions in the Urals (Breivel et al., 1980). The Upper Cambrian can be distinguished in the lower part of the transgressive Ordovician complex in the Middle and Polar Urals, but is without fauna and only weakly correlated with the South Urals.

In the Belsk-Yeletsk Zone, the lowest part of the Ordovician, bearing the brachiopod *Angarella* (supposedly Tremadocian-Arenigian), is known only in the Polar Urals. The deposits rest unconformably on pre-Ordovician rocks.

The most complete succession close to the boundary occurs in the Sakmara-Lemva Zone. In the Polar Urals, in the Paga, Pogureisk-Lemva and Sangur areas, Lower Ordovician deposits are divided into supposedly lower Tremadocian (Pagata and Pogurey formations) and upper Tremadocian-Arenigian with *Apatokephalus* and other taxa (Kibata and Grubeya formations). In the Paga area near the Pokoinitza Shor stream, the oldest Ordovician conodonts in the Urals are found (*Acodus*, *Acontiodus*, *Cordylodus*, *Drepanodus*, *Oistodus*, *Oneotodus*, and *Scandodus*) (Nasedkina, 1985). In the middle part of the Sakmara-Lemva Zone, undivided Tremadocian and Arenigian formations (Kolpakov, Telposs and others) that yield brachiopods (*Tritoechia* and others) are known.

Well documented, highly fossiliferous Tremadocian sandstone and mudstone containing the brachiopods *Tritoechia*, *Alimbella*, *Altorthis*, *Medessia*, *Finkelnburgia*, *Imbricatia*, *Obolus*, and the trilobites *Dolgeuloma*, *Pseudokainella*, *Macropyge* (*Macropyge*), *Leimitzia*, and others, as well as mudstone members with concretions bearing typical upper Tremadocian trilobites, such as *Ceratopyge forficula* Sars, *Shumardia oelandica* Moberg, and *Apatokephalus serratus* (Boeck) are known to the south of the Sakmara-Lemva Zone (Balashova, 1961; Varganov et al., 1973). Stratotypes of regional Uralian Kidryas (lower Tremadocian) and Kolnabuk (upper Tremadocian) horizons were established here (Varganov et al., 1973).

Graptolites in the boundary interval are not known. Conodonts are very rare (Nasedkina, 1985) and cannot be used for the division and correlation of the boundary interval. Hence, the base of the thick transgressive Ordovician complex in the Urals is approximately near the Cambrian-Ordovician boundary. More exact definition of the level is currently impossible. The question of whether the base of the complex is sychronous over all the Urals is not resolved. If the transgression over all the territory of the East European Platform and the Urals had the same cause, it may be possible to correlate the base of the transgressive Ordovician complex of the Urals with the base of the Pakerort Horizon, or roughly with the base of the *C. proavus* Zone.

PAY-KHOY, VAIGACH, NOVAYA ZEMLYA

The Lower Ordovician successions of Pay-Khoy, Vaigach, and South Novaya Zemlya are well documented and have been studied in detail (Bondarev et al., 1970). A clastic-carbonate, rather shallow water, complex (80-100 m thick) rests trangressively and unconformably on pre-Ordovician rocks. The regional Sokolii Horizon corresponds to the complex. The succession is divided into five beds (Bondarev et al., 1970): 1) basal beds; 2) *Syntrophopsis magna* and *Dikelokephalina* beds with *Imbricatia* sp. cf. *I. lesnikovae* Andreeva, *Syntrophopsis magna* Ulrich and Cooper, *Altorthis* sp., *Nyaya paichoica* Bursky, *Hystricurus* sp., *Nyaya* sp., *Tersella*? sp., *Dikelokephalina* sp., *Pliomeroides* sp.; 3) *Nyaya* and *Tersella* beds with *Imbricatia* sp. cf. *I. russanovi* Bondarev, *Syntrophopsis magna* Ulrich and Cooper, *Altorthis* sp., *Apatokephalus pecten* Wiman, *Apatokephalus*, *Tersella*, *Asaphellus*, *Ceratopyge*, *Pliomeroides*, *Nyaya*; 4) *Apatokephalus serratus* beds with *A. serratus* (Boeck), *Nileus limbatus* Brögger, *Pliomeroides primigenus* (Angelin), *Geragnostus*, *Geragnostella*; and 5) *Megalaspides* beds with species of *Megalaspides*, *Apatokephalus*, *Nileus*, *Niobella*, *Cybelurus*.

Bondarev et al. (1970) correlated the basal beds with the Pakerort Horizon of the East European Platform. Beds 2-4 inclusive are correlated with the *Ceratopyge* Horizon of the East European Platform. *Medalaspides* beds are compared with the lower Arenigian. Hence, the transgressive complex with a typical Tremadocian fauna rests on an erosional surface of the folded pre-Ordovician, both here, and in the East European Platform and the Urals. It is evident that the main tectonic and sedimentological processes were generally similar over the entire East European Platform.

A different situation exists in northern Novaya Zemlya. According to R.F. Sobolevskaya (pers. comm., 1988), at the very north of the island, 70 km southeast of the Zhelaniya Cape, the monotonous deep water Olen Formation occurs. It consists of dark shale with carbonate concretions, with thin limestone strata in the lower part of the succession. The thickness of the formation is 450 m. In the lower 70 m, the trilobites *Lotagnostus trisectus* (Salter), *L. hedini* (Troed.) and others of the uppermost Cambrian (*Lotagnostus hedini* Zone of Kazakhstan and China) were found. About 120 m above, the graptolites *Dictyonema angelini* Bulman, *Clonograptus tenellus* Linnarson and others of the middle Tremadocian are known. In the uppermost part of the succession, Arenigian graptolites are present.

Thus, the Cambrian-Ordovician boundary in the section can be drawn roughly in the lower part of the barren interval.

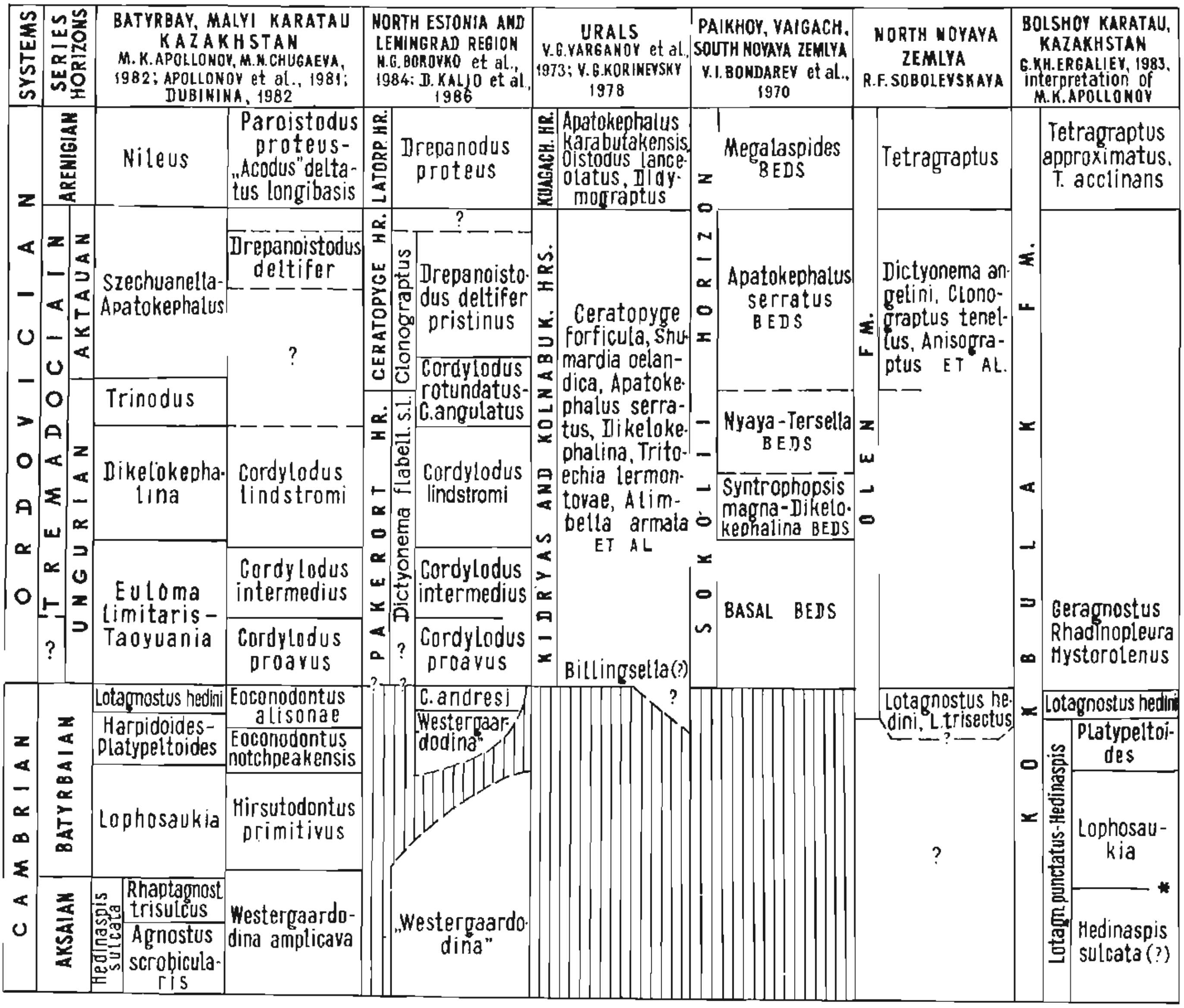
KAZAKHSTAN

In Kazakhstan, the Cambrian-Ordovician boundary beds probably were deposited over vast territories. However, tectonic deformation and sparse faunas in some facies have resulted in few good fossiliferous and uninterrupted boundary successions (Nikitin, 1972; Apollonov et al., 1981; Apollonov and Chugaeva, 1983; Ergaliev, 1983; Apollonov et al., 1988).

Some tectonic facies zones are distinguished according to facies types of Upper Cambrian and Lower Ordovician sediments and tectonic environments (Nikitin, 1972, 1973; Apollonov and Patalakha, 1981; Nikitin et al., 1986). The author considers that the overall framework of the Upper Cambrian and Lower Ordovician structure of Kazakhstan comprised two large ancient sialic plates (terrains): the Kokchetav-Chu and Balkhash massives. The plates were divided by the Erementau-Chu-Ili linear Zone (eugeocline) and bounded by the Ishim-Karatau Zone (miogeocline) on the west and by the Chingiz-Tarbagatai Zone (eugeocline) on the east.

In the Ishim-Karatau Zone, deep water limestone, shale, and chert were deposited during the Cambrian-Ordovician boundary interval (Nikitin, 1972, 1973). In the sections of the Ulutau and Bolshoy Karatau ranges (Table 1), the Upper

Table 1. Correlation chart for the Cambrian-Ordovician boundary beds of the U.S.S.R. Asterisks mark the levels proposed as the Cambrian-Ordovician boundary by the various authors



SYSTEMS	SERIES	HORIZONS	BATYRBAY, MALYI KARATAU KAZAKHSTAN M. K. APOLLONOV, M.N. CHUGAEVA, 1982; APOLLONOV et al., 1981; DUBININA, 1982		NORTH ESTONIA AND LENINGRAD REGION N.G. BOROVKO et al., 1984; D. KALJO et al., 1986			URALS V.G. VARGANOV et al., 1973; V.G. KORINEVSKY 1978		PAIKHOY, VAIGACH, SOUTH NOYAYA ZEMLYA V.I. BONDAREV et al., 1970		NORTH NOYAYA ZEMLYA R.F. SOBOLEVSKAYA		BOLSHOY KARATAU, KAZAKHSTAN G.KH. ERGALIEV, 1983, interpretation of M.K. APOLLONOV		
ORDOVICIAN	ARENIGIAN		Nileus	Paroistodus proteus-„Acodus"deltatus longibasis	LATORP HR.	Drepanodus proteus		KUAGACH. HR.	Apatokephalus karabutakensis, Oistodus lanceolatus, Didymograptus	SOKÖLII HORIZON	Megalaspides BEDS	OLEN FM.	Tetragraptus	KOKBULAK FM.		Tetragraptus approximatus, T. acclinans
	TREMADOCIAN	AKTAUAN	Szechuanella-Apatokephalus	Drepanoistodus deltifer ?	CERATOPYGE HR.	Clonograptus	? Drepanoistodus deltifer pristinus	KIDRYAS AND KOLNABUK. HRS.	Ceratopyge forficula, Shumardia oelandica, Apatokephalus serratus, Dikelokephalina, Tritoechia lermontovae, Alimbella armata ET AL		Apatokephalus serratus BEDS		Dictyonema angelini, Clonograptus tenellus, Anisograptus ET AL.			
		UNGURIAN	Trinodus		PAKERORT HR.	Dictyonema flabell. s.l.	Cordylodus rotundatus-C.angulatus				Nyaya-Tersella BEDS					
			Dikelokephalina	Cordylodus lindstromi			Cordylodus lindstromi				Syntrophopsis magna-Dikelokephalina BEDS					
			Euloma limitaris-Taoyuania	Cordylodus intermedius			Cordylodus intermedius				BASAL BEDS					Geragnostus Rhadinopleura Hystorolenus
	?			Cordylodus proavus		?	Cordylodus proavus		Billingsella(?)							
CAMBRIAN	BATYRBAIAN		Lotagnostus hedini	Eoconodontus alisonae	?	?	C. andresi		?				Lotagnostus hedini, L. trisectus ?			Lotagnostus hedini
			Harpidoides-Platypeltoides	Eoconodontus notchpeakensis			Westergaardodina"								Lotagn. punctatus-Hedinaspis	Platypeltoides
			Lophosaukia	Hirsutodontus primitivus									?			Lophosaukia — *
	AKSAIAN	Hedinaspis sulcata	Rhaptagnost trisulcus	Westergaardodina amplicava			„Westergaardodina"									Hedinaspis sulcata (?)
			Agnostus scrobicularis													

Cambrian is represented by a dark, thin-bedded limestone member, bearing trilobites of the *Lotagnostus punctatus-Hedinaspis* and *Lotagnostus hedini* zones (Ergaliev, 1983). This member is replaced above by an interbedded limestone and shale member. In Bolshoy Karatau, in the lower part of the member, the trilobite *Hysterolenus* is recorded (Nikitin et al., 1978; Ergaliev, 1983). The Cambrian-Ordovician boundary is drawn at the level of the first appearance of *Hysterolenus*.

One of the best Cambrian-Ordovician boundary successions is in the Malyi Karatau Range, south Kazakhstan (Apollonov and Chugaeva, 1982, 1983; Chugaeva and Apollonov, 1982; Apollonov et al., 1988). The boundary beds are part of the carbonate Shabakty Formation, formed on the north slope of the Aisha-Bibi Seamount, reconstructed by V.G. Zhemchuzhnikov and H. Cook and studied in 1987 by a Soviet-U.S.A. scientific team. The conodont and trilobite zonal successions have been established for the section (Table 1).

Trilobites are abundant and diverse and represented by taxa of Asiatic or cosmopolitian aspect (Apollonov et al., 1988). Conodonts are known at many levels (Dubinina, 1982; this volume). The *Eoconodontus (E.) notchpeakensis* Zone, *E. (E.) alisonae* Zone and *E. (Cambroistodus) minutus* Zone can be correlated well with the same zones in North America, China, and some other regions. The base of the *C. proavus* Zone is marked by the appearances of *C. proavus* and *C. oklahomensis* at

Table 1 (cont'd.)

KENDYKTAS KAZAKHSTAN B.M. KELLER, T.B. RUKAVISHNIKOVA, 1961	SARYKUM, KAZAKHSTAN M.K. APOLLONOV, V.G. ZHEMCHUZHNIKOV, S.V. DUBININA	OLENTY-SHIDERTY, KAZAKHSTAN N.K. IVSHIN et al., 1972, TZAY, 1983	NORTH TAIMYR A.M. OBUT, R.F. SOBOLEVSKAYA 1962; R.F. SOBOLEVSKAYA et al., 1978	ALTAY-SAYAN REGION Z. YE. PETRUNINA, 1968	SIBERIAN PLATFORM A.V. ROZOVA, 1968, 1970, 1986; SOKOLOV, 1982	NORTH-EAST OF USSR M.M. ORADOVSKAYA, 1974, 1984; M.N. CHUGAEVA et al., 1964
[KURDAY FM.] Kayseraspis kurdaicus, Apatokephalus [AGALATAS FM.] Apatokephalus replicare, Symphysurus perangustus, Ceratopyge, Tritoechia kendiktasica, Clarkella transversa [KENDYKTAS FM.] Dichelepyge bicornis, Harpides rugosus Staurograptus dichotomus Anisograptus richardsoni Symphysurus Niobe homphrai Shumardia Hysterolenus UNFOSSILIFEROUS SANDSTONE ?	Paracordylodus gracilis [CHAZHOGAY FM.] Drepanoistodus deltifer Rossodus manitouensis-Chosonodina herfurthi ? Cordylodus lindstromi ? Cordylodus proavus Lotagnostus hedini [SARYKUM FM.] Eoconodontus notchpeakensis, Lotagnostus punctatus ?	ZHELDYADYR FM. [OLENTY FM.] Clarkella supina, Nanorthis hamburgensis, Ceratopyge, Harpides, Niobe [SATPAK FM.] Staurograptus dichotomus, Anisograptus richardsoni, Bienvillia tetragonalis * Dictyonema asiaticum, Callograptus olentensis, Dendrograptus yini UNFOSSILIFEROUS SANDSTONE * SHIDERTIAN HORIZONE Lotagnostus trisectus, Hedinaspis, Charchaqia	Tetragraptus, approximatus, T. acclinans Clonograptus limatus, Cl. milesi, Temnograptus novoberacensis, Triograptus canadensis Dictyonema graptolithinum, D. taimyrense Anisograptus richardsoni Dictyonema sociale, D. parabola * [GRUSTNAYA FM.] ? Hapalopleura asiatica, Proceratopyge sp., Agnostus	LEBED HORIZON Tetragraptus approximatus TAYANZY HORIZON Tersella strobilata, Niobe zhulanica ZONE Amzasskiella mirabilis, Shumardia pusillina ZONE Apatokephalus sibiricus, Nyaya oriens ZONE DOBRYI HORIZON Koldinioidia, Onchonotellus, Macropyge (Aksapyge), Charchaqia, Harpidoides, Ivshinaspis, Callograptus, Dendrograptus halianus * ZOLOTOKITAT HORIZON Lotagnostus sp., Asiocephalus regaliformis	UGOR HORIZON Tetragraptus approximatus NYAYA HORIZON Glaphurus, Platypeltoides, Nyaya nyaensis, Apatokephalus nyaicus, Cordylodus aff. proavus ? LOPAR HORIZON Plethopeltides magnus, Pseudacrocephalites, Loparella, Archaeorthis, Dictyonema kulumbense, Callograptus MANSI HORIZON Dolgeuloma, Saukiella, Mansiella, Apheorthis, Tetralobula * KETI HORIZON Ketyna ketyensis, K. glabra, Kaninia aff. quadrata, Furnishina furnishi	KHITIN HORIZON Biolgina, Apatokephalus, Pseudomera, Drepanodus proteus INANYA HORIZON Clarkella calcifera, C. aff. obata, Nanorthis multicostata, Pliomeridae, Bathyuridae Nyaya, Dolgeuloma, Loparella Apheorthis, Tetralobula

109 m. Sixteen species disappear immediately below the base of the *C. proavus* Zone. The *C. intermedius* and *C. lindstromi* zones are distinguished above.

Inarticulate and articulate brachiopods and chitinozoans have been studied from the boundary section. The great changes in the taxonomic composition of all faunal groups coincides in the succession with the base of the *C. proavus* Zone (Apollonov et al., 1988).

Conodont alteration indices (CAI) are 1-1.5. The paleomagnetic signature has a reversed polarity to just below the base of the *C. proavus* Zone. Above this horizon, the polarities are normal (J. Kirshvink, pers. comm., 1988). Distinctive geochemical changes also take place at the base of the *C. proavus* Zone (J. Wright, pers. comm., 1988). The section was proposed as a Cambrian-Ordovician boundary stratotype (Apollonov et al., 1988).

A succession with a good typical Tremadocian fauna is known near Agalatas River in the Kondyktas Range, south Kazakhstan (Keller and Rukavishnikova, 1961). The succession consists of the clastic Kendyktas Formation (400 m thick) and carbonate Agalatas Formation (360 m thick). The lower part of the Kendyktas Formation is represented by brown, coarse, unfossiliferous sandstone. The lowest fauna in the succession, in the upper member of mudstone and siltstone, includes *Hysterolenus*, *Niobe homphrai* (Salter), *Araiopleura stephani* Rushton, *Shumardia*, *Harpides*, and *Euloma* (*Proteuloma*). The graptolites *Staurograptus dichotomus* Emmons and *Anisograptus richardsoni* Bulman were discovered by D.T. Tzaj several metres above. The trilobite *Symphysurus* is found together with graptolites, whereas *Dichelepyge bicornis* Lisogor and *Harpides rugosus* Lisogor are found in the uppermost part of the formation. Conodonts are not found in the section. The Agalatas

Formation, containing the trilobites *Apatokephalus replicare* Lisogor, *Symphysurus perangustus* Lisogor, *Ceratopyge* sp. and the brachiopod *Tritoechia kendyktasika* Rukavishnikova, rests conformably on the Kendyktas Formation. The Cambrian fauna is not known in the section. The assemblage with *Hysterolenus*, *Niobe homphrai* (Salter), *Araiopleura stephani* Rushton and others is very similar to trilobites from the *Acerocare* Zone of North Wales (Rushton, 1982). The Cambrian-Ordovician boundary in Kendyktas is drawn at the base of the beds with *Hysterolenus*.

In the Erementau-Chu-Ili Zone, uppermost Cambrian conodonts are recorded from the lower part of the widespread deep water siliceous beds (jaspers), the clastic Akdym Formation (Dvoichenko and Abaimova, 1987). Older Cambrian deposits are not known in most of the zone, except in the northern part of the zone (Selety River) where shallow water limestone, tuff, and clastic rocks of Late Cambrian age occur.

In the western part of the Balkhash Massive, in the northwestern Balkhash region, the Cambrian-Ordovician boundary was studied in the lower part of the Chazhagay Formation, in dark, deep water chert with thin limestone layers in the lower part. The best section was studied by M.K. Apollonov and V.G. Zhemchuzhnikov near the Sarykum railway station. The Chazhagay Formation gradationally replaces the upper carbonate member of the Sarykum Formation which bears the trilobite *Lotagnostus punctatus*. In the lowermost part of the Chazhagay Formation, *Lotagnostus hedini* and *L. punctatus* of the uppermost Cambrian were found in the limestone layers. Conodonts of the *Eoconodontus* (*E.*) *notchpeakensis* Zone were discovered together with trilobites. Conodonts of the *Cordylodus proavus* and *C. lindstromi* zones were extracted from the thin layers of limestone from above. The *C. intermedius* Zone has not yet been discovered.

One of the best known Cambrian-Ordovician boundary sections in northeastern Kazakhstan is located in the Olenty-Shiderty rivers area in the northeast of the Erementau-Chu-Ili Zone (Ivshin et al., 1972; Nikitin, 1972). For many years, the section was considered as the stratotype for the regional Lermontov and Shiderty horizons (Cambrian) and Satpak and Olenty horizons (Ordovician). The Shidertian Stage of the international scale of the Upper Cambrian was proposed after studying the section (Ivshin and Pokrovskaya, 1968). Later it became evident that the section was too strongly disturbed tectonically and could not be used as a stratotype. *Dictyonema flabelliforme* mentioned from the section was later shown to be a different species of *Dictyonema* (Tzaj, 1983).

The Shidertian Horizon was established for the upper part of the carbonate terrigenous Tortkuduk Formation (Ivshin et al., 1972). It is characterized by a diverse complex of typical Upper Cambrian trilobites with *Lotagnostus trisectus* (Salter), *Hedinaspis*, *Charchaqia*, and others. The conodonts *Westergaardodina*, *Furnishina* and others are also known. The Satpak Formation occurs above. Above a 5 m thick unfossiliferous interval, D.T. Tzaj (1983) erected two successive graptolite complexes: a lower one with *Callograptus olentensis* Tzaj, *Dendrograptus yini* Mu, and *Dictyonema asiaticum* Hsu; and an upper one with *Staurograptus dichotomus* Emmons and *Anisograptus richardsoni* Bulman. The lower complex also includes the trilobites *Bienvillia tetragonalis* (Harrington) and *Geragnostus* sp. (Ivshin and Pokravskaya, 1968). The base of the Ordovician is drawn by D.T. Tzaj in the base of the *Dendrograptus yini* beds.

The Olenty Formation containing *Clarkella supina* Nikitin, *Nanorthis hamburgensis* (Walcott), *Ceratopyge*, *Harpides*, and *Niobe*, which is supposedly of late Tremadocian age, rests on the Satpak Formation (Nikitin, 1956). The rocks are strongly folded.

In the territory of the Chingiz-Tarbagatay Zone, complete successions of the Cambrian-Ordovician boundary interval are unknown. In the western part of the zone, silicic (jasper) volcanogenic deposits of Arenig age rest unconformably on folded Lower and Middle Cambrian rocks. In the eastern part of the zone, tuffaceous limestone of the Mamat Formation containing *Clarkella*, *Ceratopyge* and others (supposedly upper Tremadocian) rest transgressively on folded Cambrian strata (Nikitin, 1972).

To the east, in the Zaisan Zone, with its siliceous-volcanogenic lower Paleozoic deposits, the Upper Cambrian and Lower Ordovician are not known with certainty.

In Kirgizia, uninterrupted fossiliferous Cambrian-Ordovician successions are known. In some places, limestone containing Upper Cambrian *Lotagnostus asiaticus* Troedsson, *Acrocephalina armata* Troedsson, *Charchaqia norini* Troedsson, *Hedinaspis regalis* Troedsson, *Norinia convexa* Troedsson and others are recorded (Goncharova, 1972).

MIDDLE ASIA

In Middle Asia, Cambrian-Ordovician boundary beds are known from two areas (Nikitin et al., 1986). In the central part of the Turkestan Range, the boundary is drawn at the base of Apchapkan Formation of dark shale, mudstone, and sandstone with the upper Tremadocian(?) graptolites *Callograptus* sp. cf. *C. sinicus* Mu, *Dictyonema uniformis* Mu, *Clonograptus*, *Anisograptus*?, and *Bryograptus*?. The formation rests conformably on supposed Upper Cambrian rocks.

In the Central Pamir, the systemic boundary is located somewhere near the boundary between the Zorabat and Kozyndy formations, which consist of limestone, dolomite, and shale (Balashova, 1966). No fauna occurs in the upper part of the Zorabat Formation. In the lower part of the Kozyndy Formation, the upper Tremadocian trilobites *Apatokephalus serratus pamiricus* Balashova, *Dikelokephalina*, *Shumardia* and others are recorded (Balashova, 1966).

ALTAY-SAYAN REGION

The Altay-Sayan region is an interesting and important region for studying the Cambrian-Ordovician boundary. It is considered to be the miogeoclinal system along the edge of the Siberian Platform. The evolution of structures, environments, and sedimentation in the Late Cambrian and Early Ordovician have not been studied sufficiently. Upper Cambrian and Lower Ordovician stratigraphy and trilobites were studied by Petrunina (e.g., 1968, 1973). This review is based mainly on her data.

Upper Cambrian and Lower Ordovician deposits are widespread in the territories of Salair, Kuznetsk Alatau, Gornaya Shoria and in the eastern part of the Gornyi Altay-in Biya-Katun and Uiem-Lebed zones. Boundary beds consist mainly of sandstone and mudstone with rare lenses and beds of limestone. Volcanics are widespread only in the Lower Ordovician. Abundant and diverse trilobites are known at many levels (Petrunina, 1968, 1973; Krasnov et al., 1983). Other fauna are rare. The unfossiliferous, thick (several kilometres), deep water Gornoaltay Series consists of graded-bedded sandstone and mudstone (Goosyev, 1986) and is distributed over the western part of the Gornyi Altay Zone. The upper part of the series is correlated tentatively with Upper Cambrian and Lower Ordovician (Krasnov et al., 1983).

At the boundary interval, the Zolotokitat, Dobryi, Tayanza (earlier known as Algain) and Lebed horizons are distinguished (Krasnov et al., 1983). Petrunina drew the Cambrian-Ordovician boundary at the base of the Dobryi Horizon, and correlated the Tayanza Horizon with the upper Tremadocian (e.g., Petrunina, 1968; Bessonenko and Petrunina, 1972; Yermikov et al., 1979). This point of view became almost universally recognized by Siberian stratigraphers (Krasnov et al., 1983), but now requires radical revision.

The Zolotokitat Horizon was established based on the succession along the Dobryi stream in Kuznetsk Alatau. Typical Upper Cambrian trilobites with *Pseudagnostus simplex* Lermontova, *Lotagnostus* sp., *Asiocephalus regaliformis* Ivshin, *Proceratopyge* sp. cf. *P. grabaui* Troedsson, *Plicatolina* sp. and others (Krasnov et al., 1983) are recorded from the horizon.

The type succession of the Dobryi Horizon was also described among the Dobryi stream. It is characterized by abundant and diverse trilobites (Petrunina, 1968; Krasnov et al., 1983). The *Archaeodictyota* Zone is erected on the basis of rare graptolites *Archaeodictyota* sp. aff. *A. dragunovi* Obut and Sobolevskaya, *Dendrograptus* sp., and *Callograptus* sp. (Krasnov et al., 1983). According to Petrunina (1968, 1973) more than 60 trilobite species representing over 30 genera are known from the Dobryi Horizon. She believes that the horizon correlates with the lower Tremadocian. Well known trilobites *Leiagnostus*, *Psedagnostus*, *Euloma* (*Proteuloma*), *Hospes*, *Koldinioidia*, *Parabolinella*, *Onchonotellus*, *Acrocephalina*, *Macropyge* (*Promacropyge*), *Niobella*, *Tropidopyge*, *Charchaqia*, *Harpidoides* (=*Sibiriopleura*), *Eocheirurus*, and *Diceratopyge* are very important for correlation from her point of view. New genera *Bilacunaspis*, *Ivshinaspis*, and *Skljarella* were published from the horizon (Petrunina, 1973). Some paleontological data from the Horizon have not yet been published (Petrunina, pers. comm., 1988). Two zones - *Skljarella lidiae-Tropidopyge tevipis* and *Euloma limata-Harpides* (=*Harpidoides*) *ulumandensis* - are distinguished in the Dobryi Horizon by Petrunina (1968).

The deposition of the Tayanza Horizon is widespread over the territory of the Altay-Sayan region. Petrunina (1968) correlated the horizon with the upper Tremadocian. The type succession was studied near the Tayanza River in Gornaya Shoriya. The trilobite assemblage according to Petrunina (1968) consists of more than 100 species representing about 50 genera. *Euloma* (*Euloma*), *Shumardia*, *Hysterolenus*, *Nyaya*, *Tersella*, *Apatokephalus*, *Platypeltoides*, *Niobe*, *Promegalaspides*, *Dikelokephalina*, *Ceratopyge*, *Harpides*, *Orometopus*, *Rhadinopleura* (*Sibiriopleura*), *Cyrthometopus*, *Pliomeroides*, *Amzasskiella*, *Kaltykelina*, and *Ottenbyaspis* are most characteristic. Three zones - *Apatokephalus sibiricus-Nayaya oriens*, *Amazasskiella mirabilis-Shumardia pusillina* and *Tersella strobilata-Niobe zhulanica* - are distinguished in the horizon (Petrunina, 1968). Two zones, *hyperboreus* and *osloensis*, are erected on the basis of rare graptolites. The brachiopods *Apherothis*, *Nanorthis*, *Notorthis*, *Punctolira*, the alga *Nuia sibirica*, and a single conodont species *Oneotodus datsonensis* Druce and Jones are known. The presence of Chitinozoa is recorded (Yermikov et al., 1979). The Lebed Horizon with Arenigian graptolites including *Tetragraptus approximatus* is noted above.

The presence of conodonts and graptolites in the Cambrian-Ordovician boundary interval is very rare in the Altay-Sayan region, and the boundary can be drawn only on the basis of trilobites.

Analyzing the composition of the trilobite associations of the Dobryi and Tayanza horizons, Petrunina (1968) gave special significance to the typical Ordovician (according to her) genera *Hospes*, *Tropidopyge*, and *Harpides* and to the "Ordovician appearance" of the new genera *Bilacunaspis*, *Ivshinaspis*, and *Skljarella*. However, *Hospes* and *Tropidopyge* are well known in the Aksay Horizon, deep water Upper Cambrian strata, in Malyi Karatau (Ergaliev, 1980) and in Cambrian strata in many regions of China. "*Harpides*" from Altay-Sayan region proved to be the new genus *Harpidoides*, which is common in the Batyrbay Horizon of the Upper Cambrian (*Eoconodontus notchpeakensis* Zone) in Malyi Karatau (Apollonov and Chugaeva, 1983; Apollonov et al., 1988). *Ivshinaspis* is characteristic only for the lower part of the Batyrbay Horizon.

The co-occurrence of such genera as *Pseudagnostus*, *Koldinioidia*, *Onchonotellus*, *Macropyge* (*Promacropyge*), *Charchaqia*, *Harpidoides*, and *Ivshinaspis* permits the correlation of the Dobry Horizon with the Batyrbay Horizon of Malyi Karatau, which is one of the typical units of the upper Upper Cambrian (Apollonov et al., 1988).

The Ordovician starts from the base of the Tayanza Horizon with well known typical Tremadocian trilobites *Hysterolenus*, *Euloma* (*Euloma*), *Shumardia*, *Nyaya*, *Tersella*, *Apatokephalus*, *Dikelokephalina*, *Ceratopyge*, *Harpides* (s.s.), and others. The exact level of the Cambrian-Ordovician boundary in any particular section cannot be defined more precisely from the published data. The Tayanza Horizon correlates well with the Nyaya Horizon of the Siberian Platform with the presence of trilobites *Apatokephalus* sp. cf. *A. nyaicus* Rozova and *Dolgeuloma turumakitica* Rozova in Gornyi Altay (Krasnov et al., 1983). The Tayanza Horizon can be correlated with the entire Tremadocian. The precise correlation of the Dobryi-Tayanza Horizon boundary with the conodont zonation needs further study.

SIBERIAN PLATFORM

The Siberian Platform basin in the Cambrian and Ordovician extended over the territory between the modern Yenisey and Lena rivers. The upper Upper Cambrian and Lower Ordovician consist mainly of shallow water carbonates - limestone and dolomite with beds and units of sandstone and mudstone. The facies variation is very characteristic. Only in North Taymyr are the Cambrian-Ordovician boundary beds represented by comparatively deep water limestone and graptolitic shale (Krasnov et al., 1983).

The regional Kety, Mansi, Lopar, Nyaya, and Ugor horizons are established for the boundary interval mainly on the basis of the study of trilobites (e.g., Rozova, 1968; Krasnov et al., 1983). The problem of drawing the Cambrian-Ordovician boundary in the Siberian Platform is difficult. Quite different levels have been proposed as the systemic boundary by different stratigraphers and each author believed the level proposed can be correlated with the base of the Tremadocian. Some authors (Yadryonkina, 1974; Ogienko, 1977; Krasnov et al., 1983) drew the boundary at the base of the Mansi Horizon. Rozova (1968, 1986) placed the boundary at the base of the Nyaya Horizon. Endemism of trilobites and brachiopods, and the limited study of conodonts, graptolites and other groups does not permit Siberian stratigraphers to agree on a common, precise level.

Trilobites in every horizon are strongly endemic. Naturally, those genera and species known beyond the Siberian Platform attract the greatest attention. For the Kety Horizon, *Ketyna ketiensis* Rozova, *K. glabra* Rozova, *Kaninia* sp. aff. *K. quadrata* Lazarenko are of interest as they are also known in the Aksay Horizon in Malyi Karatau (Lisogor, 1977; Ergaliev, 1980; Apollonov and Chugaeva, 1982, 1983; Rozova, 1986).

A sharp change in faunal composition is recorded between the Kety and Mansi horizons (Rozova, 1970, 1986). Endemic *Dolgeuloma*, *Pseudokoldinia*, *Mansiella*, and *Pseudacrocephalites* are characteristic for the Mansi Horizon. *Saukiella* is well known in North America, Kazakhstan, and China and occurs in the south of the platform in the Lena Zone. *Apatokephalus* (possibly the oldest occurrence) is recorded here also (Ogienko, 1977). The brachiopods *Tetralobula*, *Eoorthis*, *Apheorthis*, *Finkelnburgia*, and *Syntrophopsis* are characteristic of the horizon (Rozova, 1968, 1970; Krasnov et al., 1983). The Mansi Horizon is present for certain only in the Igarka-Norilsk and Lena zones. In the other areas it is recognized only tentatively.

The fauna of the Lopar Horizon is similar to that of the Mansi Horizon. *Pseudacrocephalites* became abundant, and the species diversity increases appreciably. *Tetralobula*, *Finkelnburgia*, and *Archaeoorthis* are most common among the brachiopods. In the middle part of the Lopar Horizon the conodont *Oneotodus nakamurai* Nogami occurs, and the graptolite *Dictyonemna kulumbense* Obut and Sobolevskaya is recorded from the upper part of the horizon (Rozova, 1986). The resemblance of fauna of the Mansi and the Lopar horizons permitted Rosova (1970, 1986) to combine the horizons into the Khantaika Superhorizon.

The fauna of the Nyaya Horizon is distinguished from that of the Lopar Horizon (Rozova, 1968, 1970). Among the trilobites, *Nyaya*, *Apatokephalus*, *Glaphurus*, and *Platypeltoides* (=*Ursibia*) are to be noted as they are well known in the Ordovician in many regions. Rare *Cordylodus proavus* Müller is recorded from lower part of the horizon (Sokolov, 1982; Rozova, 1986). Conodonts of the *Cordylodus angulatus*-*C. rotundatus* Zone occur in the upper part of the horizon (Krasnov et al., 1983). Gastropods and monoplacophorans are common.

The Ugor Horizon is characterized by the appearance of the inarticulate brachiopod *Angarella* as the background of a generally poor fauna. Thicknesses of every horizon are from tens to 150-180 m. Greatest thicknesses are recorded from the successions along the Kulyumbe River and in the south of the platform in Lena Zone (Krasnov et al., 1983).

The Cambrian-Ordovician boundary succession of North Taymyr is distinctly different (Obut and Sobolevskaya, 1962; Sobolevskaya et al., 1978; Krasnov et al., 1983). The Grustnaya Formation consists of deep water limestone (250 m thick) with trilobites *Hapalopleura asiatica* Lazarenko, *Proceratopyge* sp., and *Agnostus* sp. of Late Cambrian age, conformably overlain by a formation of dark shale and limestone, which is divided into three members. The graptolites *Dictyonema sociale* (Salter) and *D. parabola* Bulman occur in the lowest member (25 m thick); *Anisograptus richardsoni* is recorded from the lower part of this member and *D. graptolithinum* Kiaer and *D. taimyrense* Obut and Sobolevskaya are present in the upper part of the middle member (60 m thick); *Clonograptus limatus* Obut and Sobolevskaya is known from the upper member (130 m thick). *Tetragraptus approximatus* is distinguished above.

It is practically impossible to correlate precisely the shallow water boundary beds of the Siberian Platform with the deep water boundary interval of Taymyr.

Up to now, two main points of view on the Cambrian-Ordovician boundary in Siberia have been advanced (Krasnov et al., 1983). Rosova (1968, 1970) located the boundary at the base of the Nyaya Horizon with *Apatokephalus nyaicus* and *Nyaya nyayensis*, until recently correlating this level with the base of the upper Tremadocian. More recently she has decided to correlate the base of the Nyaya Horizon with the base of the Tremadocian (Rozova, 1986).

Yadryonkina (1974) and Ogienko (1977) drew the Cambrian-Ordovician boundary at the base of the Mansi Horizon on the basis of the sharp change in faunal composition. The Mansi Horizon fauna is endemic and correlation of the horizon with the lower Tremadocian is only tentative. The graptolite *Dictyonema kulumbense* Obut and Sobolevskaya recorded from the Mansi Horizon probably belongs to the ancient, rooted *Dictyonema*, which was common in the Cambrian (Rozova, 1986).

Yadryonkina et al. (1984) reported *Dictyonema flabelliforme* Zone graptolites with *Bryograptus patens* Matthew and *Staurograptus*(?) sp., and probably belonging to the *D. parabola* Zone and *D. norvegicum* Zone, from the Oldondin Formation in the Aikhal area, Vilyui Zone. Graptolites are discovered in subsurface borings; no other fauna is found. The member with graptolites is tentatively attributed to the Mansi Horizon.

Currently, the Cambrian-Ordovician boundary problem on the Siberian Platform cannot be resolved from conodont studies. The data on the group are fragmentary and controversial. Yadryonkina and Abaimova (1980) recorded the discovery of the conodont *Cordylodus proavus* Müller in the Mansi and Lopar horizons. Later, in the special paper on the Cambrian-Ordovician boundary on the Siberian Platform, Abaimova (1984) summarized all data on conodonts in Siberia. She stated that in the south of the Siberian Platform more than 20 localites of conodonts from *Proconodontus tenuiserratus*, *Proconodontus*, and *Cordylodus proavus* zones are known. The Mansi Horizon is correlated with the *Proconodontus* Zone, the Lopar Horizon with the *Cordylodus proavus* Zone, and the Nyaya Horizon with the *Cordylodus angulatus-C. rotundatus* Zone. Stratotypes for the horizons are not documented by conodonts. The known data on conodonts, graptolites, and trilobites do not compare perfectly. The position of the *C. intermedius* and *C. lindstromi* zones was not taken into consideration.

The potential Cambrian-Ordovician boundary interval (*Cordylodus proavus-C. lindstromi* Zone) is located probably somewhere near the base of the Nyaya Horizon or in the upper part of the Lopar Horizon. Hence, Rozova's (1986) position for the Cambrian-Ordovician boundary in the Siberian Platform is preferable.

NORTH AND NORTHEASTERN U.S.S.R.

In northeastern U.S.S.R., Lower Ordovician deposits cover vast territories from the Omulyovian Mountains to the Chukotka Peninsula. The transgressive cycle of sedimentation including the lower and middle Paleozoic began in the earliest Ordovician (Oradovskaya, 1974, 1984).

Cambrian deposits are known from small isolated areas. They are represented by multicoloured dolomite and marl (500 m thick). Uninterrupted Cambrian-Ordovician boundary successions are unknown (Oradovskaya, 1974, 1984). Cambrian fossils are practically absent. Only in the Nelchan Formation near Rassokha River do the inarticulate brachiopods *Linnarsoniidae* (*Neoteta*?) and *Acrotretinae* occur. Upper Cambrian trilobites in Lower Ordovician conglomerate boulders near the Rossokha River are recorded (Oradovskaya, 1974, 1984).

The interrelationships of the Lower Ordovician with pre-Ordovician deposits (Riphean-Middle or lower Upper Cambrian) were studied recently in some remote areas (Rassokha, Omuljovka, Karkodon, Kedon rivers) (Oradovskaya, 1984). The Ordovician is represented mainly by shallow water carbonate or fine grained clastic-carbonate deposits. Basal conglomerates are present in most localities.

The Inanja Horizon is distinguished in the lowest Ordovician in the northeastern U.S.S.R. (Chugaeva et al., 1964). The horizon unites the Inanya Formation in the Elgenchak Mountains, the Sekudun Formation in the Selennyakh Range, and the Luksk Formation on the Omolon Massive. The fauna in the Inanya Horizon includes the brachiopods *Clarkella calcifera* (Billings), *C.* sp. aff. *C. lobata* Ulrich and Cooper, *Nanorthis multicostata* Ulrich and Cooper, *Apheorthis*, and *Tetralobula*. Trilobites are rare and represented in some places by *Pliomeridae* and *Bathyuridae*. In the Luksk Formation, *Dolgeuloma*, *Loparella*, and *Nyaya*, characteristic for the Nyaya Horizon in the Siberian Platform, are known (Oradovskaya, 1974).

The Cambrian and Ordovician at the Chukotka Peninsula are distinguished only tenatively along the shore of the Chukchi Sea (Oradovskaya and Obut, 1977). Gently sloping (0-30 m thick) series of phyllitic shale rest unconformably on Proterozoic gneiss. The series is subdivided into Seman (375 m thick) and Ikoluvroon (250 m thick) formations, distinguished mainly by their colour. Fauna are absent from both formations. They are considered to be undivided Cambrian-Ordovician. It is possible the whole series belongs to the Ordovician. Hence, the Cambrian-Ordovician boundary in the northeastern U.S.S.R. is probably located everywhere near the base of the transgressive Ordovician complex.

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Subdivisions of the Ordovician System in Australia

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Abstract

The Victorian Ordovician stage and graptolitic zonal scheme is the main basis for correlation in the Tasman Fold Belt successions of eastern Australia and in New Zealand, and it is now recognized as one of the most useful global Ordovician biostratigraphic schemes available. The main disadvantages are that key Victorian sections have not been adequately described, and have a relatively sparse record of associated shelly faunas, though recent discoveries by I. Stewart of conodonts of North Atlantic aspect provide some useful tie points. An independent stage and zonal scheme based mainly on conodonts has been applied to the lowermost part of the Lower Ordovician continental platform sequences of northern Australia, but other shelly faunal schemes used in Australia are only informally designated. A number of major faunal events in the graptolite record are recognized as useful for international correlation, and this is also true of the Lower Ordovician shelly faunal and conodont record of the platform sequences in northern Australia. The British series can be correlated with varying levels of ambiguity with the Victorian graptolite succession. Other overseas subdivisions are less easily applied although the North American series and stage subdivision is the most useful in correlating shallow water Tasmanian sequences. Nevertheless, it is recommended that the Australasian Ordovician be divided broadly into two subsystems at the appearance of the worldwide zonal indicator Nemagraptus gracilis *(i.e., at the base of the Gisbornian Stage).*

Résumé

La corrélation des successions dans la zone de plissement de Tasman, dans l'est de l'Australie et en Nouvelle-Zélande, se fonde principalement sur l'étage ordovicien du Victorien et ses zones de graptolites. Ce plan est maintenant reconnu comme étant l'un des plus pratiques pour la biostratigraphie mondiale de l'Ordovicien. Toutefois, il présente certains désavantages, notamment l'absence d'une description adéquate de certaines coupes clés du Victorien et la rareté relative de la faune coquillère associée; or, I. Stewart a découvert récemment des conodontes qui rappellent ceux de l'Atlantique Nord et qui sont des points de rattachement utiles. On a appliqué un plan indépendant d'étage et de zones, qui se fonde principalement sur des conodontes, à la partie basale des séquences continentales de plate-forme de l'Ordovicien inférieur dans le nord de l'Australie; toutefois, d'autres plans qui ont trait aux faunes coquillères et qui sont utilisés en Australie n'ont qu'une désignation informelle. Un certain nombre d'événements fauniques majeurs dont attestent les données sur les graptolites sont utiles aux fins de la corrélation mondiale; cela est vrai aussi pour la faune coquillère et les conodontes de l'Ordovicien inférieur dans les séquences de plate-forme dans le nord de l'Australie. On peut établir une corrélation plus ou moins ambiguë entre la série britannique et la succession de graptolites du Victorien. D'autres subdivisions d'outre-mer s'appliquent moins facilement, bien que celle de l'étage et de la série nord-américaine soit la plus utile pour la corrélation des séquences tasmaniennes d'eau peu profonde. Néanmoins, on recommande de subdiviser l'Ordovicien australasien en deux grands sous-systèmes à l'apparition du marqueur zonal mondial Nemagraptus gracilis *(c.-à-d. à la base du Gisbornien).*

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INTRODUCTION

At the business meeting of the Subcommission on Ordovician Stratigraphy held at the Fourth International Symposium on the Ordovician System in Sundvollen, Norway, in August, 1982, it was agreed to establish several regional working groups to tackle issues such as clarification of the nature of chronostratigraphic subdivisions used in the particular regions, as a basis for developing an internationally accepted chronostratigraphy for the Ordovician System. An Australasian Ordovician Chronostratigraphy Working Group (AOCWG) was established in September, 1983 under the chairmanship of R.A. Cooper and A.H.M. VandenBerg, and proceeded to focus on a number of chronostratigraphic and biostratigraphic problems in the region. A questionnaire was circulated by the co-chairmen to a large number of stratigraphers with an interest in the Ordovician geology of Australia and New Zealand, and the authors listed above all contributed responses to this questionnaire. Issues canvassed included 1) the nature of biostratigraphic zonal schemes, 2) the stage classifications used for both internal and external correlations, 3) the nature of successions with diverse faunas and major events, and 4) the desirability of adopting a single local or international stage and series classification for Australasia. Cooper and VandenBerg collated the initial responses on the graptolite facies, and B.D. Webby, the shelly faunal data. A preliminary presentation of results was given at a symposium entitled "Progress Towards an Australian Geological Time Scale" at the Ninth Australian Geological Convention in Brisbane in January-February, 1988 (Webby et al., 1988). The final report (herein) was compiled by Webby from a draft circulated to members in June, 1988.

As a part of a separate Bureau of Mineral Resources project, Webby and Nicoll (1989) have also assembled an Ordovician biostratigraphic chart and explanatory notes for Australia.

BIOSTRATIGRAPHIC ZONAL SCHEMES CURRENTLY IN USE

In the graptolite facies only one zonal scheme is in general use through the Ordovician, that is, the one formulated by T.S. Hall, Keble, and Harris and Thomas between 1899 and 1938, and summarized by Harris and Thomas (1938), Thomas (1960), VandenBerg (*in* Webby et al., 1981), and Cas and VandenBerg (1988). In Victoria it is the only subdivision used (Fig. 1, Table 1); it is also applicable at the stage level, less so at the zonal level, in graptolite facies of New South Wales and Queensland. In New Zealand a slightly less subdivided but closely similar zonal scheme is employed (Cooper, 1979). Only at a comparatively few localities in southeastern Australia have conodonts of North Atlantic provincial affinity been found in graptolite-bearing associations (Nicoll, 1980; Stewart, *in* Cas and VandenBerg, 1988) providing essential tie points between graptolite and North Atlantic-type conodont schemes (Table 1). Commonly in both eastern Australia and New Zealand, it is necessary to correlate between basinal sequences with graptolite zones, and platform-type successions with shelly faunas, through the mixed graptolite-shelly sequences of Britain, and/or North America.

In platformal areas of northern Australia (Fig. 1), biostratigraphic zonal schemes are based mainly on conodonts (Druce and Jones, 1971; Jones et al., 1971; Druce et al., 1982). They are currently used for correlating the lower part of the Lower Ordovician successions in the Georgina, Amadeus, and Bonaparte Gulf basins, but nowhere is a complete zonation yet described (Table 2). There is a preliminary mixed conodont, graptolite and trilobite scheme applied to the Canning Basin (McTavish and Legg, 1976). Microfloral and nautiloid biostratigraphic schemes have also been used in the Canning Basin.

No formally designated biostratigraphic zonal schemes exist for use in the shelly facies of eastern Australia. Four informal schemes, however, have been introduced, as follows:

1. The coral-stromatoporoid-based subdivisions of Webby (1969, 1975, *in* Webby et al., 1981, and *in* Strusz et al., 1988). The scheme includes four main Upper Ordovician subdivisions (Faunas I-IV) differentiated chiefly on the basis of incomings of new groups at a generic or higher taxonomic rank. The subdivisions have proved useful for correlating the limestones of central and northeastern New South Wales and more generalized relationships with Tasmanian "coralline" successions.
2. The trilobite-based faunule subdivision of the Upper Ordovician (Webby, 1974, 1975) is more like a conventional zonal scheme with constituent name-bearing species, but has proved to be less widely applicable than the coral-stromatoporoid scheme. The four stratigraphically distinct faunules are of little use beyond central New South Wales.
3. The brachiopod-based scheme of Percival (1980) with its Faunas A, AB, B, C and D applies to the same interval in central New South Wales covered by the trilobite faunules, but has limited use elsewhere.
4. The "OT" assemblages of Tasmania span almost the entire Ordovician (Table 2), and are based on many different taxa including trilobites, brachiopods, conodonts, nautiloids, corals, and stromatoporoids (see Banks and Burrett, 1980). Graptolites are only useful at a few horizons. Work is in progress to produce a more formal zonation.

STAGE CLASSIFICATION

Internal subdivision and correlation

In Victoria, local stage names (Tables 1, 2) based on groupings of graptolite zones have been used for correlating the graptolite facies since 1899, first introduced as series divisions by T.S. Hall (1899), and later modified to stages in the 1930s by Thomas and Keble (1933), Thomas (1935), and Harris and Thomas (1938). Their nomenclature and definition have been well summarized by VandenBerg (*in* Webby et al., 1981) and Cas and VandenBerg (1988). They are easy to apply locally and can generally be correlated with most other graptolitic sequences on a global scale. Although lacking formally designated stratotypes and thus not being defined in terms of rock sequences (they are, strictly speaking, "superzones"), the stages are based on a graptolite zonal sequence that has survived more than 50 years of testing, and must be regarded as the standard for correlation in the Pacific graptolite province, indeed the most detailed and firmly established Ordovician graptolite zonal scheme in the world. Of significance is the use of the Victorian stages in the Lower Ordovician (Bendigonian-Yapeenian) of Spitzbergen (Cooper and Fortey, 1982) and the establishment by Williams and Stevens (1987) of a zonal scheme in the Lower Ordovician Cow Head Group of antipodean Newfoundland, in part based on this Australian standard. Lenz (1988) also adopted the Victorian stages for correlating the Ordovician graptolite sequences in northern Yukon, Canada. It gives the Victorian stages a high degree of reliability and enhances their status in comparison with stage schemes of other regions (it should be noted that the British series boundaries are not all yet based on clearly defined boundary stratotypes, either). A particular disadvantage is the comparative lack of other faunas – few conodonts and only sparse shelly fossils are associated with the graptolites.

Victorian stage divisions and nomenclature are also recognized in the New Zealand sequence (Cooper, 1979). In New South Wales, the Victorian stage names are applied, where shelly faunal correlations can be established in relation

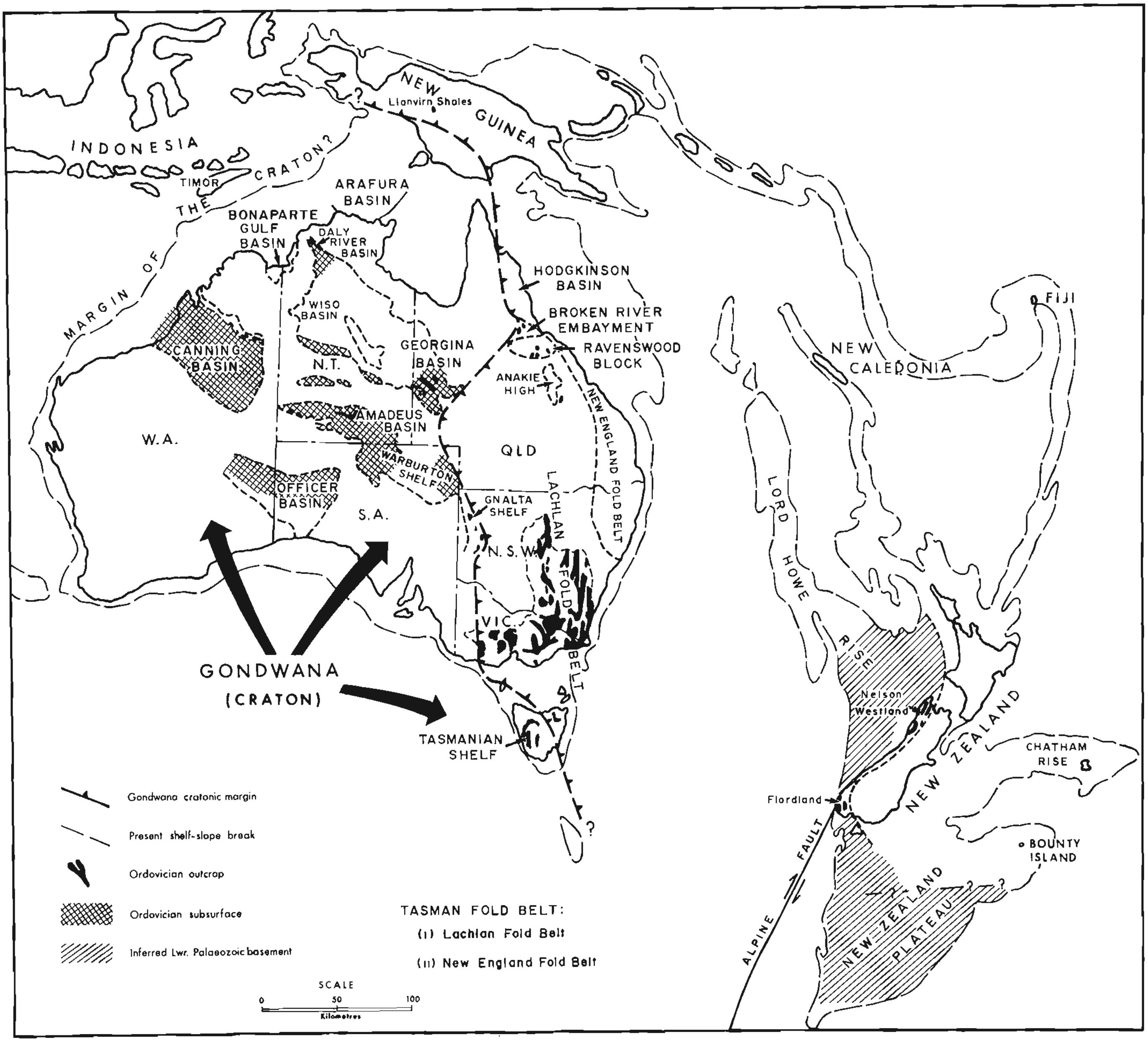


Figure 1. Map showing the distribution of Ordovician rocks in Australasia.

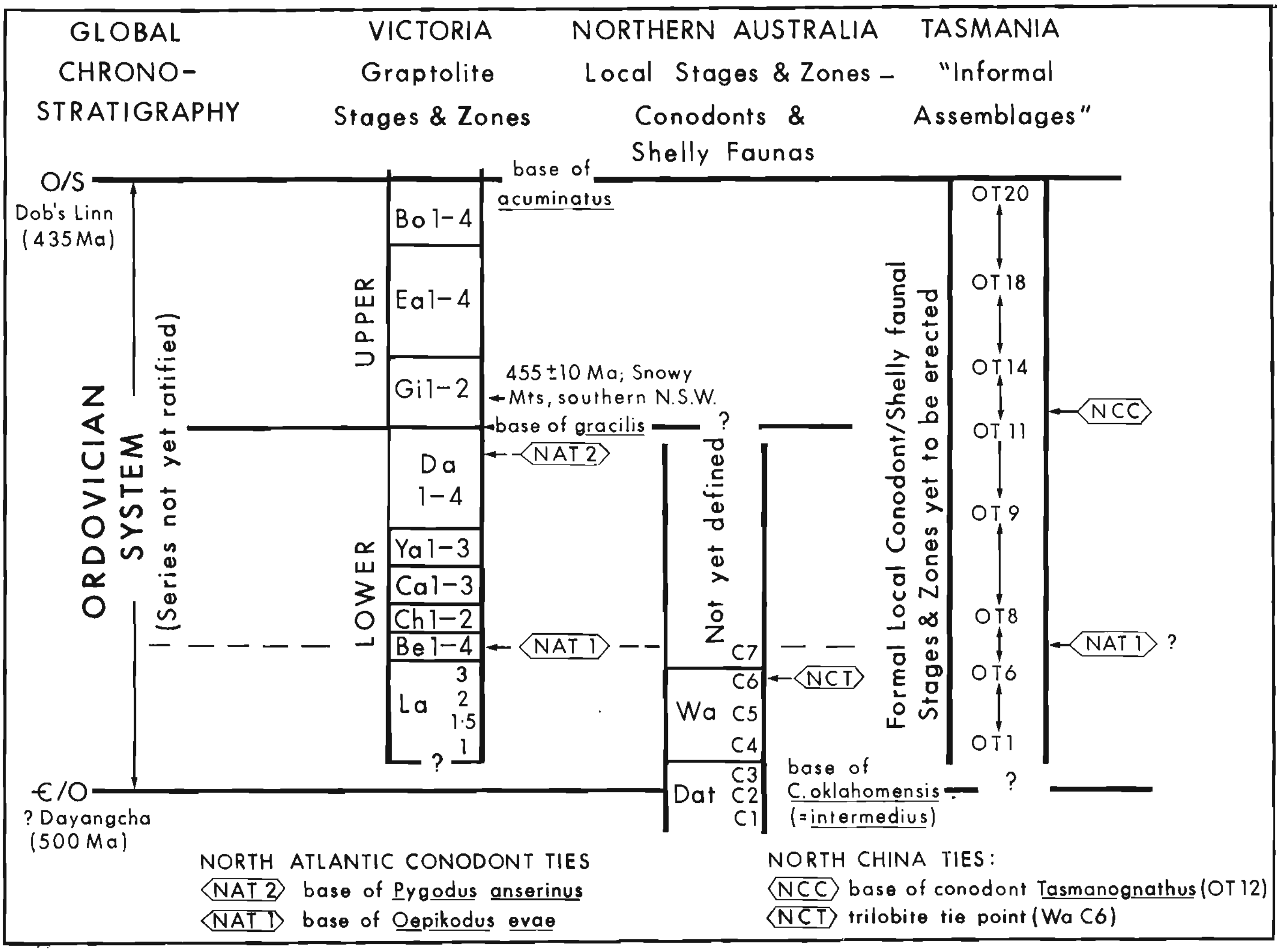
Table 1. Chart depicting the significant events in the graptolite record and relationships to the stage and zonal scheme. Also shown are the important North Atlantic conodont tie points established by I. Stewart, and correlations with the British series

SUBSYSTEMS	BRITISH SERIES CORRELATIONS	AUSTRALIAN (VIC.) STAGES & GRAPTOLITE ZONES		EVENTS IN THE GRAPTOLITE SUCCESSION	CONODONT TIES (N. ATLANTIC PROVINCE)	
UPPER ORDOVICIAN		BOLINDIAN	Bo4	← Appearance of *Glyptograptus persculptus*		
			Bo3	← Appearance of *Climacogr. ? extraordinarius*		
			Bo2	← Appearance of *Dicellogr. ornatus* *Climacogr. latus* + *Paraorthogr. pacificus*		
	ASHGILL		Bo1			
		EASTONIAN	Ea4	← Appearance of *Dicellogr. gravis*		
	base of Ashgill →		Ea3			
			Ea2	← Appearance of *Climacogr. spiniferus*		
	CARADOC		Ea1			
		GISBORNIAN	Gi2			
	base of Caradoc? →		Gi1	← Appearance of *Nemagraptus gracilis*	← *Pygodus anserinus* *Pygodus serra*	*P. anserinus* Zone
LOWER ORDOVICIAN	LLANDEILO	DARRIWILIAN	Da4			*P. serra* Zone
	base of Llandeilo →		Da3			
	LLANVIRN		Da2		← *Polonodus, Eoplacognathus, Spinodus spinatus, 'Cordylodus' horridus* + *Periodon aculeatus*	
	base of Llanvirn I →		Da1	← Appearance of *Oelandogr. austrodentatus*		
	base of Llanvirn II →	YAPEENIAN	Ya2			
			Ya2			
			Ya1			
		CASTLEMAINIAN	Ca3	↑ ↑ ↑		
			Ca2	Evolutionary continuum – *Isograptus victoriae* lineage		
	ARENIG		Ca1	↓ ↓ ↓		
		CHEWTONIAN	Ch2		*Oepikodus evae* *Bergstroemognathus extensus, Protoprioniodus aranda* + *Scolopodus rex*	*O. evae* Zone
			Ch1			
		BENDIGONIAN	Be4			
			Be3			
			Be2		*Prioniodus elegans, Paracordylodus gracilis* *Oepikodus communis,* + *Paraoistodus proteus*	*P. elegans* Zone
	base of Arenig I →		Be1	← Appearance of *Tetragraptus (Pendeogr.) fruticosus* + diverse dichograptid fauna	←	
	base of Arenig II →	LANCEFIELDIAN	La3	← Appearance of *Tetragr. approximatus*		*P. proteus* Zone
			La2	← Appearance of diverse anisograptid fauna		
	TREMADOC		La1·5	← Appearance of *Adelograptus* + *Psigraptus*		
			La1			
		?		? ?		

to the graptolite sequence. Despite L. Sherwin's misgivings, VandenBerg has had no difficulty in applying the Victorian subdivisions in New South Wales.

Subdivisions of the "Tremadoc" Series into the local stages, Datsonian and Warendian (Table 2), of northern Australia are based on conodonts from the Georgina Basin (Jones et al., 1971). Type sections and definitions are adequately published. It is possible to define younger conodont faunas apparently postdating the Warendian in the Amadeus Basin (Nicoll, in prep.). The Warendian and post-Warendian trilobite assemblages of the Amadeus Basin are also in preparation (Shergold, in prep., for the former, and Laurie, in prep., for the latter). The work in progress on the Pacoota Sandstone and Horn Valley Siltstone of the Amadeus Basin should, when completed, add greatly to our ability to apply conodont based zonations. For the present it should be noted that the

Table 2. Chart showing the two-fold subsystemic classification for the Australasian succession based on the Victorian graptolite record (stages and zones). Note also the relationships with platformal successions of northern Australia and with Tasmania; also the main conodont and trilobite tie points, and the radiometric determination of 455 Ma, just above the base of *N. gracilis*



succession of conodont faunas in the Pacoota Sandstone is dominated by a restricted fauna of nearshore elements like *Drepanodus, Scolopodus,* and *Acontiodus.* In the roughly correlative interval in the Georgina and Canning basins, genera such as *Cordylodus, Hirsutodontus, Chosonodina,* and *Acodus* dominate the faunas and indicate deeper water, or at least offshore, mixed carbonate/shale or carbonate environments. This same pattern holds for the Nora Formation of the Georgina Basin and Stairway Sandstone of the Amadeus where the Nora has a fauna with *Histiodella* but the Stairway has forms like *Erraticodon,* which is more abundant in the nearshore facies.

According to Burrett, local stage names will be introduced in Tasmania in the future. (See further discussion in a later section.)

External correlation

The British subdivisions are relatively easily applied to the graptolite Victorian sequences, though the comparatively poor documentation of the British Lower Ordovician (but see Fortey et al., this volume), and of the Victorian Upper Ordovician, provides some serious obstacles. Recent publications on the Arenig–Llanvirn of South Wales (Fortey and Owens, 1987) and the Tremadoc–Arenig interval of the Lake District (Molyneux and Rushton, 1988) provide some help, but the successions exhibit a limited graptolite record. The Gisbornian–Eastonian and Eastonian–Bolindian stage boundaries have now been sampled in some detail (although the work is not yet published) and the only weak points are within the Darriwilian and Gisbornian. The Eastonian is especially useful, providing a link with eastern and western North American, and Scottish, successions, and is better developed than the comparable interval in China and Russia. The Bolindian is also useful, filling what appears to be a gap in the Scottish sequence. The revisions of the British Upper Ordovician now allow correlation at almost zone level with confidence. This is much less so, however, in the Lower Ordovician where British graptolite faunas are less diverse and include endemic elements.

The coralline and shelly faunas of central New South Wales show provincial features which make it difficult to establish direct external correlation with British, North American or other regions. There is a strong link between "Caradoc" trilobite and brachiopod faunas of Kazakhstan and central New South Wales at the generic level, and in the brachiopods with the Scoto–Appalachian province, but all 49 brachiopod species described in Percival's (1980) Ph.D. thesis are new, and more than one-third of the genera are endemic. This means that correlation can be established in broad terms with North American stages (Blackriveran, Trentonian, Edenian, and Maysvillian) and with the Kazakhstan stages (Erkebidaik, Anderken, and Dulankara horizons).

All the trilobite species are new and a few genera are endemic. The closest faunal links are with Kazakhstan and southeast Asia. A less close connection exists with North America and northern Europe, making correlations with stages of these regions difficult. Similar problems exist with the corals and stromatoporoids.

The Tasmanian successions are mainly correlated with the North American cratonic sequences. The terms Ibexian and Whiterockian (including Chazyan) are relatively easily applied. Cincinnatian (Series) can also be used but there are problems in the Blackriveran-Trentonian part of the sequence where macrofauna and microfauna tend to be endemic Australian/North Chinese, or even endemic Tasmanian (Burrett et al., 1983). With further documentation it may be possible to apply Chinese stages. Of particular interest, the Tasmanian conodont *Tasmanognathus* (Table 2) occurs also in northern China, and several Chinese conodont genera and species erected by An et al. (1983) are also found in the Amadeus, Canning, and Tasmanian sequences. However, the Tasmanian species of *Tasmanognathus, T. careyi*, which occurs with *Chirognathus monodactylus* and *Phragmodus* sp. nov. in the "Lower Limestone Member" of the Benjamin Limestone, and occupies a position above *Histiodella sinuosa* and *Phragmodus flexuosus*, and below *P. undatus*, has not been confirmed outside Tasmania. Only different species of *Tasmanognathus*, possibly with somewhat different ranges, have so far been reported from North China.

The external correlation of the central and northern Australian platform sequences has depended on which fossil group is used. For example, for conodont-dominated assemblages, the Baltoscandian stages are employed; for graptolite-dominated horizons, the Victorian stages; for nautiloid occurrences, the North American subdivisions; and for the trilobite sequences, the informal subdivisions of the British Arenig and Llanvirn Series. However, it should be pointed out that trilobites in the lowermost Ordovician(Tremadoc through lower Arenig) of these platform sequences are essentially similar, even to the species level in some instances, to faunas of eastern Asia, particularly the Sino–Korean Platform. Conodonts represent a mixture of North American Midcontinent and similar East Asian affinities at this time.

MAJOR FAUNAL EVENTS OF USE FOR EXTERNAL CORRELATION

In the Australasian (mainly Victorian) graptolite facies, the following faunal events may be regarded as having significance for international correlation (Table 1); the first six events are recorded in Victoria and New Zealand successions, the remaining five events in Victoria only:

1. Appearance of *Adelograptus* and a diverse dendroid fauna in the La2 Zone of the Lancefieldian. Distinctive elements of this fauna useful for correlation include *Dictyonema (Araneograptus) pulchellum, D. (A.) macgillivrayi, Adelograptus (Kiaerograptus) antiquus, A. (K.) pritchardi, Temnograptus*, large clonograptids including the species *magnificus, rigidus, tenellus*, and *flexilis*, and *Tetragraptus decipiens*. The precise order of appearance of these forms has not been documented but they all make their first appearance over a relatively narrow stratigraphic interval. Similar faunas are recognized in western and eastern Canada, Norway, and Sweden.
2. Appearance of *Tetragraptus (Paratetragraptus) approximatus*. This distinctive species has a worldwide distribution and is generally taken (e.g., Skevington, 1963) to mark the base of the Arenig Series. Often associated with *T. (P.) approximatus* is *T. (P.) acclinans* (or a related form).
3. Appearance of *Tetragraptus (Pendeograptus) fruticosus* and a diverse dichograptid fauna at the base of the Bendigonian (Be1 Zone). This is possibly the most easily recognized horizon in the entire Australasian Ordovician, as it marks the appearance of the following genera: *Goniograptus, Loganograptus, Sigmagraptus, Trichograptus, Schizograptus, Trochograptus, Dichograptus*, and *Pendeograptus*. In addition, several distinctive species of *Didymograptus* (s.l.), *Tetragraptus* (s.l.), *Phyllograptus* (s.s.), and *Clonograptus* make their appearance. The influx marks a major event in the development and evolution of Ordovician graptolites, although the precise horizon of entry of *T. (P.) fruticosus*, with respect to that of other genera, is not yet documented.
4. Appearance of *Isograptus victoriae* or other datum within the *I. victoriae* lineage. The appearance of isograptids is a prominent feature of Lower Ordovician graptolite successions around the world. The succession of subspecies of *I. victoriae, (lunatus–victoriae–maximus–maximodivergens)* is one of the few evolutionary bioseries that has been statistically analyzed, making it exceptionally valuable for precise correlation.

5. The appearance of *Oelandograptus austrodentatus* and other biserial forms at the base of the Darriwilian is a prominent event in Australasia and takes place in an interval during which there is a major change from a dichograptid-dominated fauna (Bendigonian to Yapeenian) to a diplograptid (s.l.) dominated fauna (upper Darriwilian to Bolindian). However, the appearance of "biserial" scandent forms is not a sudden event. Some, such as glossograptids (*Apiograptus, Paraglossograptus*, and *Pseudisograptus*) appeared earlier, in the Yapeenian. *Glossograptus* itself does not appear quite at the base of Da1, and it is only in Da3 (a poorly documented interval probably divisible into several zones) that there is a real profusion of new diplograptaceans (*"D". decoratus, "A". confertus, "A". differtus, "A". modicellus*, and probably others). Nevertheless, the base of the Darriwilian can be correlated on the basis of the appearance of the *austrodentatus* group (including the overseas subspecies *O. austrodentatus americanus, O. a. oelandicus, O. a. sinodentatus*, etc.).
6. The appearance of *Nemagraptus gracilis* at the base of the Gisbornian. Species of *Dicellograptus* and *Dicranograptus* appear at the top of the preceeding Darriwilian Zone Da4. In both Australia and New Zealand the appearance of dicellograptids is stratigraphically a well marked event, although relatively few species are involved. The appearance of *Dicellograptus* together with *Dicranograptus* has traditionally been taken as defining the base of the Gisbornian Stage. In both Victoria (VandenBerg, *in* Webby et al., 1981) and New Zealand (Cooper, 1979), however, the worldwide zone fossil *Nemagraptus gracilis* appears at a later horizon, (contrary to the opinions of most earlier workers in the two regions), and Cas and VandenBerg (1988) have redefined the base of the Gisbornian Stage to coincide with the base of the *N. gracilis* Zone. The appearance of *Dicellograptus* and *Dicranograptus* is the most readily recognized event in the Australasian sequence but the appearance of *N. gracilis* is more reliable for precise international correlation. This has been discussed at length by Finney and Bergström (1986).
7. The appearance of *Climacograptus spiniferus* within the Eastonian (useful for correlation with northeastern North America and Scotland).
8. Appearance of *Dicellograptus gravis (=D. alector)* in the Ea4 Zone of the Eastonian (useful for correlation with western North America and Scotland).
9. Appearance of *Paraorthograptus pacificus* in a diverse fauna with *Dicellograptus ornatus* and *Climacograptus latus* (Bo2 Zone of the Bolindian).
10. The appearance of *Glyptograptus extraordinarius* in an impoverished fauna (Bo3 Zone of the Bolindian).
11. Appearance of *Glyptograptus persculptus* (now defining the topmost Bo4 Zone of the Bolindian).

There are a number of important conodont tie points that further enhance the usefulness of the Victorian graptolite scheme for external correlation (Table 1). Firstly, Stewart (*in* Cas and VandenBerg, 1988) has reported a mixed association of Bendigonian (Be1 Zone) graptolites at Dargo in Victoria with North Atlantic-type conodonts – occurrences of *Paracordylodus gracilis, Oepikodus communis, Prioniodus elegans*, and *Paroistodus proteus* – allowing the Be1 Zone to be equated with the *P. elegans* Zone of the North Atlantic conodont province (Bergström, 1986). Secondly, there are Victorian graptolitic sequences ranging in age from Bendigonian (Be3 Zone) to Chewtonian (Ch2 Zone) with conodonts of the North Atlantic *Oepikodus evae* Zone – the fauna includes *O. evae, Bergstroemognathus extensus, Protoprioniodus aranda*, and *Scolopodus rex*. Thirdly, *Pygodus anserinus*, which partially overlaps the range of *P. serra* in the Victorian sequences, defines the North Atlantic-type conodont *P. anserinus* Zone, and extends from the upper part of the *Hustedograptus* (= *Glyptograptus*) *teretiusculus* Zone (Darriwilian, Da4 Zone) to the middle of the *Nemagraptus gracilis* Zone (Gisbornian, Gi1 Zone). It is associated with short-ranging graptolites including *Dicellograptus divaricatus* and *Reteograptus geinitzianus*, as well as *N. gracilis* (Stewart, *in* Cas and VandenBerg, 1988).

The major faunal incomings recognized in the Upper Ordovician successions of central New South Wales – for example, first appearances of particular coral and stromatoporoid groups – seem, in contrast, to have limited usefulness in external correlation. There is evidence to suggest that some faunal elements, like the clathrodictyid stromatoporoids, appeared first in shallow waters of the offshore, volcanic island arc complex of central New South Wales, and then dispersed slowly globally to other shallow water habitats, including the continental platforms (Webby, 1980, 1985).

There do not appear to be any major faunal events in the Tasmanian succession that prove to be useful for external correlation. However, there are some conodonts like *Tasmanognathus*, as mentioned previously, which may prove to be of particular use for precise correlation (Table 2).

In the platform sequences of northern Australia there are also a series of faunal events which appear to be of particular value, as follows:

1. The incoming of the conodont *Cordylodus proavus* fauna, which initiates the Datsonian, can be widely correlated. However, it is now likely that the Cambrian–Ordovician boundary will be ratified internationally at a higher level, close to the base of, or within the upper Datsonian (C3) Zone of *Cordylodus oklahomensis-C. lindstromi*; i.e., associated with a slightly later event, either the appearance of *Cordylodus intermedius* (within the *C. oklahomensis* Subzone of Miller, 1984) or of *C. lindstromi* (*C.lindstromi* Subzone of Miller, 1984).
2. The appearances of the scolopodids and chosonodinids, marking the beginning of the Warendian.

3. Although not adequately documented, but possibly of importance, are the first appearances of *Histiodella*, ambonychiid bivalves, and the trilobite *Carolinites*. This latter is known to occur in central and western Queensland (Henderson, 1983; Fortey and Shergold, 1984), Tasmania (Jell and Stait, 1985), central Australia (Laurie, work in progress), northwestern Australia (Legg, 1976), and western New South Wales (Warris, 1969). Although more work needs to be done, it appears that most of the species of this genus are pandemic and will be useful for detailed intra- and intercontinental correlation.
4. The trend to gigantism in the faunas apparently occurs close to the "Arenig-Llanvirn" boundary. This is most noticeable in the bivalves, nautiloids, and trilobites of the middle Stairway Sandstone of the Amadeus Basin, and the Mithaka Formation of the Georgina Basin (Shergold, 1986). However, possible ecological factors need to be further assessed, as well as occurrences in other regions.

Unfortunately, the conodont faunas of central and northern Australia are most closely related to North American Midcontinent faunas, not to North Atlantic provincial types. Consequently, the conodont zonation scheme developed in the graptolite facies of Victoria will not be that easily applicable to the Amadeus, Canning or Georgina basin successions.

APPLICABILITY OF OVERSEAS SERIES NAMES IN AUSTRALASIA

The British series are the most widely used of the overseas series schemes for correlation, particularly in the graptolite facies. For instance, they can be used with varying levels of ambiguity in the Victorian and New Zealand successions (Table 1):

1. The base of the Tremadoc cannot be located because it lies below the lowest planktic graptolites (Assemblage La2) of the region.
2. Correlation of the base of the Arenig depends on definition of the Arenig in Britain. It is either at the base of the *Tetragraptus approximatus* Zone La3, where it is generally taken (e.g., Skevington, 1963; Strachan, 1972; Molyneux and Rushton, 1988), or approximating the base of the Bendigonian (Fortey, 1984; Fortey and Owens, 1987).
3. The base of the Llanvirn has been defined recently at Llanfallteg, near Whitland in South Wales (Fortey and Owens, 1987), at the incoming of pendent didymograptids. They are preceded in the uppermost Arenig by the first biserial scandent forms, including *Undulograptus austrodentatus*, which defines the base of the Darriwilian, so the best correlation is probably with the base of the Da2 Zone (see also Skevington, 1963), although the base of the Da1 Zone has also been suggested (Cooper and Fortey, 1982, Figs. 2, 3).
4. The base of the Llandeilo is customarily taken at the base of the *Hustedograptus* (= *Glyptograptus*) *teretiusculus* Zone, that is, at the base of the Darriwilian Da4 Zone.
5. The base of the Caradoc may correlate with the base of the Gisbornian as redefined by Cas and VandenBerg (1988) – however, the base of the Caradoc does not coincide, apparently, with the appearance of *N. gracilis* in British successions.
6. The improved correlation between the shelly and graptolite facies in Britain now allows much more accurate correlation of the Caradoc–Ashgill boundary with Victoria. Whittington et al. (1984) place this within the *Pleurograptus linearis* Zone. In Victoria the lowest occurrence of *P. linearis* is in the Ea3 Zone of *Dicranograptus hians kirki*. The Ea4 index fossil *Dicellograptus gravis* occurs at Girvan (Scotland) in limestones with a mixture of Pusgillian (lowermost Ashgill) shelly fossils and *D. complanatus* Zone graptolites. Clearly, the base of the Ashgill correlates with a level within the Ea3 Zone of Victoria.
7. Traditionally the top of the Bolindian has been drawn below the level of the *Glyptograptus persculptus* Zone. However, now that the base of the Silurian (and base of the Llandovery) is established internationally at the base of the *Akidograptus acuminatus* Zone (Cocks, 1985), the *G. persculptus* Zone may be included as the topmost (Bo4) Zone of the Bolindian (VandenBerg and Webby, 1988).

The North American graptolite subdivisions established by Berry (1960) and Riva (1974) can also be applied in Victoria but unfortunately they cannot be correlated easily with the standard North American series, which are based on shelly faunas. Correlations with China and Scandinavia are only of limited use because of endemism in the respective graptolite faunas of these regions.

The British series are preferred in New South Wales, but can only be applied easily where associated graptolites occur. For instance, it is possible to establish the approximate position of the Caradoc–Ashgill boundary in the graptolitic Malongulli Formation of the Cliefden Caves area of central New South Wales (Webby and Packham, 1982; Rigby and Webby, 1988).

The North American series (and their subdivisions) are the most useful means of establishing external correlation of Tasmanian successions. Little direct correlation can be made with British series subdivisions. The Tremadoc, and less confidently, the Arenig and Ashgill can be used, but the Llanvirn through Caradoc interval cannot be applied as yet. This is mainly because none of the more useful trilobites and brachiopods, and few graptolites and conodonts, are in common. The deeper slope Surprise Bay sequence in southern Tasmania can be correlated, but it cannot be closely related to other Tasmanian successions, which are of mainly shallow water aspect.

The British series are also not applied so easily to platform successions of northern Australia except where graptolites provide a datum. The Canning Basin has the most continuous graptolite record, through the Arenig to Llanvirn interval, but the sequences are of limited accessibility, since they are preserved mainly in the subsurface.

SUBSYSTEMIC DIVISIONS IN AUSTRALASIA

A tripartite subdivision of the Ordovician has been widely used, following Harris and Thomas (1938). It has restricted the Middle Ordovician to the Darriwilian stage only, and consequently has meant that the terms Lower, Middle and Upper do not coincide with any of the several other subsystemic schemes in use around the world. The disparity in use between these tripartite schemes has commonly led to confusion and is most undesirable. VandenBerg (*in* Webby et al., 1981, and *in* Cas and VandenBerg, 1988) noted that prior to Harris and Thomas' (1938) subdivision, Middle Ordovician was not used in Victoria, and he has recommended a reversion to the former two-fold subdivision of Lower and Upper, the boundary being placed at the base of the Gisbornian, that is, at the base of the *N. gracilis* Zone (Tables 1, 2). This subdivision is also recommended for New South Wales, Queensland, and New Zealand successions. It is important that the subsystemic division of the Ordovician be standardized around the world, and we recommend that this horizon and subdivision be adopted globally.

Although it has been traditional to use a tripartite subdivision in the platform sequences of northern Australia, the usage has depended on what fossil groups were being employed as controls. It is now suggested that a two-fold subdivision should also apply. At present the faunal record suggests that only Lower Ordovician (formerly Lower and Middle Ordovician) deposits are represented, although there are, in some sections (e.g., in the Amadeus Basin, Shergold, 1986), substantial thicknesses of unfossiliferous beds above the last age-diagnostic fossil assemblage, which may imply that the sedimentary record continues partly into the Upper Ordovician.

In Tasmania, Lower, Middle, and Upper have been used approximately as in North America, that is, the Lower = Canadian (now Ibexian), the Middle = Champlainian (Whiterockian and Mohawkian), and the Upper = Cincinnatian. This subdivision is therefore quite different from that employed in Victoria, and elsewhere in Australia and New Zealand. It seems essential therefore that if an acceptable, unified, bipartite subsystemic subdivision of the Ordovician of Australasia is to be achieved, with its boundary at the base of *N. gracilis*, then this level must be recognized also in Tasmania. Perhaps this can be done 1) by establishing closer zonal ties through the boundary interval, with other sequences exhibiting mixed graptolite, shelly and coralline facies, like those in central New South Wales, or 2) by recognizing links between the Tasmanian conodont succession, especially through the interval containing *Tasmanognathus*, and equivalent mixed conodont and graptolite facies of North China.

LOCAL STAGES OR A SINGLE STAGE CLASSIFICATION FOR AUSTRALASIA

A single stage classification is thought by most respondents to be desirable but probably impractical because the cratonic sequences have so little in common with the mainly graptolitic successions of the Tasman Fold Belt, and New Zealand. One contributor, however, argues that a single stage classification is not desirable since it is likely to obscure geological events. At this present stage of development of the Ordovician biostratigraphy of Australasia, it is probably useful to establish a number of local stage schemes in order to correlate the various platformal, as well as basinal, successions. Local conodont-based stages (the Datsonian and Warendian) have been introduced in northern Australian platform areas, and this scheme will probably be extended upward to cover almost the entire Lower Ordovician (Tremadoc–Llanvirn) interval. There is also a proposal by Burrett to introduce an independent set of local stages, based on the Florentine Valley sequence, for correlation of the "middle Tremadoc-Llandovery" interval in Tasmania. It is to be hoped that the definitions of these new local stages will be accompanied by the fullest possible documentation of the contained biotas, and that they will have more than local application.

If, alternatively, we choose only to maintain a single stage classification, it would undoubtedly be based on the Victorian graptolite scheme. However, this could not be seriously contemplated unless we were able to establish many more faunal ties to the shelly faunal and conodont-based platform schemes, even if the Victorian scheme became adopted as the international standard for the areas of the world exhibiting graptolitic successions of Pacific provincial type.

SINGLE INTERNATIONAL OR LOCAL SERIES CLASSIFICATION FOR AUSTRALASIA

If the British series are more adequately defined than at present, they may provide a workable subdivision for use in Australasia. However, they must be designated within boundary stratotypes in continuous sections, so that they can be clearly and unambiguously applied, through the graptolite facies, to successions in Australasia. It seems highly desirable that a single series classification be adopted in Australasia, and the British subdivisions are on balance probably the best available. The only viable alternative could be to establish new Australasian series.

PHYSICAL CRITERIA FOR USE AS TIME MARKERS

There is little stratigraphically controlled geochronological data presently available for the Australasian region. The K/Ar age of 455 ± 10 Ma on the Doubtful River Gabbro in the Snowy Mountains area of southern New South Wales (Owen and Wyborn, 1979) is one of the few useful determinations (Table 2). This sill-like structure intruded the upper Darriwilian-lower Gisbornian Temperance Formation prior to its complete lithification. It is related to the overlying Nine Mile Volcanics, which in its upper part has an upper Gisbornian graptolite assemblage and a solitary rugose coral, establishing its upper limits, and thus suggesting that the intrusive event occurred about the middle of the Gisbornian. This is about one graptolite zone younger than the level of the proposed

Lower–Upper Ordovician subsystemic boundary. There is currently no available magnetostratigraphic or chemostratigraphic data for the Australasian region.

GENERAL RECOMMENDATIONS

1. That we need to formally designate the various stage boundary *stratotypes* for the Victorian graptolite succession.
2. That we should focus immediate attention on the *graptolites* and the *conodonts,* particularly their ranges in Australasian successions.
3. That we should aim to establish precise *ties* between zonal indices – between North Atlantic-type conodonts and Pacific-type graptolites, and between North Atlantic and North American Midcontinent province conodonts in Australia and New Zealand, and secondarily to achieve the closest possible conodont and graptolite ties with China and other regions of central and southeastern Asia.
4. That we recommend establishing a two-fold subsystemic division of the Ordovician, with the base of *Nemagraptus gracilis* defining the base of the upper division.
5. That following definition of stratotypes and establishment of closest possible ties with conodont zonal indices, we propose the Victorian succession as the international standard for use in regions of the world exhibiting Pacific provincial-type graptolites.
6. That in establishing additional local stage schemes, for the post-Warendian interval of platform areas of northern Australia, and possibly through the Ordovician of Tasmania, we should aim at providing fullest possible documentation of biotas, and of correlation with existing local stage schemes.

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New graptolite and lithostratigraphic evidence from the Cincinnati region, U.S.A., for the definition and correlation of the base of the Cincinnatian Series (Upper Ordovician)

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Abstract

Recent study of several drill cores and surface sections has produced significant new biostratigraphic and lithostratigraphic data from the uppermost Middle Ordovician (Mohawkian Series) and lowermost Upper Ordovician (Cincinnatian Series) in the Cincinnati region, the type area of the Cincinnatian Series. A succession of diverse graptolite faunas recently discovered within the Sebree Trough permits correlation with the New York-Quebec Utica Shale graptolite zonation. The base of the Cincinnatian lies within the Climacograptus (Diplacanthograptus) spiniferus *Zone, 23 to 26 m below the base of the superjacent* Geniculograptus pygmaeus *Zone in sections along a 150 km traverse from eastern Indiana to southern Ohio and northern Kentucky. The conodont biostratigraphy is in general agreement with that based on graptolites.*

Examination of temporal morphological trends exhibited by Geniculograptus typicalis *and* G. pygmaeus *through two sections provides supporting evidence for the proposed correlations. Detailed study of geophysical logs suggests that it is possible to recognize, and trace across the Cincinnati region, stratigraphic intervals as little as 5 to 10 m thick. These lithological correlations produce units that parallel a series of K-bentonite beds and correlations based on graptolites and conodonts. Thus, the sequence studied along much of our traverses has a ''layer-cake'' stratigraphy of considerable lateral uniformity, particularly in the Kope Formation and the Lexington Limestone.*

Résumé

De nouvelles données biostratigraphiques et lithostratigraphiques importantes sont tirées de l'étude récente de plusieurs carottes de forage et coupes de surface provenant de la partie sommitale de l'Ordovicien moyen (série du Mohawkien) et de la partie basale de l'Ordovicien supérieur (série du Cincinnatien) près de Cincinnati, région type de la série du Cincinnatien. Grâce à la découverte récente d'une succession de faunes à graptolites variées dans la cuvette de Sebree, il est possible d'établir une corrélation avec la zonation des graptolites des schistes argileux d'Utica dans l'État de New York et la province de Québec. La base du Cincinnatien se situe dans la zone à Climacograptus (Diplacanthograptus) spiniferus, *soit 23 à 26 m sous la base de la zone susjacente, à* Geniculograptus pygmaeus, *dans des coupes le long d'une traverse de 150 km allant de la partie est de l'État d'Indiana jusque dans le sud de l'Ohio et le nord du Kentucky. La biostratigraphie des conodontes s'accorde généralement avec celle des graptolites.*

L'examen des tendances morphologiques temporelles que présentent Geniculograptus typicalis *et* G. pygmaeus *dans deux coupes, fournit des indices à l'appui des corrélations proposées. L'étude détaillée de diagraphies géophysiques porte à croire qu'il est possible de reconnaître des intervalles stratigraphiques*

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qui n'ont que 5 à 10 m d'épaisseur, et de les retracer à travers la région de Cincinnati. Ces corrélations lithologiques produisent des unités qui sont parallèles à une série de lits de bentonite potassique et d'unités fondées sur des graptolites et des conodontes. La séquence étudiée sur une grande partie de nos traverses présente donc une stratigraphie «étagée» qui a une grande uniformité latérale, notamment dans la formation de Kope et le calcaire de Lexington.

INTRODUCTION

In recent years, considerable effort has been devoted to the establishment of a set of global series for the Ordovician System. Although some progress has been achieved, the many problems associated with building a standard set of divisions from a disparate group of regional series remain to be solved. Providing precise definitions for bases of the existing Ordovician series in terms of zonations useful for intercontinental correlation is one of the fundamental steps that must be taken. This step is particularly difficult in the case of those series, such as the Cincinnatian Series, that are defined in shallow water epicratonic sequences dominated by largely endemic shelly fossil assemblages. This paper presents new data obtained in our efforts to integrate the Cincinnatian Series, which is the North American standard for the Upper Ordovician, with regionally recognizable graptolite and conodont zonal units.

Although no formal stratotype has been designated for the Cincinnatian Series, its base coincides with that of the Edenian Stage. Traditionally, the base of the Edenian is placed at the base of the Kope Formation (formerly known as the Eden Shale) in the vicinity of the city of Cincinnati, Ohio. However, this lithostratigraphic boundary is not readily definable in terms of shelly fossil zones useful for interregional correlation. Furthermore, this level is no longer exposed in or near Cincinnati. Nevertheless, work on conodonts and other fossils is improving the situation. Sweet and Bergström (1971, 1976) and Sweet (1979, 1984, 1988) suggested, on the basis of conodont studies, that the base of the Edenian Stage is correlative with the base of the Cobourg Limestone, as developed in the Mohawk Valley of New York, the region of the reference standard for the upper Middle Ordovician Trenton Group and the Mohawkian Series. Riva (1969a, 1974) and Rickard (1973) suggested that the base of the *Climacograptus* (*Diplacanthograptus*) *spiniferus* Zone in the Utica Shale of New York and Quebec also is probably coeval with the base of the Cobourg Limestone in the Mohawk Valley. This conclusion is further supported by Cisne and Chandlee (1982), based on tracing of bentonite beds between Trenton Falls and the Mohawk Valley. In a recent regional subsurface study, Bergström and Mitchell (1989) concluded that the base of the Cincinnatian is coeval with a level in the basal 6.5 to 10 m of the Cobourg Limestone of New York. Thus, the position of the base of the Cincinnatian Series is still not precisely known within the eastern North American graptolite succession. In order to facilitate intercontinental correlation, we would prefer to have a more direct biostratigraphic definition of the base

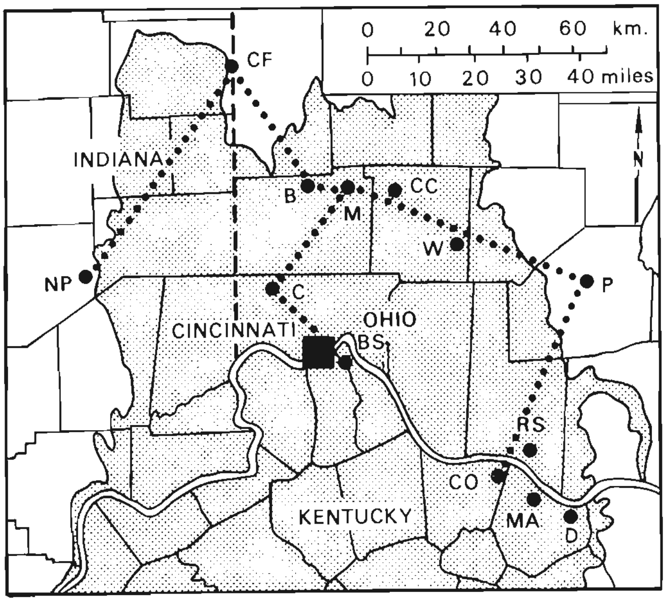

Figure 1. Sketch-map showing location of drilling sites and surface sections examined in the present study. Stippled pattern denotes outcrop area of Ordovician rocks in the Cincinnati region. Straight dotted lines between localities are traverses illustrated in Figures 10 and 11. Localities are as follows: **NP**, New Point well (Indiana Geological Survey Drill Hole 124), Decatur Co., Indiana (Gray, 1972, Fig. 9; Sweet, 1979, p. G24); **CF**, Cook Farm well (Indiana Geological Survey Drill Hole 57), Wayne Co., Indiana (Gray, 1972, Fig. 7); **B**, Beiser well, Butler Co., Ohio (Stith, 1986, Pl. 2 [Permit No. 2]); **M**, Middletown well, Butler Co, Ohio (Bergström and Sweet, 1966; Sweet, 1979, p. G23; Stith, 1986, Pl. 2 [Permit No. 4]); Crosby well, Hamilton Co., Ohio (Ohio Geological Survey, unpublished); **BS**, Brent section, Campbell Co., Kentucky (Hoge, 1985); **CO**, Cominco American well, Mason Co., Kentucky (Sweet et al., 1974; Sweet, 1979, p. G24; Stith, 1986, Pl. 3); **MA**, Maysville section, Mason Co., Kentucky (Carpenter and Ory, 1961; Evans, 1981; Bergström and Mitchell, 1986); **D**, Dravo well, Alpha Portland Cement Company, Mason Co., Kentucky (Ohio Geological Survey, unpublished); **RS**, section along Hwy 68 just northeast of Ripley, Brown Co., Ohio (Rytel, 1982); **P**, Penn well, Highland Co., Ohio (Stith, 1986, Pl. 2, section D [Permit No. 4]); **W**, Washington well, Warren Co., Ohio (Stith, 1986, Pl. 2 [Permit No. 8]); **CC**, Clear Creek well, Warren Co., Ohio (Stith 1986, Pl. 2 [Permit No. 7]). The Indiana cores are kept at the Indiana Geological Survey, Bloomington, Indiana, and the Ohio cores at the Ohio Division of the Geological Survey, Columbus, Ohio. The Cominco core and one half of the New Point core are stored at the Department of Geological Sciences, The Ohio State University, Columbus, Ohio.

of the Cincinnatian Series in its type area, particularly one based on graptolites or other fossils with an intercontinental distribution in this interval.

In the present paper we summarize new data obtained from an ongoing study of graptolites and graptolite biostratigraphy of the Cincinnatian and upper Mohawkian strata in the Cincinnati region. Space limitations make it necessary to omit much detailed information now available. In addition, some of the conclusions presented here are preliminary. A more detailed account, including descriptions of the graptolite species encountered and a more rigorous statistical treatment of their stratophenetic patterns, will be presented in a planned monograph.

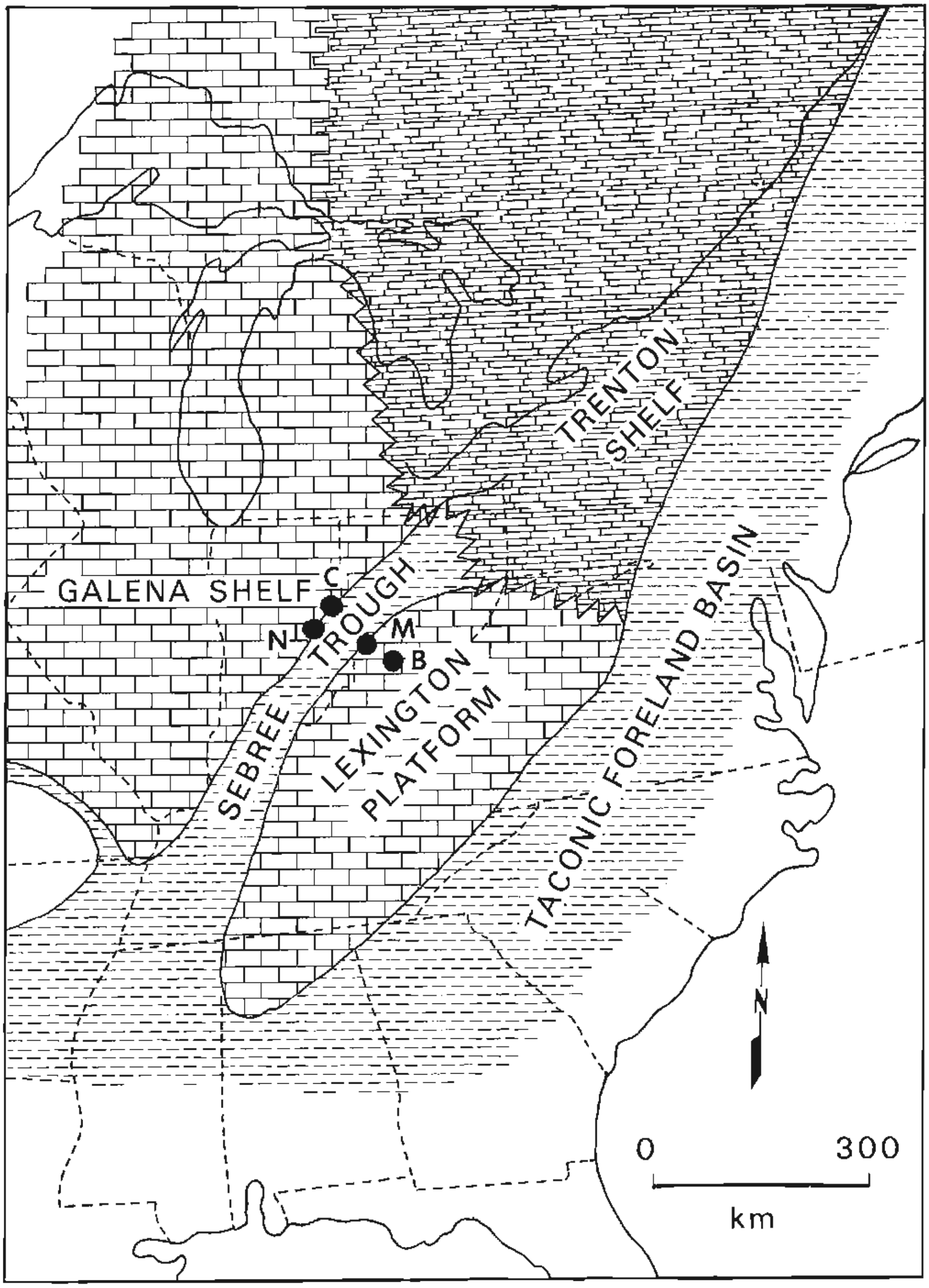

Figure 2. Sketch-map showing major paleogeographic features in east central United States during late Middle Ordovician time (modified from Keith, 1985, Fig. 5). Note location of four of our key sections relative to the Sebree Trough. Letter designations are as follows: N, New Point well; C, Cook Farm well; M, Middletown well; B, Brent section.

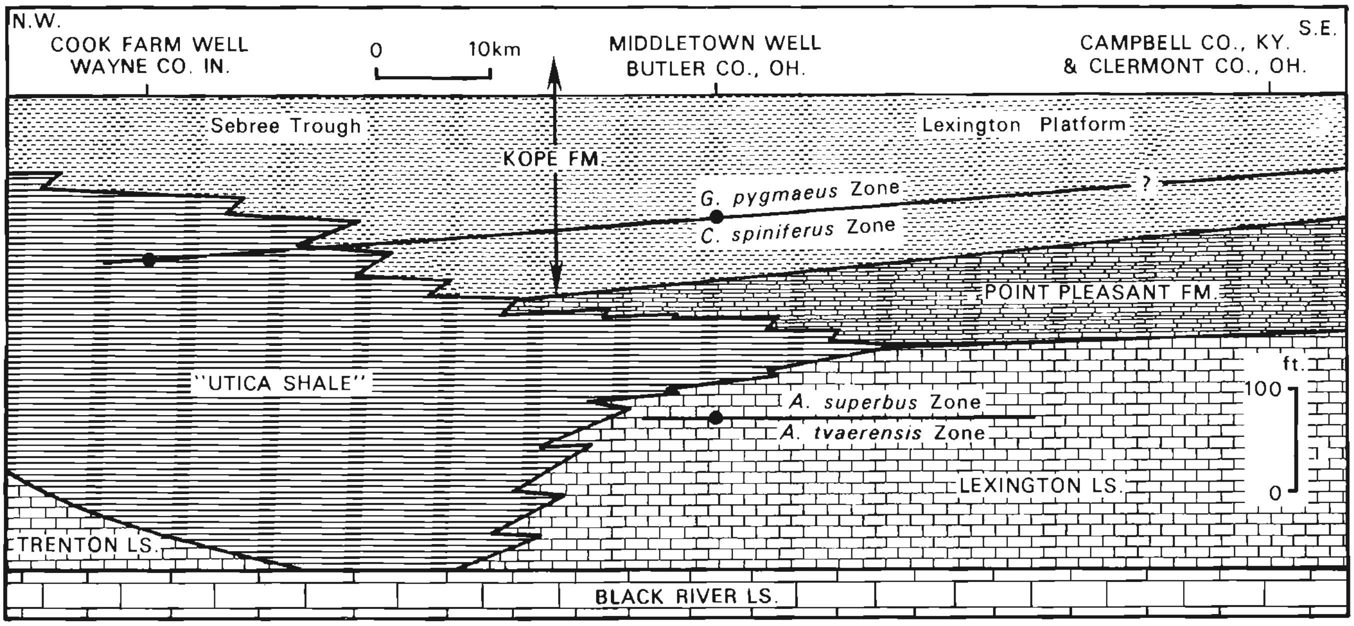

Figure 3. Schematic cross-section from the Cook Farm well to a composite of the Brent section and the Pierce well, Clermont Co., Ohio (Sample 2080 of Stith, 1986, Pl. 1). Note relations between the Sebree Trough "Utica Shale" succession and the Lexington Limestone and Point Pleasant Formation. As indicated in the figure, the distinctive depositional regime of the Sebree Trough was established in post-Blackriveran time and disappeared in early *Geniculograptus pygmaeus* Zone time.

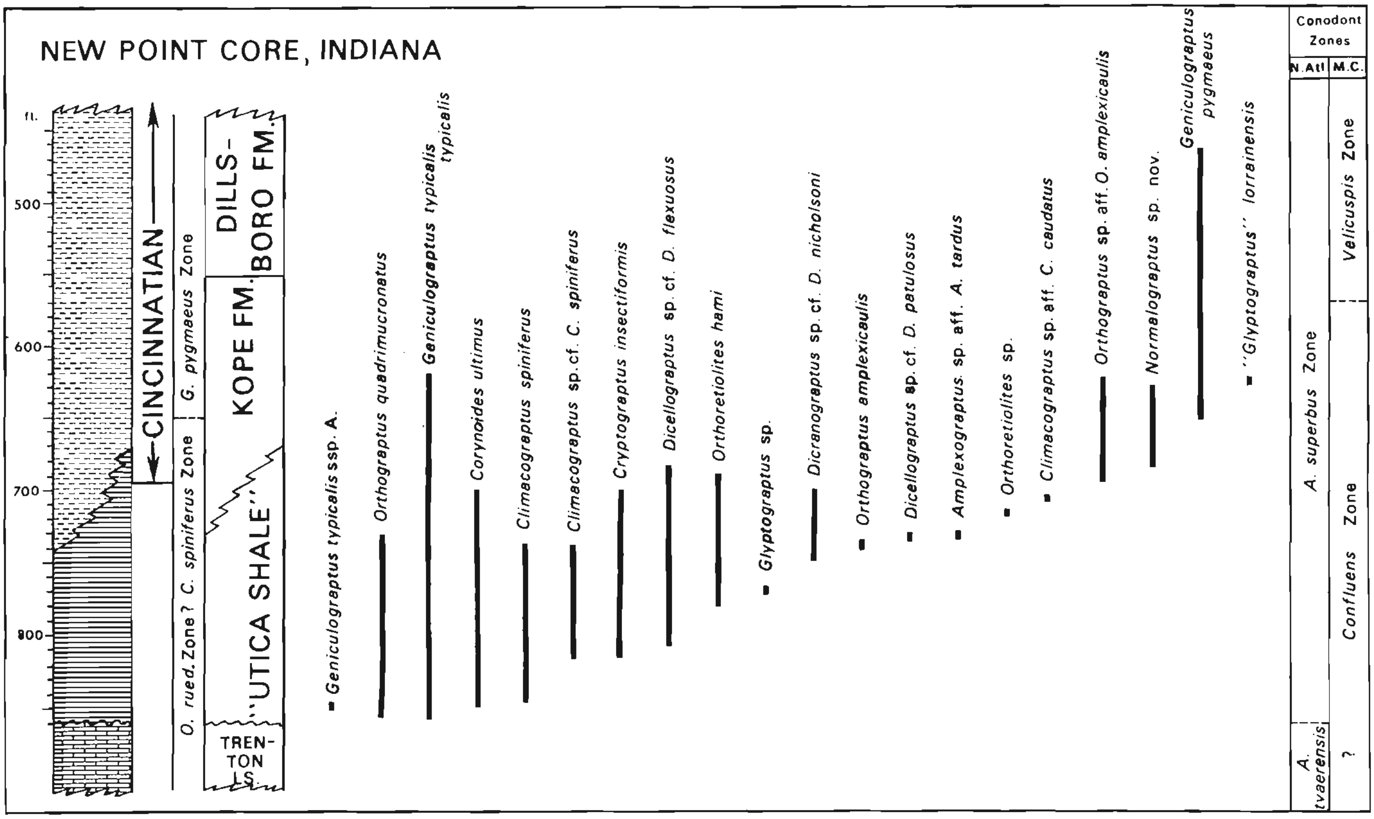

Figure 4. Ranges of graptolites in the "Utica Shale", Kope, and Dillsboro formations in the New Point well. Note that the base of the *Geniculograptus pygmaeus* Zone is in the lower Kope Formation. Projected level of the base of the Cincinnatian is according to Sweet (1979, pers. comm.). Conodont chronozones in righthand column are those of Sweet (1984).

GRAPTOLITES AND CONODONTS FROM THE SEBREE TROUGH

We have discovered a series of varied and well preserved graptolite faunas in a number of drill cores (Fig. 1) from the Cincinnati region. Most of these graptolites occur in several cores that penetrate the Sebree Trough (named the "Sebree Valley" by Schwalb, 1980, and referred to as the "Kope Trough" by Keith, 1985). The Sebree Trough (Fig. 2) is a narrow, and approximately linear, zone of dark shale located in the subsurface on the northwest side of the present Cincinnati Arch. Through their regional lithofacies studies, Cressman (1973) and Keith (1985, 1989) concluded that this feature was a topographic trough in which mud deposition prevailed from early Mohawkian through late Edenian time. At its peak expression in the late Shermanian and early Edenian, the Sebree Trough extended in a northeast-southwest direction through central Ohio, southeastern Indiana, and western Kentucky to western Tennessee. It may have connected with the Ouachita Geosyncline to the south and the Appalachian Basin to the east (Keith, 1989). The origin of this feature is uncertain and several different interpretations have been proposed (e.g., Rooney, 1966; Schwalb, 1980).

Blackriveran age carbonates underlie the Sebree Trough (Fig. 3). Contrary to Rooney's (1966) suggestions, structure contours on the top of the Black River Limestone in Indiana do not exhibit evidence of significant structural offset (Keith, 1986). From these and other features, Keith (1985) inferred that the development of the trough postdates the Blackriveran. Pre-Kope Formation sediments of the Sebree Trough are dominantly medium brown to dark grey, weakly to strongly laminated, silty mudstone and shale with a few thin, shelly layers, and dark, argillaceous limestone and siltstone interbeds. Local drillers have long referred to these sediments as the "Utica Shale", and this designation has been used by several recent authors (e.g., Bergström and Sweet, 1966; Cressman, 1973; Sweet, 1979). The most common shelly fossils are smooth ostracodes and orbiculoid brachiopods, which occur in abundance on occasional bedding planes together with rare conodonts (mostly *Phragmodus* and *Plectodina*). With the exception of chitinous hydroids, and sparse

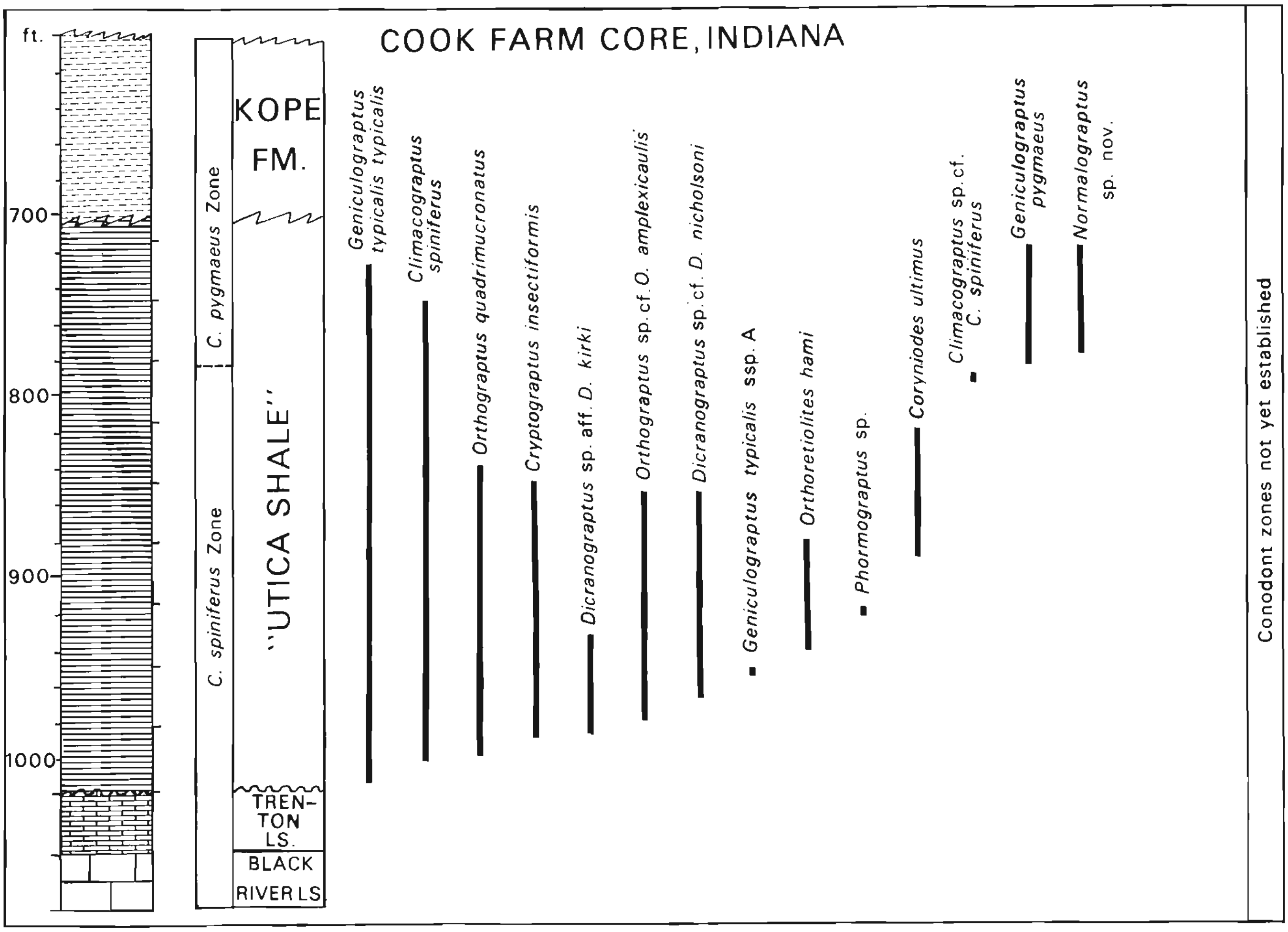

Figure 5. Ranges of graptolites in the "Utica Shale" and Kope Formation in the Cook Farm well. Note that the base of the *Geniculograptus pygmaeus* Zone is within the "Utica Shale", well below the base of the Kope Formation.

Triarthrus and *Cryptolithus* trilobite fragments, benthic fossils are nearly absent. Graptolites occur in moderate abundance throughout the unit and in the lower part of the overlying Kope Formation.

Southeastern Indiana

In most of southeastern Indiana, the "Utica Shale" overlies the calcarenitic grainstones of the Trenton Limestone (see Keith, 1985) with obvious disconformity. We have studied the "Utica Shale" as well as both older and younger units in two cores from the northwestern part of the Sebree Trough in Indiana: the New Point and Cook Farm cores (Fig. 1).

Sweet (1979, locality 31) reported on conodonts from the New Point core and, on the basis of fluctuations in the relative abundance of a suite of conodont species, projected the base of the Shermanian Stage to a level a few metres below the top of the Trenton Limestone in this core. Sweet (1984, 1988) also used this core as one of the reference sections to construct the standard reference section (SRS) for his graphic correlation network. Using Sweet's (1979) projections of the base of the Edenian and Maysvillian stages as a guide to the correspondence between the New Point core and the composite SRS, we have calculated the projected levels of the *Belodina compressa* and *Oulodus velicuspis* chronozones in the New Point core. These are shown in the righthand column of Figure 4. We have made additional conodont collections from the Trenton Limestone in this section. On the basis of these new collections, we place the base of the *A. superbus* Zone about 0.6 m below the top of the Trenton Limestone. Thus, the base of the "Utica Shale" in this section is probably no older than mid-Shermanian.

Graptolite collections from the "Utica Shale" in the Cook Farm and New Point cores (Figs. 4, 5) are taxonomically diverse and specimens are moderately abundant. The lower "Utica Shale" yields graptolites typical of the *Climacograptus*

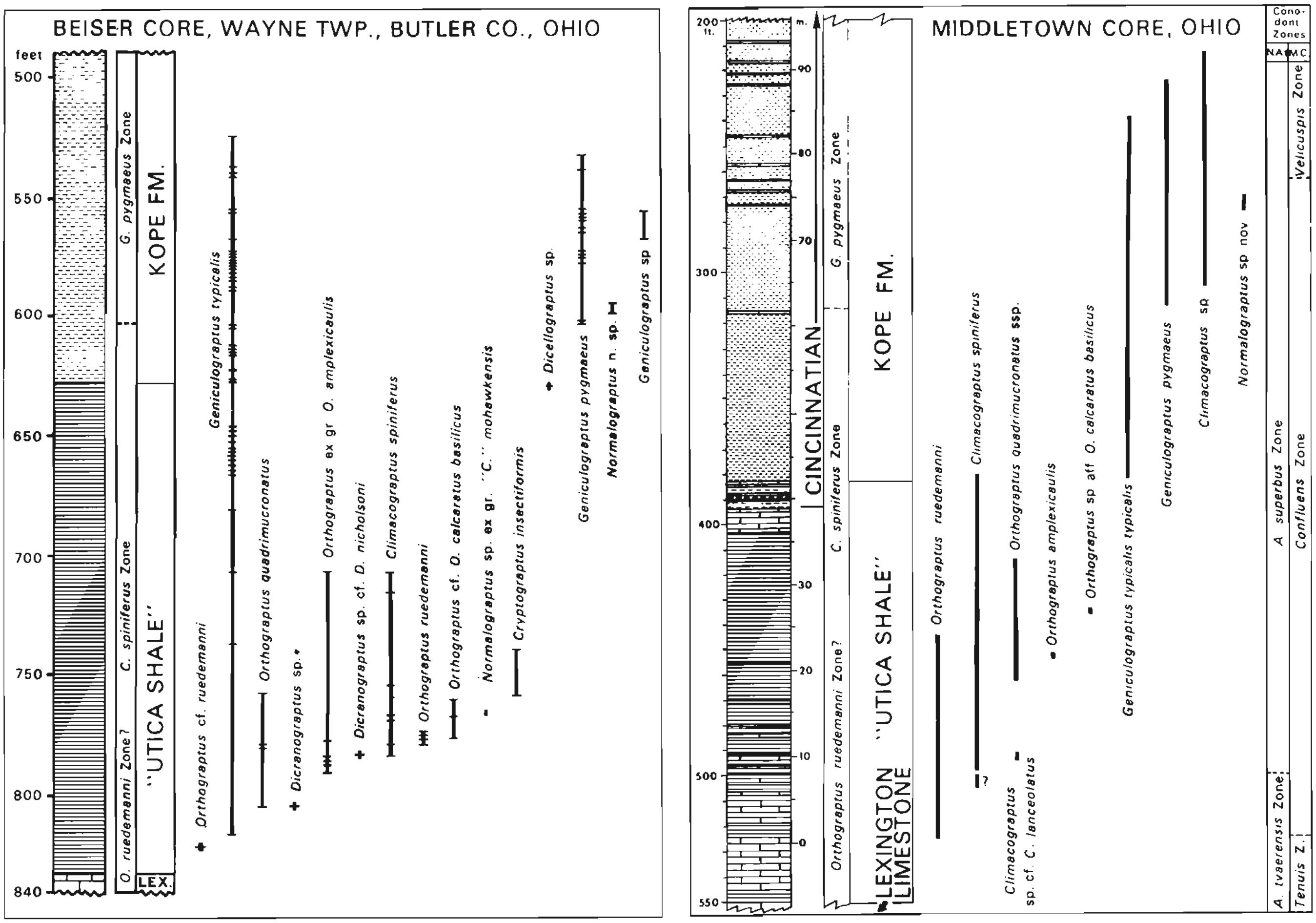

Figure 6. Vertical ranges of graptolite species in the "Utica Shale" and lower Kope Formation in the Beiser and Middletown wells. Note that the base of the *Geniculograptus pygmaeus* Zone is well above the base of the Kope Formation in these wells. Note added in proof: Re-examination of *Climacograptus* specimens in light of VandenBerg (1990) indicates that of the specimens in the Middletown core only those at 43.6 m are certainly *C.* (*D.*) *spiniferus*. Specimens from 6.1 m in the Middletown core are *C.* (*D.*) sp. cf *C.* (*D.*) *spiniferus* and those at 9.9 m are *C.* (*D.*) sp. cf *C.* (*D.*) *lanceolatus* VandenBerg. The range of *C.* (*D.*) *spiniferus* in the Beiser well is accurate as shown, but specimens of *C.* (*D.*) sp. cf *C.* (*D.*) *lanceolatus* occur at 758 feet (231 m) in this core.

(Diplacanthograptus) spiniferus Zone (Fig. 6) in New York and Quebec, including *C. (D.) spiniferus*, *Geniculograptus typicalis*, and *Orthograptus quadrimucronatus*. These species continue upward into the upper "Utica Shale" and are quickly joined by other typical *C. (D.) spiniferus* Zone elements, such as *Orthoretiolites hami*, *Dicellograptus flexuosus*, and *Dicranograptus nicholsoni*.

In addition to the usual *C. (D.) spiniferus* Zone species, this fauna also includes a few specimens of *G. typicalis* that bear a prominent mesial spine on the first theca (Plate 1, figs. 36, 37). Specimens of the spine-bearing *G. typicalis* ssp. A are restricted to the lower 19 m of shale. This interval also yields specimens of *Corynoides ultimus* Ruedemann, which ranges upward into the upper "Utica Shale" (Plate 1, figs. 18, 34). In the Mohawk Valley this species has been found only in the *Geniculograptus pygmaeus* Zone (Ruedemann, 1925; Riva, 1974), but the Sebree Trough and other occurrences indicate that *C. ultimus* first appears in older rocks outside New York (Mitchell, 1989).

In both the Cook Farm and New Point cores (Figs. 4, 5), *Geniculograptus pygmaeus* makes its first appearance in the upper part of the "Utica" interval. *Geniculograptus pygmaeus* makes its debut at 65 m above the top of the Trenton Limestone in the New Point core, which equates to 24 m above the base of the Cincinnatian Series as projected by Sweet (1979, Fig. 2), or 1087 m above the base of Sweet's (1988) composite SRS. *Geniculograptus pygmaeus* and *G. typicalis* range upward through an interval of gradational rock types into the soft, fissile, grey-green shale and argillaceous limestone of the Kope Formation. Although initially rare, *G. pygmaeus* becomes abundant upward in the lower Kope Formation. In the Sebree Trough sequence, as in the Utica Shale of New York and Quebec (Riva, 1969a, p. 520), a minute species with variably glyptograptid to climacograptid thecae (*Normalograptus* sp. nov., Pl. 1, figs. 9-12; see discussion of *Normalograptus* in Melchin and Mitchell, *this volume*) is present in an interval spanning the boundary between the *C. (D.) spiniferus* and *G. pygmaeus* zones. Evans (1981) discovered three-dimensional specimens of this species (identified by him and listed by Bergström and Mitchell, 1986, as *Orthograptus* sp.) together with *G. pygmaeus* in the upper Kope Formation at Maysville, Kentucky. Other noteworthy elements of the *G. pygmaeus* fauna are *Corynoides ultimus*, "*Glyptograptus*" *lorrainensis* (Pl. 1, figs. 7, 8), and specimens of an *Orthoretiolites* species similar to those figured by Riva (1969a, Figs. 4k,l) from this zone in Quebec.

Southwestern Ohio

In southeastern Indiana, the Trenton Limestone thins rapidly to the southeast, and in the centre of the trough shale may rest directly on Blackriveran strata (Fig. 3). On the east side of the trough, in extreme southeastern Indiana and southwestern Ohio, the "Utica Shale" interfingers with dark argillaceous wackestone of the Lexington Limestone (Fig. 10). This relationship is displayed well in two cores that we have studied: the Beiser and Middletown cores (Fig. 6). Conodonts from the Middletown core were described by Bergström and Sweet (1966). On the basis of projections of the Cincinnatian stages into this section supplied to us by Sweet, we have calculated the position of the base of the *Belodina confluens* and *Oulodus velicuspis* chronozones in the Middletown core, and indicated these in Figure 6.

Graptolites occur less frequently in the Middletown and Beiser cores than in the Indiana cores and represent fewer species. Stratigraphically useful faunas are present, however. The middle and upper Kope Formation contains a *Geniculograptus pygmaeus* Zone fauna dominated by the same species as in the more westerly cores discussed above. The first appearance of *G. pygmaeus* in the Middletown core is 62 m above the top of the Lexington Limestone, and 23 m above the projected base of the Cincinnatian Series, or 1086 m above the base of the composite SRS. Thus, the difference in level of first appearance of *G. pygmaeus* in the New Point and Middletown cores (1 m) is substantially less than the present resolution (W = 6 m) of Sweet's graphic correlation network (Sweet, 1984, 1988), and these two occurrences, therefore, are effectively isochronous.

The lower Kope Formation and "Utica Shale" yield a sparse fauna indicative of the *Climacograptus (Diplacanthograptus) spiniferus* Zone (Fig. 6). Near the base of the brown shale interval, and well below Cincinnatian rocks, we find *Geniculograptus typicalis typicalis* and *C. (D.) spiniferus* in the Beiser core, and *Climacograptus (Diplacanthograptus)* sp. cf. *C. (D.) spiniferus* in the Middletown core together with numerous specimens of *Orthograptus ruedemanni* (Pl. 1, figs. 29, 30, 39). We have also discovered specimens in this interval of the Middletown and Beiser cores that resemble *Climacograptus (Diplacanthograptus) lanceolatus* VandenBerg, a member of the *C. (D.) spiniferus* lineage recently recognized in Australia (VandenBerg, 1990, see Pl 1, fig. 43).

The graptolite zonal correlation of these lowest "Utica Shale" beds is problematic. The *Orthograptus ruedemanni* and *Climacograptus (Diplacanthograptus) spiniferus* zones, as defined by Riva (1969a, 1974), are assemblage zones. At numerous localities in the New York-Quebec region, ranges of characteristic elements of the *O. ruedemanni* Zone, such as *O. ruedemanni* and *Normalograptus mohawkensis*, significantly overlap those of several elements of the lower *C. (D.) spiniferus* Zone (e.g., *C. (D.) spiniferus*, *Dicranograptus nicholsoni*, and *Amplexograptus praetypicalis*) (Ruedemann, 1912, 1925; Riva, 1969a,b, 1972). Riva (1969a,b) placed the base of the *C. (D.) spiniferus* Zone within this zone of overlap, essentially at the base of the ranges of *D. nicholsoni* and *Dicranograptus ramosus*. The sequence developed in the Beiser core (Fig. 6) is similar to that seen in the Utica Shale, so we have tentatively identified the presence of the *O. ruedemanni* Zone in this and the nearby Middletown core. However, *Geniculograptus typicalis typicalis*, which in New York and Quebec is reported only from the middle and upper portions of the *C. (D.) spiniferus* Zone, first appears in the Beiser core within the possible *O. ruedemanni* Zone.

Limestone interbedded with the lower "Utica Shale" of the Middletown core yields amorphognathid conodonts. These indicate that the base of the range of *C. (D.) spiniferus* (and possibly also that of *Geniculograptus typicalis typicalis*) lies within the upper part of the *A. tvaerensis* Zone.

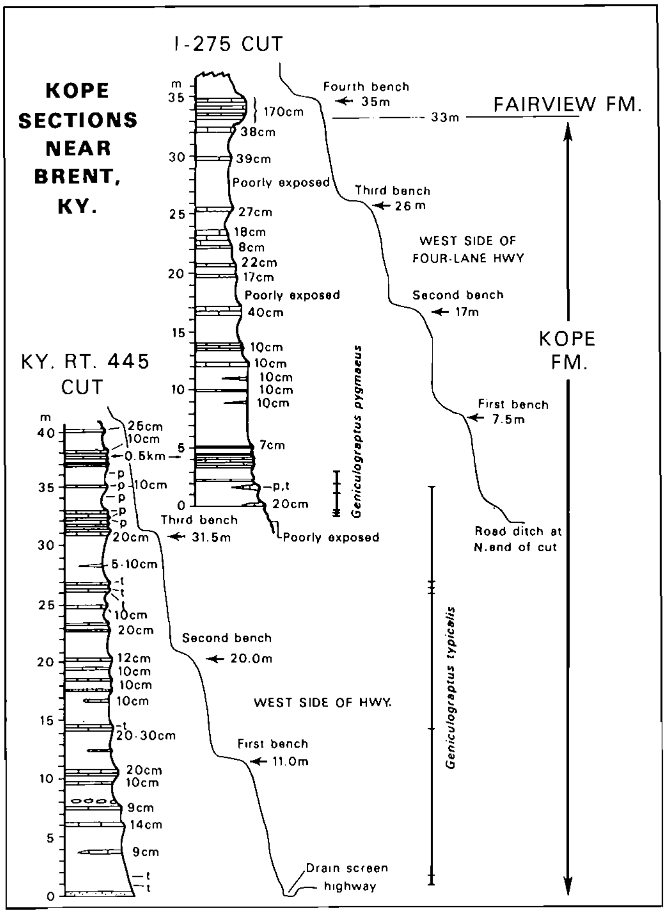

Figure 7. Vertical ranges of graptolite species in the Kope Formation at the Brent section. The correlation between the two partial sections is based on leveling and comparison of the lithological succession. Cm-figures refer to thickness of the 5cm thick limestone beds, and the letters "t" and "p" indicate occurrences of *Geniculograptus typicalis* and *G. pygmaeus*, respectively.

Orthograptus ruedemanni continues upward to 16 m above the base of the *A. superbus* Zone in this core. Thus, *C. (D.) spiniferus* and *G. typicalis typicalis*, which are characteristic of the *C. (D.) spiniferus* Zone, make their debut in the Sebree Trough succession in rocks that appear to be substantially older than the portion of the Utica Shale in which they first appear in the New York-Quebec region. Finney (1986) has described a similar situation in the Lexington Limestone from Kentucky and the lower Viola Springs Formation in Oklahoma, although Riva and Malo (1988) dispute his interpretations. We examine these problems further in our discussion of the correlation of the Sebree Trough succession with that of the Mohawk Valley.

GRAPTOLITES FROM THE KOPE FORMATION NEAR BRENT, KENTUCKY

Bergström and Mitchell (1986) reviewed the data available at that time on the graptolite succession in the Upper Ordovician outcrop belt of the Cincinnati region. In order to try to establish the temporal equivalents of the Sebree Trough succession in the shelly facies of the Cincinnati Arch, we have collected additional data in southwestern Ohio and northern Kentucky from both outcrop sections and drill cores.

Well exposed sections through the entire Kope Formation are rare. The early Edenian Fulton Shale, undoubtedly a tongue of the Sebree Trough "Utica Shale", is no longer exposed (see Bergström and Mitchell, 1986, p. 248, 249). One of the best Kope Formation outcrops near Cincinnati occurs in two adjacent and overlapping road cuts near Brent, Kentucky, approximately 13 km southeast of the centre of the city (Fig. 1). Some 60 m of Kope Formation are exposed although, as usual, the lower 8 to 12 m of the unit are covered (Fig. 7). Exact correlation between the base of the upper section and a level in the lower section is difficult to establish, but the 0 m level in the upper section corresponds approximately to the 33 m level in the lower section. Hoge (1985) studied the distribution of conodonts and brachiopods in this section, provided a detailed description of the lithological succession, and discussed its graphic correlation with Sweet's (1984) composite SRS.

So far, we have found only two graptolite species in the Kope shale and limestone at the Brent sections. *Geniculograptus typicalis typicalis* occurs at 0.9 m above the base

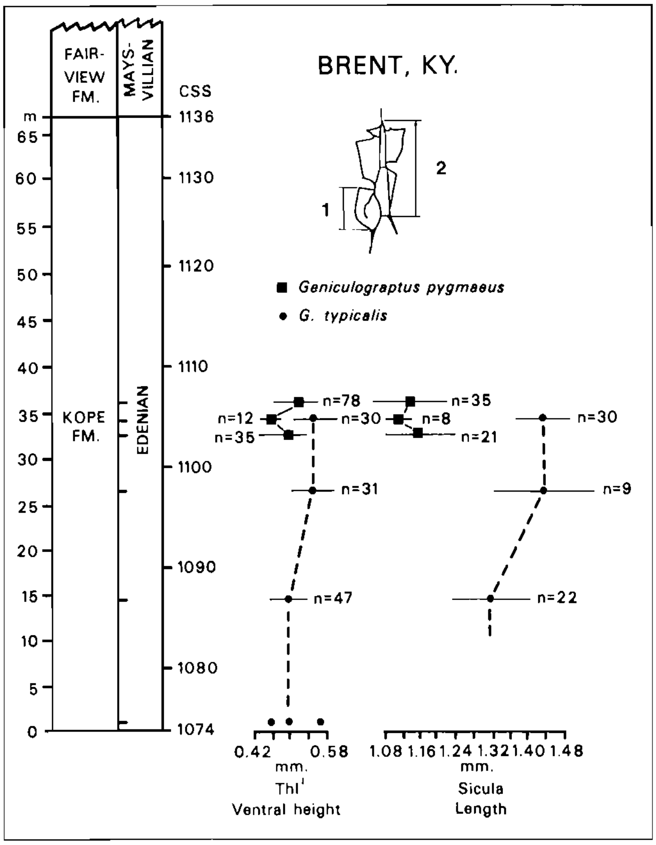

Figure 8. Stratophenetic variation of morphologic features in *Geniculograptus typicalis* and *G. pygmaeus* in the Kope Formation of the Brent section. For variables measured, see inset figure, in which "1" denotes ventral height of $Th1_1$ and "2" the length of the sicula.

and at several widely spaced horizons up to 35 m. *Geniculograptus pygmaeus* appears at 33 m above the base of the section, which corresponds to a level 43 m above the base of the Cincinnatian, or 1106 m in the composite SRS. The species occurs profusely in several collections from just above this level in both sections. Much of the interval between 20 and 31.5 m in the lower section is exposed only in a steep, not easily accessible slope and we have not yet carried out bed by bed collecting from these strata. Given the infrequent occurrence of graptolites in much of the Brent section, it is likely that further work will extend the range of *G. pygmaeus* downward at this locale.

CORRELATION AMONG CINCINNATI REGION SECTIONS

In the sections and cores that we have studied between the Middletown well in Butler County, Ohio and the sections at Maysville, Kentucky (see Fig. 1), the Kope Formation maintains a consistent thickness of about 76 m. Likewise, the first appearance of *Geniculograptus pygmaeus* occurs consistently at a horizon about 50 m below the top of the Kope Formation along a traverse from the Beiser well on the east flank of the Sebree Trough to the Cominco well near Minerva, Kentucky and sections near Ripley, Ohio and Maysville, Kentucky – a distance of some 125 km.

In the Brent sections, however, *Geniculograptus pygmaeus* appears 30 m below the top of the Kope Formation, which is about 20 m higher than expected. This occurrence is also about 20 m higher than expected based on its projected level in the composite SRS (1106 m versus 1086 m from the New Point and Middletown cores). This deviation may represent no more than incomplete collecting, or, alternatively, it might indicate either a delayed immigration of *G. pygmaeus* to this locality or a time-transgressive pattern of deposition for the Kope and Fairview formations. On the basis of the relative abundance patterns of several conodont species, Sweet (1979, Fig. 3) suggested that these units are indeed diachronous within the Cincinnati region. However, the uniformity in thickness of the Kope Formation and in the location of the first appearance of *G. pygmaeus* relative to the top of the formation indicate that this diachroneity may be less than some authors have suggested. For instance, Tobin and Pryor (1981), in an attempt to explain the occurrence of cyclical sedimentation patterns within the Cincinnatian Series, consider the Kope and Fairview formations and the Bellevue Limestone to be a set of contemporaneously deposited facies. To further examine this question, we have taken two different and independent approaches: 1) study of the stratophenetic pattern exhibited by the *Geniculograptus* species, and 2) study of the relationship between K-bentonite lithochronostratigraphy and lithostratigraphic patterns revealed by correlation of geophysical logs.

Stratophenetic correlation

We have gathered data on the length of the sicula and the first theca of some 410 graptolite specimens from the New Point core and the Brent sections. Because early growth stages are much more common than large rhabdosomes in our core samples, we have used these in our morphometric study. Previous studies (Mitchell, 1986) indicate that the early growth stages of *Geniculograptus typicalis* and *G. pygmaeus* differ significantly in several features. Of these features, we chose to use only length because it is affected less by flattening than width is; cylinders and cones become wider rather than longer when flattened. The sketch inset in Figure 8 shows the location of the variables measured. Figures 8 and 9 present sample means and standard deviations for the two variables plotted against their position in the stratigraphic section at the Brent locality and in the New Point core. To provide a common scale, we projected the sampled levels into

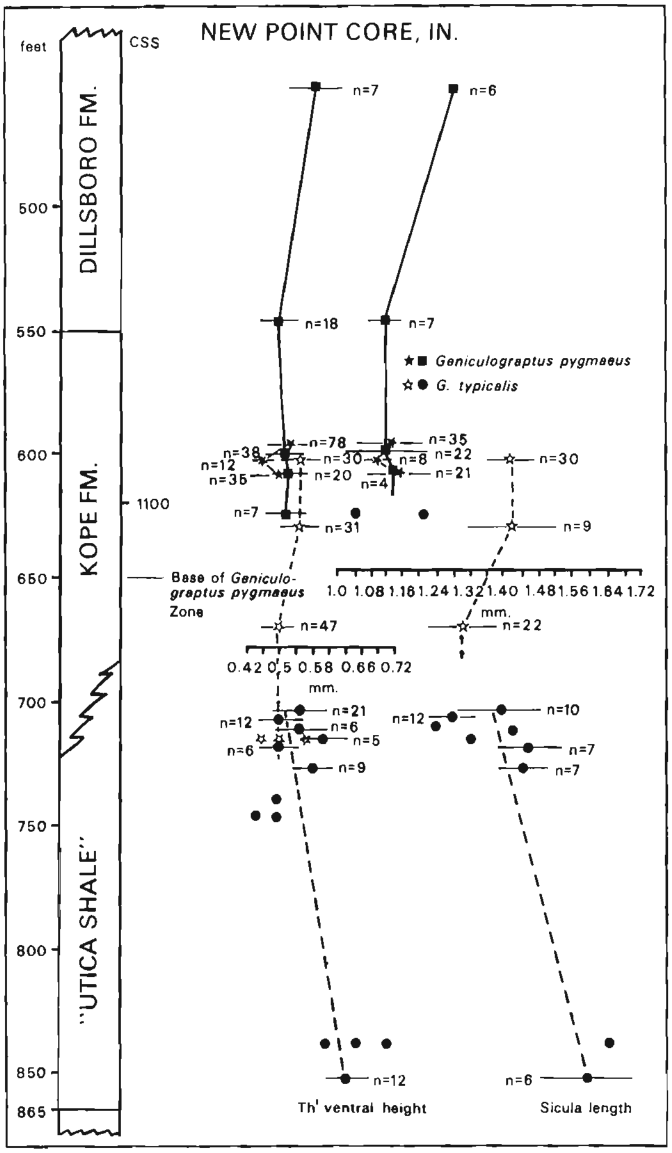

Figure 9. Comparison of stratophenetic variation patterns of measured morphological characters of *Geniculograptus pygmaeus* and *G. typicalis* in the New Point well (dots and squares) and the Brent section (stars). Note that based on estimated best fit, the first appearance of *G. pygmaeus* in the Brent section corresponds to a level well above the base of the *G. pygmaeus* zone in the New Point well.

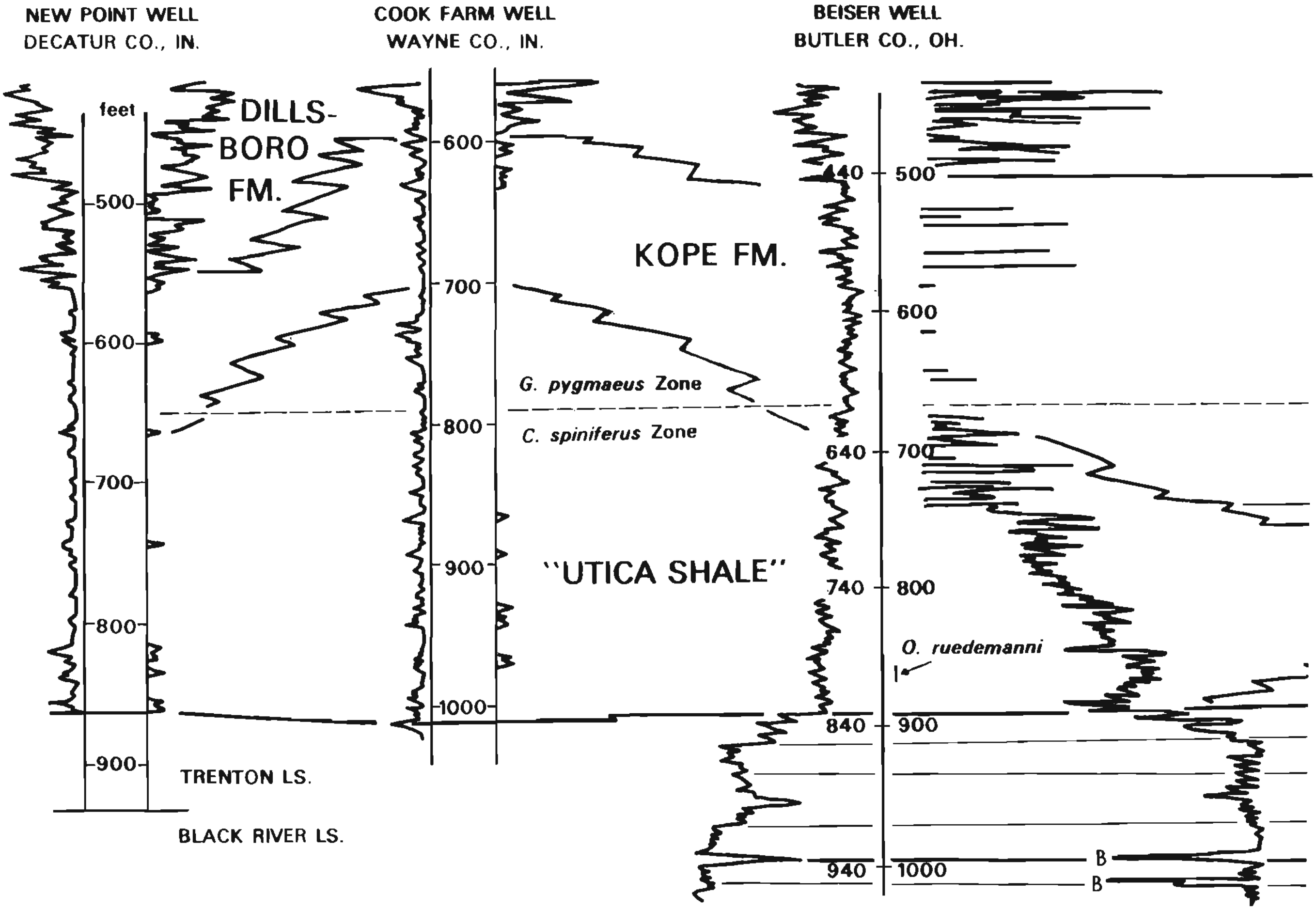

Figure 10. Correlation based on geophysical logs, shale percentage logs, and biostratigraphic evidence along a traverse from the New Point well to the Brent section. Logs lacking shale percent scale are geophysical logs (only the upper 275 feet [83.8 m] of the Middletown well is represented by a shale percentage log). For traverse location, see Figure 1. As in Figure 11, the gamma ray-neutron logs of the Middletown well are projected from the Permit No. 4 well (cf. Stith, 1986, Pl. 2). The latter was drilled at Middletown very close to the site of the Permit No. 1 well (the Middletown well of Bergström and Sweet, 1966) and has a virtually identical rock succession. Left footages on the Middletown well column refer to Permit No. 1, right to Permit No. 4. Left footages of the Beiser well refer to the Ohio Geological Survey Core No. 2537 (company hole No. 09-2082) examined by us, right footages and geophysical logs are from Permit No. 2 well, which was drilled 0.85 miles (1.37 km) from the No. 2537 well (cf. Stith, 1986, Pl. 2) and has a closely similar rock succession. Note facies relations between the Sebree Trough "Utica Shale", Kope Formation, Lexington Limestone, and Point Pleasant Formation. Part of the diagram is after Stith (1986, Pl. 2).

Sweet's (1984, 1988) composite SRS based on information provided in Sweet (1979) and Hoge (1985). Both *G. typicalis* and *G. pygmaeus* exhibit changes in the length of the sicula and first theca. In *G. typicalis*, there was an initial decrease in size, most of which was later reversed. During the latter, reversal phase, *G. pygmaeus* appeared. It too underwent an increase in these lengths, but was markedly smaller in size (among other differences – see Riva, 1974; Mitchell, 1986) than *G. typicalis* throughout its range.

Figure 9 also depicts our estimated best fit between the stratophenetic patterns exhibited by the two species in the Brent and New Point successions. We must emphasize that this is only fit by eye; we plan additional quantitative work on this problem. Although a number of alternative alignments are possible, these do not differ greatly. The pattern illustrated here suggests that the first appearance of *Geniculograptus pygmaeus* at Brent is stratigraphically above (by about 15 m) and hence temporally later than, its appearance in the New Point core. This figure is in close agreement with the relative position of the base of the range of *G. pygmaeus* vis-a-vis the base of the Fairview Formation and its projected level in the composite SRS. Accordingly, we interpret the general uniformity of level of first occurrence of *G. pygmaeus* across the Sebree Trough and at least into the western edge of the Lexington Shelf in the Cincinnati region as an indication of a similar uniformity in age for this horizon, i.e., for the base of the *G. pygmaeus* Zone. On the basis of the stratophenetic pattern, we project the base of this zone to a level about 18 m

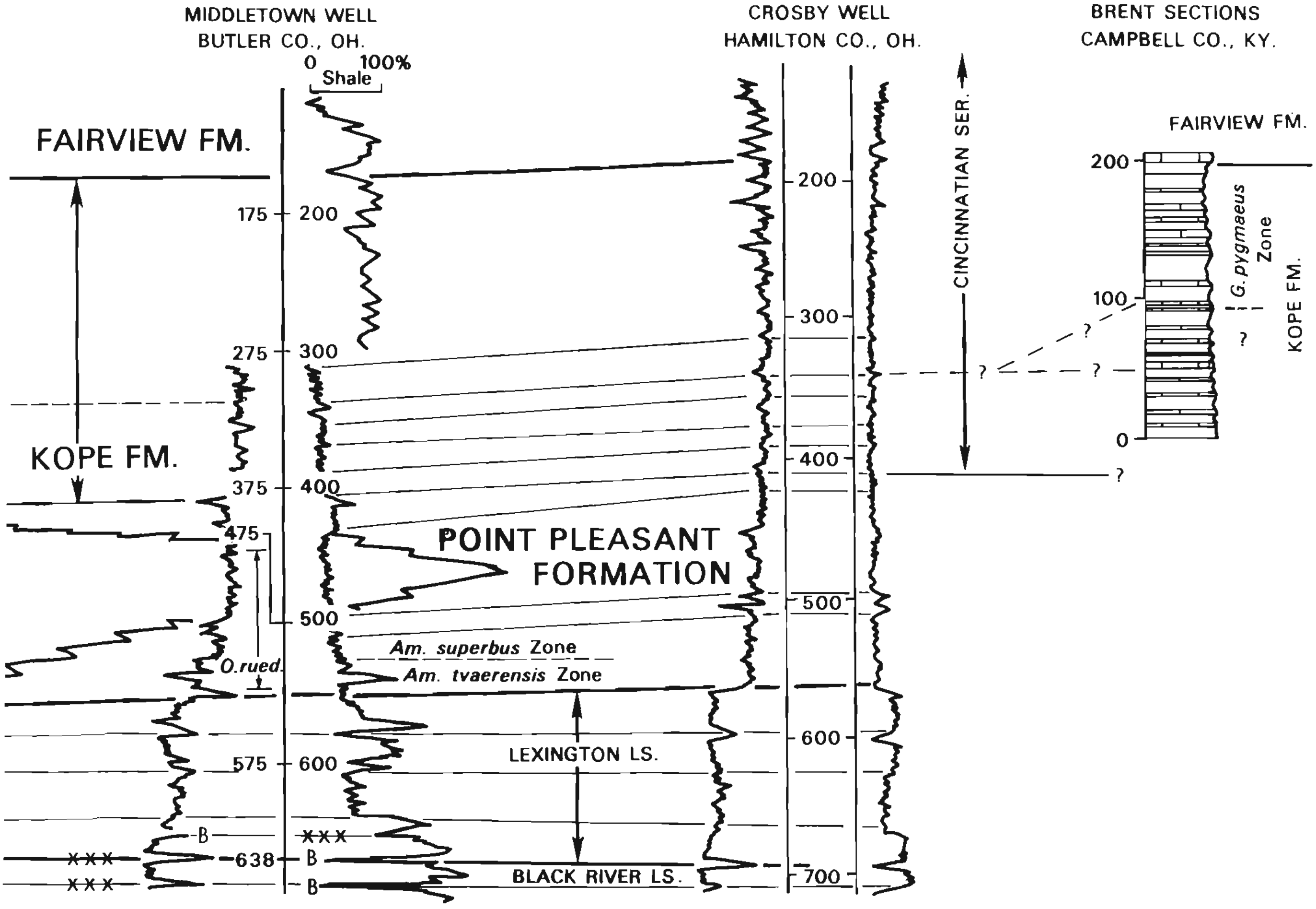

Figure 10 (cont'd.).

above the base of the lower Brent section and about 26 m above the base of the Edenian Stage and Cincinnatian Series in their type area.

Lithostratigraphic correlation

Despite the general agreement between the stratophenetic pattern exhibited by the *Geniculograptus* species and the level of first appearance of *G. pygmaeus* relative to the top of the Kope Formation and the projected level of its debut in the composite SRS, it is still conceivable that the lithological units, the morphological changes in the graptolites, and the lines of correlation produced within Sweet's (1979) conodont relative abundance log network are time transgressive, and that they are time transgressive in a similar way. This situation might arise if all were reflecting a slowly shifting pattern of marine environments. To assess this possibility, we have sought another means of judging the synchroneity of the Kope Formation and associated lithostratigraphic units and the base of the *G. pygmaeus* Zone across the region.

The Black River and Trenton groups in the Cincinnati region, as in most of eastern North America (Fetzer, 1973; Rickard, 1973; Bergström, 1989; Bergström and Mitchell, 1989), contain several prominent K-bentonite beds. Individual beds often can be traced over wide areas. K-bentonites are easily identifiable in cores and on geophysical logs. Those in this area can be correlated among numerous logged sections with considerable confidence (Stith, 1986; Schumacher and Carlton, 1989). Figure 10 presents a lithostratigraphic correlation between the New Point core and the Brent sections based on features of the logs. This diagram shows that the tops of the Black River Limestone, Lexington Limestone, and the Kope Formation are, by and large, parallel to the isochronous K-bentonites. Thus, these surfaces are likely to be correspondingly isochronous. It is also evident that the base of the *Geniculograptus pygmaeus* Zone parallels the bentonites as well – both from the New Point to the Middletown wells (in both of which the base of the zone is well controlled) and beyond to its projected level in the Brent composite. Figure 11 shows a similar pattern of parallelism for an additional set of localities from the Middletown well to the Cominco well, including several intervening wells. In fact, close inspection of logs from these and other wells throughout the region convinces us that surprisingly thin units (some as little as 8 m or less in thickness) can be traced across the region (Bergström and Mitchell, 1989). This unexpectedly layer-cake-like lithostratigraphic pattern further

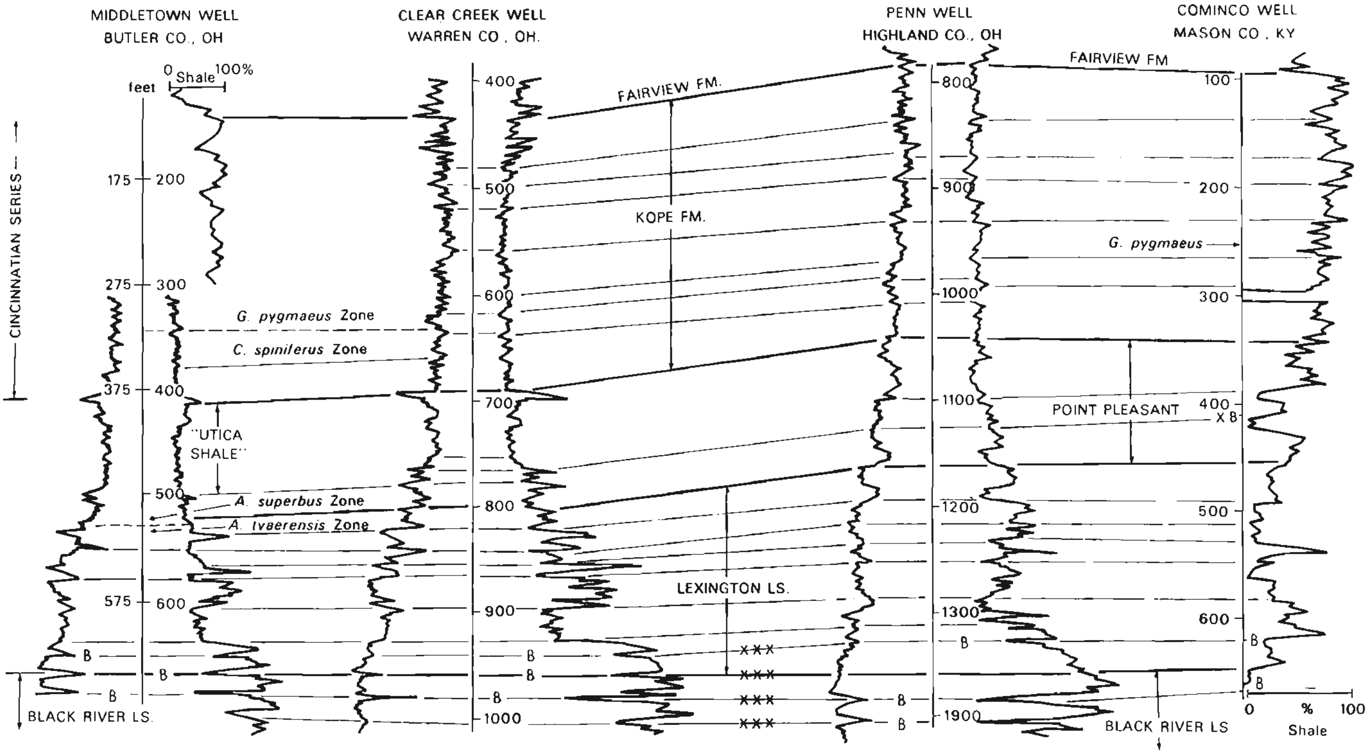

Figure 11. Correlation based on geophysical logs, shale percentage logs, and biostratigraphic evidence along a traverse from the Middletown well to the Cominco well. For traverse location see Figure 1. Note surprising lateral persistence of several relatively thin rock units in the Lexington Limestone and the Kope Formation, which are parallel to bentonite beds (B) and the base of the *Geniculograptus pygmaeus* Zone.

supports the conclusion that the base of the *G. pygmaeus* Zone does approximate an isochronous surface and that the unusually high appearance of this species in the Brent section is probably an artifact of sampling or preservation.

CORRELATION WITH THE TRENTON GROUP, MOHAWK VALLEY, NEW YORK

Any discussion of the definition and correlation of the base of the Upper Ordovician Cincinnatian Series should also consider its relationship to the Middle Ordovician Mohawkian Series as it is developed in its reference area in the Mohawk Valley, New York. At the classic Trenton Falls section, the Denmark and Cobourg limestones of the Trenton Group contain no graptolites useful for precise correlation. These rocks do yield conodonts, however (Schopf, 1966; Sweet, 1984). The Utica Shale, 40 to 60 km to the east, contains four graptolite zones: the *Corynoides americanus*, *Orthograptus ruedemanni*, *Climacograptus* (*Diplacanthograptus*) *spiniferus*, and *Geniculograptus pygmaeus* zones (Riva, 1969a,b, 1974; Rickard, 1973; Fisher, 1977). Correlation from this shale facies into the shelly facies is difficult, but Riva (1969a), Rickard (1973), and Bergström and Mitchell (1989) suggested that the base of the *C.* (*D.*) *spiniferus* Zone is likely to approximate the base of the Cobourg Limestone northwest of the Mohawk Valley (Fig. 12). Bentonite bed tracing by Cisne and coworkers (Cisne and Rabe, 1978; Cisne and Chandlee, 1982) supported this correlation. On the basis of conodont studies, Sweet and Bergström (1971) and Sweet (1984, 1988) suggested that the latter level, in turn, corresponds to the base of the Cincinnatian. As an additional link between these regions, the level of the late Mohawkian *Amorphognathus tvaerensis*-*A. superbus* zonal boundary within the lower Denmark Limestone can be correlated with the same boundary within the Lexington Limestone in the Middletown core and elsewhere in the Cincinnati region.

The first appearance of *Climacograptus* (*Diplacanthograptus*) *spiniferus* and *Geniculograptus typicalis typicalis* in the Beiser Core falls within strata that are probably correlative with the upper *A. tvaerensis* Zone (cf. Figs. 6, 10). Accordingly, these strata may be early Shermanian in age and, hence, considerably older than the basal Cobourg of New York. If the co-occurrence of these graptolites is taken as being indicative of the middle to upper *C.* (*D.*) *spiniferus* Zone, as has been the case in New York-Quebec (Riva, 1969a,b, 1987; Riva and Malo, 1988), then the base of the Cincinnatian in its type area must lie well above the base of the *C.* (*D.*) *spiniferus* Zone. Thus, correlation of the base of the Cincinnatian with the base of the Cobourg Limestone creates a marked contradiction between the present interpretations of the available graptolite and conodont evidence.

There are several factors that probably contribute to the present contradiction between conodont and graptolite-based correlations. First, it is likely that the geochronological scope

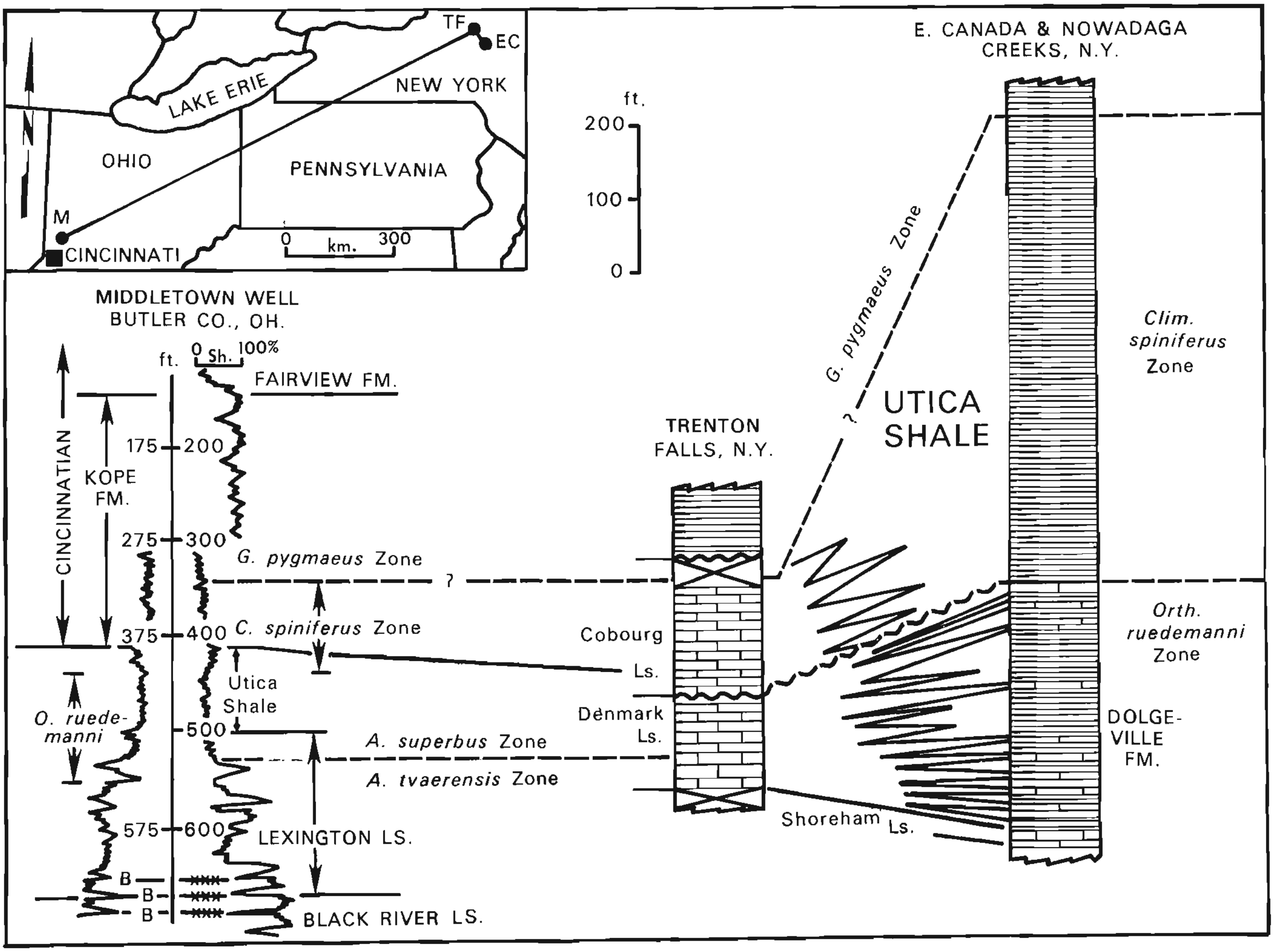

Figure 12. Stratigraphic relations between the Middletown well succession and standard carbonate and shale sequences in New York. Correlation of the Trenton Falls section with the Mohawk Valley shale sequence modified from Rickard (1973, Pl. 1, Cross-section 10). Note that the base of the *Geniculograptus pygmaeus* Zone in the Mohawk Valley is likely to correspond to a level well below the top of the Cobourg Limestone in the Trenton Falls section.

of the *Climacograptus* (*Diplacanthograptus*) *spiniferus* Zone is not the same in New York and the Cincinnati region. In the latter, the *C.* (*D.*) *spiniferus* Zone may include strata coeval not only with the *C.* (*D.*) *spiniferus* Zone in New York, but also the *Orthograptus ruedemanni* Zone (see Bergström and Mitchell, 1986, p. 253). Finney (1986) has also suggested that *C.* (*D.*) *spiniferus* and *Geniculograptus typicalis* appeared earlier in the Cincinnati region than in New York. This interpretation is supported by the fact that *C.* (*D.*) *spiniferus* and *Geniculograptus typicalis* occur within the upper *Amorphognathus tvaerensis* Zone in the Sebree Trough succession, whereas correlations between Trenton Falls and the Mohawk Valley in New York (Rickard, 1973; Cisne and Chandlee, 1982; Cisne et al., 1982) suggest that the base of the *C.* (*D.*) *spiniferus* Zone falls well above the top of the *A. tvaerensis* Zone in that region. This latter level appears, instead, to correspond to a level in the lower *O. ruedemanni* Zone based on our re-examination of Cisne and coauthors' graptolite collections and bentonite correlations (cf. the facies interpretation presented recently by Hay and Cisne, 1989, Fig. 2; Fig. 12).

A second factor possibly contributing to the lack of agreement between conodont and graptolite correlations of the Sebree Trough sequence with that of New York may be that the unconformity postulated by Rickard (1973) to be present at the base of the Cobourg Limestone has cut out equivalents of the lower *C.* (*D.*) *spiniferus* interval present in the pre-Cincinnatian strata of the Sebree Trough. This possibility is difficult to evaluate but it should be noted that Titus (1989) and other recent authors do not recognize an unconformity at this contact. Further, such an unconformity would not explain the differences in the pattern of graptolite first occurrences in the Sebree Trough succession and the Utica Shale – particularly the association of *Geniculograptus typicalis* with *Orthograptus ruedemanni*, and *Corynoides ultimus* with a *C.* (*D.*) *spiniferus* Zone assemblage.

It is also possible that the base of the Cincinnatian Series is not precisely coeval with the base of the Cobourg Limestone, but rather corresponds to a somewhat higher level within this formation. On the basis of tracing markers in geophysical logs, Bergström and Mitchell (1989) suggested that the base of the Cincinnatian correlates with a level some 6.5 to 10 m above the base of the Cobourg Limestone. The magnitude of this revision of the accepted correlation between these units is not substantially more than the stated precision in Sweet's (1984) conodont-based graphic correlation. This interpretation accounts for the presence of a *C.* (*D.*) *spiniferus* Zone fauna below the base of the Cincinnatian in the Sebree Trough, but again does not directly address the problem posed by the occurrence of *C.* (*D.*) *spiniferus* Zone graptolites in the *A. tvaerensis* Zone.

All three of these alternatives are possible and others are conceivable as well. Nevertheless, we consider a combination of the first and third of the above-mentioned interpretations to provide the most likely explanation for the observations discussed here. If this is correct, it implies that the characteristic elements of the *C.* (*D.*) *spiniferus* Zone immigrated into the Appalachian Foreland Basin in New York and Quebec later than they appeared elsewhere. Several additional lines of evidence that support this inference include:

1. Phylogenetic studies of the species of *Climacograptus* indicate that *C.* (*Diplacanthograptus*) *spiniferus* is closely allied to *C.* (*Climacograptus*) *bicornis* (Riva, 1976; Mitchell, 1987). Indeed, VandenBerg (1990) has discovered a succession of species that link *C.* (*C.*) *bicornis* and *C.* (*D.*) *spiniferus* within the Pacific Province graptolite facies of Australia, including a new species *Climacograptus* (*Diplacanthograptus*) *lanceolatus* VandenBerg. Yet in New York-Quebec, *C.* (*C.*) *bicornis* and *C.* (*D.*) *spiniferus* are separated by an interval corresponding to the *Corynoides americanus* and *Orthograptus ruedemanni* zones (Riva, 1976; Bergström and Mitchell, 1986). *Climacograptus* (*Diplacanthograptus*) *lanceolatus* is not yet known from the Utica Shale but specimens of this species do occur in the Exploits Group, central Newfoundland (see Erdtmann, 1976, Pl. 7, figs. B/3a, B/6a, B/6b, and B/6c) within the *Corynoides americanus* Zone. We have also recovered a single specimen that may be referable to *C.* (*D.*) *lanceolatus* from the *Amorphognathus tvaerensis* Zone, at 9.9 m in the Middletown core (Pl 1., fig. 43; Fig. 6), which is above the first occurrence of *C.* (*D.*) *spiniferus*. VandenBerg (1990) reports only a short interval of overlap between these two graptolites, and this occurs in the lowermost *C.* (*C.*?) *baragwanathi* Zone. VandenBerg and Stewart (1983, Table 1) equate the base of this Victorian zone with the base of the *O. ruedemanni* Zone in New York-Quebec.
2. The *Orthograptus ruedemanni* Zone of the Utica Shale in the Mohawk Valley yields a depauperate fauna of low species diversity. *Orthograptus ruedemanni* and *Normalograptus mohawkensis* are the only species that occur commonly throughout the interval and these two occur in great profusion on many bedding planes. The other five or six species known to occur in this zone occur sporadically in small numbers, and nearly all are more common in the underlying strata. Several of the species that are common in the *Corynoides americanus* Zone reappear toward the top of the *O. ruedemanni* Zone or in the succeeding *C.* (*D.*) *spiniferus* Zone (e.g., *Dicranograptus nicholsoni*, *Climacograptus* (*Climacograptus*) *caudatus*, and *Orthograptus quadrimucronatus* ssp.; see Riva, 1969b). *Corynoides americanus* also disappears in the lower *O. ruedemanni* Zone and a representative of the genus (*C. ultimus*) only reappears within the *Geniculograptus pygmaeus* Zone in the upper Utica Shale (Ruedemann, 1925; Riva, 1974; Mitchell, 1989). In contrast, *C. ultimus* appears within the lower portion of the range of *C.* (*D.*) *spiniferus* in the Sebree Trough.

The pattern of occurrence of graptolites within the Utica Shale in the Mohawk Valley suggests that the *O. ruedemanni* Zone may reflect the establishment of unusual ecological conditions in this region, producing a taxonomically impoverished and geographically restricted assemblage. These observations, taken together with the biogeographic and phylogenetic data outlined above, suggest that the evolutionary origin of *Climacograptus* (*Diplacanthograptus*) *spiniferus*, *Geniculograptus typicalis*, and *Corynoides ultimus* may have taken place during this interval and that the subsequent gradual appearance of the *C.* (*D.*) *spiniferus* zonal assemblage in the Mohawk Valley represents immigration controlled by the progressive return to conditions more favourable to a broad array of graptolite species.

One further issue regarding graptolite correlation between the Cincinnatian Series and the strata of the Mohawk Valley deserves comment here. The appearance of *Geniculograptus pygmaeus* in shales above the top of the Cobourg Limestone in New York (Rickard, 1973) is unlikely to be coeval with the base of this zone in the Cincinnati region. According to conodont data, the base of the *G. pygmaeus* Zone is likely to correspond to a level in the middle part of the Cobourg Limestone. Hence, as suggested by Hay and Cisne (1989, Fig. 2) and Bergström and Mitchell (1989), the non-graptolite-bearing, upper portion of the Cobourg Limestone may be an equivalent to the *G. pygmaeus* Zone rather than to the *C.* (*D.*) *spiniferus* Zone as indicated by Riva (1972) and Rickard (1973). However, at the present time we have no decisive biostratigraphic data with which to evaluate the isochroneity of this boundary between the two regions. In any event, the resolution of this issue only indirectly affects the correlation of the base the Cincinnatian Series with the New York succession.

CONCLUSIONS

The "Utica Shale" and Kope Formation of the Sebree Trough yield a diverse assemblage of graptolites that can be directly correlated with the classic facies of the Cincinnatian Series. The base of the series lies within the *Climacograptus* (*Diplacanthograptus*) *spiniferus* Zone as developed in this region, about 23 m below the base of the *Geniculograptus pygmaeus* Zone. We are uncertain whether the base of the *C.* (*D.*) *spiniferus* Zone in the Sebree Trough sections is actually isochronous with the base of the *C.* (*D.*) *spiniferus* Zone in the Utica Shale of the Mohawk Valley. However, on the basis of the data available, we suspect that it is not; that outside the region of the Appalachian Basin, *C.* (*D.*) *spiniferus* and at least some of the other species of this zonal assemblage (such as *Geniculograptus typicalis*) range downward into rocks synchronous with the older *Orthograptus ruedemanni* Zone in New York. This observation, if it is confirmed, will have important implications for the biostratigraphy, biogeography, and phylogeny of graptolites in this interval. Discussion of these implications, however, is beyond the scope of the present paper.

As the base of the Cincinnatian Series in its reference area does not coincide with a graptolite or conodont zonal boundary, or with any other known biostratigraphic zonal boundary, the base of the series cannot be defined precisely on the basis of these traditional and regionally useful zonations. In view of this, the Cincinnatian Series in its present scope is unlikely to be a suitable global chronostratigraphic unit. Nevertheless, the designation of a boundary stratotype is needed to provide a stable and definable horizon as a basis for correlation with other regions and series. There are several alternatives, which are as follows:

1. Define the base of the series in terms of Sweet's (1984) graphic correlation network as the horizon within the SRS that corresponds to the traditional base of the series at the base of the Kope Formation in the city of Cincinnati. This level could then be projected to a specific section where the boundary could be permanently fixed and accessible to earth scientists. Recently, Sweet (1988) did this and suggested that the Cominco Core is a suitable reference section. However, the small diameter of this core and its poor condition (it has already been sampled for several different studies) make it less useful as a reference standard than a well exposed surface section. More generally, this choice might also limit the range of correlation techniques that subsequently could be used for precise correlation of the Cincinnatian boundary stratotype with other sections. There is no guarantee that any future biostratigraphic, magnetostratigraphic, or other time-significant boundary will prove to be coincident with the age of the base of the Kope Formation in Cincinnati.
2. Move the boundary slightly to coincide with the base of a more traditional biozone such as at the base of the *Geniculograptus pygmaeus* Zone or the *Amorphognathus superbus* Zone. Both of these options also have disadvantages in addition to their obvious advantages. The most serious of these is that at the present time no surface section within the region is known to expose either zonal boundary within a continuously fossiliferous succession, and again, we do not favour use of a core as a boundary stratotype. Thus, we believe that selection of a boundary stratotype for the base of the Cincinnatian Series must await further research on the fossils and strata of its classical reference area.

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NOTE ADDED IN PROOF

Since the completion of this paper in 1988, additional information has become available concerning the graptolite correlation of the base of the Cincinnatian Series in its type area. In a recently drilled core (Ohio Geological Survey core No. 3020) from a site at Galbraith Road about 15 km north of the centre of Cincinnati, *Geniculograptus pygmaeus* appears about 20 m above the base of the Kope Formation (= the base of the Cincinnatian Series). This supports our conclusion that the known first appearance of this species in the Brent section is well above the base of the *G. pygmaeus* Zone in this region. Additional data on the graptolite correlation of the base of the Cincinnatian Series, including the position of this level in terms of the Australian standard graptolite succession, have been presented by us in a recently published paper (Journal of Paleontology, v. 64, p. 992-997).

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PLATE 1

Representative graptolite specimens from the *Orthograptus ruedemanni*, *Climacograptus spiniferus*, and *Geniculograptus pygmaeus* zones of the "Utica Shale" and Kope Formation in the Cook Farm, New Point, Beiser, and Middletown cores. Locality designations as in Figure 1; stratigraphic horizon given in metres above the top of the Trenton or Lexington limestones (e.g., "M43.5" indicates a specimen from 43.5 m above the top of the Lexington Limestone in the Middletown core).

Geniculograptus pygmaeus Zone

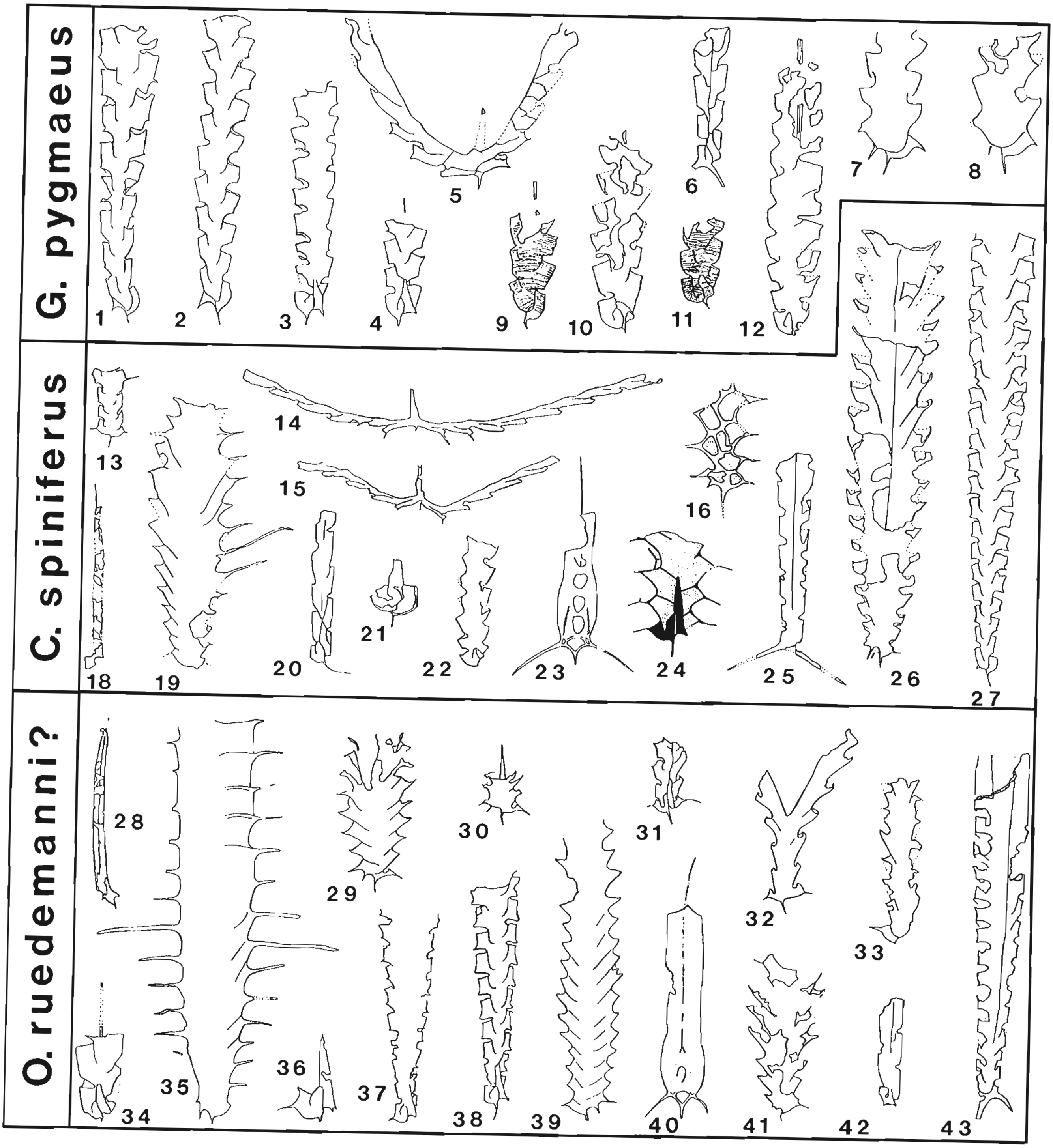
Figure 1.	*Geniculograptus typicalis typicalis* (Hall), CF68.8, x8
Figures 2-4.	*Geniculograptus pygmaeus* (Ruedemann), NP70.0, NP64.0, NP69.9, respectively, x8
Figure 5.	*Dicellograptus* sp. CF84.8, x8
Figure 6.	*Climacograptus* (*Diplacanthograptus*) *spiniferus* Ruedemann, CF79.4, x8
Figures 7, 8.	"*Glyptograptus*" *lorrainensis* (Ruedemann), NP71.6, x8
Figures 9-12.	*Normalograptus* sp. nov., figs. 9-11 from NP67.9, fig. 12 from CF88.5.

Climacograptus (*Diplacanthograptus*) *spiniferus* Zone

Figure 13.	*Dicranograptus* sp. cf. *D. nicholsoni* Hopkinson, NP47.5, x5
Figure 14.	*Dicellograptus flexuosus* Lapworth, NP48.7, x5
Figure 15.	*Dicellograptus* sp. cf. *D. patulosus* Lapworth, NP 38.7, x5
Figure 16.	*Orthoretiolites hami* Whittington, CF39.4, x8
Figure 18.	*Corynoides ultimus* Ruedemann, CF57.7, x8
Figure 19.	*Orthograptus quadrimucronatus* (Hall), CF33.3, x5
Figure 20.	*Climacograptus* (*Diplacanthograptus*) sp. cf. *C.* (*D.*) *spiniferus* Ruedemann, NP56.4, x8
Figure 21.	*Climacograptus* (*Climacograptus*) sp. cf. *C.* (*C.*) *caudatus* Lapworth, NP46.7, x14
Figure 22.	*Normalograptus* sp. nov., NP56.4, x5
Figure 23.	*Cryptograptus insectiformis* Ruedemann, NP49.4, x5
Figure 24.	*Orthoretiolites* sp., NP43.6, x8
Figure 25.	*Climacograptus* (*Diplacanthograptus*) *spiniferus* Ruedemann, CF37.8, x5
Figure 26.	*Amplexograptus* sp. aff. *A. tardus* (T.S. Hall), NP38.7, x5
Figure 27.	*Geniculograptus typicalis typicalis* (Hall), CF64.4, x5.

Orthograptus ruedemanni? Zone

Figure 28.	*Corynoides ultimus* Ruedemann, NP15.0, x8
Figures 29, 30, 39.	*Orthograptus ruedemanni* (Gurley), figs. 29 and 30 from M31.2 and fig. 39 from M30.2, x5
Figure 31.	*Dicranograptus* sp. cf. *D. nicholsoni* Hopkinson, CF26.7, x5
Figure 32.	*Dicranograptus* sp. cf. *D. kirki* Ruedemann, CF7.3, x5
Figure 33.	*Orthograptus* sp. aff. *O. calcaratus basilicus* Lapworth, M33.3, x5
Figure 34.	*Normalograptus* sp. cf. *N. mohawkensis* (Ruedemann), NP26.9, x14
Figure 35.	*Orthograptus quadrimucronatus* (Hall), CF34.0, x5
Figures 36, 37.	*Geniculograptus typicalis*
Figure 40.	*Cryptograptus insectiformis* Ruedemann, NP26.8, x5
Figure 41.	*Orthograptus amplexicaulis* (Hall), NP36.5, x5
Figure 42.	*Climacograptus* (*Diplacanthograptus*) sp. cf. *C.* (*D.*) *spiniferus*, NP 13.2, x8
Figure 43.	*Climacograptus* (*Diplacanthograptus*) sp. cf. *C.* (*D.*) *lanceolatus* VandenBerg, M9.9, x5.


G. pygmaeus
C. spiniferus
O. ruedemanni?
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Contributions of 19th century Canadian geologists to the adoption of the Ordovician System

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Abstract

Ever since Billings established the correct stratigraphic position of the Quebec Group, Ordovician rocks have been important to Canadian geologists. The first usage of the term Ordovician in the New World was by Hunt in 1883. In several of his papers, he publicized this new division of the geological column. Although there was no international consensus on the usage, the Geological Survey of Canada recognized the time interval in about 1888; it was designated as Cambro-Silurian. In 1891, Sir William Dawson was the first to use the term Ordovician in an American textbook. These actions by Canadian geologists collectively influenced general acceptance of that new system.

Résumé

Depuis l'établissement, par Billings, de la position stratigraphique correcte du groupe de Québec, les roches ordoviciennes revêtent une importances pour les géologues canadiens. Hunt est le premier à utiliser le mot «Ordovicien» dans le Nouveau Monde en 1883. Il mentionne cette nouvelle division de l'échelle stratigraphique dans plusieurs de ses rapports. Malgré l'absence de consensus international sur l'usage, la Commission géologique du Canada reconnaît cet intervalle de temps, appelé «Cambro-Silurien», vers 1888. En 1891, sir William Dawson est le premier à se servir du mot Ordovicien dans un manuel américain. Ces mesures prises collectivement par les géologues canadiens incitent à l'acceptation générale de ce nouveau système.

INTRODUCTION

The important contributions made by Canadian geologists to acceptance of the Ordovician System, proposed by Charles Lapworth in 1879, have not been summarized. By almost any measure one can apply, these workers were quicker to recognize the merits of this new system and to use it than were their far more numerous colleagues south of the border. It is not evident why this should be. One may speculate that during the last half of the 19th century, Canadian geologists were more preoccupied with the eastern part of their country than were their counterparts in the United States. Thus, they were more influenced by the rocks which eventually came to be designated as Ordovician in age, but it seems almost too facile an explanation.

THE SIGNIFICANCE OF BILLINGS (1820-1878)

Elkanah Billings has never received proper recognition for his insight in stratigraphic paleontology. He was the equal of James Hall in this regard, and in one important event was right when Hall was wrong. In the history of American geology, the age of the Quebec Group forms an important episode (Merrill, 1924, p. 412). Although these Quebec rocks were "Silurian" in the 1860s, the sequence was intimately involved in the arguments over the validity of the Taconic System. Sir William Logan, originally Provincial Geologist and subsequently the first Director of the Geological Survey of Canada, had established a sequence of rocks along the St. Lawrence River. However, after carefully studying the fossils

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which had been collected, Billings, paleontologist of the Survey, concluded that the sequence, as given by Logan, was in error. Billings worked with a small number of fossils, a paucity of literature, and the still limited knowledge of the European succession; his conclusion was brilliant.

Logan was swayed by the paleontological evidence that Billings had marshalled. As a result, he produced a structural solution to explain an inverted sequence (Eagan, 1987). This episode was one of the first in the New World whereby paleontological data was critical in identifying a major tectonic feature. Logan's Line stems directly from biostratigraphic investigation of Ordovician fossils. The problem of the age of the Quebec Group indicated the importance of correlation of relatively small divisions of geological time. A generation earlier, this structure would have passed unnoticed within the thick mass of enigmatic ''transition'' rocks.

Although Billings died a year before the term Ordovician was introduced, his investigations had laid a strong foundation for that systemic division. In spite of the complexities of the Quebec Group problem, it soon became evident that rocks of Murchison's ''Lower Silurian'' were widespread and commonly quite fossiliferous in Canada. Both in quantity and what might be called quality, they outshadowed the outcrop areas of Great Britain and the Continent.

THE ROLE OF HUNT (1826-1892)

T. Sterry Hunt, geochemist and synthesizer, wrote extensively, both in his particular specialty and in many other aspects of geology; by the same token, reams could be written about him. Hunt began his professional life in 1847 as a chemist and mineralogist with the Geological Survey of Canada, remaining with the organization for a quarter of a century (Boyle, 1971). He had an illustrious career as the first geochemist in Canada, though a bit of the lustre he cast has faded. Hunt was a leading light in the new Royal Society of Canada when it was formed, and presented a long paper on May 23, 1883, which was printed the same year. A small part is quoted below.

''The name of Ordovician (sometimes contracted to Ordovian) which we have introduced in this table was proposed by Lapworth in 1879 to designate the group of paleozoic rocks found in Wales between the base of the Lower Llandovery and the base of the Lower Arenig .... This new division is convenient in American geology from the fact that it includes the group of strata between the base of the Silurian (Oneida) sandstone and the base of the Chazy limestone; the latter together with the Trenton, Utica and Loraine divisions being equivalent to the Ordovician'' (Hunt, 1883, p. 223). In the table mentioned in the quotation, Ordovician is listed with (Upper Cambrian) in smaller letters below it, presumably indicating use as a synonym. Cambrian is below Ordovician and Taconian is below that (see Table 1).

For some years, the annual reports of the Smithsonian Institution included accounts of the latest developments in various fields of science. Hunt was asked to prepare the summary on geology for both 1882 and 1883. In his first summary for the Smithsonian, Hunt also used the term Ordovician. After discussing Lapworth's work briefly and the application of his new term in Europe, Hunt (1884, p. 334) noted, ''The base of the Ordovician, as thus defined, is marked by a great stratigraphical break, attendant on continental movements, in eastern North America .... Similar strata of Ordovician age, as long since pointed out, are found in eastern Canada, apparently dipping beneath the older Cambrian (Quebec group), in which they were by Logan included .... Researchers in Canada and in Vermont have long since shown that in this greatly disturbed and involved belt are included fossiliferous strata, holding all three of the great lower Paleozoic faunas, Cambrian, Ordovician, and Silurian.''

In the second Smithsonian report, Hunt (1885, p. 451) wrote: ''While in the Adirondack area there is a break, both paleontological and stratigraphical, between the Cambrian and the Ordovician, which begins in some places with the Chazy and in other with the Trenton, we have, according to the late studies of Walcott in Nevada, a gradual passage from the Cambrian to the Ordovician (Lower Silurian or second fauna of Barrande).'' He then went on to quote one of Walcott's statements, preliminary to Walcott's Eureka [Nevada] Monograph of 1884. This part of Hunt's survey is followed by a text centre heading of ''Silurian, Cambrian, and Taconian,'' and within that section of the paper, the term Ordovician is used in the same table published in 1883.

Four years after his first use of the term Ordovician, Hunt (1887) referred to his original table and stated: ''In the tabular view above noticed the writer attempted to introduce for the first time into American geology the term Ordovician proposed by Lapworth in 1879 for the group of strata between the proper Silurian of Murchison and the undisputed Cambrian of Sedgwick, which, though by the latter named Upper Cambrian, has been alternately called by others Lower Silurian, Siluro-Cambrian, and Cambro-Silurian, and includes the Chazy, Trenton, Utica, and Loraine subdivisions of the New York system. The name of Ordovician has since been adopted by many British and European geologists, and is now used by Mr. Charles D. Walcott, of the United States Geological Survey.''

Had it not been for the traditional delay of the United States Government Printing Office, the Smithsonian Annual Report might have come out before the Transactions of the Royal Society of Canada. However, whether it was in 1883 or 1884, or whether it was published in Canada or the United States that Ordovician was first used in America, are small matters; the credit for first usage of the term Ordovician in the New World without question goes to T. Sterry Hunt.

Employment of the general term Ordovician in the literature and the boundaries that were associated with this new unit are necessarily interrelated, but are two distinct issues. For Hunt, the upper boundary of the system was not generally different from that recognized today. However, the Calciferous, loosely the Early Ordovician of current usage, was considered part of the Cambrian. Hunt suggested adoption of the term Ordovician at first on the notion of a major stratigraphic discontinuity in the eastern United States; almost immediately more continuous sections were noted in the

Table 1. Table showing the stratigraphical relations of the taconic rocks

Eaton's Nomenclature (1832.)				Later Names.	
I. Primitive.	3.	Corniferous Lime-rock..........................		Upper Helderberg...............	**Devonian.**
		Geodiferous Lime-rock..........................		Niagara.........................	**Silurian.**
	2.	Millstone-grit...	**Second Graywacke.**	Oneida — Medina...............	
	1.	Graywacke-slate		Utica — Loraine...............	**Ordovician.** (Upper Cambrian)
II. Transition.	3.	Metalliferous Lime-rock........................		Chazy — Trenton..............	
		Sparry Lime-rock. Calcifer. Sand-rock.			
	2.	Millstone-grit...	**First Graywacke.**	**Upper Taconic.** (Quebec Group)	**Cambrian.** (Middle and Lower Cambrian)
		Graywacke-slate			
	1.	Argillite......................................		**Lower Taconic.** (Itacolumitic Group)	**Taconian.**
III. Lower Secondary.	3.	Granular Lime-rock.............................			
	2.	Granular Quartz-rock........................			
	1.	Gneiss and other Crystalline Rocks.........		**Huronian.** **Laurentian.**	**Montalban.** **Norian.** **Arvonian.**

Modified from Hunt (1883)

western United States, but the new data did not cause him to abandon the concept of this system. He continued to use the term Ordovician as appropriate in his writings, and his advocacy must have had a strong effect on the geological community. The boundaries that Hunt originally indicated seem to have been those that prevailed when the term Ordovician came into general use; only later was the lower boundary moved downward, and that mainly was because the Upper Cambrian faunas became better known.

THE INACTION OF THE CONGRÈS GÉOLOGIQUE INTERNATIONAL

The first international assembly of geologists stemmed directly from a meeting at the American Association for the Advancement of Science in 1876; Hunt formally introduced the motion for an international congress. As a member of the American Committee, he served as a "sparkplug," in promoting the 1878 Congress in Paris. The success of this original international gathering of scientists served as an inspiration for international meetings in other disciplines and provided a model for their organization. Ordovician rocks were not of particular interest to Hunt, and it might have been his exposure to European and British geologists at the sessions of the Congress and his aim to be known as an internationalist which led to his use of the term in his 1883 monograph.

From its inception, the Congress was concerned with divisions of the geological column and their appropriate nomenclature (Hughes, 1891). As is so common with many bodies, the Congress appointed a Commission to consider and report later. At the second Congress, in Bologna in 1881, it was decided that a geological map of Europe should be compiled and there was general agreement on the standardization of map colors to use for certain rocks. At the Berlin Congress of 1885, considerable discussion ensued concerning the units to be recognized on the European map. The term Ordovician was discussed, but it was finally agreed that for this map, Silurian in a broad sense was the most appropriate term. If various other views that were presented in 1884 had prevailed, the oldest parts of the Paleozoic geological column might have been named Taconic, Cambrian, and Bohemian.

At the 1888 London Congress, one of the three principal topics of discussion was the issue of classification of the Lower Paleozoic. The deaths of both Sedgwick and Murchison had not clarified the situation as much as one might have expected. There was extended discussion, but ultimately President J. Preswick (1891, p. 53) summed it up by indicating that the status quo of Cambrian, Lower Silurian and Upper Silurian would be maintained. The Congress also continued its international commission on nomenclature; it was early recognized by the Congress that the principle should be one vote per country. Robert Bell was the representative for Canada, but seemingly took no part in the discussions concerning the lower Paleozoic.

The fifth session of the Congress was in 1891 in Washington. The word Ordovician does not appear in the Compte Rendu. Perhaps even more interesting than this official account is a local guide for the Congress visitors — later republished in the proceedings — which included a section on geology of the region (McGee, 1891); in that, Lower Silurian is used. In fact, it was not until the Copenhagen meeting of 1960 that the International Geological Congress finally adopted the term Ordovician as it is used today. The various working groups and subcommissions of the present-day Commission on Stratigraphy move slowly, but one hopes not that slowly!

THE ACTION OF THE GEOLOGICAL SURVEY OF CANADA

One important development may have stemmed from the 1888 Congress. Although Alfred R.C. Selwyn, who since 1869 had been the second Director of the Geological Survey of Canada, was not present, he might have been indirectly influenced by its action, or rather its lack of resolution, in the stratigraphic deliberations. So far as one can determine, there was tacit recognition among many of the geologists present in London, though not necessarily the most outspoken ones, that division of Murchison's Silurian into the Upper and Lower was not the most appropriate or convenient terminology. One serious problem was the lack of consensus at the Congress that adoption of Lapworth's terminology was indeed the most satisfactory solution to the nomenclature issue.

Some time shortly after the London Congress, the Geological Survey of Canada used the term Cambro-Silurian in its publications. This was not a term lumping the Lower Paleozoic, but was the formal designation for a middle division between Cambrian and Silurian. The term was not unique to Canada and had been used earlier. Nevertheless, organizational usage may carry more weight than mention in a paper. Now, it is a truism that an organization does what the head of it desires, or at least it does not do what the head is opposed to. Any way the problem is approached, Selwyn deserves credit for allowing publication of the term. The term Cambro-Silurian was in use in official Canadian publications at least until after the death of George Mercer Dawson who succeeded Selwyn; Zazlow (1975, p. 200 C) provides an interesting example in Quebec of its usage on a map.

At precisely what date the term Cambro-Silurian came into the official glossary is difficult to determine, for the Geological Survey of Canada had no special committee to render and formalize stratigraphic decisions. This term was certainly used and well known a decade or less after Lapworth's proposal. James Dwight Dana was aware of the need for greater precision in the geological column, but he was much torn on the issue of "justice" to Sedgwick and to Murchison and seemingly approved of Selwyn's approach. "With regard to the Lower Silurian, the term *Cambro-Silurian*, which is used already by the Canadian Survey, might serve, supposing another name needed, if its signification accorded with geological usage. It means, as employed in Canada, a formation that is neither Cambrian nor Silurian, but which lies between the Cambrian and Silurian formations" (Dana, 1890, p. 235). However, for some reason, Dana judged the term to be out of harmony with other compound period words then in usage. That he added this note to a paper already published shows that it was no later than 1890 when Selwyn authorized use of the term.

A particularly important Geological Survey publication during this interval was that of Ells (1888). It included a reconsideration of the Quebec Group and further focused attention on a classic American area of Ordovician rocks and fossils. This probably was the first Canadian work to use the term Cambro-Silurian and it may have been the one which caused the Director to make a decision on nomenclature. In a review of this report, Walcott, (1890, p. 102) equated the Cambro-Silurian with the Ordovician. [When Walcott touched on this term in his own writings during the 1880s and early 1890s, he rendered it as "Lower Silurian (Ordovician)."]

One may contrast the approach of the Geological Survey of Canada under Alfred Selwyn with that of the United States Geological Survey under John Wesley Powell. On several occasions, there were efforts in Washington to standardize the use of stratigraphic nomenclature. In the spring of 1888, among other dicta issued on stratigraphy, was the note that Cambrian was to be followed by Silurian. The Ordovician System was not accepted by the U.S. Geological Survey until 1903, a year after it was adopted by the Geological Survey of Great Britain.

THE INFLUENCE OF DAWSON (1820-1899)

If Dana is one of the patriarchs of geology in the United States, especially in the academic field, John William Dawson has to be accorded the same role in Canada. Like Dana, Sir William wrote a major text which also went through four editions. Technically, Dawson's book really went through two editions; the third included new data in an appendix of about 100 pages and the fourth, with a minor change in title, added another shorter appendix to the first. In the second edition this work could be said to be comparable to Dana's *Manual of Geology*, but it was rather more of a compendium than a classroom book.

The large size of Canada and the few geologists in the country might lead one to expect that investigations would have lagged behind those of Great Britain. This is not correct, for by the time of his second edition in 1868, Dawson already had a fossiliferous base to his geological column, consisting of the rocks in Newfoundland and their presumed unfossiliferous equivalents in New Brunswick. In the next edition, in 1878, Dawson used the heading "Cambrian" in an even more modern sense, whereas Dana (1879) placed the Cambrian as a subdivision of the Lower Silurian.

Dawson also made interesting comments under the heading of "Lower Silurian." "In the second edition of *Acadian Geology*, following, though under protest, the Murchisonian nomenclature, then current in both England and America, and adopted by the Canadian Geological Survey, I included under this head all the fossiliferous rock older than the Upper Silurian. If, however, we restrict the term Lower Silurian to that geologic group of which the great Trenton formation in America and the Bala in England are the main and typical members, and which contain what Barrande has called the 'second fauna,' then we have as yet no certainly determined fossiliferous group of this age in Acadia; and there seems little doubt that the great Lower Silurian fossiliferous limestones are absent, as they appear also to be in Newfoundland" (Dawson, 1878, p. 78 appendix). At least, this gives both a clear idea of the correlations of the time in Canada and an indication that the second fauna of Barrande marked a distinct interval.

Dawson, like Dana, could keep up with the times, though his viewpoint was different. In his supplement, Dawson (1891, p. 24) follows a summary of "The Silurian" by two pages on "The Cambro-Silurian, or Ordovician" and this in turn is followed by "The Cambrian." Lapworth is not mentioned, and what reason there is for the change in nomenclature is given in two sentences. "With the incoming of this age, a more marked distinction occurs in America between the marginal and plateau-deposits. This had previously appeared in the Cambrian, and became more distinct in the Erian [Devonian] and Carboniferous, but it is somewhat peculiar as between the marginal and submarginal area and those inland in the period on which we now enter." In a table of Acadian rocks and correlations to England, Dawson (1891, p. 36) uses Ordovician as a centre heading, with Silurian above and Cambrian below. For this, he should receive credit as the first in the New World to introduce Lapworth's term into a geological textbook. In the final edition of his textbook, published the year after his death, Dana (1895) still did not use the term Ordovician in the stratigraphic column.

DISCUSSION

One searches in vain through the journal literature of the 1890s in North America for any extended discussions either championing or rejecting the concept of Ordovician. A few United States geologists indicated they saw no need for it, but in the main, the term was ignored. It is nearly as fruitless to look for its use in a stratigraphic paper. The first decade of the *Bulletin of the Geological Society of America* indexes this term only three times. The *Journal of Geology* records it little more during the same interval. None of these usages are for rocks in Canada. The *American Geologist* was even later than these two journals to publish an article using the term Ordovician in the text.

During the 1880s, geologists in North America devoted a great deal of discussion to the issue of acceptance or rejection of the Taconic System. It seems to have worn out the subject of stratigraphic nomenclature for some time. That might explain why, during the 1890s, there was a dearth of argument as to the merits or disadvantages of utilizing the term Ordovician. The absence of any public discussion or strong proponent during this time casts a veil over the process of acceptance of this term. In North America, acceptance of the term Ordovician "just happened." This lack of even a basis for speculation regarding the process of acceptance is exceedingly frustrating, for all later proposals to significantly modify the geological column have been unsuccessful. It would be interesting to know if there was any logical process as to why it was accepted and why Erian, Ozarkian, and a host of other proposed systems have been generally rejected by the profession.

The term Ordovician was not so much a new scientific idea, in the same way that plate tectonics might be judged a new idea, as it was a change in general terminology. Lower Silurian rocks were distinct and were well known in North America long before the term Ordovician appeared. In a sense, with that geological basis, it required only a bit of extra attention to raise these rocks to the prominence they deserved as a separate system. If there were any persons who should be singled out to deserve our appreciation for leading this effort, they are Hunt and Dawson.

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Stratigraphic correlation of the Appalachian Ordovician using advanced U-Pb zircon geochronology techniques

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Abstract

The determination of accurate and precise U-Pb zircon ages for Ordovician felsic igneous rocks has been difficult due to the presence of invisible cores of inherited Proterozoic zircon in newly grown magmatic zircon. Techniques pioneered at the Royal Ontario Museum, including hand picking of each zircon to be analyzed, low Pb-blank chemistry to permit a significant reduction in sample size, and abrasion to minimize or eliminate secondary Pb-loss have been invaluable in dealing with multiple age zircon populations. U-Pb zircon ages for rhyolites of 473 +3/-2 Ma for the Buchans Group (with latest Arenig-earliest Llanvirn conodonts), 473 +/-2 Ma for the Roberts Arm Group and 469 +5/-3 Ma for the Cutwell Group (with early to middle Llanvirn conodonts) have been obtained. These demonstrate age correlation of the groups and the potential for isotopic-fossil age correlation in volcanic arc sequences.

Tonalite of a major batholith in southwestern Newfoundland is 456 +/-3 Ma old, exactly coeval with isotopic ages of early Caradoc bentonites in Sweden and Kentucky. The La Poile Group, a major felsic volcanic sequence, is of Silurian age (420 +8/-2 Ma), and contains inherited Proterozoic zircon. It is not Ordovician in age as previously reported.

Résumé

La datation exacte et précise par la méthode U-Pb sur zircons de roches ignées felsiques ordoviciennes s'est avérée difficile en raison de la présence de zircons protérozoïques hérités sous forme de noyaux invisibles dans le zircon magmatique à croissance récente. Les nouvelles techniques mises en oeuvre au Musée royal de l'Ontario, notamment le prélèvement manuel de chaque zircon à analyser, une réduction importante des échantillons rendue possible par le faible nombre de blancs de Pb et l'abrasion pour minimiser ou éliminer la perte secondaire de Pb, se sont révélées inestimables pour traiter les populations de zircon d'âges multiples. On a daté par la méthode U-Pb sur zircons des rhyolites de 473 +3/-2 Ma du groupe de Buchans (avec des conodontes de la toute fin de l'Arénigien et du tout début du Llanvirnien), des rhyolites de 473 +/-2 Ma du groupe de Roberts Arm et des rhyolites de 469 +5/-3 Ma du groupe de Cutwell (avec des conodontes du Llanvirnien inférieur à moyen). Ces datations permettent de corréler ces groupes et d'établir la possibilité de corrélations de datations isotopiques dans les séquences d'arc volcanique.

La tonalite d'un important batholite dans le sud-ouest de Terre-Neuve date de 456 +/-3 Ma; elle est donc exactement contemporaine avec les bentonites du Caradocien inférieur en Suède et au Kentucky. Le groupe de La Poile, une importante séquence volcanique felsique, remonte au Silurien (420 +8/-2 Ma) avec du zircon protérozoïque hérité, et non pas à l'Ordovicien comme on l'a précédemment indiqué.

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INTRODUCTION

Calibration of the geological time scale by isotopic dating of igneous units interstratified with fossiliferous sedimentary rocks is of fundamental importance to the earth sciences. In the assessment of orogenic processes operative in the history of the Appalachian-Caledonian Mountain Belt, it is critical that the age of igneous and metamorphic rock units can be compared directly to the stratigraphic age of sedimentary rocks.

Rb-Sr age determinations have in general been unreliable, likely due to the migration of Sr and Rb in late magmatic or alteration fluids. The resulting ''mixichrons'' can be as much as 50 or 100 m.y. older or younger than the true age of crystallization, based on a comparison of results from rocks in Newfoundland, Nova Scotia, Scotland, and Norway (e.g., Dunning et al., 1987, 1990). Ar-Ar ages record the age at which the analyzed mineral cooled through its blocking temperature and thus may be significantly younger than the age of crystallization in plutonic rocks. These provide valuable data on the cooling history of an intrusion or metamorphic assemblage, but not the primary age. Recently reported Ar-Ar (sanidine, biotite) results from bentonite beds in Kentucky and Sweden are in agreement with concordant zircon ages from the same beds (Kunk et al., 1988) and it is clear that Ar-Ar data provide reliable ages where suitable fresh minerals are preserved in eruptive units.

U-Pb ZIRCON GEOCHRONOLOGY

The U-Pb zircon geochronological technique has several advantages over other methods used to date crystallization of igneous rocks, and has become the standard against which results from other dating techniques are judged. These advantages include the resistance of the mineral zircon to alteration and metamorphism and the fact that, during crystallization, U is accepted into the crystal lattice but Pb is not. This means that the zircon is an ''empty box'' at the time of setting of the isotopic clock, and the Pb content measured in high quality zircon is virtually entirely radiogenic; the ideal case for geochronology. A further advantage is the existence of two independent chronometers; ^{235}U - ^{207}Pb and ^{238}U - ^{206}Pb. The resulting ages can be compared and, if they differ (are discordant), a determination of the degree of Pb-loss or disturbance of the system can be made.

The accuracy and precision of a U-Pb age determination are dependent upon:

1. The accuracy and precision of the measurement of the isotopic ratios. Accuracy is assessed by repeated measurement of standards; precision is a function of the amount of the element present, the stability of the ion beam, and other physical factors. It is assessed by calculations of means, standard deviations, and reproducibility throughout a determination.
2. The ratio of radiogenic Pb to common Pb. A high ratio minimizes the uncertainty in the correction for contaminant common Pb in the analysis. This common Pb occurs in altered domains, inclusions and on the surfaces of the zircon grains, and as common Pb contributed through the laboratory procedures, especially in reagents. Measurement and laboratory procedures will not be covered in this paper except as the latter directly relate to the determination of reliable ages for the Ordovician rocks discussed below.
3. The distribution of data points on the concordia diagram. The distribution is a function of several properties of the zircons analyzed, and the interpretation of many data sets is not straightforward.

Interpretation of U-Pb data arrays

In the simplest case of high quality zircon that has not undergone Pb-loss since crystallization, data points will fall on the concordia curve (Fig. 1; example A), the locus of all concordant ages on the common concordia diagram (Faure, 1986). This simple result is rarely attained from the analysis of large fractions. Improved techniques for the selection of the highest quality grains using a magnetic pin under the microscope (Krogh, 1982a), and abrasion of these grains, which might have undergone Pb-loss, to remove outer surfaces and altered domains (Krogh, 1982b), have resulted in a larger proportion of concordant analyses.

To obtain an analysis of the smaller amount (about 0.20 to 0.05 mg at Ordovician age) of high quality zircon obtained with these procedures, improvements have been made in chemistry and measurement techniques. Laboratory Pb blanks have been reduced to the 2-5 picogram (pg = 10^{-12}g) range so that high ratios of radiogenic Pb:common Pb are maintained despite the analysis of smaller amounts of sample

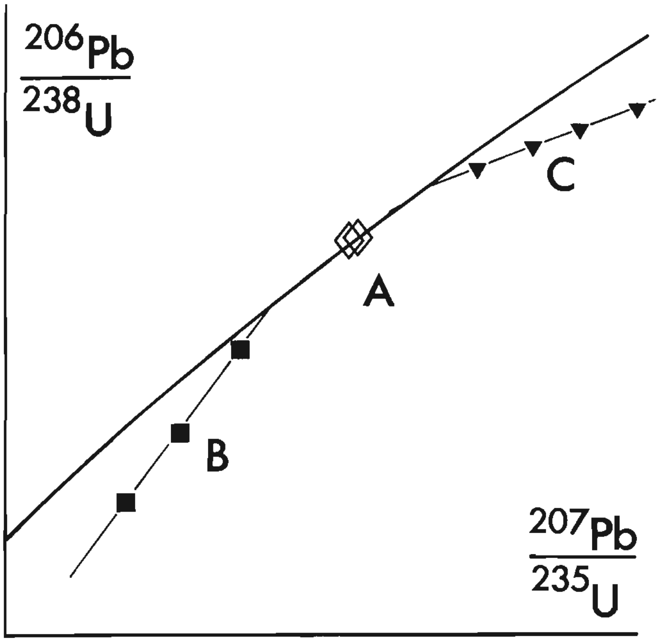

Figure 1. Generalized concordia diagram showing three types of data arrays that yield valid ages. A. duplicated concordant points; B. simple discordia or "Pb-loss" line; C. mixing line. See text for discussion.

Pb. At the same time, the development of commercial mass spectrometers with extended geometry has led to a four- or five-fold increase in ion transmission. This has, to a certain degree, counterbalanced the decrease in the amount of sample Pb measured so that an acceptable measurement precision is maintained.

Analyses from a single age igneous zircon population in some cases yield a linear array defining a Pb-loss trajectory (Fig. 1, example B). The upper intercept of this discordia line with the concordia curve indicates the age of crystallization. In these cases, the uncertainty regarding the age is related to the measurement uncertainty of each analysis, because these propagate through to uncertainty on the slope of the line. A discordia line without a point near concordia will yield a larger uncertainty on the upper intercept age. A fraction of highest quality abraded zircon is always analyzed to provide a data point as close to concordia as possible in order to reduce the uncertainty that results from projection to the concordia curve. This has been carried out in all the examples presented below.

Because of the refractory nature of zircon, many magmas produced by partial melting of continental crust or sediment contain an inherited zircon component. In many Ordovician rocks of the Appalachian-Caledonian Mountain Belt, a component of inherited Proterozoic zircon is present as discrete rounded cores enclosed in newly grown euhedral magmatic zircon. These must be recognized and properly dealt with if a valid crystallization age is to be obtained. It is now clear that, because of the failure to recognize inherited zircon, many published U-Pb zircon ages of Paleozoic rocks are invalid. If core material is present, it is critical that all fractions analyzed be of highest quality zircon, with no cracks or inclusions, that has been strongly abraded to minimize or eliminate secondary Pb-loss. The data points will define a mixing line (Fig. 1, example C) between the average age of inherited zircon at the upper intercept with concordia, and the age of magmatic overgrowths at the lower intercept. If unabraded zircon or poor quality grains are included in the analysis, they may cause the data point to fall below the correct line as a result of secondary Pb-loss. If included in the linear regression, an invalid lower intercept age would result. Where inheritance is present in a sample, it may be possible to select needle-like zircon free of inherited cores for separate analysis. As an alternative, magmatic monazite or titanite can be analyzed, and these should plot on concordia at the lower intercept of the mixing line. However, it must be considered that under certain geological conditions these minerals could yield ages truly younger than the correct zircon age. Whatever the complexity of the zircon population in an igneous rock, it should be possible to determine an accurate and precise age of crystallization by paying attention to the details of morphology and selecting for analysis only the highest quality zircon.

U-Pb zircon ages of Ordovician rocks: Newfoundland Appalachians

Determination of the ages of stage boundaries in the Ordovician is hampered by the fact that the most complete fossiliferous sections in sedimentary basins have few interstratified volcanic horizons, while the volcanic arc sequences only rarely preserve strata containing diagnostic fossil assemblages. As part of a program of U-Pb zircon age determinations of volcanic and plutonic rocks from allochthonous remnants of the Iapetus Ocean, ophiolites and volcanic rocks that are interbedded or associated with fossiliferous strata were dated. These data provide some information to constrain the Ordovician time scale.

Analytical procedures

The results presented below are all relevant to a discussion of the Ordovician System. Some have been presented previously, along with details of procedures. Our analytical techniques have improved significantly since 1982 and this section presents only a generalized discussion that applies to all results. Further details of our present procedures can be found in Corfu and Ayres (1984) and Dunning and Pedersen (1988). Zircons were separated under clean conditions using standard techniques of crushing and mineral separation with a Wilfley table, Frantz isodynamic separator, and heavy liquids. They were then processed into magnetic and size fractions for which the nomenclature is explained in Table 1. Zircons were dissolved and the Pb and U separated by the method of Krogh (1973), with modifications. A mixed ^{205}Pb-^{235}U isotopic tracer (Krogh and Davis, 1975) was introduced into the Teflon capsules with the zircon and 0.5-1.0 ml HF and 5 drops 6N HNO_3 for dissolution, which was accomplished in an oven at 220°C in five days. U and Pb were separated on standard (1982-1984) or miniaturized ion exchange columns (one-tenth of previous resin and reagent volumes) using standard procedures, and loaded on Re filaments using H_3PO_4 and silica gel. Analyses were performed on a MM 30 or VG 354 mass spectrometer with ratios measured in the temperature range 1400 to 1600°C. A single Faraday cup was used for larger samples, except for the ^{207}Pb-^{204}Pb ratio. However, for several small samples, a Daly detector was used for all measurements.

For the Cutwell and La Poile group analyses, uncertainties in the isotopic ratios were calculated using the error propagation procedure of Ludwig (1980). The other analyses were performed in 1982-1984 and were assigned uncertainties based on estimated overall reproducibility. The discordia lines, intercept ages and uncertainties (95% confidence interval) were calculated using the program of Davis (1982).

Betts Cove Complex-Snooks Arm Group

Zircons separated from gabbro of the Betts Cove ophiolite complex (Fig. 2) were clear, coarse, dark to light brown anhedral fragments. Four fractions analyzed defined a short colinear array, with the best abraded fraction only 1% discordant (Fig 3A; Table 1) and yielded an age of 489 +3/-2 Ma (Dunning and Krogh, 1985). Immediately overlying the complex are tuffs, lavas, and sedimentary rocks of the Snooks Arm Group, from which a limited graptolite fauna was obtained that indicates a middle Arenig age (Snelgrove, 1931). This age was challenged as being based on "slight" paleontological evidence (Ross et al., 1978) in a paper that reviewed evidence that could constrain the Ordovician time scale. A new collection of material from this sample locality (Williams, 1989) has yielded a graptolite fauna characteristic of the lower part of the *D. bifidus* Zone of Williams and Stevens (1988). Although these lower middle Arenig strata must be slightly younger than the dated underlying gabbro of the ophiolite, it is considered that they are close in age.

Buchans-Roberts Arm-Cutwell groups

These three stratigraphic packages of volcanic rocks with interbedded sedimentary rocks occur from central to north-central Newfoundland and are allochthonous and structurally imbricated remnants of an island arc of the Iapetus Ocean. The Buchans, Roberts Arm, and Cutwell groups were considered

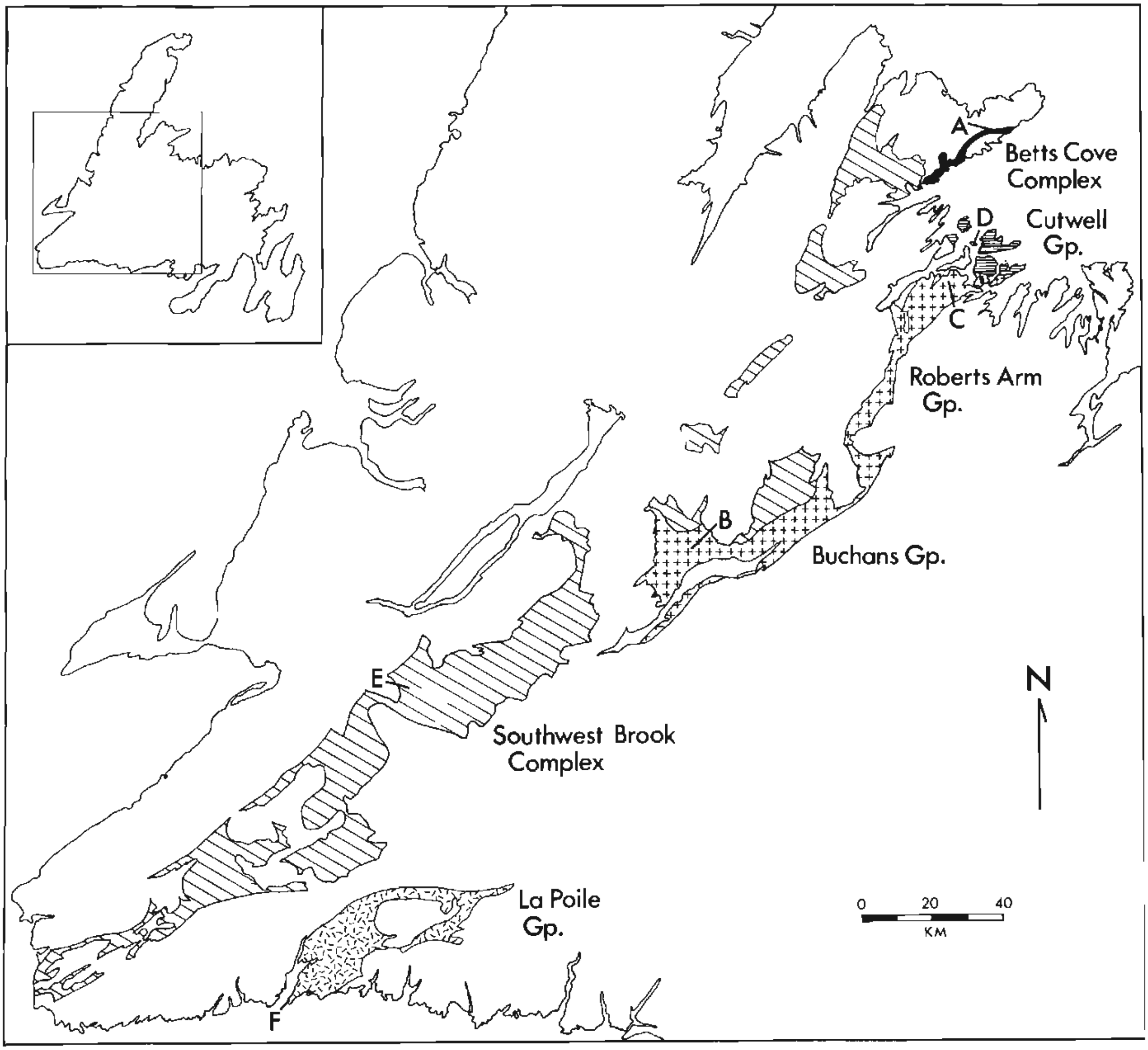

Figure 2. Map of Newfoundland showing distribution of the units discussed in the text and sample locations. Inset map shows area of Newfoundland covered in main map.

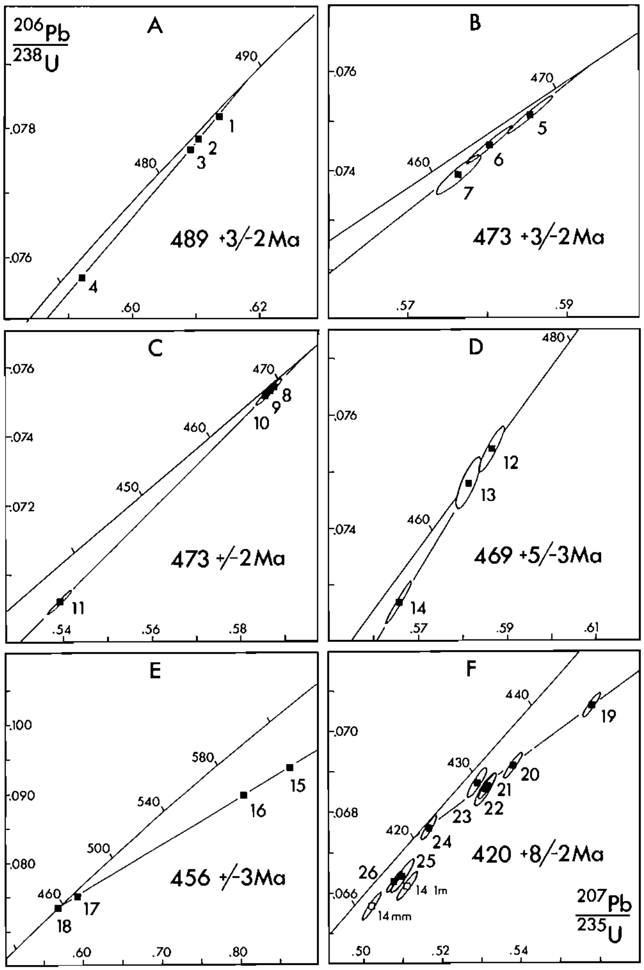

Figure 3A. Concordia diagram for the Betts Cove gabbro. **B.** Concordia diagram for the Buchans Group rhyolite. **C.** Concordia diagram for the Roberts Arm Group felsic tuff. **D.** Concordia diagram for the Cutwell Group rhyolite. **E.** Concordia diagram for the Southwest Brook batholith tonalite showing mixing line with concordant titanite. Error ellipses are smaller than the symbols shown. **F.** Concordia diagram for the La Poile Group felsic tuff. Analyses 14 1m and 14 mm were reported by Chorlton and Dallmeyer (1986) for the same outcrop.

to be of Late Ordovician or Silurian age (Strong, 1977; Kean et al., 1981) based on regional correlation and the requirements of tectonic models. The Cutwell Group was cited as the best example of a post-Caradoc arc sequence because rocks of the group were interpreted as unconformably overlying a dated Caradoc black shale (Dean, 1978). Nowlan and Thurlow (1984) obtained a late Arenig-early Llanvirn age from conodonts in limestone blocks in a breccia within the Buchans Group and suggested that it, and the correlative Roberts Arm Group, were therefore of that age.

Zircons from a rhyolite of the Buchans Group are sparse euhdral to subhedral fine grains. Three fractions analyzed (Fig. 3B; Table 1) show a limited range of discordance from about 1.8 to 4% discordant and yield an upper intercept age of 473 +3/-2 Ma (Dunning et al., 1987). Zircons from a rhyolitic tuff from the upper section of the Roberts Arm Group are abundant, coarse, euhedral grains and the three high quality abraded fractions plot 1.1 to 1.2% discordant. These, together with an unabraded fraction of fine grains, 9.4% discordant, define a line with an upper intercept age of 473 +/-2 Ma (Fig. 3C). These U-Pb zircon ages of rhyolites from the Buchans and Roberts Arm groups confirm their correlation and age. As well, the proximity of the zircon and conodont samples within the sequence at Buchans means that they can be considered coeval (Dunning et al., 1987), so that 473 +3/-2 Ma is a latest Arenig or earliest Llanvirn age.

Resampling of the sedimentary sequence that is overlain by rhyolite of the Cutwell Group was undertaken to check the paleontological age (Dean, 1978), and a U-Pb zircon age was determined for the rhyolite from the same outcrop on the Oil Islands (Fig. 2; locality D). A sparse conodont fauna indicates an early to middle Llanvirn age of deposition (Meyer et al., 1988), refuting the Caradoc age reported earlier.

Zircons separated from the Cutwell Group rhyolite are abundant, coarse euhedral grains. Two abraded and one unabraded fraction analyzed define a line with a 31% probability of fit and yield an upper intercept age of 469 +5/-3 Ma (Fig. 3D; Table 1). This line is anchored on concordia with a concordant point (analysis 12). Graptolites collected from shales of the Cutwell Group at other localities and further collections of conodont samples from limestone lenses in the volcanic rocks have yielded latest Arenig or earliest Llanvirn ages (Williams, 1988, 1989; O'Brien and Szybinski, 1988). These clearly demonstrate that this group is of early Ordovician age everywhere.

The general lithological correlation, overlapping U-Pb zircon ages of rhyolites, and close similarities of the conodont faunas (O'Brien and Szybinski, 1988) strongly support the correlation of the Buchans, Roberts Arm, and Cutwell groups. All three are late Arenig and/or early Llanvirn sequences, based on the available faunal and U-Pb age data; there is no evidence for the presence of a post-Caradocian submarine volcanic arc sequence in the Newfoundland Appalachians. Indeed, it is only because of an artifact of the scale of mapping (1:50,000) and previous invalid radiometric and fossil age determinations that these rocks were assigned to different groups.

Southwest Brook Complex, Newfoundland

A major element of the geology of southwestern Newfoundland is an early Paleozoic batholith with plutons that vary in composition from hornblende gabbro and diorite to tonalite, monzonite, and granodiorite. Components of this batholith include the Cape Ray Granite, the Southwest Brook Complex, Hungry Mountain Complex, Burlington Granodiorite, and other smaller units correlated with them (Dunning and Chorlton, 1985; Dunning et al., 1989). Hornblende and biotite-bearing tonalite plutons make up the largest proportion of the batholith. These rocks intrude and locally engulf mafic ophiolitic rocks throughout southwestern Newfoundland. Coarse clastic sedimentary rocks, correlated with the Eocambrian-Cambrian continental margin sequence of the Fleur de Lys Supergroup, also occur as xenoliths and large rafts in the batholith. A biotite and hornblende-bearing

Table 1. U-Pb data

FRACTIONS			CONCENTRATIONS		MEASURED[b]	ATOMIC RATIOS CORRECTED FOR BLANK AND COMMON Pb[c]				AGE [Ma]
No.	Properties	Wt. [mg][a]	U [ppm]	Pb rad [ppm]	$^{206}\mathrm{Pb}/^{204}\mathrm{Pb}$	$^{208}\mathrm{Pb}/^{206}\mathrm{Pb}$	$^{206}\mathrm{Pb}/^{238}\mathrm{U}$	$^{207}\mathrm{Pb}/^{235}\mathrm{U}$	$^{207}\mathrm{Pb}/^{206}\mathrm{Pb}$	$^{207}\mathrm{Pb}/^{206}\mathrm{Pb}$
A	**BETTS COVE COMPLEX GABBRO**									
1	NO -100 clear Abr	21.20	129	10.3	7652	.125	.0782	.6137	.05693	489
2	NO -100 bulk Abr	11.61	128	10.2	5337	.133	.0779	.6104	.05687	486
3	NO -70 white frosted Abr	2.55	48	3.8	839	.109	.0777	.6093	.05688	487
4	M3 -70 brown cracked frags	2.57	131	10.2	640	.143	.0757	.5923	.05676	482
B	**BUCHANS GROUP RHYOLITE**									
5	N10 -100 clear Abr	0.09	605	58.8	4136	.443	.0751	.5854	.05651	472
6	N10 -200 equant clear Abr	0.09	196	16.7	1078	.271	.0745	.5803	.05650	472
7	N10 -200 clear prisms Abr	0.02	204	16.9	482	.248	.0739	.5764	.05660	476
C	**ROBERTS ARM GROUP FELSIC TUFF**									
8	NO -70 clear Abr	0.28	408	33.6	5046	.214	.0753	.5872	.05654	473
9	NO -100 clear Abr	0.14	400	34.3	13429	.265	.0753	.5867	.05653	473
10	NO -100 clear subhed Abr	0.30	412	35.2	4482	.262	.0753	.5868	.05649	472
11	M2 -200 cracks subhed	0.45	539	44.6	2156	.329	.0692	.5390	.05648	471
D	**CUTWELL GROUP RHYOLITE**									
12	M1 -100 clear euhed Abr	0.20	263	21.0	13017	.176	.0754	.5865	.05644	470
13	M1 -100 clear euhed Abr	0.07	262	20.7	6497	.175	.0748	.5813	.05633	466
14	MO -100 clear equant	0.67	291	22.8	1063	.195	.0727	.5656	.05645	470
E	**SOUTHWEST BROOK COMPLEX TONALITE**									
15	NO -70 clear Abr	2.38	231	22.8	5666	.433	.0939	.8643	.06673	829
16	NO -100 clear Abr	3.11	266	25.3	10943	.389	.0897	.8036	.06498	774
17	NO -325 bulk Abr	0.39	563	46.3	8925	.270	.0750	.5924	.05725	501
18	Titanite brown Abr	4.63	389	28.3	743	.526	.0734	.5685	.05618	459
F	**LA POILE GROUP FELSIC TUFF**									
19	NO -100 Abr	0.27	204	15.8	15285	.218	.0706	.5577	.05727	502
20	NO -100 euhed Abr	0.50	173	13.4	20002	.247	.0691	.5377	.05641	468
21	M1 -325 bulk Abr	0.24	363	27.7	4583	.237	.0687	.5317	.05614	458
22	NO -70 7 euhed grains Abr	0.03	140	10.9	1204	.259	.0686	.5307	.05616	459
23	M1 -325 needles Abr	0.05	340	25.5	4071	----	.0687	.5286	.05580	444
24	MO -100 prism tips Abr	0.10	169	12.9	4882	.249	.0676	.5163	.05539	428
25	M1 -325 euhed needles	0.22	417	30.5	1270	.223	.0664	.5096	.05566	439

Notes: NO, M1 = non-magnetic, magnetic fractions; 0 and 1 indicate degrees of tilt on a Frantz isodynamic separator; -100 = grain size between 100 and 200 mesh; see Krogh (1982) for details of sample prep.

[a] Error in weight +/- 0.005 mg (1 sigma).

[b] Corrected for fractionation and common Pb in the spike

[c] Corrected for fractionation, common Pb in the spike, 5 to 20 pg Pb blank and common Pb at the age of the sample calculated from the model of Stacey and Kramers (1975) and 2 pg U.

tonalite was dated from the Southwest Brook Complex in southwestern Newfoundland (Fig. 2). Three fractions of clear euhedral abraded zircon define a mixing line that is anchored on concordia by an abraded titanite fraction (analysis 18). The line through all points has a 95% probability of fit and yields a lower intercept age of crystallization of 456 +/-3 Ma with an average age of inherited zircon cores of 1430 +18/-17 Ma (Fig. 3E). This age is earliest Caradocian and identical to that reported for the "Big Bed" at Kinnekulle, Sweden (Kunk et al., 1988) and correlative units in Kentucky and elsewhere. It is suggested that plutons of this batholith, others of which have yielded ages of 463 +/-6 Ma (Burlington Granodiorite; Mattinson, 1975) and 467 +/-8 Ma (Hungry Mountain Complex; Whalen et al., 1987), may have been the magma sources for eruptive units now preserved in the Ordovician sections in North America. Another batholith of similar age and with a similar range of rock types is present in the Smola-Hitra Terrane of the western Trondheim district, Norway (R.D. Tucker, pers. comm., 1988), near the Swedish sections.

La Poile Group

The La Poile Group is a subaerial volcano-sedimentary complex that is exposed in the western part of southwestern Newfoundland (Fig. 2). In most areas its northern boundary is the Bay d'Est Fault; it is bounded to the south by either the Burgeo intrusive suite or, as is the case in the La Poile Bay type area, by mylonite.

Cooper (1954) established a tripartite division in the La Poile Bay area that included a basal conglomerate, an intermediate, mainly sedimentary unit, and an upper widespread unit of subaerial ash-flow tuff, rhyolite, lahar, and associated finer grained epiclastic rocks (Chorlton, 1980; S. O'Brien, 1983; B. O'Brien, 1987). Cooper (1954) originally assigned these rocks to the Devonian, based on a lithological correlation with fossiliferous Devonian strata to the west, whereas Chorlton (1980) subsequently assigned an Ordovician age to the La Poile Group. Chorlton and Dallmeyer (1986) reported a U-Pb age of 452 +51/-13 Ma for the La Poile Group, based on analysis of bulk zircon fractions from two different tuff horizons.

A quartz and feldspar-phyric tuff collected from the east side of La Poile Bay at locality 14 of Chorlton and Dallmeyer (1986)(Fig. 2; locality F) yielded abundant clear, euhedral grains. Eight fractions with differing grain size and magnetic properties were analyzed (Table 1), including six abraded fractions that define a linear trend (Fig. 3F) from analysis 24, closest to concordia, to analysis 19. The lower intercept of 420 +8/-2 Ma (probability of fit 12%) is taken as the age of eruption of this rhyolite (= Silurian). The abraded tips of euhedral prismatic zircons (24) contain the least amount of inherited core material because cores typically occur centrally in the grain. The upper intercept of 1376 +/-120 Ma is the average age of the cores, but these could have actual ages of 1 and 1.6 Ga for example. Analyses 5 and 6, comprising cracked or very fine grains, have undergone significant secondary Pb-loss and approximately duplicate bulk analyses (14-mm, 14- 1m) reported for the same outcrop by Chorlton and Dallmeyer (1986). These authors grouped their two data points with two from another sample of the La Poile Group collected 7 km away (loc. 18; Chorlton and Dallmeyer, 1986) and fitted a discordia line to all four points to determine an age of 452 +51/-13 Ma, noting that three of four fractions give similar $^{207}Pb/^{206}Pb$ ages of 448 to 455 Ma (each +/-7 m.y.) as additional justification for their age. However, they failed to recognize that this sample contained a mixed-age zircon population.

This is a clear-cut example emphasizing the need for careful selection of the highest quality zircons, followed by abrasion to eliminate altered portions of grains, prior to analysis. In a sample with inherited zircon, it is only such fractions that define a valid mixing line projecting to the true age of crystallization. Analyses of large fractions or poor quality zircon yield *no* valid age information.

DISCUSSION

The U-Pb zircon ages presented here demonstrate different types of data sets in rocks of Ordovician or, in the case of the La Poile Group, previously interpreted Ordovician age. The importance of selection of the highest quality grains for analysis is emphasized in all cases, as these yield less discordant, or truly concordant, data points, which provide more accurate and precise ages. The potential for matching isotopic and fossil ages is demonstrated with the Buchans and Cutwell group examples and further U-Pb work is currently underway on other fossiliferous Ordovician strata. The previously reported U-Pb zircon age for the La Poile volcanics is typical of many in the literature, where large zircon fractions were analyzed without attention being paid to the complexity or quality of the zircon population. The data points can appear consistent (and colinear) because the large sample weights of zircon analyzed lead to an averaging of the mixed magmatic and inherited component in each fraction. In such a case it would not be possible to get a concordant fraction, but because concordant points are not commonly obtained without careful sample selection and abrasion, this might not arouse the suspicions of the analyst. The result is then published and believed because of the high degree of confidence placed in zircon ages in general. Although both of the present authors are enthusiastic U-Pb geochronologists, they hope to have made the point here that all zircon ages are not created equal. Many published ages are invalid as a result of the analysts' failure to recognize the complexity of some zircon populations and deal with them effectively (see examples in Rogers et al., 1989; Rogers and Dunning, in press). It is important that geologists become aware of possible pitfalls as well as the inherent advantages of the U-Pb geochronological technique. They can then critically assess U-Pb data sets and judge the quality and validity of the ages reported. Most credence should be attached to U-Pb ages based on concordant data points and on analyses with a high ratio of radiogenic Pb to common Pb. Discordia or mixing lines involving three or more analyses of carefully selected, high quality fractions, including abraded fractions, are the most reliable. In the final analysis, consistency, reproducibility and compatibility with clear field relationships are important tests of any isotopic age determination.

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Ordovician vertebrates

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Abstract

The fossil record for the early evolution of vertebrates consists of the remains of Agnatha (jawless vertebrates) from the Ordovician of Australia, North America, and Bolivia. Although vertebrate fragments have been reported from the Upper Cambrian and Lower Ordovician, none can be conclusively identified as vertebrate and their status remains uncertain at present. The oldest agreed vertebrates occur in the Stairway Sandstone of central Australia (Arandaspis, Porophoraspis) *of early Llanvirn age, whereas slightly younger remains occur in the Harding Sandstone and equivalents of the U.S.A. and Canada* (Astraspis, Eriptychius) *and the Anzaldo Formation of Bolivia* (Sacabambaspis) *of early Caradoc age. All occur in very nearshore, intertidal to subtidal, marine environments. Present knowledge of the anatomy of these animals shows that the Australian and Bolivian forms are more closely related to each other than to the North American genera. Although the Ordovician vertebrates had previously been classified with the Heterostraci because of their external armour of bony plates formed of aspidin, it has now been shown that they possess primitive features such as numerous branchial openings and superficial sensory grooves that would exclude them from that group. Understanding of relationships with later vertebrates is hampered by a gap in the fossil record between the Caradoc and the upper Llandovery that correlates with Late Ordovician glaciation.*

Résumé

Les premières traces fossiles de l'évolution des vertébrés sont des poissons fossiles sans mâchoires (''Agnathes'') récoltés dans les roches ordoviciennes d'Australie, d'Amérique du Nord, et de Bolivie. Bien que des microfragments de tissue phosphaté du Cambrien supérieur ou de l'Ordovicien inférieur aient été signalés, il n'existe aucune certitude permettant de désigner ces restes comme appartenant à des vertébrés. Les plus vieux vertébrés reconnus sont donc ceux des grès de Stairway du centre de l'Australie (Arandaspis *et* Porophoraspis, *du Llanvirn inférieur), des grès de Harding et de leurs équivalents américains et canadiens* (Astraspis *et* Eriptychius) *ainsi que de la formation Anzaldo de Bolivie* (Sacabambaspis *du Caradoc inférieur). Tous les gisements appartiennent à un milieu marin de type intertidal à infratidal. L'étude des relations de parenté des divers taxons permet de rapprocher le genre bolivien Sacabambaspis du genre australien Arandaspis, les genres nord-américains étant phylogénétiquement plus éloignés. Par ailleurs, les vertébrés ordoviciens étaient, jusqu'á présent, classés parmi les Hétérostracés du fait de leur armure osseuse externe formée d'os acellulaires (apidine) cependant des caractères primitifs découverts sur les spécimens boliviens, telle que la présence de plusieurs ouvertures branchiales et de sillons sensoriels superficiels, les excluent de ce groupe à moins d'en modifier la définition. La compréhension des relations phylogénétiques entre les premiers vertébrés et les formes plus tardives est rendue difficile par l'absence de fossiles entre le Caradoc et le Llandovery supérieur, un intervalle de temps qui est en corrélation avec la glaciation qui eut lieu au cours de l'Ordovicien supérieur.*

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INTRODUCTION

For many years there has been considerable interest in the earliest vertebrates due to the light that they might shed on questions of early vertebrate relationships, and the environment in which the earliest vertebrates developed (Romer, 1946; Denison, 1956; Boucot and Janis, 1983). Although for many years there was little information on vertebrates below the Silurian (Walcott, 1892), in recent years there has been a sudden increase in our knowledge of these forms due to new material being available (Ritchie and Gilbert-Tomlinson, 1977; Gagnier et al., 1986) or new interpretations being made (Elliott, 1987). In addition, the search for the earliest vertebrates has resulted in a number of Upper Cambrian and Lower Ordovician occurrences of supposed vertebrate material that are of doubtful significance (Bockelie and Fortey, 1976; Repetski, 1978).

In this paper we attempt to distinguish the earliest vertebrates and review what is known of their relationships, ecology, and distribution.

WHAT ARE THE EARLIEST VERTEBRATES?

During the last 130 years, a number of Cambrian and Ordovician phosphatic fossils have been identified as vertebrate. However, not all of these determinations have been accurate and it is now clear that many of these fossils do not show vertebrate morphology or histology. Most of the earlier designations have now been corrected. For instance, Pander (1856) considered that Ordovician and Silurian conodonts were fish teeth. Though there have been many different opinions about its affinity, nothing has been known about the gross morphology of the conodont animal until recently, when a well preserved specimen was described (Briggs et al., 1983). Though opinion seems to be divided as to whether conodonts are chordates (Aldridge et al., 1986; Briggs and Clarkson, 1987) or members of an invertebrate group such as the aplacophoran molluscs (Tillier and Cuif, 1986; Briggs et al., 1987; Janvier, 1988), it is at least clear that they are not vertebrates. Other fragments such as *Archeognathus* (Cullison, 1938; Miller et al., 1947) also have affinities with conodonts (Klapper and Bergström, 1984).

The main feature that distinguishes lower Paleozoic vertebrates from conodonts is the histology of the hard tissues. In conodonts this is basically lamellar, whereas in vertebrates it is typically a honeycomb or reticulated structure of the cancellous layer of the dermal bone. This means that conodont-bearing animals or other described enigmatica cannot be considered vertebrate.

In 1976, Bockelie and Fortey named a new taxon, *Anatolepis heintzi*, from phosphatic fragments found in the Valhallfonna Formation of Spitsbergen. The fragments were dated as Arenig or possibly early Llanvirn age, and were interpreted as the earliest vertebrates known at that time. *Anatolepis* has since been reported from the Upper Cambrian and Lower Ordovician of the U.S.A. (Wyoming, Montana, Oklahoma, Washington, Alaska, Nevada, Arkansas, New York, Utah, Texas), Greenland, and northern Germany (Peel and Higgins, 1977; Repetski, 1978, 1980; Boucot and Janis, 1983; Schallreuter, 1983). However, there is as yet no definite proof that these fragments are from vertebrates. Though they appear to be composed of hydroxyapatite as in vertebrates, this is not in itself proof of a vertebrate origin. The surface is ornamented with scale-like tubercles, unlike surface structures seen in accepted early vertebrates. In addition, the fragments are extremely small and thin (70-100 μm thick) so that nothing is known of the gross morphology of the animal. Though Bockelie and Fortey (1976) and Repetski (1978) state that their fragments show a microstructure similar to that found in vertebrates from the Silurian and Devonian, their illustrations are not convincing and they did not show any of the features that would identify the tissues as acellular bone (aspidin), a tissue characteristic of the earliest vertebrates. Peel and Higgins (1977) have suggested that the material actually consists of fragments from a merostome arthropod, some of which have a phosphatic, three-layered shield with a scaled surface ornament. This has been refuted by Briggs and Fortey (1982). There is clearly insufficient evidence at present to designate *Anatolepis* as a vertebrate (see also Peel, 1979).

Other phosphatic microremains have been designated as possible vertebrates in recent years (e.g. *Hadimopanella* [Bengtson, 1977; Gedik, 1977; Harper, 1979; see also Märss, 1988]) but as yet it has not been shown conclusively that any of them have vertebrate characteristics.

In addition, Repetski (1981, p. 399) has suggested a close similarity between *Hadimopanella* and *Utaphospha* Müller and Miller (1976) (see also Peel and Larsen, 1984), however, Bendix-Almgreen and Peel (1988) have suggested that both genera represent the sclerites of urochordates, a group that is known down to the Lower Cambrian (Müller, 1977; Zhang, 1985; Bendix-Almgreen and Peel, 1988).

It should be pointed out here that most papers describing phosphatic microremains feature SEM photographs of external surfaces (often weathered) of hard tissues. This is insufficient to distinguish typical vertebrate or nonvertebrate histology, however, and thin sections of good material are an absolute necessity.

Remains that could be conclusively designated vertebrate were first described from the Ordovician by Walcott (1892), who reported three new species from the Harding Sandstone of Colorado. The first two of these, *Astraspis desiderata* and *Eriptychius americanus*, consisted of numerous fragments of bony elements and scales that covered the animals with a flexible armour. Study of the histology of the individual elements and gross morphology of the few partially articulated specimens proves conclusively that these animals were vertebrates. The third of Walcott's species, *Dictyorhabdus priscus*, was described as the notochordal sheath of a chimaeroid fish. Although it has subsequently been assigned to various invertebrate groups, both Denison (1967) and Spjeldnaes (1979) have pointed out that it cannot be unequivocally assigned at present to any known group.

In 1977, Ritchie and Gilbert-Tomlinson described vertebrate fragments from the Stairway Sandstone (late Arenig or early Llanvirn age) of central Australia. Of the two species described, *Arandaspis prionotolepis* is represented by a number of shield fragments and scales preserved as external moulds. As no original material is present, the microstructure and composition of the hard tissues is unknown. Despite that, and the incompletely known morphology of the animal, it was identified as a heterostracan, a diverse group of Paleozoic agnathans, and placed within its own family, the Arandaspididae. The second species, *Porophoraspis crenulata*, is known only from a few fragments and is assumed to be an agnathan similar to *Arandaspis*.

Most recently, a new agnathan was described from the Anzaldo Formation of Bolivia (Gagnier et al., 1986). Named *Sacabambaspis janvieri*, this animal comes from rocks of early Caradoc age and almost complete specimens are known (Gagnier, 1989). The gross morphology of *Sacabambaspis* is known and its microstructure is compatible with that of other vertebrates.

It seems, therefore, that though a number of occurrences of vertebrates have been reported from the lower Paleozoic, only some of them have been substantiated by details of their gross morphology and histology. These are: 1) *Astraspis* and *Eriptychius* from the lower and middle Caradoc of North America; 2) *Arandaspis* and *Porophoraspis* from the upper Arenig or lower Llanvirn of Australia; 3) *Sacabambaspis* from the early Caradoc of central Bolivia.

ORDOVICIAN VERTEBRATES

North American vertebrates

Vertebrates from the Ordovician were first described by Walcott (1892). More detailed descriptions were provided by Bryant (1936). Walcott described two taxa still accepted as vertebrate, *Astraspis desiderata* and *Eriptychius americanus*, mostly from fragmentary remains. Since then, these same vertebrates have been reported from a large number of localities of similar age in the U.S.A. and Canada (Ross, 1957; Ørvig, 1958; Sinclair, 1958; Sawin, 1959; Denison, 1967; Eliuk, 1973; Lehtola, 1973, 1983; Nitecki et al., 1975; Ossian and Halseth, 1976; Darby, 1982).

Astraspis desiderata (Figs. 1B; 2A; 3A) is normally found as isolated tesserae ornamented with small stellate tubercles. The histology of these tesserae has been discussed by Denison (1967), who has shown that they are three-layered. A basal laminated layer of acellular bone (aspidin) is succeeded by a middle layer, which consists of a complex meshwork of canals surrounded by concentrically laminated aspidin. The surface layer of each tubercle consists of a pulp cavity surrounded by orthodentine through which minute tubules radiate, and a clear capping of durodentine.

Two articulated specimens are known. The first of these is the external mould of an articulated head shield, first described by Eastman (1917) though Walcott mentioned it in a footnote to his 1892 paper. Ørvig (1958) identified the position of the lateral line canals and showed that the rostral area of the shield was lacking. The shield is composed of a large number of interlocking tesserae bearing small tubercles; as larger tubercles are common on isolated tesserae, it may be that this specimen represents a young individual. A central ridge is flanked by a pair of median ridges and further ridge plates form the lateral margin of the shield. There is no indication of the position of the orbits; a group of high-crowned tubercles in an anterior median position may indicate the position of the pineal body.

The second specimen was collected in 1968 and consists of most of the body and tail of one individual, together with a partial headshield. It was originally described as a poorly preserved mass of scales (Lehtola, 1983), but subsequent work (Elliott, 1987) has shown that the margin of the headshield is preserved and that the orbit and a series of eight branchial openings are present. This specimen also shows an entire caudal fin, which consists of a clump of large triangular scales. No internal structures are present with either specimen, but pieces of globular calcified cartilage present in collections of fragmentary *Astraspis* material probably represent part of the calcified endoskeleton.

Ørvig (1958) described *Pycnaspis splendens* from the Bighorn Mountains of Wyoming. He distinguished it from *Astraspis* on the basis of large mushroom-shaped tubercles present in the adult, though in immature specimens the ornamentation was similar to that in *Astraspis*. Denison (1967) referred *Pycnaspis splendens* to *Astraspis*, arguing that a size difference was apparent in the Wyoming specimens when compared to the original material from Colorado, but that it was of specific significance only.

The other Harding Sandstone vertebrate, *Eriptychius americanus* (Fig. 3B) is similarly found normally as disarticulated tesserae and scales. These are ornamented with coarse ridges and are thus easily distinguishable from those of *Astraspis*. The histology of the plates and scales differs from that found in *Astraspis* by the presence of large ascending canals in the basal layer of aspidin, and the presence of elongated pulp canals beneath the dentine ridges. Denison (1967) has also reported the presence of an enameloid layer on the surface of some of the ridges.

Only one partially articulated specimen is known. This shows part of the rostral area of the shield, and Denison (1967) was able to show that a number of pieces of globular calcified cartilage were present. As the plates were applied closely to the surface of the cartilage, these clearly were part of the internal skeleton of *Eriptychius* and presumably represent the rostral and orbital cartilages. Isolated pieces of globular calcified cartilage from localities yielding *Eriptychius* plates do show a consistent morphology, but as yet no attempt has been made to identify the position of these endoskeletal elements. Ørvig (1958) described a perforated plate of *Eriptychius* that might be a branchial plate; but without further articulated material it is not possible to say if *Eriptychius* had a branchial structure similar to that of *Astraspis* or not. Denison (1967) raised an additional species, *E. orvigi*, for plates from Wyoming that were thicker and showed coarser ornament than those of *E. americanus*.

Although it has not been possible to reconstruct *Eriptychius*, it has been assumed that it was a similar animal to *Astraspis* and probably closely related to it. However, differences in the histology of the bone and in the shape and appearance of the body scales may indicate that the relationship is a more distant one. Some cellular bone fragments have been figured from the Harding Sandstone (Denison, 1967), but for now they cannot be referred to any known group.

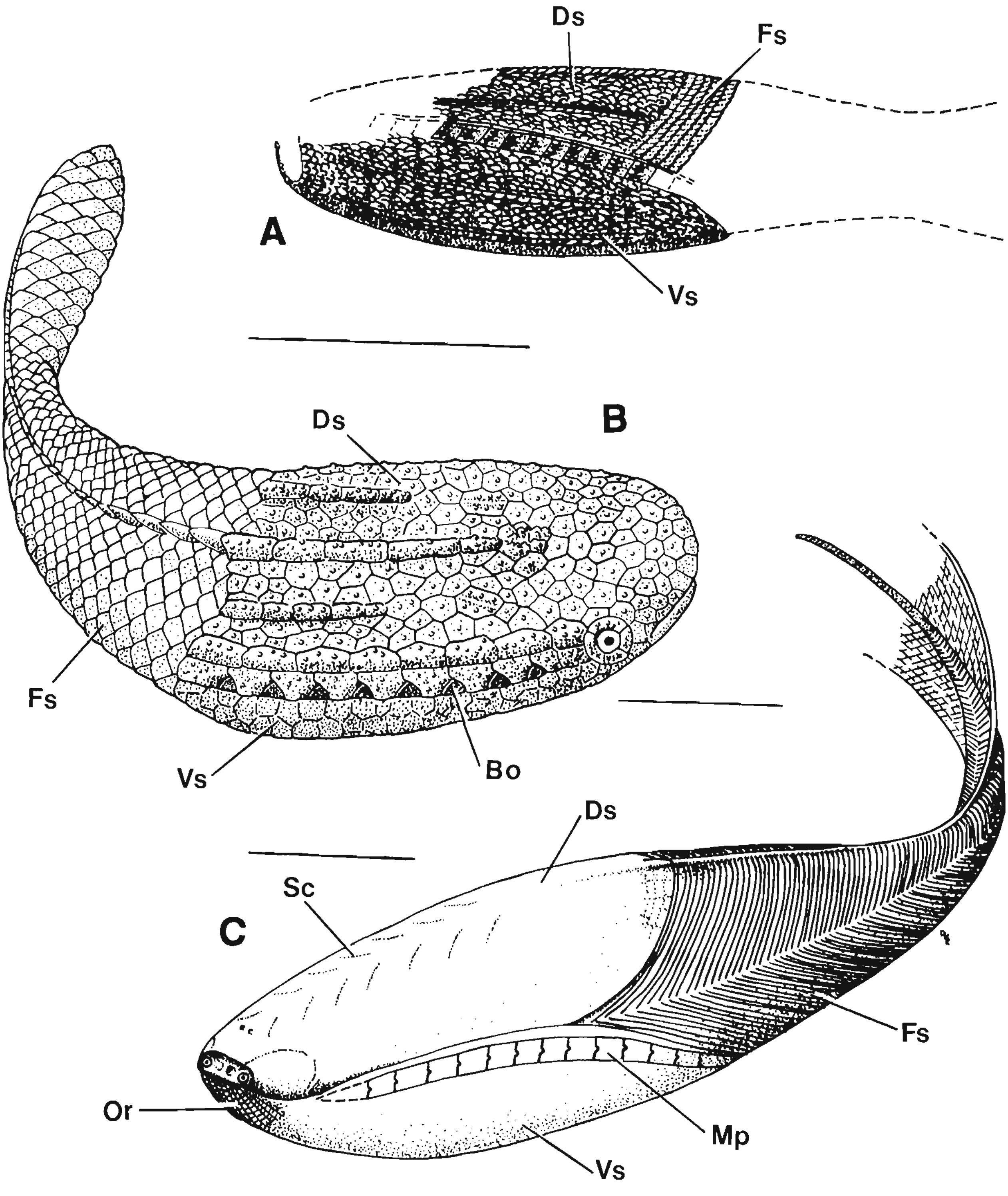


Figure 1. Reconstructions of Ordovician vertebrates. *A. Arandaspis prionotolepis* (after Ritchie and Gilbert-Tomlinson, 1977). *B. Astraspis desiderata. C. Sacabambaspis janvieri.* Scale bars = 5 cm. Abbreviations: Bo, branchial opening; Ds, dorsal shield; Fs, flank scale; Mp, marginal plate; Or, oral area; Sc, sensory canal; Vs, ventral shield.

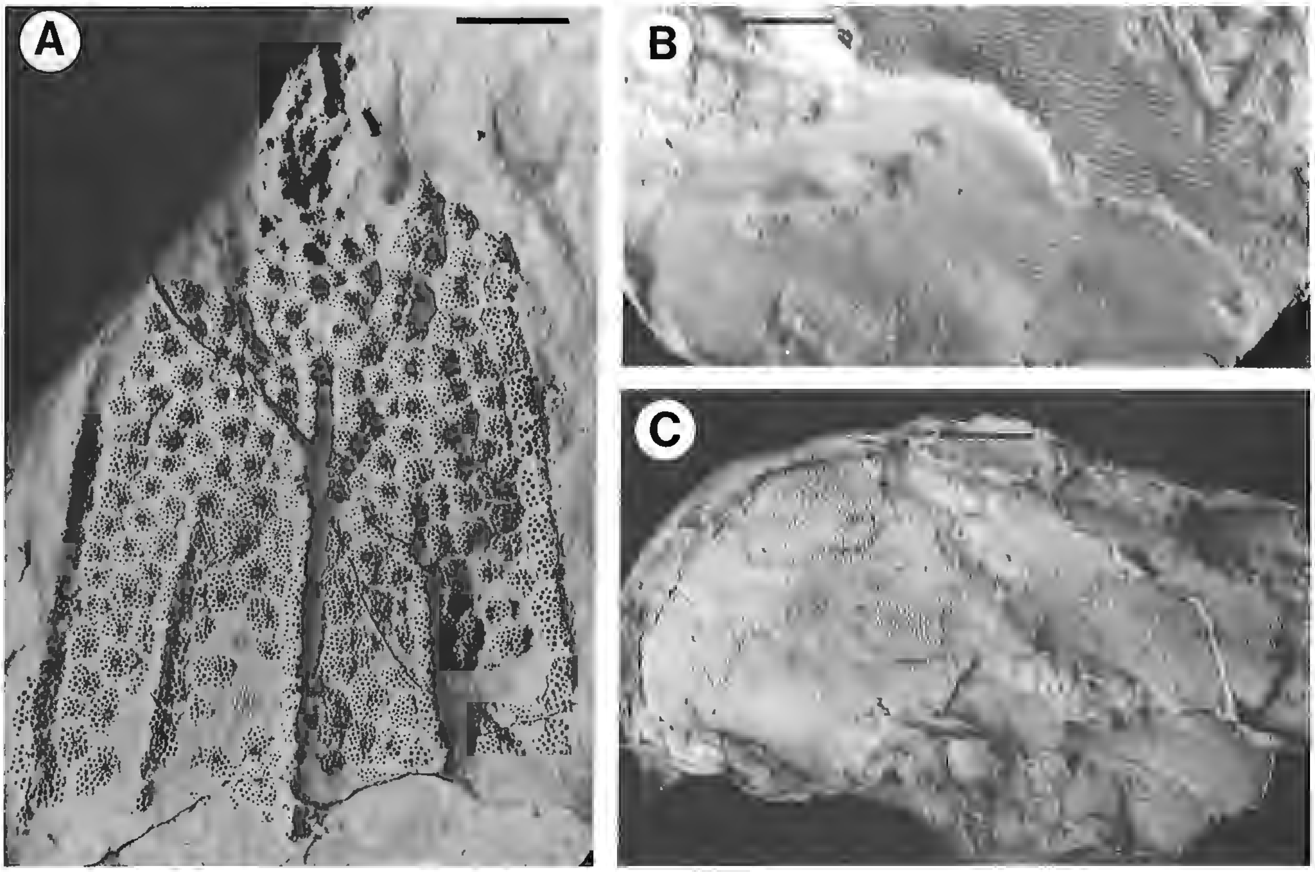


Figure 2. Shields of Ordovician vertebrates. *A. Astraspis desiderata*, dorsal shield, USNM8121. *B. Arandaspis prionotolepis*, shield fragment with isolated scale, CPC15469a. *C. Sacabambaspis janvieri*, dorsal shield, YPFB PAL 6221. Scale bars = 1 cm.

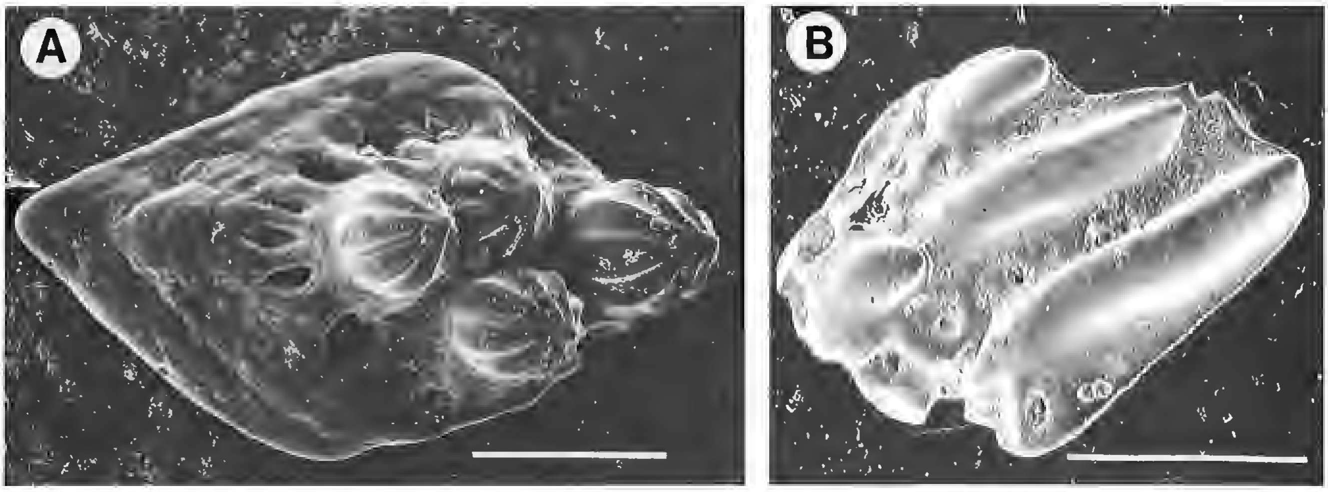


Figure 3. SEM photographs of tesserae of *Astraspis* (**A**) and *Eriptychius* (**B**). Scale bars = 0.5 mm.

Australian vertebrates

Vertebrates were first described from the Ordovician of Australia in 1977 (Ritchie and Gilbert-Tomlinson). The material comes from the Stairway Sandstone in the Amadeus Basin, Northern Territories, and consists of moulds of plates and body scales. No bone is present so no details of the histology or composition are known. One species, *Arandaspis prionotolepis* (Figs. 1A; 2B), is the most common, and though no complete specimens are present, enough partially articulated material is known to enable a reconstruction to be made (Ritchie, 1985). Large dorsal and ventral median plates were separated by a slanting row of square (possibly branchial) plates. No branchial openings can be discerned, however. The ornamentation consists of closely packed oval or rhomboidal tubercles. Dorsal shields show a parallel pair of longitudinal grooves, bordered by regularly arranged denticles, which are probably the sensory canal grooves.

It has recently been shown (Ritchie, 1985) that there is a double opening anteriorly in the midline of the dorsal shield that may represent the position of a pineal-parapineal structure. No specimen shows the rostral part of the animal, however, and so nothing is known about the position or structure of the oral and orbital areas.

The trunk appears to have been covered by a series of long thin scales with parallel rows of ornamentation. These were arranged in two slanting rows on each side of the body. No details of the tail structure are known.

The second species from the Stairway Sandstone, *Porophoraspis crenulata*, is only known from a few plate fragments, but its very distinctive ornament of low-rounded tubercles with crenulate margins indicate that it is generically distinct.

Bolivian vertebrates

The most recent discovery of Ordovician vertebrates was made in the Anzaldo Formation of Bolivia (Gagnier et al., 1986). One new species, *Sacabambaspis janvieri*, has been described, initially from fragments, but more complete material is now known (Gagnier, 1989). This new agnathan shows many points of similarity with *Arandaspis*, although the ornamentation is "oak-leaf" rather than rhomboidal in shape. Median dorsal and ventral plates are separated by a slanting row of twenty square plates that are probably branchial, though no branchial openings can be seen (Figs. 1C; 2C). On the dorsal shield, the sensory canal system consists of two rows of short, paired grooves in an en echelon pattern in the midline of the shield. Anteriorly, a double median pineal-parapineal opening is present as in *Arandaspis*. The anterior margin is embayed and delimits an elliptical space that contains the orbits and (possibly) olfactory capsules. The anterior margin of the ventral shield has a series of small square platelets in rows, which form an exoskeletal mouth apparatus. Though no details are available as yet, it is reported that the hard tissues of *Sacabambaspis* are composed of aspidin, and it is also probable that the shields are formed of small tesserae that have fused (Gagnier, 1989).

The body is covered by rows of thin elongated scales ornamented with "oak-leaf" tubercles. The scales are organized in two main rows arranged in a chevron pattern down each side of the body and separated dorsally and ventrally by rows of median ridge scales. The tail is incomplete but it is covered by small scales and seems to have a long chordal extension.

RELATIONSHIPS OF ORDOVICIAN VERTEBRATES

The first Ordovician vertebrates to be described, *Astraspis* and *Eriptychius*, were considered to be placoderm fishes (Walcott, 1892), though this was challenged by Cope (1893) who pointed out that they were more likely to be Agnatha. This view has been accepted and these forms have traditionally been attributed to the Heterostraci, and even in some schemes (Obruchev, 1964) placed in separate orders, the Astraspidida and Eriptychidida. Workers on the histology of their hard tissues (Ørvig, 1958; Denison, 1967) have demonstrated an apparent similarity to the structures seen in typical heterostracans. Due to this initial designation it was assumed that *Arandaspis* and *Porophoraspis* were also heterostracans (Ritchie and Gilbert-Tomlinson, 1977), although their histology was unknown and the major morphological characters could not be demonstrated.

A similar assumption has been made by Gagnier et al. (1986) in relating *Sacabambaspis* to the heterostracans on the basis of the presence of two large shields and the gross histology of the hard tissue.

In order to decide if these forms are heterostracans, it is first necessary to assess the characters of that group. The Heterostraci are a group of jawless vertebrates that were separated from the Osteostraci by Lankester (1868-70) on the basis of the external ornamentation and histological structure of the bony plates that were composed of acellular bone or aspidin. A further diagnostic feature described on all well preserved heterostracans is the presence of one common, external branchial opening on each side of the dorsal shield. Both these features are considered to be synapomorphies (shared derived characters) of the heterostracans and a further apomorphy (derived character) may be the presence of a single median growing plate above and below the head (Janvier and Blieck, 1979; Janvier, 1981; Forey, 1984). However, though these characters can be demonstrated in the better known groups of heterostracans, there are a number of poorly known groups whose designation as heterostracans relies on somewhat limited evidence. This problem has recently been discussed (Blieck et al., in press) and it has been concluded that a considerable amount of work is needed before many of the groups can be adequately defined and included within a phylogenetic scheme.

This is particularly true of the tessellated forms in which the shields are formed of small tesserae or platelets. As the armour tends to disintegrate on death, little is known of the gross morphology of these organisms and their attribution to the Heterostraci has rested almost entirely on the histology of the plates (see Elliott and Loeffler, 1990, for a discussion of this problem).

Recent work on *Astraspis* (Elliott, 1987) has demonstrated that this animal possessed a series of eight external branchial openings, despite having a histology apparently similar to that found in heterostracans. Clearly *Astraspis* and probably also *Eriptychius* cannot be considered heterostracans sensu stricto as they now only possess the synapomorphy of hard tissues composed of aspidin. It also seems that *Arandaspis* and *Sacabambaspis* cannot be considered heterostracans as, although they apparently possessed aspidin, they also appear to have a series of external branchial ducts.

What characters are available to assess the phylogenetic position of the Ordovician forms if they are not heterostracans? Clearly both *Arandaspis* and *Sacabambaspis* are very similar organisms that form a distinct group. They share a suite of distinctive characters including the presence of a slanting row of branchial plates, shields formed of small fused tesserae, sensory canals in open grooves on the surface, double anterior dorsal openings in the shield, and elongated scales arranged in a chevron pattern.

The North American forms are quite distinct, however. *Astraspis* and *Eriptychius* are forms in which the shields are formed of tesserae that fuse in some areas only; the branchial ducts are open, eight in number and form a horizontal row (known only for *Astraspis*). No pineal opening is visible, scales are thick and polygonal in *Astraspis*, and more elongated in *Eriptychius*.

These two groups are clearly distinct. The only characters that relate them to each other are the presence of numerous branchial ducts, sensory canals borne in open surface grooves, and hard tissues composed of aspidin. The first two characters are almost certainly primitive for vertebrates and differ in detail in the two groups. Branchial ducts are slanting, closed, and twenty in number in the *Arandaspis-Sacabambaspis* group; open, horizontal and eight in number in the *Astraspis-Eriptychius* group (in so far as this is known). Sensory canals are borne in continuous grooves in the *Arandaspis-Sacabambaspis* group, and in occasional short grooves between tubercles in the *Astraspis-Eriptychius* group.

As the hard tissue of *Sacabambaspis* is aspidin, it appears that it shares this character with *Astraspis*, *Eriptychius*, and the heterostracans. However, acellular bone is also present in anaspids, another agnathan group. This may indicate that acellular bone is primitive among agnathans and among vertebrates; however, the acellular bone of anaspids is lamellar and without a cancellous layer (Gross, 1958), which might suggest that acellular bone has developed more than once. The acellular bone present in some osteostracans, acanthodians, and teleost actinopterygians is certainly considered to be secondary (Ørvig, 1967). If this is the case, then its presence cannot be used as a character linking these early groups. It has been suggested (Blieck et al., in press) that the presence of long thin flank scales arranged in a chevron, and a slanting row of branchial ducts might be characters that could indicate a relationship between *Sacabambaspis* and anaspids. However, this raises other problems of relationship with osteostracans, and it is probable that these similarities are not significant.

At present, therefore, though at least two distinct groups of Ordovician vertebrates can be distinguished, it is difficult to determine their relationship to each other or to the later agnathan groups. This difficulty is mostly due to the lack of agnathan specimens through the Upper Ordovician and Lower Silurian, which means that there are no known intermediates between the early forms and the more numerous and diverse later groups.

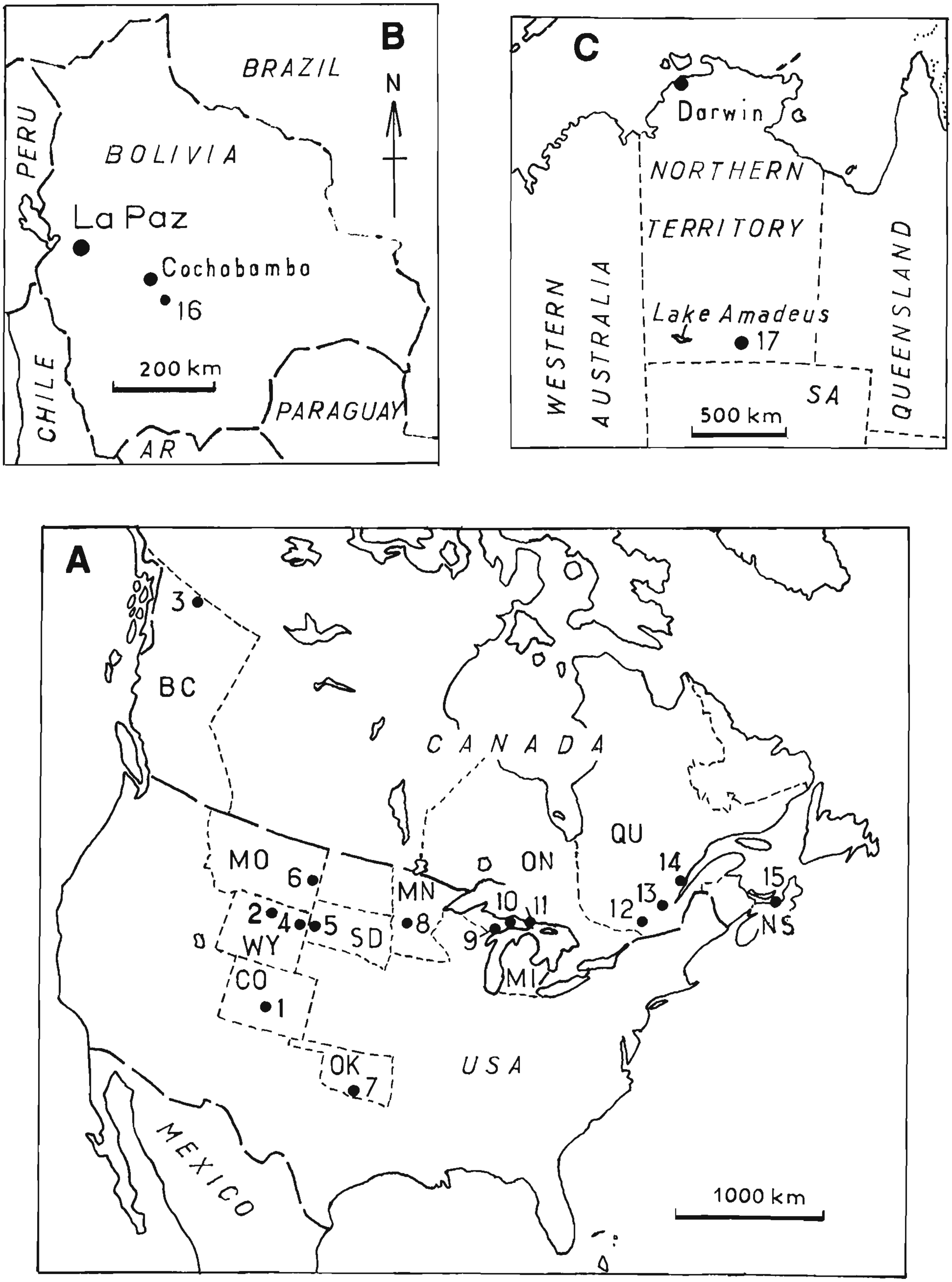
ENVIRONMENT AND ECOLOGY

The environment in which the early vertebrates lived has been the subject of some dispute over the years due to its bearing on the environment in which the vertebrates originated. Romer (1946) used the then freshwater interpretation of the Harding Sandstone in combination with ideas on the way in which the glomerular kidney had developed (Smith, 1932; Romer and Grove, 1935) to support the view that vertebrates had developed in fresh water. These ideas were questioned later, however (Denison, 1956; Robertson, 1957, 1963; Darby, 1982) in the light of new paleontological and sedimentological evidence and new approaches to physiology. It is now generally accepted that the earliest vertebrates originated in the marine realm and invaded freshwater later (Thomson, 1971; Halstead, 1973; Blieck, 1985).

Of course, direct evidence from the Ordovician vertebrates attests only to the environment that they were living in at that time and not to the environment that their ancestors inhabited. It is interesting to note, however, that the Ordovician localities represent very similar environments. The western North American Harding Sandstone, though originally considered to be a freshwater deposit, is now thought to represent sediment laid down in shallow, nearshore conditions (Denison, 1956; Fischer, 1978; Darby, 1982). Spjeldnaes (1967, 1979) concluded that restricted marine conditions may have been present, with salinity as the restricting factor, and that changes occurred from brackish to normal marine in a lagoon-like or estuarine-deltaic habitat. Certainly the presence of *Cruziana* and other trace fossils indicates a shallow, nearshore environment and the presence of conodonts and other marine organisms points toward a marine habitat.

Boucot and Janis (1983) concluded that the North American Ordovician vertebrate localities correspond to their marine benthic assemblages BA1 (marine intertidal), BA2 or BA3 (shallow subtidal).

The environment in which the Australian vertebrates *Arandaspis* and *Porophoraspis* lived seems to have been very similar to that postulated for the Harding Sandstone. The Stairway Sandstone was deposited as part of a transgressive event, not far from shore (Ritchie and Gilbert-Tomlinson, 1977), and there may have been specialized local conditions as shown by an unusual invertebrate fauna in some areas. The relationship with *Cruziana* once again points to a nearshore, shallow marine habitat (benthic assemblage BA2 of Boucot and Janis, 1983). A similar habitat is inferred for the Bolivian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TERRITORY
WESTERN
AUSTRALIA
Lake Amadeus
17
QUEENSLAND
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Figure 4 (opposite). Ordovician vertebrate localities of North America (**A**), Bolivia (**B**), and Australia (**C**). Abbreviations: BC, British Columbia; ON, Ontario; QU, Quebec; NS, Nova Scotia; MO, Montana; WY, Wyoming; CO, Colorado; OK, Oklahoma; SD, South Dakota; MN, Minnesota; MI, Michigan; AR, Argentina; SA, South Australia.

Localities: **1.** a) Harding Sandstone Quarry, near Cañon City, Harding Sandstone Formation type locality (Walcott, 1892; Eastman, 1917; Bryant, 1936; Ørvig, 1965; Spjeldnaes, 1967, 1979; Yochelson, 1983; Lehtola, 1983; Elliott, 1987); b) Sturbaum Quarries, north of a, west of Cañon City, Harding Sandstone Formation (Denison, 1967); c) other localities in the Harding Sandstone Formation include St. John's Quarry, the Cripple Creek Shelf Road location, both in the vicinity of Cañon City, the west side of Sawatch Range, a locality near Taylor Park, the Indian Springs Ranch in Phantom Cañon, a locality in Helena Cañon (Bryant, 1936; Sawin, 1959; Fischer, 1978; Spjeldnaes, 1979). *Astraspis desiderata, Astraspis* sp., *Eriptychius americanus*, vertebrate indet. A. **2.** a) South fork of Rock Creek section, Bighorn Mountains, Harding Sandstone equivalent (Eastman, 1917; Ørvig, 1958; Denison, 1967): *Astraspis splendens, Astraspis*? sp., *Eriptychius orvigi*, undet. fragments; b) other localities in the same formation include the north fork of Crazy Woman Creek, the south fork of Piney Creek, Little Tongue River, Tongue River, Bear Rocks, all on the eastern slope of Bighorn Mountains (Denison, 1967). **3.** Dease River region, "Middle Ordovician beds" (Sinclair, 1958): "*Astraspis* sp." **4.** Sheep Mountain, in the Bear Lodge Mountains, top of Icebox Shale (Denison, 1967): "*Astraspis*", "*Eriptychius*.' **5.** Whitewood Creek section, Black Hills, Icebox Shale (Eastman, 1917; Ørvig, 1958; Denison, 1967): *Eriptychius* sp., "*Pycnaspis* sp.", *Astraspis* sp., undet. fragments. **6.** Shell Pine Unit No. 1 core, Williston Basin, Winnipeg Formation (Ørvig, 1958): *Eriptychius* sp., *Astraspis* sp., "*Pycnaspis* sp.", *Eriptychiida* indet. **7.** Interstate Highway 35, Arbuckle Mountains, north of Ardmore, near the Viola-Bromide formations boundary (Ossian and Halseth, 1976): *Eriptychius* sp. **8.** ?, Glenwood Formation (Stauffer, in Darby, 1982): "*Eriptychius*." **9.** Near Escanaba, Black River Group (Hussey, **in** Lehtola, 1973): supposed vertebrate fragments. **10.** Near Shingleton, Black River Group equivalent (Darby, 1982): "*Astraspis*." **11.** a) East coast of St. Joseph Island, Gull River Formation (Lehtola, 1973): *Astraspis desiderata*; b) north coast of St. Joseph Island, Swift Current Formation (Darby, 1982): *Astraspis, Eriptychius*? **12.** Ouareau River section, upper Pamelia Formation (Eliuk, 1973): *Astraspis desiderata*. **13.** Pont-Rouge section, Lowville and Leary formations (Eliuk, 1973): *Astraspis desiderata*. **14.** Near La Malbaie, basal Trenton beds (Sinclair, 1958), "*Astraspis* sp." **15.** Cape George, quartzite blocks in a fault zone, cf. School Brook Cove beds (Ørvig, **in** Boucot et al., 1959): undet. fish fragments. **16.** a) Challaque River, Sacabamba, upper Anzaldo Formation (Gagnier et al., 1986; Gagnier, 1987): *Sacabambaspis janvieri;* b) Sacabambilla, north of a. **17.** a) Mt. Watt, southeastern Amadeus Basin, Stairway Sandstone Formation: *Arandaspis prionotolepis, Porophoraspis crenulata*, undescribed taxon; b) Mt. Charlotte, north of a, Stairway Sandstone: *A. prionotolepis*; c) localities in the same area, Carmichael Formation: vertebrate fragments (Ritchie and Gilbert-Tomlinson, 1977; Ritchie, 1979).

vertebrate *Sacabambaspis* (Gagnier et al., 1986). The Anzaldo Formation contains a variety of shallow marine invertebrates and the vertebrates are again associated with the trace fossil *Cruziana*.

Ordovician vertebrates normally occur as fragments, however, and because so little work has been carried out on fish taphonomy (Elder and Smith, 1984) little can be inferred from them. It has been suggested that *Astraspis* and *Eriptychius* fragments might have been carried into the Harding Sandstone depositional site by streams, but the great lateral extent of the formation with its ubiquitous marine invertebrates makes this unlikely. The carcasses were probably disarticulated by scavengers and then distributed across the current-swept nearshore environment (Boucot and Janis, 1983). Darby (1982) has shown that in the Gull River Formation of Ontario, Canada, a limestone that is age equivalent to the Harding Sandstone, a correlation can be seen between the presence of sand in the limestone and the presence of vertebrate fragments. A study on the equivalent to underlying Pamelia Formation (Eliuk, 1973) showed a similar correlation and it might be inferred from this that both sand and vertebrate material were transported from elsewhere. Nevertheless, the distribution of well preserved vertebrate material was used (Eliuk, 1973) to propose a carbonate tidal flat habitat for the vertebrates. Darby (1982) has used this data to suggest that the vertebrates preferentially inhabited sandy sediments and filter fed from the interstitial waters. This mode of life is similar to that of the modern cephalochordate amphioxus (*Branchiostoma*), which is restricted to nearshore marine environments where the substrate is a coarse sand or shell gravel. They are unable to live in sandy sediment if the silt content is over 1.5% (Webb and Hill, 1958). Although the North American Harding Sandstone and Australian vertebrate-bearing beds are sandstones and the Quebec and Ontario localities are limestones with some sand, the Anzaldo Formation is reported as being a lutite and it is interesting to note that this is the only horizon at which complete animals have been found. It has been suggested (Gagnier, 1989) that the vertebrates may have been killed by an influx of fresh water. Possibly an influx of finer sediment may also have contributed to the mass mortality of *Sacabambaspis*.

There is little direct evidence of the mode of life of these animals. Though the trace fossil *Agnathichnus* has been named from the Harding Sandstone (Fischer, 1978) there is no evidence to support its designation as a vertebrate feeding trace. Mouthparts are known for *Sacabambaspis* only and appear to consist of a ventral series of flexible elongated structures (Fig. 1B) attached to each other by connective tissue (Gagnier, 1989). A structure of this type could have formed a suctorial mouth adapted for removing small particles from the sediment. It has been proposed (Mallatt, 1984) that, based on protochordates and extant fish, the earliest Paleozoic fish were microphagous suspension feeding animals. If that were the case then, like modern ammocoete larvae, they would have required highly concentrated suspensions of a type present in benthonic aquatic environments but not in open-water habitats. It therefore seems probable that these early vertebrates were chiefly benthonic deposit feeders, though it is unlikely that they actually lived in the sediment.

AGE AND DISTRIBUTION OF ORDOVICIAN VERTEBRATES

The stratigraphic and geographic record of Ordovician vertebrates is rather restricted. They are known from either near the Lower/Middle Ordovician boundary in Australia, or from the middle Middle Ordovician of North America and Bolivia.

Australia

Vertebrates have been collected from near the southeastern extremity of the Amadeus Basin, Northern Territory, mainly in two localities, Mt. Watt and Mt. Charlotte (Fig. 4, locality 17). *Arandaspis prionotolepis* has been recorded from both localities, *Porophoraspis crenulata* from Mt. Watt only. Both species have been collected in the lower Stairway Sandstone Formation (Ritchie and Gilbert-Tomlinson, 1977). A third taxon at least is present in Mt. Watt material (Ritchie, 1985) but has not yet been described. The age of the vertebrate horizons is stated to be earliest Middle Ordovician (i.e., probably early Llanvirn) by Ritchie and Gilbert-Tomlinson (1977), which makes these the earliest vertebrate records yet available. A few fragmentary skeletal plates have been reported from the younger, upper Llanvirn, Carmichael Formation of the same area (Ritchie and Gilbert-Tomlinson, 1977).

Bolivia

Sacabambaspis *janvieri* was originally described from fragmentary plates from the upper Anzaldo Formation at Sacabamba on the Challaque River, department of Cochabamba, in central Bolivia (Gagnier et al., 1986; Gagnier, 1987) (Fig. 4, locality 16). The horizon was dated as earliest Caradoc (Gagnier et al., 1986). Subsequently, a second vertebrate locality was discovered at Sacabambilla, province of Punata and north of Sacabamba. This is also in the Anzaldo Formation (Gagnier, 1989; Aceñolaza and Baldis, 1987).

North American

Ordovician vertebrates are classically known from the Cordilleran region; however, the numerous reports of their occurrence also include localities in the Great Plains, the Arbuckle Mountains of Oklahoma, the Great Lakes region of the U.S.A. and Canada, the St. Lawrence Lowlands of Quebec, as well as Nova Scotia and as far north as northern British Columbia. Nevertheless, only three localities have been studied in detail:

1. The original type locality of the Harding Sandstone Formation; the Harding Sandstone Quarry (= Old Harding Quarry), near Cañon City, southwest of Colorado Springs, Colorado (Fig. 4, locality 1a) (Walcott, 1892; Vaillant, 1902; Eastman, 1917; Stetson, 1931; Bryant, 1936; Ørvig *in* Stensiö, 1958, 1964; Foss, 1960; Ørvig, 1965; Spjeldnaes, 1967, 1979; Yochelson, 1983; Lehtola, 1983; Elliott, 1987). The vertebrate faunal assemblage includes *Astraspis desiderata* Walcott, *Eriptychius americanus* Walcott, and perhaps other as yet undescribed taxa (plus *Dictyorhabdus priscus* Walcott, whose vertebrate affinities have not yet been demonstrated). From its associated conodonts the Harding Sandstone is now correlated with the middle Trenton Limestone, Kirkfieldian Stage of the Mohawkian Series, Zone of *Orthograptus truncatus intermedius*. This correlates with the *Climacograptus wilsoni* Zone in the British succession (Lehtola, 1973; Finney *in* Ross et al., 1982).
2. The Sturbaum quarries, north of the Harding Quarry, west of Cañon City, Colorado (Fig. 4, locality 1b) (Denison, 1967). The vertebrate assemblage comprises *A. desiderata*, *E. americanus* and vertebrate indet. A Denison. It comes from the Harding Sandstone (correlated with the middle Trenton Limestone [see above], not the Black River Group as indicated by Denison, 1967, p. 132).
3. The south fork of the Rock Creek section, Johnson County, east slope of the Bighorn Mountains, northwest of Buffalo, near Lake DeSmet, north-central Wyoming (Eastman, 1917; Ørvig, 1958; Denison, 1967) (Fig. 4, locality 2a). The fauna consists of *Astraspis splendens* (Ørvig), *Astraspis*? sp., *Eriptychius orvigi* Denison and undetermined fragments. The sedimentary sequence they come from is considered to be a Harding Sandstone equivalent (below the Bighorn Dolomite, not above it as indicated by Ørvig, 1958) and distinct from the Lander Sandstone at the base of the Bighorn Dolomite (discussion *in* Denison, 1967; see Ross et al., 1982, log 28).

All the other records (Fig. 4) have not been adequately described or are based on fragmentary material whose determination has not been reviewed. The assemblages in Colorado and northern Wyoming are specifically distinct, with *A. desiderata* and *E. americanus* to the south and *A. splendens* and *E. orvigi* to the north. Most of the other localities are either considered to be Harding Sandstone equivalents (middle Trenton age, middle Caradoc) or older (Black River age, lower Caradoc, for the Great Lakes and St. Lawrence lowlands localities) (Barnes et al., 1981; Ross et al., 1982) (Fig. 4, localities 9-14).

DISCUSSION

From the foregoing record it seems that Ordovician vertebrates in the Americas disappear before the end of the Caradoc and, therefore, before most of the invertebrate groups whose diversity decreases at the beginning of the Ashgill (crinoids, brachiopods, graptolites, conodonts, corals, trilobites, bryozoans, acritarchs). This event can now be correlated with eustatic changes, in relation to the cooling of the climate during the Ashgill as a whole. Its acme was reached with the Tamadjert glacial epoch in the upper Ashgill (Hirnantian; see reviews in Hambrey, 1985; Barnes, 1986; Sheehan *in* Scotese and McKerrow, eds., 1988). We propose, therefore, that there is evidence for the Ordovician vertebrates suffering the effects of the Ashgill cooling. However, as concluded by Spjeldnaes (1979, Fig. 10), North American Ordovician vertebrates were restricted to regions close to the Ordovician paleoequator; this also seems to be the case with the Australian record (Figs. 4, 5). However, this does not seem to be applicable to the Nova Scotia locality (though this is based on doubtful vertebrate remains; Ørvig *in* Boucot et al., 1959) nor to the Bolivian one (Figs. 4, 5, localities 15, 16). South America and Nova Scotia were indeed in southern high latitudes in Middle and Late Ordovician times. Nova Scotia drifted from the North Gondwanan margin toward the southwest corner of Baltica (e.g., Scotese, 1986). Nova Scotia and Newfoundland suffered the Gander Bay glacial event during

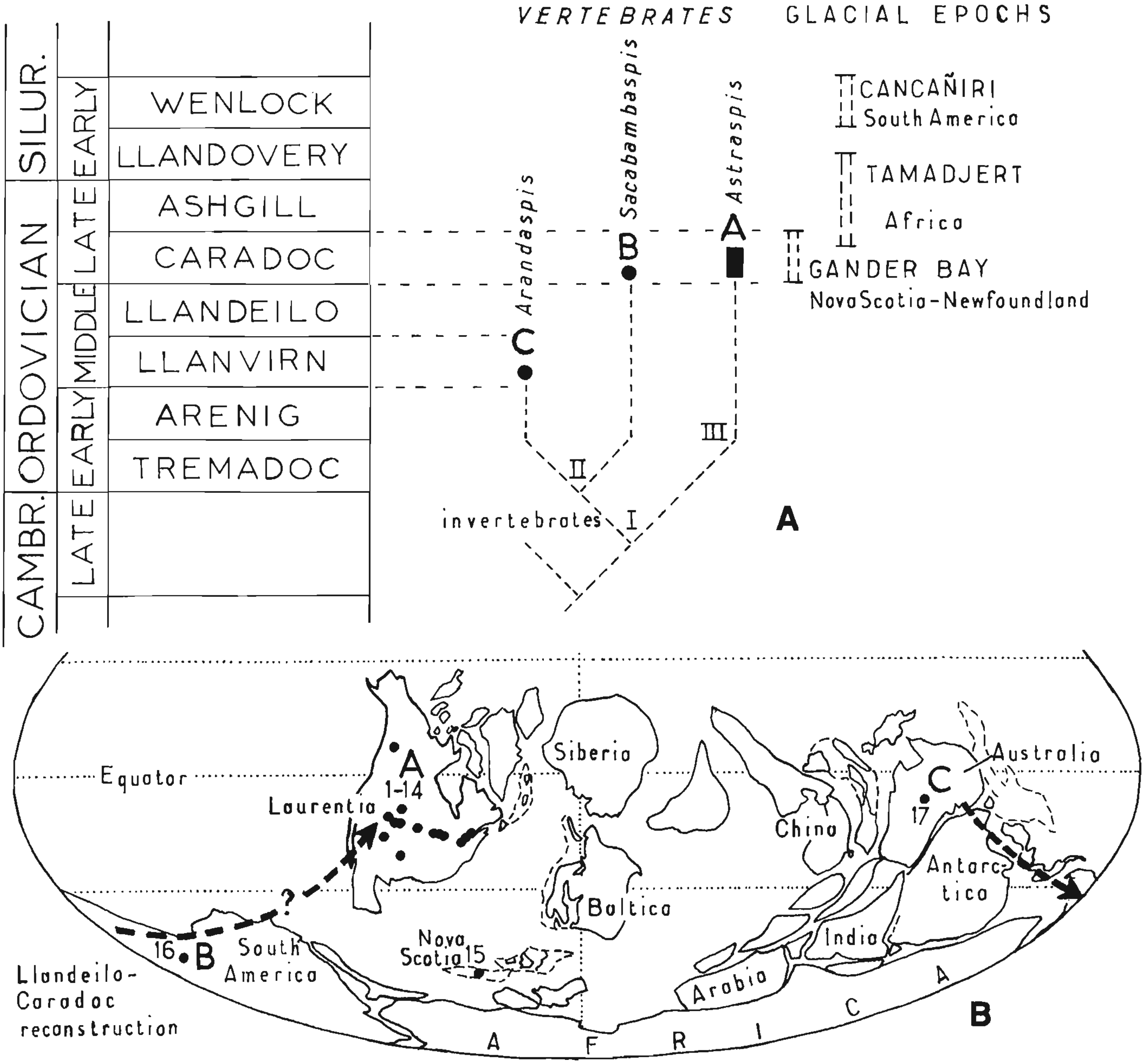

Figure 5. Interpretation of Ordovician vertebrate distribution. **A.** Hypothetical phylogenetic relationships of Ordovician vertebrates with the deduced C-B-A area cladogram, in relation to the presently known Late Ordovician-Early Silurian glacial epochs (Hambrey, 1985, Table I). I = craniate conditions - phosphatic, three-layered dermal bone with a spongy median layer of aspidin - sensory lines in open grooves; II = slanting row of numerous branchial platelets - shields formed of fused small tesserae - elongated, thin, serrated flank scales; III = eight branchial openings in a horizontal row on both sides of the cephalic carapace - shields formed of wider, partially fused tesserae - thick, polygonal or slightly elongated flank scales. **B.** Ordovician vertebrate localities 1-17 (same as in Fig. 4) on the Llandeilo-Caradoc paleocontinental reconstruction of Scotese (1986), with a supposed migratory route (arrow).

the Late Ordovician (Caradoc-Ashgill), and another glacial event took place in South America in ?Late Ordovician-Early Silurian times (Zapla-Cancañiri glacial epoch; Hambrey, 1985). This hypothesis does not explain why in Australia, far away from any glacial evidence, vertebrates are restricted to lower Llanvirn localities and it is probable that we still have vertebrate-bearing sites to discover.

Finally, from the morphological data available at present, it is clear that the Ordovician vertebrate genera are much more closely related to each other than to any other vertebrate group; it is also probable that the Bolivian genus *Sacabambaspis* is more closely related to the Australian genus *Arandaspis* than to the North American genera *Astraspis* and *Eriptychius*. It is difficult, however, to explain these phylogenetic relationships in relation to the present paleogeographic reconstructions in which Australia, Bolivia, and North America are widely separated (Fig. 5). It is possible to look at this problem from a migratory point of view or to use a vicariant scheme. Under a migratory model it is necessary to propose a migration from the older, Llanvirn, Australian block to the younger, Caradoc, Bolivian locality, and then (?) to the North American block (Fig. 5B). However, Ordovician vertebrates appear to have been restricted to subtidal, marine environments and may not have been capable of transoceanic migrations. It is thus necessary for this hypothesis to bring Australia, Bolivia, and North America closer together, or to find intermediate fossil sites.

Under a vicariant scheme it would be necessary to have a pre-Llanvirn pattern in which the ancestors of Ordovician vertebrates had a generalized trackway on a joined Australian-Bolivian-North American supercontinent (Fig. 5). Unfortunately, neither of these hypotheses can be fitted to the presently available paleogeographic reconstructions for pre-Ordovician and Ordovician times (e.g., Scotese, 1986; Scotese and McKerrow, 1988).

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Upper Cambrian and Lower Ordovician conodont associations from open ocean paleoenvironments, illustrated by Batyrbay and Sarykum sections in Kazakhstan

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Abstract

Upper Cambrian and Lower Ordovician conodont associations are reported from the carbonate facies of the Batyrbay section, Malyi Karatau, south Kazakhstan and siliceous-carbonate facies of the complete Sarykum section, central Kazakhstan. Different paleoenvironments at Batyrbay and Sarykum affected the composition of conodont associations. They are compared with associations from other regions of the world for specific time intervals (as chronological equivalents of conodont zones) using the Cambrian-Ordovician zonal scale at Batyrbay as the example.

These data distinguish (from the total warm and cold faunal realms of J.F. Miller) a so-called transitional (T) faunal realm. The extent and faunal characteristics of the T realm are considered. The significance of contemporaneously redeposited warm water euconodont elements is noted to compare different warm and T faunal realms.

The evolutionary zonal subdivisions at Batyrbay are recognized by: the earliest evolutionary stage of the Hirsutodontus lineage of the Upper Cambrian; the lineages from Eoconodontus (E.) notchpeakensis *to* E.(E.) alisoneae *and from* E.(E.) notchpeakensis *to* Cordylodus primitivus *in the uppermost Cambrian; and from* Prioniodus (P.) deltatus longibasis *to* P.(P.) elegans *in the lower Arenig.*

The beginning of J.F. Miller's Lange Ranch Eustatic Event at the Cambrian-Ordovician boundary has been given a more precise definition, i.e., from the second phase of the Cordylodus primitivus *time of the latest Cambrian.*

Résumé

Les associations de conodontes du Cambrien supérieur et de l'Ordovicien inférieur ont été localisées dans le faciès carbonaté au profil Batyrbay (Malyi Karatau, Kazakhstan du Sud) et dans le faciès siliceux-carbonaté de tout le profil Sarykum (Kazakhstan central). Différents paléo-environnements à Batyrbay et Sarykum ont affecté la composition des associations de conodontes. Elles sont comparées à des associations d'autres régions dans le monde pour des intervalles de temps spécifiques (comme équivalents chronologiques des zones à conodontes) en appliquant, à titre d'exemple, l'échelle zonale cambrienne-ordovicienne à Batyrbay.

Ces données font ressortir (basé sur les domaines fauniques chauds et froids totaux de J.F. Miller) un domaine faunique de transition (T). L'étendue et les caractéristiques fauniques du domaine T sont analysées. L'importance des éléments d'euconodontes d'eau chaude redéposés contemporainement est retenue pour la comparaison des différents domaines fauniques chauds et T.

Les sous-divisions zonales évolutives à Batyrbay se distinguent par : le stade évolutif précoce de la lignée de Hirsutodontus du Cambrien supérieur; les lignées de Eoconodontus (E.) notchpeakensis *à* E.(E.) alisonae *et de* E.(E.) notchpeakensis *à* Cordylodus primitivus *dans le sommet du Cambrien; et de* Prioniodus (P.) deltatus longibasis *à* P.(P.) elegans *dans l'Arénigien inférieur.*

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Le début de l'événement eustatique du chaînon Lange de J.F. Miller à la limite du système cambrien-ordovicien a été défini avec plus de précision, c'est-à-dire à partir de la deuxième phase de l'époque de Cordylodus primitivus *du Cambrien terminal.*

INTRODUCTION

The Upper Cambrian and Lower Ordovician conodont associations are known from carbonate and siliceous-carbonate facies in Kazakhstan (Fig. 1). The Batyrbay section of the Malyi Karatau in south Kazakhstan belongs to the carbonate facies (Fig. 2). The section has been described in numerous papers (Apollonov, Chugaeva, and Dubinina, 1981, 1984; Apollonov and Chugaeva, 1982, 1983; Chugaeva and Apollonov, 1982; Dubinina, 1982; Apollonov et al., 1985, 1988; Zhemchuzhnikov, 1986, 1987; Apollonov and Zhemchuzhnikov, 1988; Chugaeva et al., 1989) and embraces a continuous carbonate succession from the Upper Cambrian up to the lower Arenig, inclusive. The accumulation of these sediments took place in an open basin, in environments varying from the Upper Cambrian lower slope and basin to the carbonate shelf in the Arenig (Fig. 2). The recent detailed sedimentological study carried out by V. Zhemchuzhnikov (pers. comm., 1988) and completed by H. Cook (pers. comm., 1988) suggests that the Malyi Karatau in the early Paleozoic represented a carbonate seamount not connected to a continental massif. Thus, the basin had permanent and free access to the open ocean. It was also influenced by a certain system of oceanic currents. Such a paleoenvironment affected the composition of conodont associations at Batyrbay (located, as the adjacent Kyrshabakty section, on the northeastern margin of the seamount).

The Sarykum section in central Kazakhstan represents the second (siliceous-carbonate) type of deposit (Zhemchuzhnikov, 1987). The complete section consists of three parts (Fig. 3). The continuous succession of siliceous-carbonate sediments was deposited in an environment that increased in depth from Late Cambrian to early Arenig time. Thus, the deposits of the upper part of the Sarykum Formation (uppermost Cambrian; Fig. 4, beds 3-5), where the carbonates of submarine fans prevailed over cherts, probably accumulated near the lower slope of the carbonate platform. The lower member of the Chazhagay Formation (from the uppermost Cambrian to the Tremadoc; Fig. 4, beds 7-13; Fig. 3), with a lower carbonate content and prevailing chert, was formed in a hemipelagic environment. The upper member of this formation (lower Arenig; Fig. 4, bed 14; Fig. 3), represented by black chert with no carbonate component, was formed in a pelagic environment.

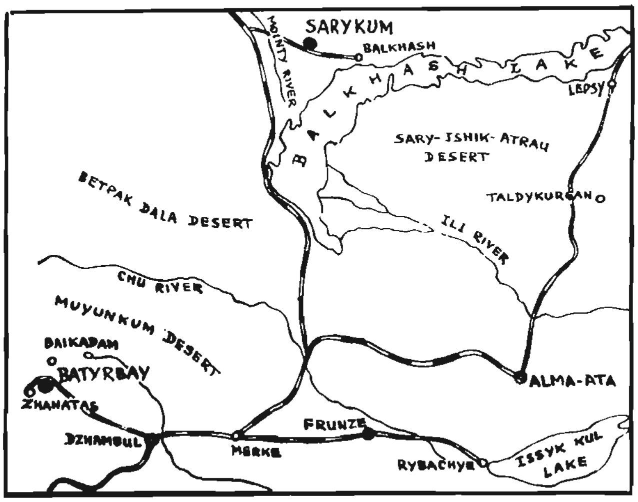

Figure 1. The Batyrbay and Sarykum sections in Kazakhstan.

The Batyrbay and complete Sarykum sections fortunately supplement each other, as the continuous succession of boundary deposits is traced herein in different facies appearing at various bathymetric levels. Thirteen conodont zonal subdivisions were established at Batyrbay, whereas at Sarykum there are six (Fig. 5). The succession of conodont assemblages distinguished in both sections does not differ in principle. The main purpose of this paper is to determine the specific features of conodont associations of Kazakhstan for certain time periods from the Late Cambrian to the Early Ordovician. This task is accomplished by comparing data from different regions of the world. This approach leads to a generalized, but far from perfect, pattern of the distribution of conodont associations from various paleoenvironments. The pattern includes the results of the previous studies by Sweet and Bergström (1974), Lindström (1976), Landing (1983), Ethington and Repetski (1984), and Miller (1984), concerning paleogeographic distribution of conodonts of these time intervals.

Distinct from the warm and cold faunal realms of Miller (1984), a so-called transitional faunal realm is proposed (Figs. 6, 6a). Herein, the cold (C) faunal realm includes various paleoenvironments of ancient seas and oceans in relatively high paleolatitudes (northwestern Europe) as well as the deepest paleoenvironments in middle paleolatitudes (central Kazakhstan). The warm (W) faunal realm includes ancient shallow water seas (excluding the outermost open shelves) in low to middle paleolatitudes, i.e., ancient epicontinental seas of North America, north China, and Australia, with restricted access to the open ocean. This W realm is confined to stable cratons and continental massifs in low to middle paleolatitudes. Finally, the transitional (T) faunal realm includes ancient open seas and oceans in low to middle paleolatitudes. It embraces outermost continental shelf with unrestricted access to the open ocean, continental slope, basinal (hemipelagic), i.e., any carbonate continental or seamount margins and can be illustrated by sections in south Kazakhstan (this paper), western Mackenzie Mountains, Canada (Landing et al., 1980), central Nevada (Taylor and Cook, 1976; Taylor, 1977; Cook et al., 1989), northwestern Vermont (Landing, 1983), western Newfoundland (Barnes, 1988; Pohler et al., 1987), and in the future it might be

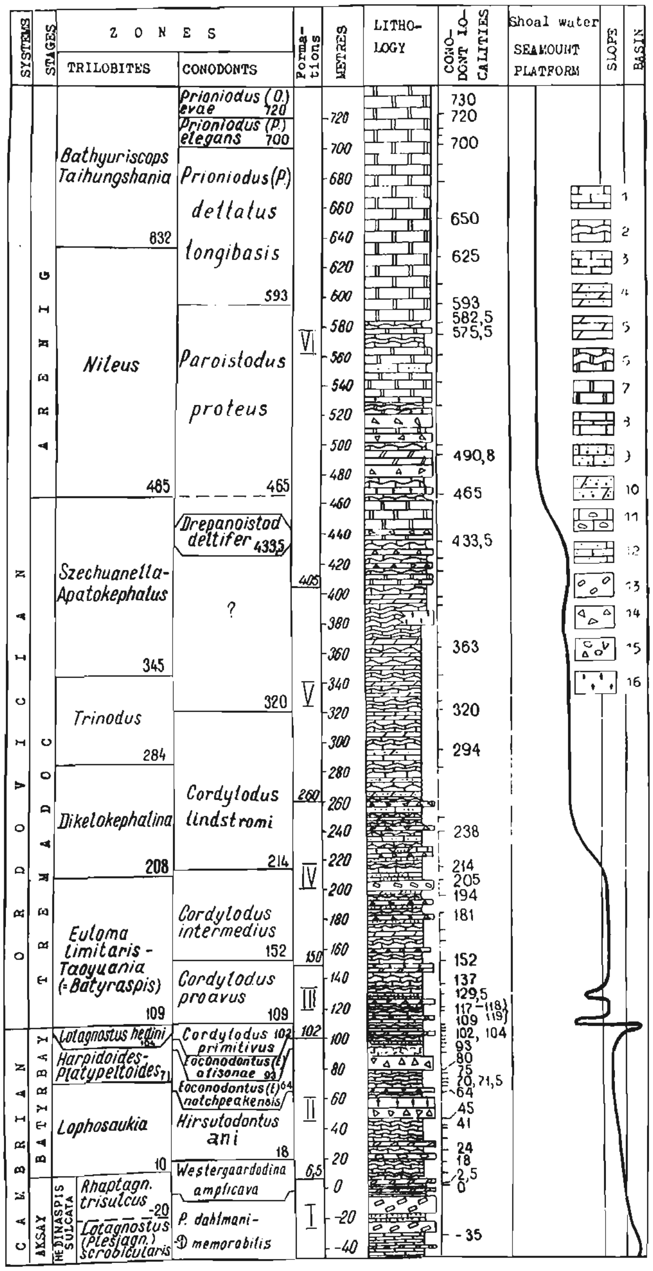

Figure 2. Stratigraphic column, conodont and trilobite zonal subdivisions in the Batyrbay section, South Kazakhstan. 1. thin and medium bedded calcarenite and calcisiltite with graded and laminated bedding; 2. dark, wavy bedded and nodular limestone (fine grain calcarenite, calcisiltite, calcilutite, and pelloidal limestone); 3. thick bedded calcarenite and calcrudite; 4. thick bedded dolarenite and dolorudite; 5. dolomite; 6. thin bedded, light grey, wavy bedded, pelloidal limestone and lime mudstone; 7. thick bedded, light lime mudstone, wackestone, and algal limestone; 8. medium bedded, fine grained calcarenite with micrograding fabric; 9. graded bedded calcarenite and calcirudite, with injection and flame structures; 10. graded bedded dolarenite and dolorudite, with injection and flame structures (bird's eye); 11. fine grained calcarenite with fenestral structures; 12. fine grained calcarenite with ripple marks; 13. flat-pebble breccia (gravity flow deposits with graded bedding); 14. breccia with rounded or irregularly shaped clasts; 15. coarse breccia with boulders of algal limestone; 16. algal bioherm.

observed in southeastern China. Thus, the T faunal realm is established mainly from the cold, and only partially from the warm, faunal realms of Miller (1984).

COMPARISON OF COEVAL CONODONT ASSOCIATIONS FROM VARIOUS ENVIRONMENTS

Pie diagrams illustrating the relationship of the main components of conodont associations from different paleoenvironments in certain time intervals are shown in Figures 7-10. The term "time" is used herein as a chronological equivalent of zones. The conodont zonal scale presented here is elaborated for the transitional faunal realm using the Batyrbay section. The lower part of this section contains various types of carbonate accumulated at the seamount base-of-slope, and in lower and upper slope environments (Fig. 2). The conodont samples in this part of Batyrbay were taken only from thinly bedded calcisiltite turbidites and from in situ calcisiltites and calcilutites. It is interesting that the conodont associations of thin-bedded turbidites and in situ calcisiltites are similar. Proto- and paraconodonts predominate in both of them, but the thin turbidite associations usually differ from those in situ by the relatively small euconodont component. The latter is considered to have been redeposited synchronously by turbidity currents from the adjacent shallow water environment. Such a pattern is noted in Batyrbay beginning with the appearance of the first primitive euconodonts at the lowest Batyrbay Stage of the Upper Cambrian. The similarity in the ratio of the main conodont components in thin turbidite and in situ associations suggests a low density of turbidity currents at Batyrbay. In fact, thin turbidites contain a small percentage of euconodonts redeposited from shallower habitats, and abundant indigenous oceanic (in situ) protoconodonts and paraconodonts. After *notchpeakensis* time (Fig. 7), relatively abundant cosmopolitan and indigenous oceanic euconodonts were transported by turbidity currents from the underlying unconsolidated sediment. In the pie diagram, the lower part of the Batyrbay section is represented only by thin turbidite conodont associations, the nature of which depends on the relative abundance of the rare redeposited euconodont elements.

Other examples of conodont faunas related to the transitional faunal realm are known from western Newfoundland, northwestern Vermont, central Nevada, and western Mackenzie Mountains. In western Newfoundland, the succession of Cambrian-Ordovician boundary conodonts has been recovered from the Cow Head Group deposited at the base of the continental slope along the northern margin of the ancient Iapetus Ocean (Barnes, 1988). The Cow Head Group has been illustrated by six key boundary sections representing proximal to distal facies accumulated at the toe of the slope. Each of these sections has yielded abundant collections of conodonts, which were subjected to numerical analysis (i.e., quantitative analysis of specimens of each species in every conodont sample). Conodont associations in distal sections (e.g., Martin Point, Green Point) connected with the pelagic, thinly bedded lime mudstone, as well as the associations in intermediate (more proximal) sections (e.g., Broom Point)

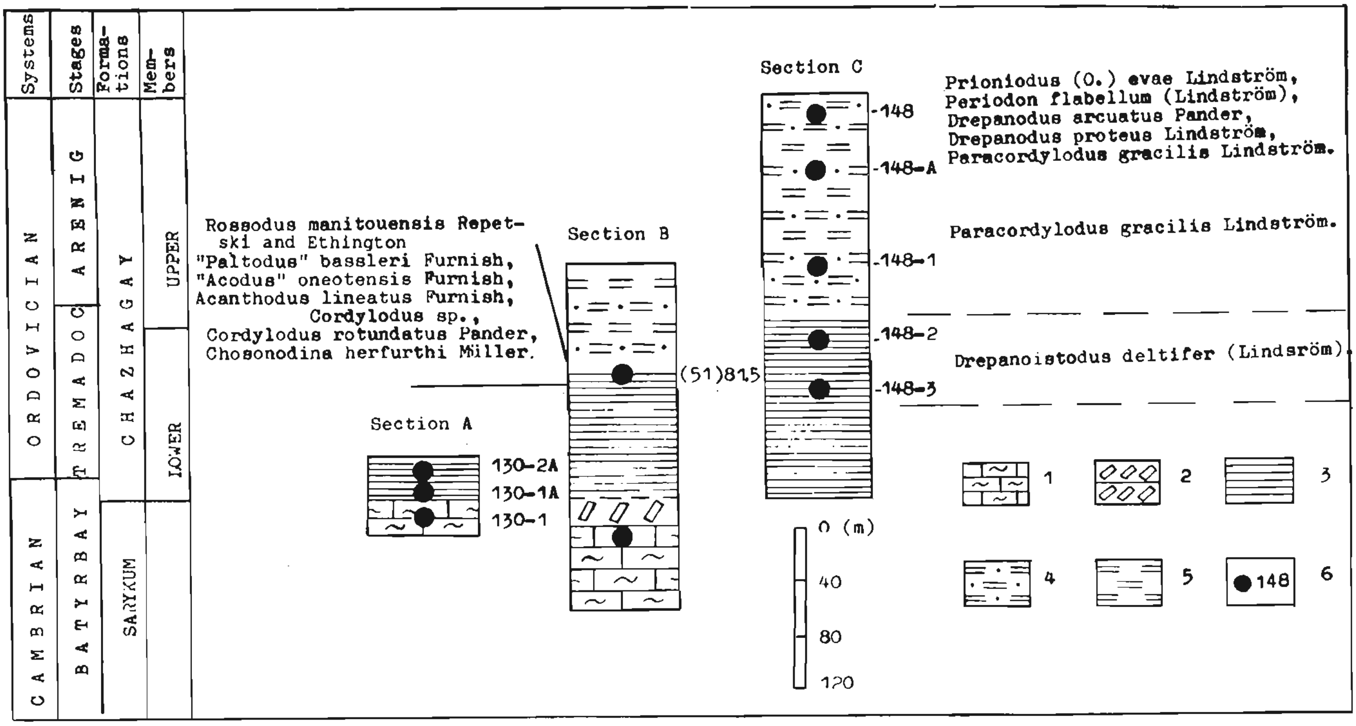

Figure 3. Three sections (A, B, C) representing the complete Sarykum section, central Kazakhstan. 1. fine grained, muddy and marly limestone; 2. breccia with limestone clasts; 3. chert interbedded with limy, cherty shale and limestone (calcarenite); 4. alevrolite chert, spongolite, radiolarite; 5. chert, cherty shale, and black carbon-rich chert; 6. conodont localities.

derived from the pelagic sediments and thin turbidites, were taken into account while comparing them with thin turbidite and in situ conodont associations at Batyrbay. In the pie diagrams, the ratios of the main components in conodont associations recovered from these facies are approximate, giving only a general relationship of the biofacies components. The conodont components of associations of western Newfoundland are separated from each other by dotted lines. The elements redeposited from the epicontinental shallow water habitats are marked by a special symbol (Figs. 7-10).

The illustrations (Figs. 7-10) show that the thinnest slope facies in western Newfoundland and in Batyrbay usually contain abundant in situ (indigenous) oceanic specimens and relatively rare elements redeposited from a shallow water environment. The oceanic (indigenous) specimens could have been incorporated by low density turbidity currents from underlying nonlithified carbonate sediment and transported downslope. Thus, the redeposition not only of platform elements, but also of oceanic (in situ) ones, is a natural process for an open ocean (slope) environment. Thin turbidites belong to the numerous slope facies that could have contained faunas derived from different habitats (i.e., mixed faunas). The oceanic elements (in thin turbidites) are abundant, but platform specimens are relatively rare. The ratio of elements (or main components) in an open ocean conodont association can help to understand the origin of elements (i.e., in situ or allochthonous origin) on one hand, and to define more exactly the peculiarities of the lithofacies containing these associations on the other. However, rare platform elements can be found in open ocean in situ associations. These elements (W' in Fig. 6), having a restricted access to the open ocean, could have been carried oceanward (by surface, wind driven currents or due to the influence of upwelling) and later fallen to the bottom. In such cases, the in situ and thin turbidite associations are similar to each other.

The probability of occurrence of platform elements in the oceanic in situ sediments seems appropriate for both the continental slope (western Newfoundland) and the seamount slope (lower part of Batyrbay) environments. Upper Cambrian and Lower Ordovician deposits in the Hot Creek Range of central Nevada are thought to represent a submarine fan toward the uppermost slope complex formed at the margin of the proto-Pacific Ocean (Taylor and Cook, 1976; Cook et al., 1989). The detailed analysis of conodont associations from various lithofacies in the Hot Creek Range is now in progress by J.F. Miller (*in* Cook et al., 1989). Redeposition of platform specimens together with in situ ones (as discrete elements or in small clasts), leading to a mixing of faunas of different habitats (and ages), is noted by Miller for continental slope facies, in particular, in central Nevada. These data of Miller, and previous studies of conodonts by Landing (1983) in continental slope facies of northwestern Vermont and in low angle slope or outermost shelf facies of western Mackenzie Mountains (Landing et al., 1980), are considered to be in agreement with the above-mentioned data from western Newfoundland and Kazakhstan. Currently, this is the only information available about the T faunal realm; not complete but sufficient to provide a general view.

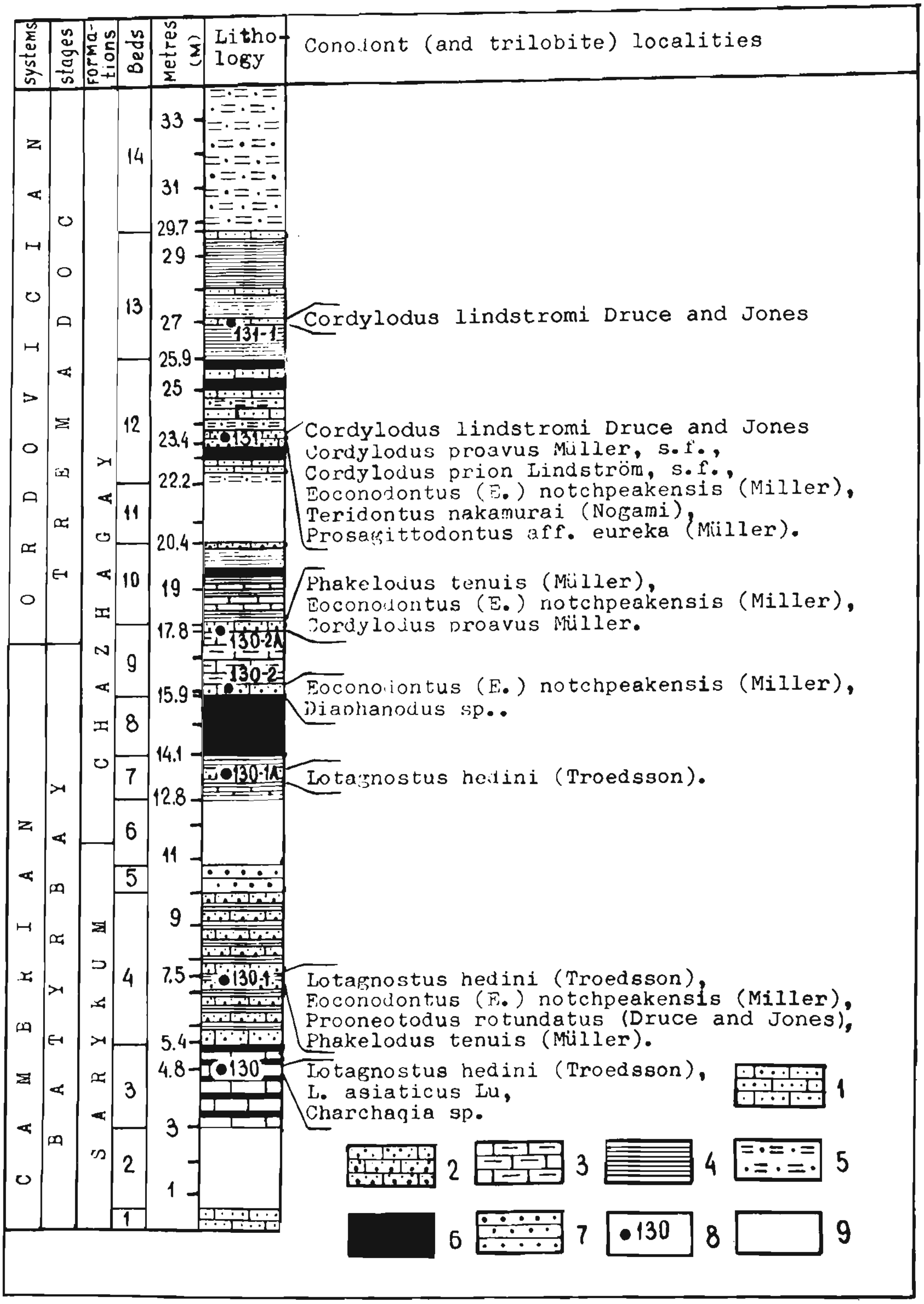

Figure 4. Section A of the complete Sarykum section. Continuous Cambrian-Ordovician boundary succession of siliceous-carbonate rocks. The conodont localities are derived from the calcarenite beds only. 1. grey and light grey fine grained calcarenite; 2. medium grained calcarenite and cherty lime sandstone with graded bedding; 3. muddy calcisiltite and calcilutite; 4. limy, cherty shale; 5. alevrolite chert, spongolite, radiolarite; 6. black carbon-rich chert; 7. limy quartz-feldspar sandstone; 8. conodont (and trilobite) localities; 9. unexposed intervals of the section.

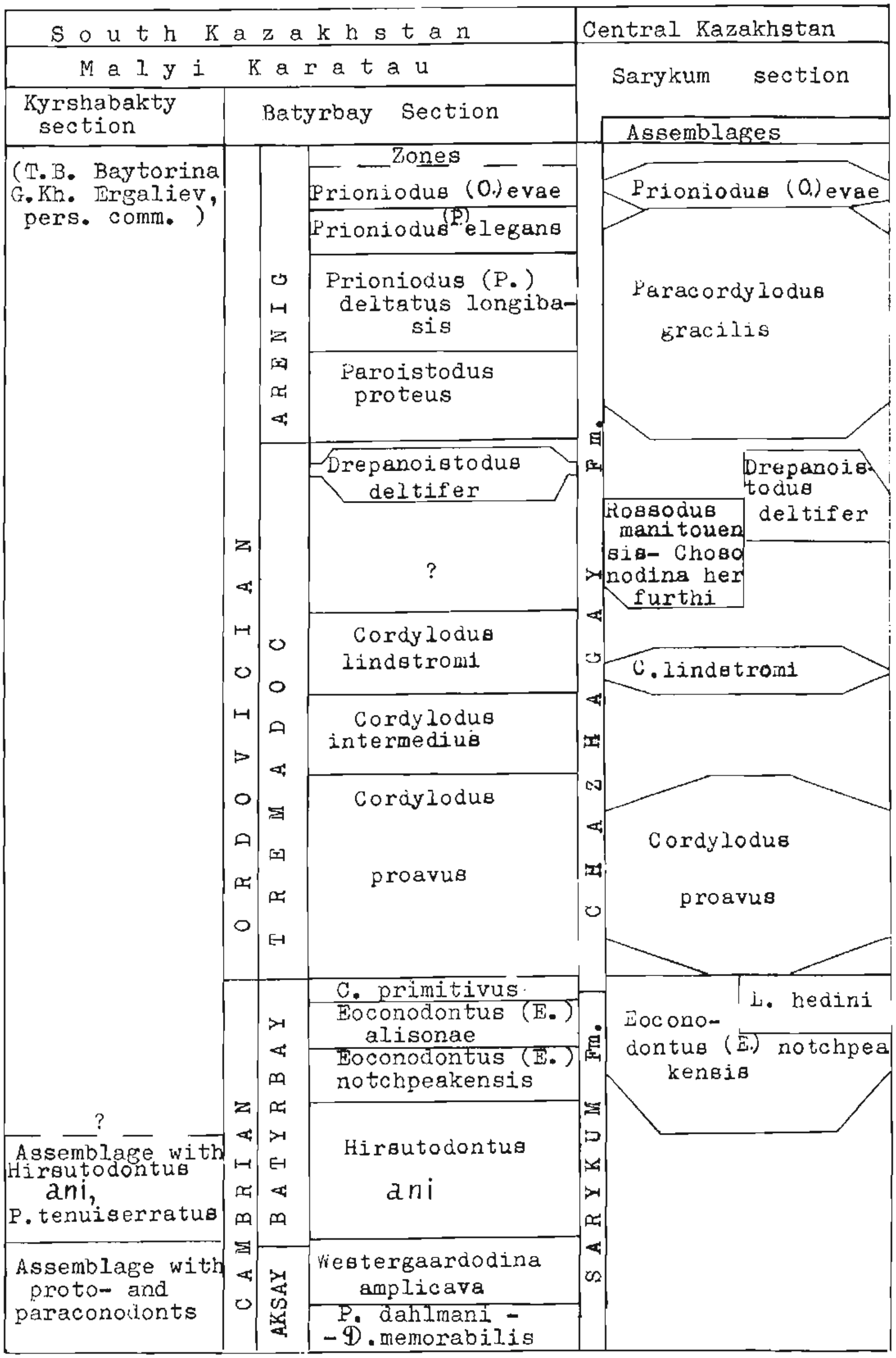

Figure 5. Correlation of Upper Cambrian and Lower Ordovician strata in south and central Kazakhstan (Batyrbay, Kyrshabakty, Sarykum sections) based on the conodont zone succession.

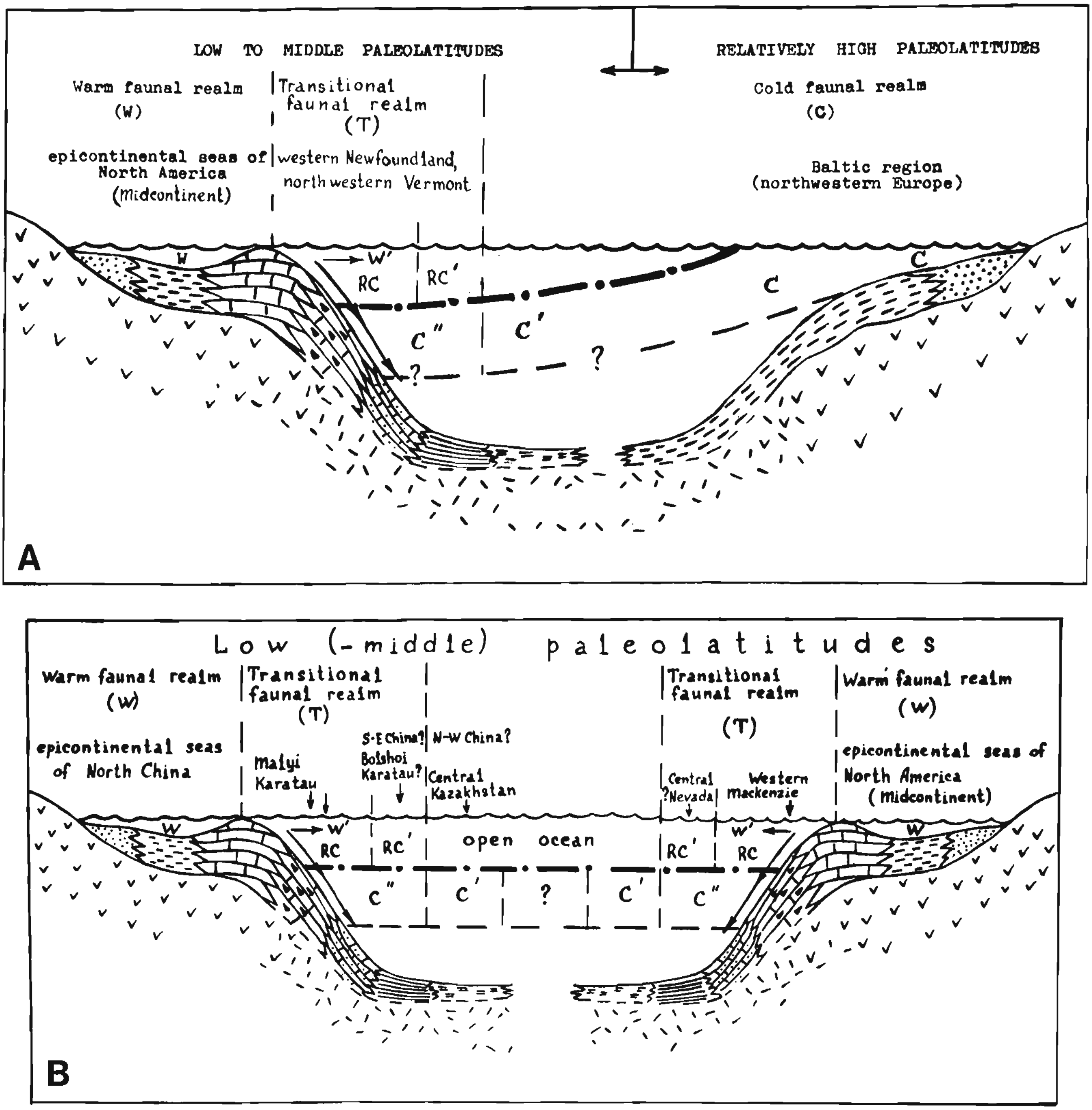

Figure 6. Inferred paleoenvironmental relationship of the warm, transitional, and cold conodont faunal realms in the Late Cambrian and Early Ordovician. Not to scale. General models by Taylor (1977, Textfig. 14) and Taylor and Cook (1976, Fig. 49) were used to determine extension of conodont realm. W, realm inhabited by warm water conodont species (populations, associations), i.e., epicontinental seas in low to middle paleolatitudes; W', warm water species starved in an open ocean environment; RC, relatively cold, (relatively deep water species [associations] living above the permanent thermocline); RC', RC species (associations) starved in an open ocean beyond a certain distance from shoreline; C' and C", variants of cold, deep water conodont associations living below the permanent thermocline; T, realm inhabited by W', RC, RC', and C" conodont associations (populations). Above-mentioned designations related to low and middle paleolatitudes. C, cold water associations in relatively high paleolatitude environments of shallow seas and ocean (—.—.—. indicates level of the permanent thermocline).

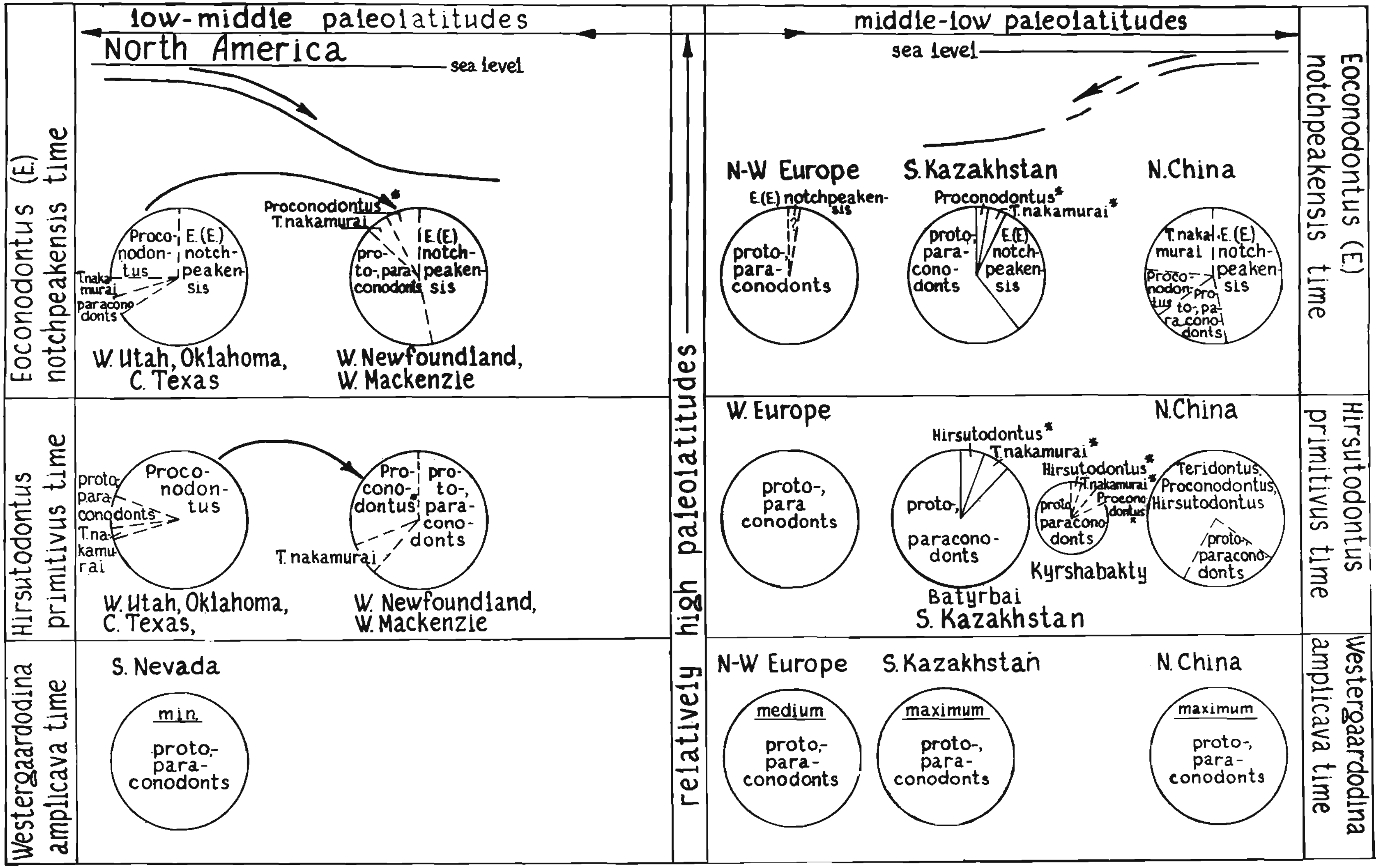

Figure 7. Comparison of coeval conodont associations in various paleoenvironments since *amplicava* to *notchpeakensis* time of the Late Cambrian. References: southern Nevada (Miller et al., 1981); western Utah, Oklahoma, central Texas (Miller et al., 1982); western Newfoundland (Barnes, 1988); western Mackenzie Mountains (Landing et al., 1980); north China (An, 1982; An et al., 1983); northwestern Europe (Kaljo et al., 1986). (* indicates euconodont elements transported from shallow water (cratonic) environment).

The contrast between T (transitional) and W (warm) faunal realms is obvious. The T realm has the allochthonous faunal component redeposited from the W realm. The re-establishing of the main faunal components in conodont associations of the W faunal realm (and the relationship of their components) in specific Cambrian-Ordovician time intervals is not difficult now with the data from North America (Miller et al., 1982; Ethington and Repetski, 1984; Miller, 1984), north and northeastern China (An, 1982; An et al., 1983; Chen and Gong, 1986), and Australia (Druce et al., 1982). The reverse relationship of coeval faunal components is typical for the W and T realm associations and can be explained by the differences of T and W conodont ecology. The absence of several T taxa is true for the W realm as well. The data concerning the W faunal realm associations are conditionally reflected in the pie diagrams, where faunal components are separated from each other by dotted lines (Figs. 7-10). The general information about C (cold) faunal realm associations was obtained from earlier studies by Müller (1959), Bednarczyk (1979), from the recent data of Kaljo et al. (1986), and from this paper (central Kazakhstan). The line between the T and C realms (Fig. 6) is tentative at this stage.

The pattern of realms develops in the Late Cambrian, at the end of the Aksay Epoch, the *Westergaardodina amplicava* time. This is the time of abundant proto- and paraconodonts (Fig. 7). Maximum taxonomic diversity of these conodont groups occurs in the W and T associations of central Asia (sections of northern China, as well as Batyrbay and Kyrshabakty in Kazakhstan). Medium and minimum taxonomic diversity is found in the cold water associations in northwestern Europe (Kaljo et al., 1986) and warm water associations of North America (Miller et al., 1981), respectively. The central Asia, Kazakhstan-China, region may have been the centre of origination and migration of the proto- and paraconodont fauna.

The first euconodonts occur in the beginning of the Batyrbay Epoch. Coeval associations are characterized by different relations between euconodont and proto-, paraconodont elements. Faunal contrast becomes sharper from the *alisonae* time of the latest Cambrian. Within this final epoch of the Late Cambrian the following associations are found for the various time intervals:

Hirsutodontus ani time

In the *Hirsutodontus ani* time, the dominant component of the W associations is represented by euconodonts of *Proconodontus* in North America (Miller et al., 1982); of *Proconodontus*, *Teridontus*, and *Hirsutodontus* in North China (An, 1982; An et al., 1983); of *Proconodontus* and *Teridontus* in Australia (Druce et al., 1982), whereas a relatively small component is composed of proto- and paraconodonts (Fig. 7). In the T associations recovered, for example, from the base-of-slope deposits of the Batyrbay section, we observe the reverse relationship of elements: proto- and paraconodonts prevail, but the stratigraphically important euconodonts make up (in thin-bedded turbidites) a minor percentage component (2-5%), which is allochthonous by nature (i.e., synchronously transported from the nearby shallow environment by turbidity currents). Euconodonts, abundant in shallow, warm water associations, arriving in the T realm, make up therein a minor component. Thus, the analysis of euconodonts from the W and T realms can help to define their origin and compare different facies. Euconodonts of this time interval belong to exclusively warm water varieties. They are absent in C assemblages, making the correlation with northwestern Europe difficult.

Eoconodontus (E.) notchpeakensis time

The *Eoconodontus (E.) notchpeakensis* time is connected with the appearance of *E.(E.) notchpeakensis*, the first cosmopolitan euconodont dominating the W realm. The average euconodont content in the T realm (up to 30% in calcisiltite turbidites in the Batyrbay section) is caused, apparently, by both in situ occurrence and transport from shallower environments. Rare elements of *Proconodontus* (1-2% from the total assemblage) most probably are transported, though no attrition or damage is observed. Evidently, this kind of preservation of elements is characteristic of this type of transport, as already reported by Ethington and Repetski (1984). In the C realm, the elements of *E. (E.) notchpeakensis* are extremely rare, for example, in the Estonia and Leningrad areas (Kaljo et al., 1986), and in Sweden (Müller, 1959) and Poland (Bednarczyk, 1979) they are either absent or as yet unknown. Disproportionate abundances between proto- and paraconodonts on one hand, and euconodonts on the other, are typical for the associations discussed for this time interval (Fig. 7).

Eoconodontus (E.) alisonae time

The *Eoconodontus (E.) alisonae* time is characterized by the strict confinement of some euconodont species to certain environments. Species of the *E. (Cambrooistodus)* subgenus are typical for the W ancient epicontinental seas, whereas *E. (Eoconodontus) alisonae* Landing characterizes only the T open ocean paleoenvironments (Fig. 8). This fact was first noted by Landing (1983). In the T associations (for example, in Batyrbay) rare elements of *E. (Cambrooistodus)* and *Proconodontus* are redeposited. The C fauna is devoid of warm water *E. (Cambrooistodus)*, and the transitional *E. (E.) alisonae* remains unchanged. Therefore, conodont zonal subdivisions of the same name cannot be established in northwestern Europe.

Cordylodus primitivus time

The most interesting, disputed, and short time interval is the *Cordylodus primitivus* time. In this brief interval, the first primitive species of *Cordylodus* occur, such as *C. andresi* Viira and Sergeeva (Viira et al., 1987) and *C. primitivus* (Bagnoli et al., 1987). However, during this period these species inhabited the C and T realms only, and did not penetrate into the warm water cratonic seas of North America,

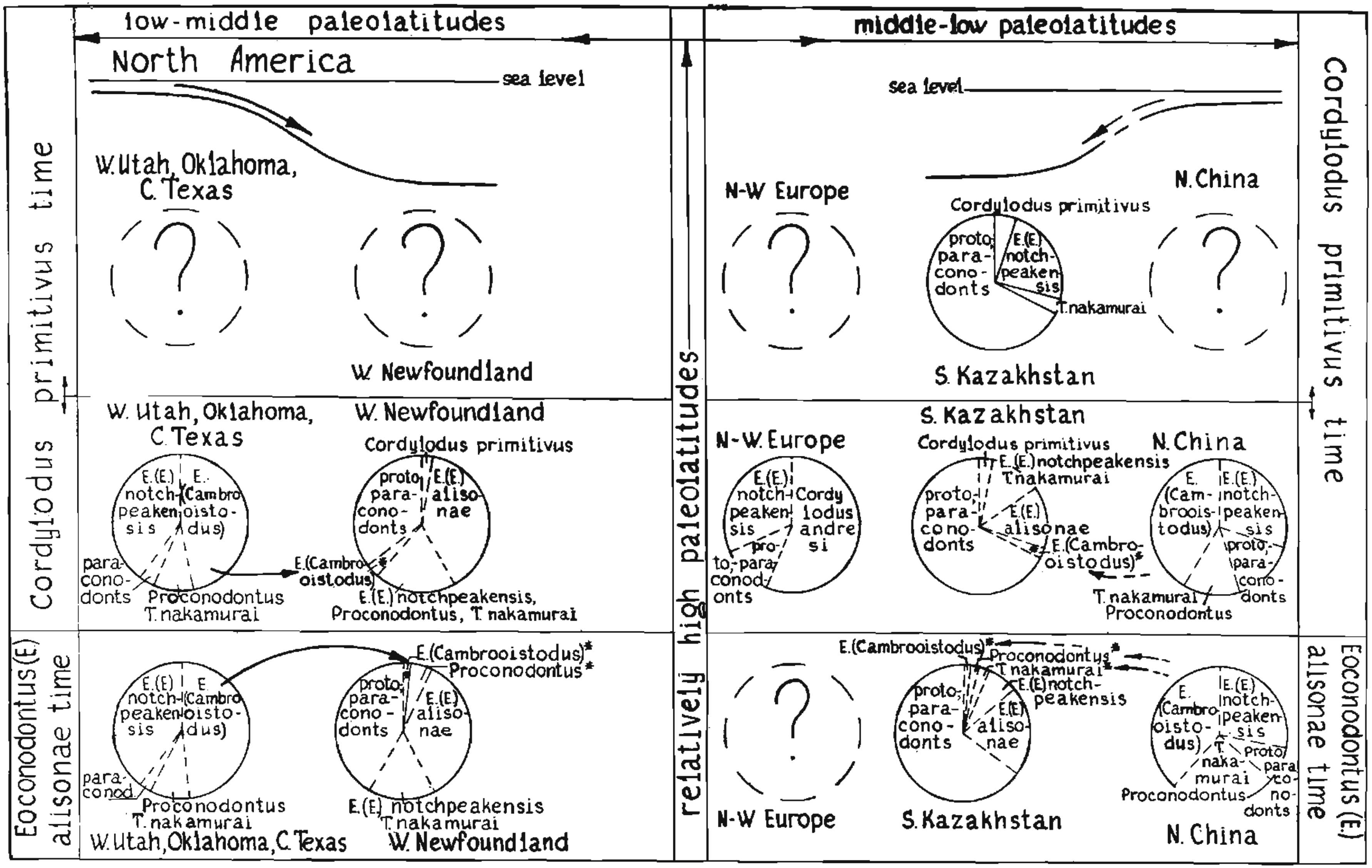

Figure 8. Comparison of coeval conodont associations in various paleoenvironments from *alisonae* to *Cordylodus primitivus* time of the latest Cambrian. References and symbols as for Figure 7.

north China, or Australia. Invasion of the primitive cordylodan species onto the cratons occurred somewhat later, at the beginning of *proavus* time.

The faunal data of the Batyrbay section show that *Cordylodus primitivus* time should be divided into two phases (Fig. 8). During the first, in the T associations (Batyrbay), primitive species of *Cordylodus* coexist with proto- and paraconodonts, *E.* (*E.*) *notchpeakensis* and with the less common *E.* (*E.*) *alisonae*. By the end of the first phase of *primitivus* time, *E.* (*E.*) *alisonae* becomes extinct and the rare transported elements of *E.* (*Cambrooistodus*) and *Proconodontus* disappear from the T associations as well, which is explained by the extinction of these species in the adjacent cratonic paleoenvironments.

Associations of the second phase of *primitivus* time can be observed in the C and T realms or only in the latter (e.g., Batyrbay section). These associations are represented by primitive elements of *Cordylodus*, still existing *E.* (*E.*) *notchpeakensis*, and proto- and paraconodonts. In shallow water cratonic environments, the second phase of *primitivus* time is probably reflected by a hiatus in sedimentation that lasts until the beginning of *proavus* time.

Thus, the beginning of the Lange Ranch Eustatic Event, marked by a sharp and episodic drop in sea level (Miller, 1984), is fixed apparently at the second phase of the *Cordylodus primitivus* time. In this case, the first hiatus in the sedimentation of the shallow water platform carbonates embraces the time interval from the end of *primitivus* time until the very beginning of *proavus* time. During this period, the conodont associations of the Batyrbay section have no analogues in shallow cratonic seas (Fig. 8) as well as in proximal and distal parts of ocean basins, for example, in western Newfoundland (Barnes, 1988), because some of the latter are characterized by intensive erosion processes resulting in a loss of coeval deposits.

Cordylodus proavus time

Cordylodus proavus time refers to early Tremadoc time. Faunal changes at the base of the *proavus* Zone, considered herein as the systemic boundary, are not equivalent in different paleoenvironments. In the C realm, these changes are manifested only by the extinction of some species of paraconodonts and in the appearance of numerous specimens of *C. proavus*. The T associations also do not have sharp changes. Near the systemic boundary, *E.* (*E.*) *alisonae* and some paraconodont species disappear and, soon after, *C. proavus* is common. Only warm water associations show significant changes. Their composition undergoes alterations due to the extinction of *E.* (*Cambrooistodus*) and *Proconodontus* species and to the appearance of *C. proavus*, species of *Hirsutodontus*, and, slightly later, species of *Fryxellodontus* and *Clavohamulus*. At this time in the carbonate shallow water environments of North America (Miller, 1984), the steady spread of *Teridontus nakamurai* and *Cordylodus primitivus* takes place.

Hirsutodontus, *Fryxellodontus*, *Clavohamulus* (*H.*, *Fr.*, *Cl.*) were unable to tolerate the open sea and oceanic environments, and together with the more eurytopic *E.* (*E.*) *notchpeakensis* and *Cordylodus*, provide the differences in the ratio of the faunal components in the W and T associations as well as the absence of the *H.*, *Fr.*, *Cl.* component in the C realm (Fig. 9). The *H.*, *Fr.*, *Cl.* genera are dominant to relatively abundant in the shallow, warm water seas (Miller, 1984). In the slope environments (thin turbidites and in situ deposits of western Newfoundland), they rank second (Barnes, 1988), which can be explained by transportation by turbidity and surface currents. In environments such as the open outermost shelf or upper slope of Malyi Karatau seamount (Batyrbay) and the low angle continental slope or outer shelf of western Mackenzie Mountains (Landing et al., 1980) *H.*, *Fr.*, *Cl.* elements may be missing. In the T associations, proto- and paraconodonts still prevail; *Cordylodus proavus*, *C. primitivus*, *E.* (*E.*) *notchpeakensis*, and *T. nakamurai* are present in smaller but consistent numbers, and the component with *H.*, *Fr.*, *Cl.* elements is minute or missing. The C fauna is characterized by numerous *Cordylodus proavus*, *C. andresi*, *E.* (*E.*) *notchpeakensis* and rare proto- and paraconodonts (Kaljo et al., 1986). Thus, the time interval connected with the appearance of new species of *H.*, *Fr.*, *Cl.* genera at the systemic boundary may be traced in the deposits of carbonate platform and adjacent slope facies. In the first case, these in situ species are numerous, whereas in the second, they are transported and relatively rare.

The same principle is valid for the ratio of the conodont faunal components of the *Cordylodus intermedius* time (Fig. 9). In this time interval, the transported elements are the new representatives of the *Teridontus* lineage (i.e., elements of the *Monocostodus*, *Utahconus*, and *Semiacontiodus* genera) and the *Hirsutodontus* lineage. A certain number of such redeposited specimens in slope associations is common, as in western Newfoundland (Barnes, 1988) and central Nevada (Cook et al., 1989). But in the transition zone, from the outermost shelf to slope environment, the probability of their meeting is small, e.g., rare, as *H. simplex* at Batyrbay or even absent, as in western Mackenzie Mountains (Landing et al., 1980). It is known that species such as *H. simplex* were adapted exclusively to the restricted marine environments of ancient epicontinental seas (Landing et al., 1980; Miller, 1984). The bypass zone from the open outermost shelf to the slope is a most unsuitable environment not only for the habitat of *H. simplex*, but also for the capture of the *H. simplex* elements for redeposition. Most probably, the few specimens of *H. simplex* at Batyrbay are indigenous (in situ) and survived in the seamount outermost shelf to upper slope environment.

Generally, the associations of the T realm include, apart from the allochthonous (transported) or starved W component, a considerable cosmopolitan component (i.e., cordylodan species, *E.* (*E.*) *notchpeakensis*, *T. nakamurai*), proto- and paraconodonts (or moderate numbers of cosmopolitans and abundant proto- and paraconodonts, as in the Batyrbay section). The content of cosmopolitan species of *Cordylodus* and *E.* (*E.*) *notchpeakensis* increases in both the C and taxonomically homogeneous assemblages.

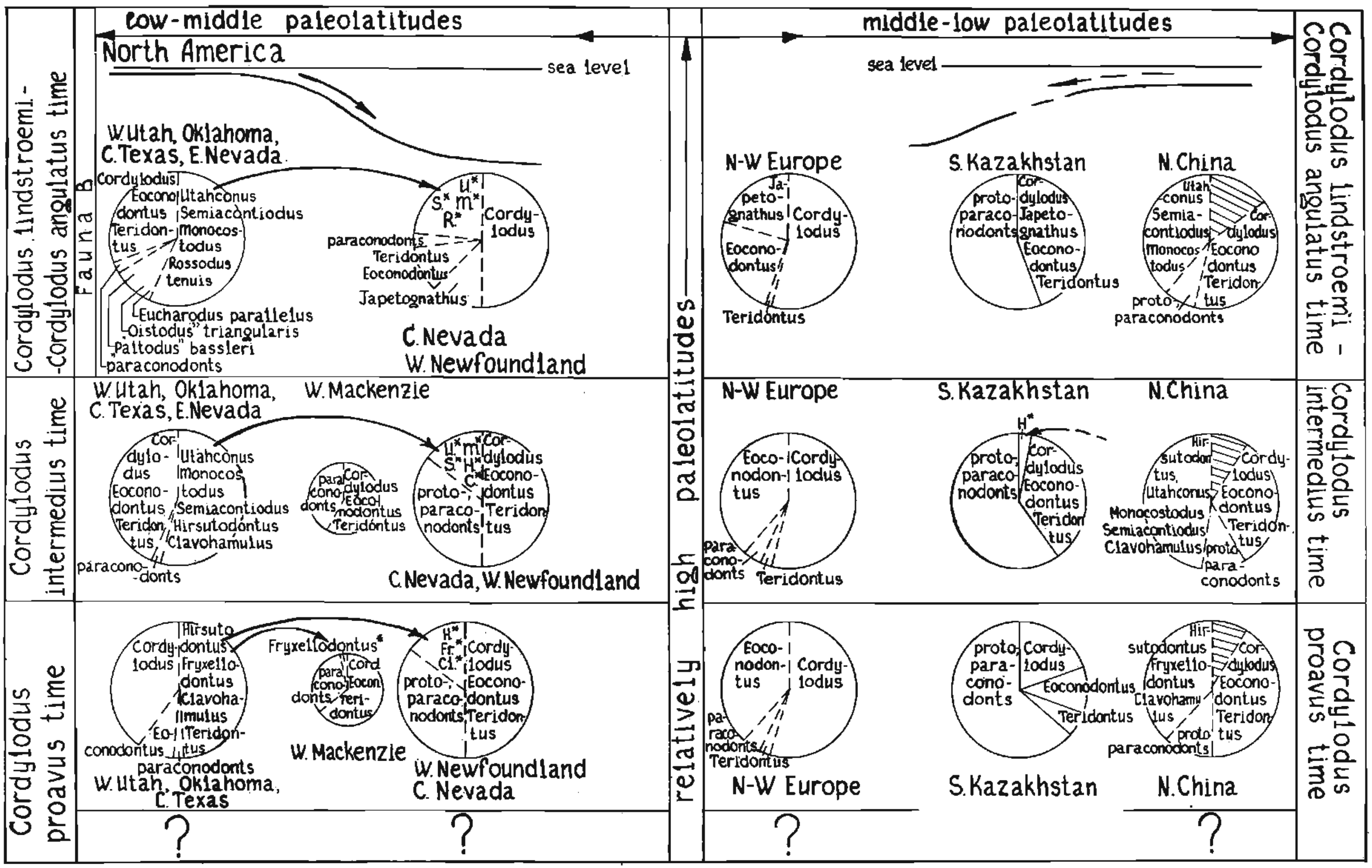

Figure 9. Comparison of coeval conodont associations in various paleoenvironments from *proavus* to *lindstromi-angulatus* time of the Early Ordovician. References and symbols as for Figure 7, except eastern Nevada (Ethington and Repetski, 1984); central Nevada (Cook et al., 1989). Faunal components endemic to central Asia symbolized by cross-hatching.

Cordylodus lindstromi-C. angulatus time

Cordylodus lindstromi-C. angulatus time is characterized by the last remaining cosmopolitan *E.* (*E.*) *notchpeakensis* and species of *Cordylodus*. In the W realm, as shown by Miller (1984) and Ethington and Repetski (1984), a small percentage of new Midcontinent species is added to the numerous distacodontiform and less abundant cordylodan species (Fig. 9). The C and T realms undergo some innovations due to the appearance of *Iapetognathus*. The same regularity in the ratio of faunal components in the adjacent W and T realms is evident. The C realm becomes progressively more different from the others. However, the contrast between the coeval faunal realms is still smoothed by cosmopolitan elements. The actual difficulties in comparing these realms increase from the end of the Black Mountain Eustatic Event (Miller, 1984), when numerous coniform taxa appear in the W and C realms but there are practically no taxa in common (Miller, 1984).

Drepanoistodus deltifer-Rossodus manitouensis time

The proof that the upper Tremadoc cold water *Drepanoistodus deltifer* fauna is coeval with the warm water *Rossodus manitouensis* fauna (or Fauna C of North America) is based on the fact that both of them replace previous associations with the last remaining cosmopolitans. To compare these sharply contrasting faunas, attention should be paid to the T realm's associations (Fig. 10), containing, together with the *deltifer* faunal elements, some shallower water transported forms, shown in the data of Landing (1983) from the continental slope deposits of the Highgate Formation in northwestern Vermont. There are also some examples of changes in the ratio of different components of the *Rossodus manitouensis* fauna discovered from the shallow to relatively deep environments of North America, described by Ethington and Repetski (1984).

Such observations are of special value for the interpretation of the upper Tremadoc associations derived from the different lithofacies of the Sarykum section. The upper part of the lower member of the Chazhagay Formation is considered herein (Fig. 3). The first conodont association at Sarykum is a typical warm water one with *R. manitouensis*, *Chosonodina herfurthi* and other species (Fig. 3, section B, sample [51] 81.5). This association, derived from the calcarenite turbidites of the Chazhagay Formation's lower member, has been transported from the neighbouring shallow water carbonate environment into the hemipelagic environment without any changes in the ratio of faunal components. This can be proved by having the same ratio of components in both the Sarykum and in the shallow water Fauna C of North America (Fig. 10). Furthermore, the invariability in the ratio of faunal components of the *R. manitouensis* association during redeposition from the shallow to hemipelagic environment at Sarykum can be evidence of the high density of turbidity currents in the Sarykum Basin. The second conodont association, a typical cold water association found in the cherts of the same member (Fig. 3, section C, samples 148-3, 148-2) was formed in situ in a relatively deep hemipelagic environment of the Sarykum Basin.

With reference to the above, the events happening at this time in south Kazakhstan are of acute interest (Fig. 10). The Batyrbay section serves as an example of the coexistence of *deltifer* fauna with proto- and paraconodonts in the shoal environment of the seamount, subjected to ocean currents and the constant influence of free access to the open ocean. Consequently, for the seamount, it is not mandatory to have a relatively deep environment (similar to the continental slope in North America) in order to have a *deltifer* component in the population. However, coeval proto- and paraconodonts are absent in the basinal Sarykum hemipelagic environment, which contains only the *deltifer* fauna, similar to that of Europe but less taxonomically diverse. Such is the difference in the upper Tremadoc biofacies in Kazakhstan.

The lower Arenig fauna, when compared with its predecessors, is more differentiated, depending on the specific environments. The complete diversity of environments and associations might be presented as a pattern, wherein various ratios of faunal components take part: at first, two, i.e., North Atlantic or (in the narrower sense of this term) Acado-Baltic (A-B) and Midcontinent (M); and later, three, i.e., A-B, M, and *Juanognathus* (Ju).

Paroistodus proteus-Prioniodus (P.) deltatus longibasis time

The distribution of these two types of faunal components from the *Paroistodus proteus* to the *Prioniodus* (*P.*) *deltatus longibasis* time is shown in Figure 10. The data reveal the similarity of the Batyrbay shallow water association of the seamount to the North American populations from the continental slope and basin environments. Likewise, the close similarity of the deep water basinal association from the black chert of the Sarykum section (i.e., from the upper member of the Chazhagay Formation; Fig. 3, section C, samples 148-I, 148-A, 148) and the cold water fauna of northwestern Europe is shown. The Sarykum association, when compared to that of Europe, is taxonomically less diverse; probably only a limited number of A-B species were adapted to the relatively deep, cold environment of the central Kazakhstan Basin.

Prioniodus (P.) elegans-P. (Oepikodus) evae time

The last time interval to be considered herein is the *Prioniodus* (*P.*) *elegans-P.* (*Oepikodus*) *evae* time, which is connected with the appearance of the *Juanognathus* component (Lindström, 1976) in the conodont fauna. This component is typical only for the T faunal realm. It is worth mentioning that if the A-B and M components under certain paleoenvironments make up the basis for the contrasting associations of the same names, the Ju component is not known as an independent unit or Ju population. Its occurrence either with M, A-B or with both components provides a solution to the problem of comparing provincially differentiated lower Arenig A-B and M faunas. In North America, Ethington and Repetski (1984) identified four coeval populations traced from inner shallow shelf to ocean basin environments (Fig. 10). The open outer shelf and continental slope environments containing, respectively, the second and the

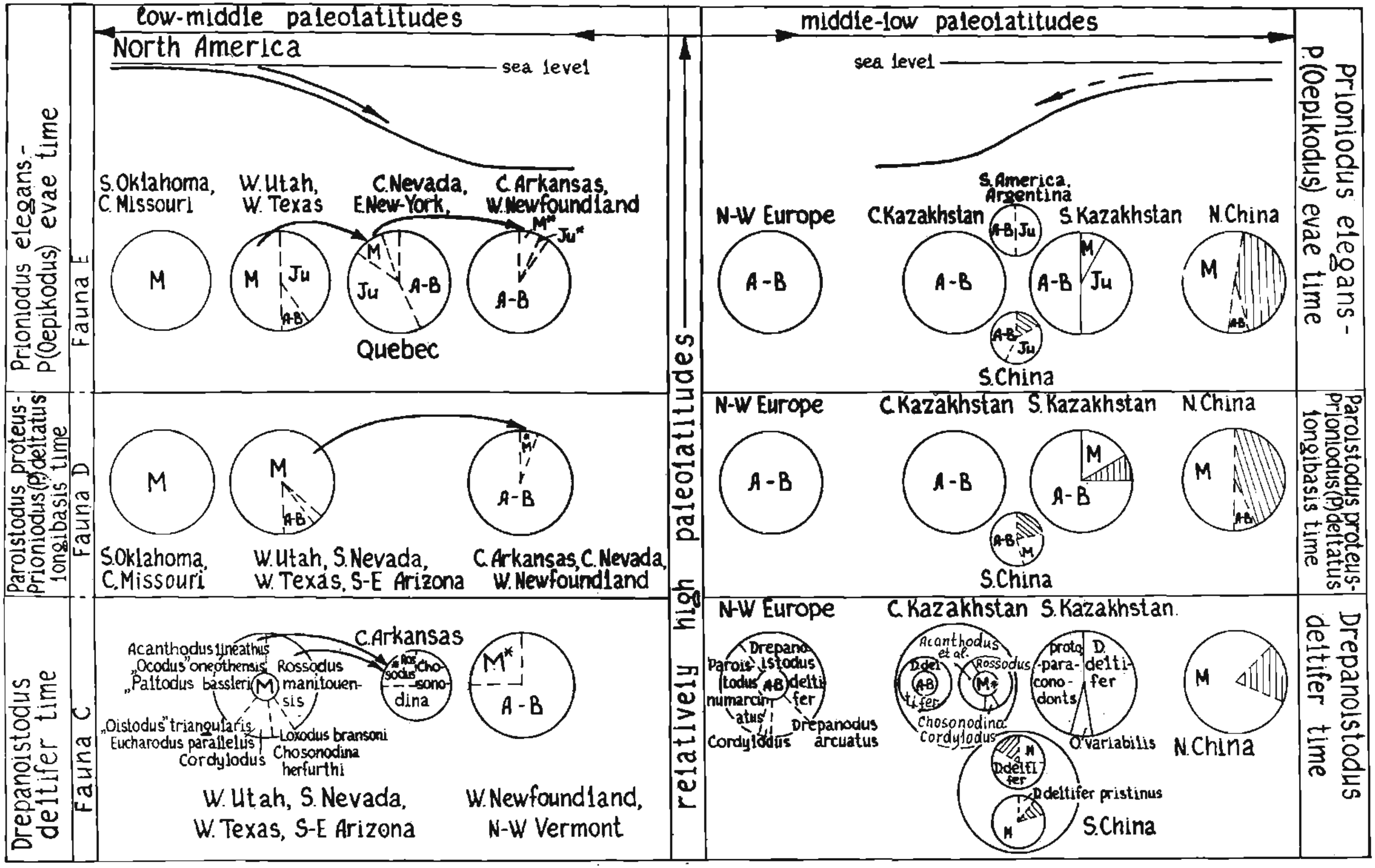

Figure 10. Comparison of coeval conodont associations in various paleoenvironments from late Tremadoc *deltifer* time to the early Arenig *evae* time. References: western Utah (Ethington and Clark, 1981); Quebec (Landing and Ludvigsen, 1984); western Newfoundland (Fåhraeus and Nowlan, 1978); northwestern Vermont (Landing, 1983); eastern New York (Landing, 1976); Argentina (Serpagli, 1974); south China (Ni et al., 1983; Zeng et al., 1983). References for southern Oklahoma, central Missouri, southern Nevada, southeastern Arizona, western Texas, central Nevada, central Arkansas - see Ethington and Repetski (1984). References for northwestern Europe, north China - see Figure 7. Symbols as for Figure 9.

third populations, are the most suitable for the Ju component. Faunas A and B of Argentina (Serpagli, 1974) and the Batyrbay shallow water association including both A-B and Ju components are most similar to the above-mentioned populations. The probability of the Ju and M components meeting in northwestern Europe and in the basinal pelagic environment of central Kazakhstan does not exist. These are the environments inhabited by A-B species exclusively. Thus, the T realm's associations with a Ju component in situ, or in transported form, could not be overestimated for the correlation of different types of lower Arenig deposits.

In conclusion, it is worth enumerating the main faunal features typical of the T (transitional) faunal realm, thus giving the reason for distinguishing it. The T realm is characterized by the following:

1. The Upper Cambrian *Eoconodontus* (*E.*) *alisonae*. This species is known in the T realm not only in situ, but also in transported form. It means that the habitat of *E.* (*E.*) *alisonae* could embrace an open ocean environment of outermost shelf and upper slope (i.e., above the permanent thermocline), but the lower continental slope and hemipelagic environment might be represented, apparently, by transported specimens of *E.* (*E.*) *alisonae*.
2. The abundance and diversity of long-lived proto- and paraconodonts continues until the late Tremadoc or early Arenig.
3. In the early Arenig, a *Juanognathus* component develops. It is known in the T realm not only in situ, but also in transported form.
4. The presence of rare (starved), warm water species.

The contemporaneously redeposited warm water euconodont elements are especially important for comparing different cratonic and oceanic biofacies (i.e., to compare the W and T faunal realms).

Depending on the bathymetric parameters, the T realm may be considered as a continuous series of paleoenvironments and their associations of lower rank. In such a series, three or more paleoenvironments of low rank can be conditionally recognized, with a continuous transition from so-called moderate warm, to moderate cold, to cold water paleoenvironments. However, all four of the above-mentioned characteristics of the T realm cannot be applied to the deepest cold water paleoenvironments (and associations). The latter, such as a pelagic basinal environment (and association) in central Kazakhstan in early Arenig time, is similar to that of the C faunal realm nearby, in the relatively high paleolatitudes.

A comparison of different types of paleoenvironments and their associations provides several advantages. Firstly, it makes the correlation more precise. Secondly, it allows a generalization of the stages in early conodont evolution. The stages might be based on the changes in the euconodont evolutionary lineages of McTavish (1973), Miller (1980, 1984, 1988), Dzik (1983), Repetski and Ethington (1983), Bagnoli and Stouge (1985), and Bagnoli et al. (1987) on the rate of their evolution and on the analysis of the proto- and paraconodonts. This must be accompanied by a generalization of the changes in the evolutionary lineages in the W, T, and C realms and on the evaluation of the heirarchy of these changes. Hence, the most significant changes in the conodont fauna at the base of the *Cordylodus proavus* Zone correspond to the change in the evolutionary stages of the first rank, i.e., between Late Cambrian and Early Ordovician time. This is regarded herein as the systemic boundary. The evolutionary stages of second rank might correspond to changes at the boundaries of series (or stage) subdivisions, and influence ideas on the duration of the Tremadoc Series. Finally, the stages of the third rank provide a basis for zonal subdivisions.

The conodont zonal scale at Batyrbay is one of the few scales elaborated for the T realm.

The following euconodont lineages have a particular significance for establishing zonal conodont subdivisions at Batyrbay:

1. Hirsutodontus lineage

The most primitive representatives of the *Hirsutodontus* lineage, such as *H. ani* (Wang), succeeded by *H. transmutatus* (Xu and Xiang), are known in the Upper Cambrian of north China (An, 1982; An et al., 1983) and of the Malyi Karatau (Batyrbay and Kyrshabakty sections). These species inhabited the Chinese epicontinental seas and were found in situ, whereas in the Malyi Karatau slope facies, they were contemporaneously transported by currents from the adjacent shallower environment. The primitive species of the *Hirsutodontus* lineage appeared practically simultaneously with the first primitive species of the *Proconodontus* lineage (i.e., *P. tenuiserratus* Miller) at the base of the Fengshan Stage of north China and, respectively, at the base of the Batyrbay Stage in Kazakhstan (Kyrshabakty section). These data indicate that the *Hirsutodontus* lineage can be traced from the Late Cambrian, i.e., from the beginning of Batyrbaian (Kazakhstan) and Fengshanian (north China) time, equivalent to latest Franconian (North America) time, rather than from the Early Ordovician *proavus* time as suggested by Miller (1980, 1984). Furthermore, the *Hirsutodontus* lineage apparently developed independently from the *Teridontus* lineage.

Miller (1988), having summarized the latest data from the Dayangcha section of northeastern China (Chen and Gong, 1986), came to similar conclusions. In south Kazakhstan, the earliest evolutionary stage of the *Hirsutodontus* lineage has been used for establishing the *Hirsutodontus ani* Zone in the lowest Batyrbay Stage of the Upper Cambrian.

2. Lineages from Eoconodontus (E.) notchpeakensis to E. (E.) alisonae or to Cordylodus primitivus

The lineages from *Eoconodontus* (*E.*) *notchpeakensis* to *E.* (*E.*) *alisonae* and, respectively, from *E.* (*E.*) *notchpeakensis* to *Cordylodus primitivus*, characteristic of the T realm, were taken into account while establishing the *Eoconodontus* (*E.*) *alisonae* and *Cordylodus primitivus* zones in the uppermost

Cambrian in south Kazakhstan (Apollonov et al., 1988; Chugaeva et al., 1989) and subzones of the same name in western Newfoundland (Barnes, 1988).

3. Lineage from Prioniodus (P.) deltatus longibasis to P. (P.) elegans

The reconstruction of the *Prioniodus* (*P.*) *elegans* lineage remains one of the interesting evolutionary problems (McTavish, 1973; Bagnoli and Stouge, 1985). New data of Pohler et al. (1987) show that the new species of *Prioniodus*, recorded in the upper slope facies of western Newfoundland, is probably the ancestor of *Prioniodus elegans*. Pohler et al. (1987) noted the similarity in the elemental composition of the *P.* n. sp. and *P. elegans* apparatuses, but noted also the more primitive (minute) denticles of the *P.* n. sp. elements compared with those of *P. elegans*.

P. (*P.*) *deltatus longibasis* (McTavish), recovered from the shallow water Malyi Karatau seamount facies (Batyrbay), has the same type of apparatus and probably the same primitive denticulation of elements as the *P.* n. sp. of Pohler et al. (1987). However, the denticulation of the *P.* (*P.*) *deltatus longibasis* (McTavish) elements is so minute, that it is often difficult to recognize. If the *P.* n. sp. in western Newfoundland and *P.* (*P.*) *deltatus longibasis* (McTavish) in Batyrbay are synonymous, the lineage, named at present as a transition from *P.* (*P.*) *deltatus longibasis* to *P.* (*P.*) *elegans* (this paper) or from *P.* n. sp. to *P. elegans* (Pohler et al., 1987), will be very important for the detailed zonation of the lower Arenig in the T faunal realm.

The lineage from *Prioniodus* (*P.*) *deltatus longibasis* to *Prioniodus* (*P.*) *elegans* was used to establish conodont zones of the same name in the continuous lower Arenig deposits in south Kazakhstan.

CONCLUSIONS

The continuous succession of Upper Cambrian and Lower Ordovician conodont associations recognized in the carbonate facies of the Batyrbay section, Malyi Karatau, south Kazakhstan, is more detailed than the conodont succession established in the siliceous-carbonate facies of the complete Sarykum section, central Kazakhstan, but they do not differ in principle.

Generally, the composition of the Batyrbay conodont associations depended on their location on the margin of the ancient carbonate seamount, where there was permanent and free access to the open ocean, and the influence of a system of ocean currents. Basinal and slope environments in the Upper Cambrian were succeeded by the Lower Ordovician upper slope and, later, the lower Arenig shallow water environments of the seamount. In contrast, the sufficiently deep oceanic environments at Sarykum changed from the base-of-slope of the carbonate platform in the latest Cambrian to Tremadoc hemipelagic and early Arenig pelagic environments. Such paleoenvironments determined the composition of conodont associations at Batyrbay and Sarykum. They were compared to associations from other regions of the world for specific time intervals, considered as chronological equivalents of conodont zones. The Cambrian-Ordovician zonal scale of Batyrbay was taken as the standard for particular time intervals.

The comparison of coeval conodont associations from different paleoenvironments permits the recognition (separate from Miller's warm and cold faunal realm) of the transitional (T) faunal realm. This included ancient open seas and oceans in low to middle paleolatitudes. The realm embraced outermost continental shelves with unrestricted access to the open ocean, continental slopes, basinal (hemipelagic) and any carbonate seamount paleoenvironments. The realm is traced along the ancient continental margins and can be illustrated by sections in south Kazakhstan, western Mackenzie Mountains (Landing et al., 1980), central Nevada (Cook et al., 1989), northwestern Vermont (Landing, 1983), western Newfoundland (Barnes, 1988), and probably occurs in southeastern China.

The latest Cambrian *Eoconodontus* (*E.*) *alisonae*, the abundance and diversity of long-lived proto- and paraconodonts, the early Arenig *Juanognathus* faunal component in conodont associations and, finally, the rare (starved) warm water species are all typical characteristics of the T faunal realm. The significance of contemporaneously redeposited warm water euconodont elements is noted to compare cratonic and oceanic biofacies (i.e., to compare markedly different warm and T faunal realms).

The zonal scale of Batyrbay is one of the few scales elaborated for the T realm, with the scale of Sarykum likewise a model for the C realm.

Four euconodont lineages had particular significance for establishing zonal subdivisions at Batyrbay. The earliest evolutionary stage of the *Hirsutodontus* lineage has been used for recognition of the *Hirsutodontus ani* Zone in the lowest Batyrbay Stage of the Upper Cambrian. The lineages from *Eoconodontus* (*E.*) *notchpeakensis* to *E.* (*E.*) *alisonae* and from *E.* (*E.*) *notchpeakensis* to *Cordylodus primitivus*, typical for the T faunal realm, helped to establish the *Eoconodontus* (*E.*) *alisonae* and *Cordylodus primitivus* zones in the uppermost Cambrian. The lineage from *Prioniodus* (*P.*) *deltatus longibasis* to *P.* (*P.*) *elegans* was used to define these zones in the lower Arenig. If this latter lineage is equivalent to the *P.* n. sp.-*P. elegans* lineage of Pohler et al. (1987), it will be important for detailing the lower Arenig zonal scales of the T faunal realm.

The faunal data of the Batyrbay section showed that the very short *Cordylodus primitivus* time interval of the latest Cambrian should be divided into two parts. The conodont association of the second part could be observed in the C and T faunal realms or in the T realm only. In shallow cratonic environments (warm realm) the second part of the *Cordylodus primitivus* time was apparently reflected by a hiatus. Earlier, the hiatus was recognized within *Cordylodus proavus* time by Miller (1984). Thus, the beginning of Miller's Lange Ranch Eustatic Event at the systemic boundary (as defined herein) has obtained a more precise definition, i.e., from the second phase of the *Cordylodus primitivus* time of the latest Cambrian.

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Conodont biostratigraphy of the upper St. George Group (Canadian to Whiterockian), western Newfoundland

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Abstract

The St. George Group of western Newfoundland consists of two megacycles of autochthonous carbonate deposition. The sequence, from high in the lower part of the Catoche Formation to the uppermost part of the Aguathuna Formation, i.e., most of the upper megacycle, may be subdivided into seven informal assemblages on the basis of conodont faunas. The lower three assemblages are dependent upon faunal lineage changes. Assemblages may intergrade, as is the case between Assemblages III and IV. Assemblages V to VII occupy the middle to upper part of the Aguathuna Formation. Although defined by changes in conodont associations, the upper Aguathuna Formation assemblages are strongly environmentally controlled and much of the fauna is dominated by an extremely shallow water association, which is unusual in its abundance of particularly enigmatic taxa. This interval is also marked by the occurrence of reworked conodonts.

The biostratigraphy was defined at Table Point and from cores from the nearby Daniel's Harbour zinc mine. When applied to lithostratigraphically similar rocks at other locations throughout western Newfoundland, stratigraphic hiatuses and faunal data indicate varied time intervals for the deposition of Aguathuna facies.

Résumé

Le groupe de St. George dans l'ouest de Terre-Neuve est composé de deux mégacycles de sédimentation carbonatée autochtone. La séquence allant du sommet de la partie inférieure de la formation de Catoche au sommet de la partie supérieure de la formation d'Aguathuna, c'est-à-dire presque tout le mégacycle supérieur, peut être subdivisée en sept assemblages non officialisés basés sur les faunes de conodontes. Les trois assemblages de la base se fondent sur les changements de lignées. Les assemblages peuvent s'entremêler, comme c'est le cas pour les assemblages III et IV. Les assemblages V à VII sont contenus entre les parties intermédiaire et supérieure de la formation d'Aguathuna. Même s'ils ont été définis par des changements d'associations des conodontes, les assemblages supérieurs de la formation d'Aguathuna se distinguent fortement par des caractéristiques environnementales et la grande partie de la faune est constituée d'une association d'eau très peu profonde où l'on trouve en abondance inhabituelle des taxons particulièrement énigmatiques. Cet intervalle est également marqué par la présence de conodontes remaniés.

Cette biostratigraphie a été définie à la pointe Table et à partir de carottes prélevées dans la mine de zinc voisine à Daniel's Harbour. Lorsqu'on les applique à des roches lithostratigraphiquement semblables situées à d'autres endroits répartis dans tout l'ouest de Terre-Neuve, les lacunes stratigraphiques et les données sur la faune révèlent divers intervalles de temps en ce qui concerne la sédimentation du faciès d'Aguathuna.

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INTRODUCTION

The St. George Group is that sequence of Lower to Middle Ordovician autochthonous rocks deposited on the stable North American platform margin of the Iapetus Ocean. It outcrops along a discontinuous sinuous belt in western Newfoundland from the Port au Port Peninsula in the south, to Cape Norman and Hare Bay in the north (Fig. 1). It is composed, in ascending order, of the Watts Bight, Boat Harbour, Catoche, and Aguathuna formations, and consists of two shallowing-upward megacycles of deposition (Knight and James, 1987).

These rocks are important as the autochthonous equivalent of the upper Cow Head Group. Within that group, the study of intercalated shallow water shelly faunas with graptolite sequences, and conodonts derived from the slope margin, has enabled direct comparison of previously unrelated biostratigraphic schemes, and facilitated international correlation (Williams and Stevens, 1988). Shallow water conodonts of the St. George Group show affinity with conodonts of the Midcontinent Conodont Province, but may be correlated with the Cow Head Group at various levels. Thus, the North American Midcontinent conodont zonation may be directly compared with graptolite and European shelly fossil biostratigraphy via the St. George and Cow Head groups.

Stable platform sedimentation in the Ordovician of western Newfoundland is last represented in the St. George Group. The overlying Table Head Group represents deposition onto a subsiding, and eventually foundered, platform (James et al., 1989). The initial effects of the Taconic Orogeny in platformal sequences are recorded in the upper St. George Group (Knight and James, 1987; James et al., 1988, 1989).

In addition, the St. George Group hosts stratabound Mississippi-Valley-type mineralization at various levels (Lane, 1984). A precise and detailed conodont biostratigraphy has enabled clarification of lithostratigraphic relationships within mine and prospective target areas.

Previous studies

This paper forms part of a sedimentological and paleontological study of the Cow Head and Ordovician platformal rocks of western Newfoundland. The St. George (Knight and James, 1987), Table Head (Klappa et al., 1980), and Cow Head (James and Stevens, 1986; James et al., 1987) groups have recently been redescribed and defined. St. George Group cephalopods (Flower, 1978), trilobites (Fortey, 1979; Stouge and Boyce, 1983), conodonts (Barnes and Tuke, 1970; Stouge, 1982; Stouge and Boyce, 1983), and Table Head Group conodonts (Stouge, 1984), and Cow Head Group graptolites (Williams et al., 1987; Williams and Stevens, 1987, 1988) and conodonts (Fåhraeus, 1970; Fåhraeus and Nowlan, 1978; Pohler et al., 1987) have been described. Comparative studies between these include Bergström et al. (1974), Ross and James (1987), and Williams et al. (1987). Syntheses of Ordovician sedimentation in western Newfoundland are also currently available (James et al., 1988, 1989).

On the basis of shelly fossils, the studied sequence correlates with Zones H to L or M of the Ibex area (Boyce, 1985). Trilobite occurrences suggest a Zone I or J age for the base of the Aguathuna Formation (Fortey, 1979; Boyce, 1985), which corresponds with Bendigonian 3 or 4 of the Australian graptolite zonation (Williams et al., 1987).

Studies of brachiopods (Ross and James, 1987) and graptolites (Williams et al., 1987) indicate that the Cow Head Group is entirely older than the Table Head Group, whereas Williams et al. (1987) concluded that correlation with the preliminary conodont zonation of the St. George Group was not possible. A more complete and precise conodont biostratigraphy for the Catoche and Aguathuna formations is developed herein.

The conodont succession was initially determined from the shore section from Freshwater Cove to Table Point. This locality was chosen because it is the type section for the Aguathuna Formation. Lithostratigraphy of the section has been well documented (Knight and James, 1987). Nearby, subsurface samples enabled extension of the conodont biostratigraphy into the lower Catoche Formation. Much of the upper megacycle of the St. George Group was sampled in a 180 m thick section. Constituent rock types ranged from subtidal limestones (lower Catoche Formation) through variously dolomitized and mineralized limestones (upper Catoche Formation) to peritidal dolostones of the Aguathuna Formation.

Within these lithotypes, faunas are of mixed affinity. The Catoche Formation contains conodonts characteristic of both North Atlantic and Midcontinent provinces in varying ratios. A primarily Midcontinent fauna is evident, with incursions of North Atlantic province faunas at various levels. A greater proportion of North Atlantic taxa occurs in the primarily subtidal Catoche Formation than in the peritidal Aguathuna Formation. The pattern of relative abundances, best seen in Assemblage III, is a response to changes in environmental parameters. Assemblage boundaries can be seen to be independent of these changes (Fig. 2).

CONODONT SUCCESSION

The conodont zonation of the Ibex area (Ethington and Clark, 1981) was not found to be applicable to faunas of the upper St. George Group, due to environmental differences at their respective sites of deposition. Different, informal assemblages are therefore proposed for the St. George Group, based upon the succession of species of *Diaphorodus* Kennedy (Fig. 3) and *Parapanderodus* Stouge (Fig. 4), consideration of concurrent ranges of a large number of species within these strata, and upon acme of abundances of certain species (Fig. 5, Table 1). These assemblages are recognizable within autochthonous platformal sequences throughout western Newfoundland, but not internationally, nor between the different environmental settings preserved in autochthonous and allochthonous sequences of western Newfoundland.

Peritidal sediments of the Aguathuna Formation contain many levels at which hardgrounds or disconformity surfaces are preserved. This is particularly common in the middle and

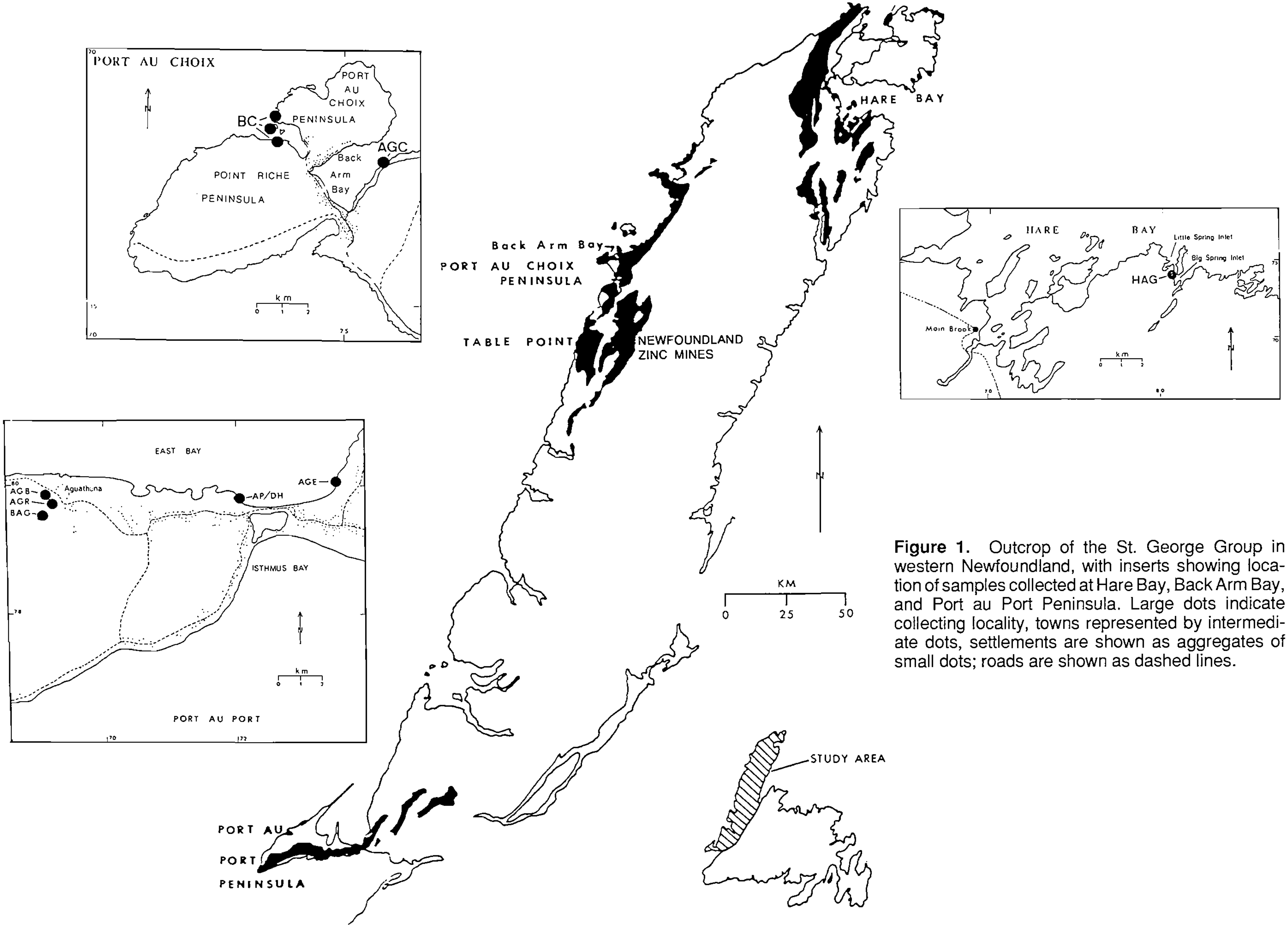

Figure 1. Outcrop of the St. George Group in western Newfoundland, with inserts showing location of samples collected at Hare Bay, Back Arm Bay, and Port au Port Peninsula. Large dots indicate collecting locality, towns represented by intermediate dots, settlements are shown as aggregates of small dots; roads are shown as dashed lines.

upper parts of the Aguathuna Formation (Knight and James, 1987). Such surfaces commonly show evidence of erosion and reworking of sediments. The anomalously late occurrence of some conodont specimens with frosted surfaces at certain levels is therefore interpreted as evidence of reworking of these conodonts. Such reworked specimens were not included within the biostratigraphic scheme: only well preserved specimens without frosted surfaces were considered in defining the assemblages.

Assemblages were therefore constructed only from taxa without severe environmental control, or from taxa displaying morphological gradients indicative of evolutionary lineages.

Assemblages I and II are described from Diamond Drill Hole (D.D.H.) A1 because the lowest strata sampled at Freshwater Cove contain an impoverished Assemblage III fauna. The contact with Assemblage II is not recorded in this section, but is found in D.D.H. 438, 715, and 1827. These confirm that there is no significant break in the faunal succession between uppermost faunas of D.D.H. A1 (Assemblage II) and the lowest sample taken from Freshwater Cove (Assemblage III).

Assemblage I

Assemblage I is dominated by *Oepikodus communis* (Ethington and Clark), which is considerably reduced in abundance in Assemblage II. *Paltodus sweeti* Serpagli, "*Scolopodus*" *filosus* Ethington and Clark, *Eucharodus parallelus* (Branson and Mehl), *Drepanoistodus forceps* (Lindström), *Diaphorodus delicatus* (Branson and Mehl), and *Drepanoistodus inaequalis* (Pander) are associated with *O. communis* in Assemblage I. Minor constituents of the fauna include *Glyptoconus quadraplicatus* (Branson and Mehl), *Parapanderodus striatus* (Graves and Ellison), *Paroistodus parallelus* (Pander), ?*Protoprioniodus papilosus* (van Wamel), *Rossodus highgatensis* Landing et al., and "*Scolopodus*" *emarginatus* Barnes and Tuke. The long-ranging *Oneotodus costatus* Ethington and Brand, *Semiacontiodus asymmetricus* (Barnes and Poplawski), ?*Semiacontiodus cordis* (Hamar), n. gen. 2 n. sp. A, and *Walliserodus ethingtoni* (Fåhraeus) are found in this assemblage.

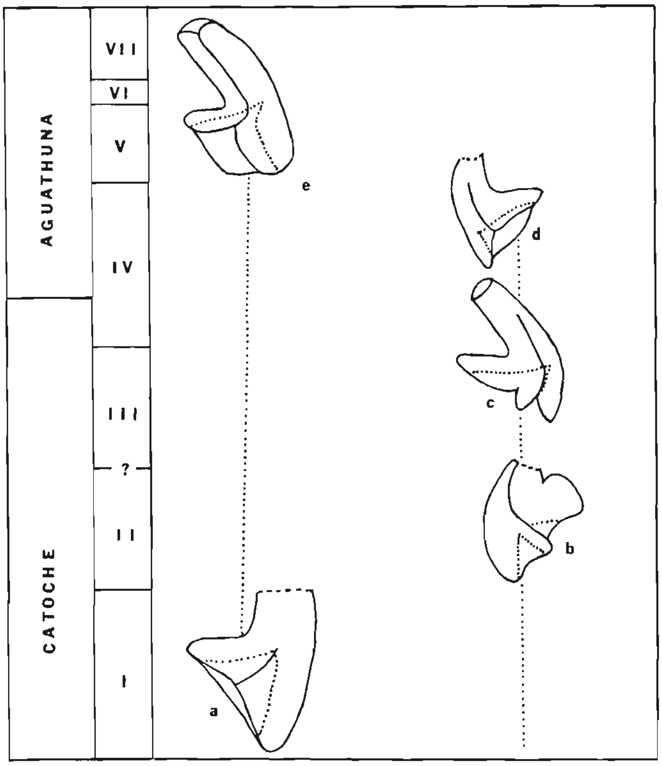

Figure 3. Species of *Diaphorodus* Kennedy found in the St. George Group, showing morphological changes in diagnostic prioniodontiform element with stratigraphic position. a, *Diaphorodus delicatus* (Branson and Mehl), ranges from Assemblage I to early Assemblage IV; b, ?*Diaphorodus russoi* (Serpagli), Assemblages II to III; c, *Diaphorodus emanuelensis* (McTavish), Assemblages III and IV; d, *Diaphorodus* n. sp. B, found in Assemblage IV (outside the Table Point area) to early Assemblage V; e, *Diaphorodus* n. sp. A, Assemblage V to Table Point Formation. The two successions are separated according to whether or not processes are developed.

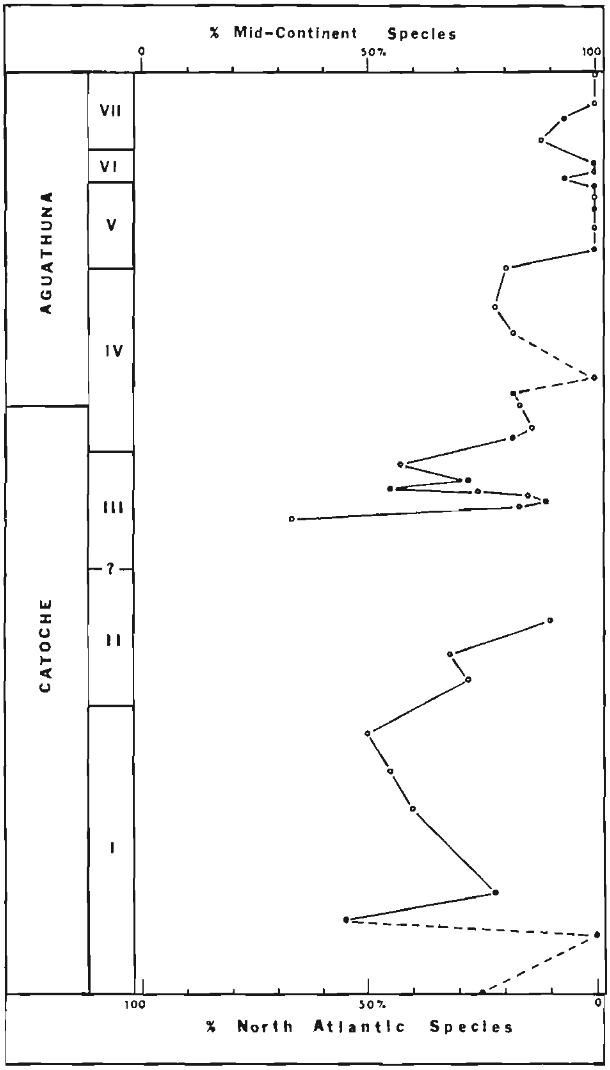

Figure 2. Relative abundance of Midcontinent and North Atlantic province conodonts plotted as a function of stratigraphic height for the D.D.H. A1 and Freshwater Cove to Table Point section. Dashed lines indicate trends based upon samples with only a few conodont species or specimens.

Within Assemblage I, alternating sequences of faunas show affinity with either Midcontinent Fauna D or E. Thus, for example, samples containing *Glyptoconus quadraplicatus* and *Rossodus highgatensis* (Midcontinent Fauna D) are interspersed with those containing *Paltodus sweeti*, *Microzarkodina*? *marathonensis* (Bradshaw), *Diaphorodus delicatus*, and *Drepanodus concavus* (Branson and Mehl), which are associated with Midcontinent Fauna E. As a consequence, many of the taxa dominating Assemblage II are present as sporadic, isolated elements throughout Assemblage I, and *Glyptoconus quadraplicatus* and *Rossodus highgatensis* are not found in populations younger than Assemblage I.

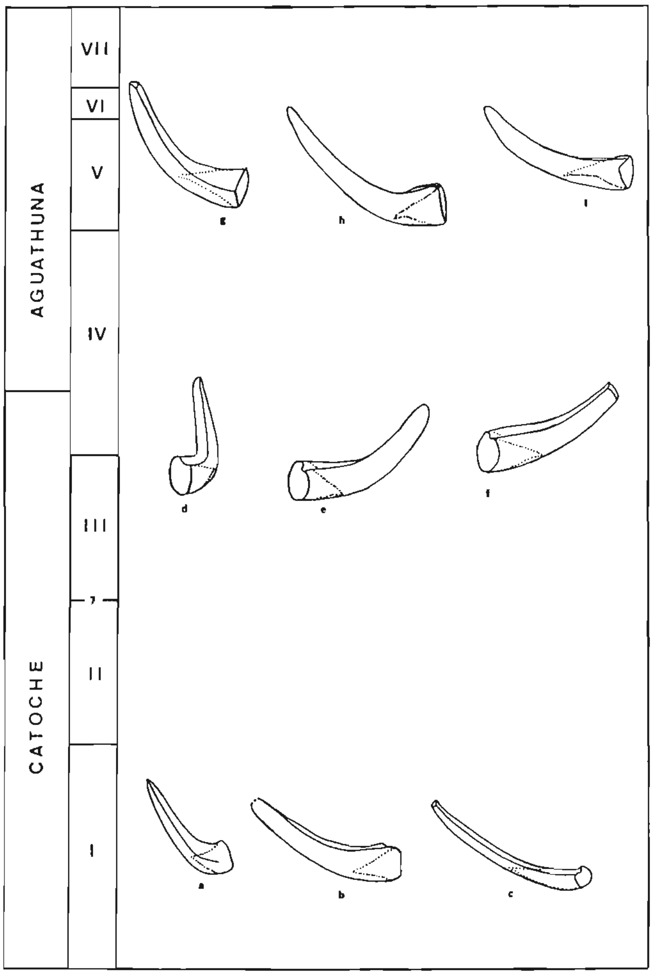

Figure 4. Species of *Parapanderodus* Stouge found in the St. George Group and overlying Table Point Formation, showing morphological changes in all elements with stratigraphic position. a-c, *Parapanderodus striatus* (Graves and Ellison); d-f, Para*panderodus* sp. aff. *P. gracilis* (Ethington and Clark); g-i, *Parapanderodus striolatus* (Harris and Harris); a,d,g, triangulariform elements; b,e,h, short based elements; c,f,i, long based elements. *Parapanderodus striatus,* characterized by round cusp section and deep basal cavity parallel to anterior margin, is present through Assemblages I to III. *Parapanderodus* sp. aff. *P. gracilis* of Löfgren (1978) is found in Assemblages III and IV; it has reduced a basal cavity and the cusp is moderately laterally compressed. *Parapanderodus striolatus* (Harris and Harris) continues from Assemblage IV into the Table Point Formation. This species shows further basal cavity reduction, and a greater degree of lateral compression.

With the first occurrence of *Oepikodus communis* defining the base of Midcontinent Fauna E (Ethington and Repetski, 1984), the presence of *Glyptoconus quadraplicatus* and an abundance of *Parapanderodus striatus* in Assemblage I indicates a correlation with lower Fauna E of Ethington and Repetski (1984).

The lower limit of Assemblage I is unknown. The faunas of this assemblage, however, are found in the lower to middle Catoche Formation.

Assemblage II

Assemblage II is recognized by the abundance of Midcontinent Fauna E species, particularly *Parapanderodus striatus*, *Microzarkodina*? *marathonensis*, *Bergstroemognathus extensus* Serpagli, *Fryxellodontus corbatoi* Serpagli, and *Diaphorodus delicatus.* It commences with the incoming of *Bergstroemognathus extensus*, *Diaphorodus russoi* (Serpagli), *Fryxellodontus corbatoi*, *Rossodus* n. sp., *Tripodus laevis* Bradshaw, and n. gen. 1 n. sp. A. Of these species, *B. extensus* and *F. corbatoi* are restricted to this assemblage; *D. russoi* and *Paroistodus parallelus* are only found in abundance in Assemblage II. *Drepanoistodus inaequalis* (Pander), *Clavohamulus* n. sp., ?*Protoprioniodus costatus* (van Wamel), ?*Protoprioniodus papilosus* (van Wamel), and *Protopanderodus gradatus* Serpagli are last found in Assemblage II.

This assemblage first appears in the middle Catoche Formation, and continues to the lower part of the upper strata of this formation.

The lowest conodont faunas recovered from the section north of Table Point are of Assemblage III affinity. These are 25 m stratigraphically above the highest levels of D.D.H. Al. The transition from assemblages II to III, recorded in drill core from the nearby Daniel's Harbour mine, corresponds with that described above, with Assemblage III close to the upper limits of this unsampled interval.

Assemblage III

Assemblage III is recognized by the occurrence of *Drepanodus concavus*, *Scalpellodus* n. sp., and both *Drepanoistodus basiovalis* (Sergeeva) and *D. forceps* (Lindström) as major components, together with elements of *Oistodus bransoni* Ethington and Clark.

Drepanoistodus basiovalis and *Oistodus bransoni* first occur early in Assemblage III; the incoming of *Parapanderodus* n. sp. (*Scolopodus* sp. aff. *S. gracilis* of Löfgren, 1978), and of *Scalpellodus* n. sp. occurs later in this assemblage. *Paltodus sweeti*, "*Scolopodus*" *filosus*, *Paroistodus parallelus*, *Microzarkodina*? *marathonensis*, *Parapanderodus striatus*, n. gen. 1 n. sp. B, *Diaphorodus russoi*, *Paracordylodus gracilis* Lindström, and n. gen. 1 n. sp. A are all last found within Assemblage III.

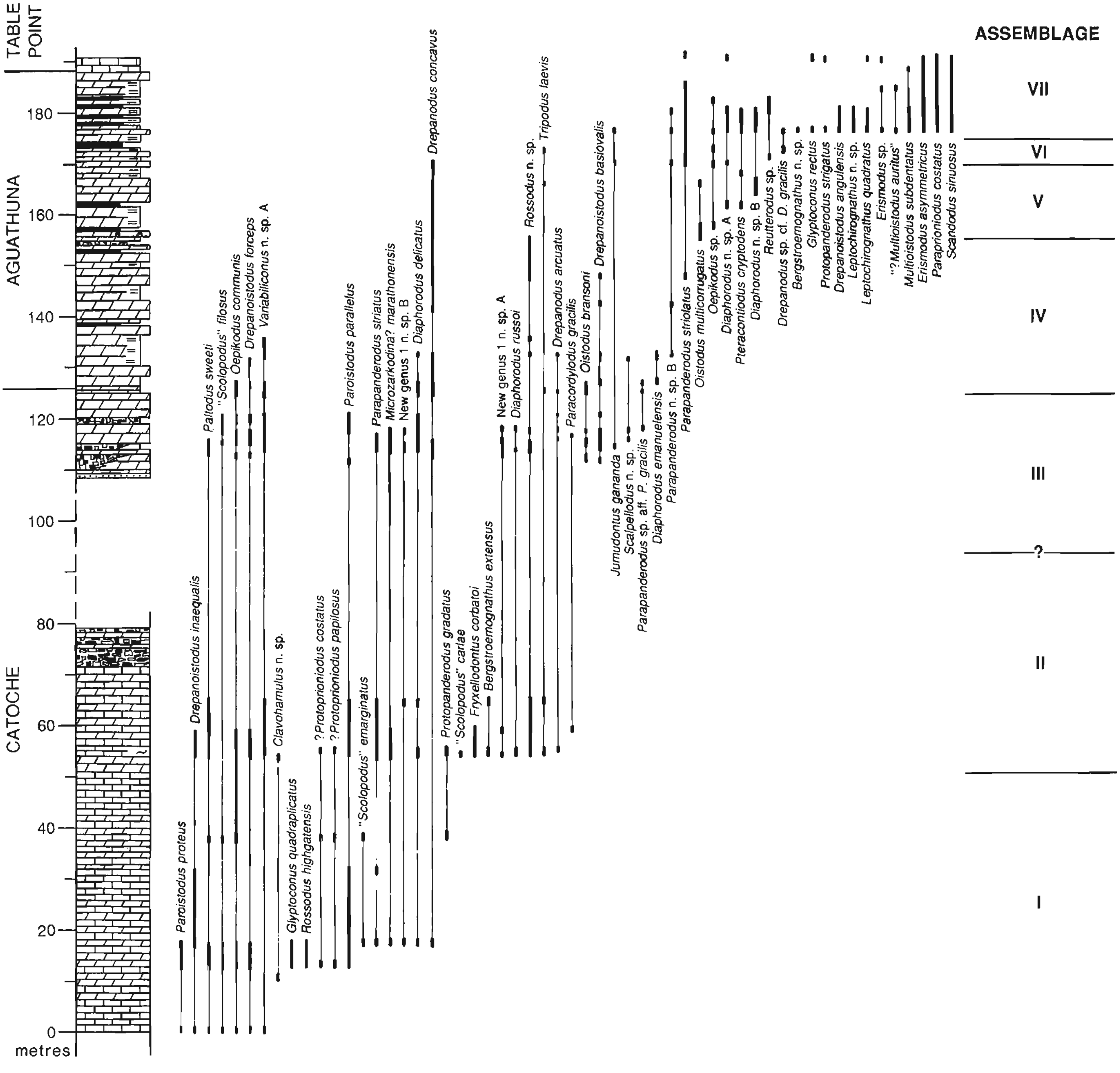

Figure 5. Range chart of upper St. George conodont species plotted against lithotypes (to left) and assemblages (to right of diagram). Lithological symbols are standard, with the exception of the mottled pattern used to indicate the level of pseudobreccias. Dashed line between upper and lower section represents stratigraphic gap between these two sequences. Total range of species within the St. George Group is indicated by a thin line; thick lines represent occurrence in consecutive samples. Squares at level of Table Point Formation are used to denote occurrence at some level within the Table Point Formation, but not necessarily at the level indicated.

Table 1. Fauna, stratigraphic level and correlation of upper St. George Group assemblages

Assemblage	Strat. level	Fauna	Correlatives
I	Lower to Middle Catoche Formation	**Oepikodus communis* *Paltodus sweeti* "*Scolopodus*" *filosus* *Drepanoistodus inaequalis* *Diaphorodus delicatus* *Glyptoconus quadraplicatus* **Parapanderodus striatus* *Rossodus highgatensis* "*Scolopodus*" *emarginatus*	Early M/C Fauna E; SF Zone H
II	middle to low in upper Catoche Formation	*Bergstroemog. extensus* *Diaphorodus russoi* *Paroistodus parallelus* *Fryxellodontus corbatoi* ?*Clavohamulus* n. sp. **Drepanoistodus forceps* *Microz.? marathonensis* **Diaphorodus delicatus*	M/C Fauna E; SF Zone H; Faunas A and B of San Juan Fm. of Argentina
III	upper Catoche Formation	*Parapanderodus* aff. *gracilis* *Oistodus bransoni* *Drepanoistodus basiovalis* *Scalpellodus* n. sp. *Variabiliconus* n. sp. A	M/C Fauna E; SF zone I; u. Jefferson City Dolomite
IV	uppermost Catoche and lower Aguathuna formations	*Diaphorodus emanuelensis* **Parapand.* n. sp. B #*Eucharodus parallelus* #*Oneotodus costatus* #*Drepanodus concavus* *Diaphorodus* n. sp. B *Parapanderodus striolatus*	M/C Fauna E/1
V	middle Aguathuna Formation	*Pteracontiodus cryptodens* *Oistodus multicorrugatus* *Oepikodus* sp. *Diaphorodus* n. sp. A *Diaphorodus* n. sp. B	M/C Fauna 1/2
VI	low in upper Aguathuna Formation	*Drepanodus* sp. cf. *D. gracilis* *Reutterodus* sp. (R) #*Oneotodus costatus*	M/C Fauna 1/2
VII	Upper Aguathuna and lower Table Point formations	*Paraprioniodus costatus* *Scandodus sinuosus* *Leptochirogn. quadrata* *Multioistodus subdentatus* *Drepanoistodus angulensis*	M/C Fauna 4; Kanosh and Lehman fms. of Utah

Conodont species are listed in order of abundance for each assemblage.
Species marked with an asterisk are also important in the succeeding assemblage.
M/C, Midcontinent conodont zonation; S/F, shelly fossil zones of the Ibex section;
#, acme of long-ranging species; (R), listed species was possibly reworked in assemblage.

Assemblage III is known from the upper Catoche Formation, and does not extend into the Aguathuna Formation. Transition to Assemblage IV faunas is gradual, commonly occurring within several metres of the base of the Aguathuna Formation.

Assemblage IV

Assemblage IV is characterized by lower conodont diversity and abundance than previous assemblages. Long ranging taxa dominate, particularly later in the assemblage. These include unusually high abundances of *Eucharodus parallelus*, *Oneotodus costatus* Ethington and Brand, *Drepanodus concavus*, ?*Semiacontiodus cordis* (Hamar), n. gen. 2 n. sp. A, and *Semiacontiodus asymmetricus* (Barnes and Poplawski).

Few species are first found in Assemblage IV. *Diaphorodus emanuelensis* (McTavish) is only known from early to middle Assemblage IV; *Parapanderodus* n. sp. B first occurs in mid-assemblage. Earliest specimens of *Parapanderodus striolatus* (Harris and Harris) immediately precede Assemblage V.

Many of the species continuing from Assemblage III disappear by mid-assemblage. These include *Drepanoistodus forceps*, *Oepikodus communis*, *Variabiliconus* n. sp. A, *Diaphorodus delicatus*, *Drepanodus arcuatus* Pander, *Oistodus bransoni*, *Parapanderodus* n. sp. (aff. *gracilis*), and *Scalpellodus* n. sp. *Drepanoistodus basiovalis* also disappears from the Table Point section before the emergence of Assemblage V.

Assemblage IV is recorded from the uppermost Catoche Formation and the lower Aguathuna Formation of the shore section north of Table Point. It immediately underlies a prominent solution surface; strata above this surface contain Assemblage V faunas.

Assemblage V

The first occurrence of *Diaphorodus* n. sp. A, together with the presence of *Pteracontiodus cryptodens* (Mound), *Oistodus multicorrugatus* Harris, and *Oepikodus* sp. are indicative of Assemblage V. These are joined by *Diaphorodus* n. sp. B at the Table Point section. Conodont faunas are of low diversity and abundance.

With the exception of *Rossodus* n. sp., long ranging forms of Assemblage IV continue through the Assemblage V into Assemblage VI, although many of these are not represented within Assemblage V.

An early Whiterockian age (Midcontinent Fauna 1 of Sweet et al. [1971] as discussed by Ethington and Repetski [1984, p. 97]) is indicated by the first appearance of species of *Diaphorodus* with *Pteracontiodus cryptodens* (Ethington and Repetski, 1984).

This assemblage is known from the middle of the Aguathuna Formation.

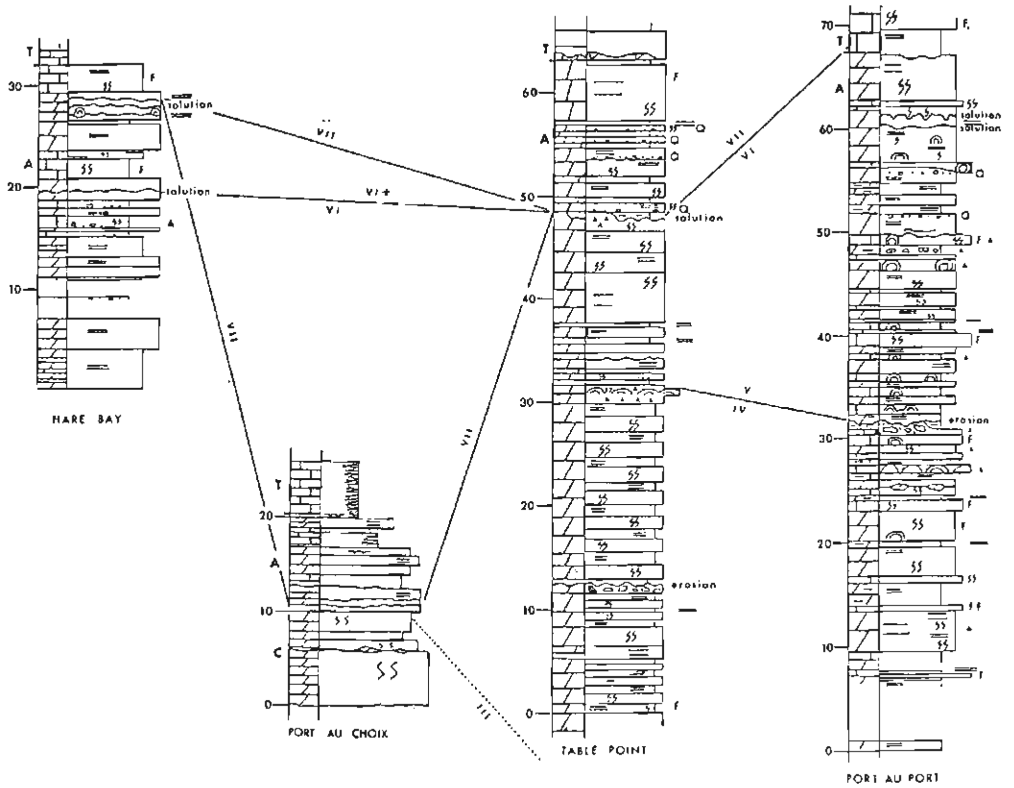

Figure 6. Biostratigraphic correlations between Port au Port, Table Point, Port au Choix, and Hare Bay sections. Coeval strata are joined by solid lines, with important assemblages marked. Dotted line is used to show correlation with a stratigraphic level lower than those shown in the diagram. Assemblage VI+ represents the faunal succession omitted from the Table Point sequence between Assemblage VI and Assemblage VII. Table Point-Aguathuna formation boundary shown by dashed line. Stylized sections of Port au Port and Table Point sections modified from Knight and James (1987).

Assemblage VI

This asemblage is distinguished from previous assemblages by the first appearance of *Drepanodus* sp. cf. *D. gracilis* and *Reutterodus* sp., and last occurrence of *Drepanodus concavus* and *Tripodus laevis*. This indicates an age equivalenat at least to that of Midcontinent Fauna 2. The base of this assemblage records the last appearance of conodonts of Canadian (Ibexian) aspect. Conodonts remain low in abundance and diversity, with faunas dominated by *Oneotodus costatus*.

Reworking of the conodonts is suggested by the occurrence of *Reutterodus* sp. and *Oneotodus costatus* with latest *Tripodus laevis*. It is, however, possible that the range of *O. costatus* is extended in very shallow environments.

Assemblage VI is found within a brief interval, low in the upper part of the Aguathuna Formation.

Assemblage VII

The incoming of a highly abundant, diverse fauna marks the beginning of Assemblage VII. *Drepanoistodus angulensis* (Harris), *Erismodus asymmetricus* (Branson and Mehl), *Leptochirognathus quadratus* (Branson and Mehl), "*?Multioistodus auritus*" (Harris and Harris) sensu Ethington and Clark, *M. subdentatus* Cullison, *Paraprioniodus costatus* Ethington and Clark, *Protopanderodus strigatus* Barnes and Poplawski, *Scandodus sinuosus* Mound, and endemic species are all found in the lowest sample of this assemblage, and continue through to the uppermost part of the Aguathuna Formation or into the Table Point Formation.

Jumudontus gananda Cooper, *Oepikodus* sp., *Oneotodus costatus*, *Parapanderodus* n. sp., *Pteracontiodus cryptodens* (Mound), *Reutterodus* sp., and n. gen. 2 n. sp. A do not continue into the Table Point Formation (Stouge, 1984). Anomalously high ranges of *J. gananda*, *O. costatus*, *P. cryptodens*, and *Reutterodus* sp. indicate reworking of conodonts into the upper Aguathuna Formation from lower strata.

This assemblage is found in the uppermost 15 m of the Aguathuna Formation and basal Table Point Formation. It frequently overlies an omission surface, and the lowest level of this assemblage marks a distinct change in conodont fauna.

Conodont species known from the Ibex section are indicative of a Midcontinent Fauna 4 age, with the exception of *Erismodus asymmetricus*, which is known from Midcontinent Fauna 5 and younger strata. The presence of *Multioistodus subdentatus* is suggestive also of Midcontinent Fauna 5, but this is known from older rocks. It is likely that the range of *E. asymmetricus* is extended downward in western Newfoundland.

REGIONAL ASPECTS

In order to test the validity of this conodont biostratigraphy, samples of primarily Aguathuna Formation rock types were collected from Big Spring Inlet of Hare Bay, the mainland shore of Back Arm Bay, at the Gravels section at the southernmost point of East Bay on Port au Port Peninsula, and road and quarry sections at Aguathuna (Fig. 1). Port au Port and Hare Bay sections were deposited in settings outboard of the setting recorded at Table Point (Knight and James, 1987).

Big Spring Inlet samples contained Assemblage IV to VII faunas, although the incoming of Assemblage VII species is not abrupt as north of Table Point. The sequence of first appearance of Assemblage VII conodonts indicates that some of the strata missing from the section north of Table Point are preserved at Big Spring Inlet.

At Back Arm Bay, Assemblage VII overlies Assemblage III: most of the Aguathuna Formation is missing at Back Arm Bay, with only that part deposited above the uppermost erosion level at Table Point preserved.

Strata from a section west of The Gravels at Port au Port Peninsula commenced within Assemblage IV, in the lower part of the Aguathuna Formation. Uppermost strata of this formation contain Assemblage VI. Assemblage VII is not found until three metres into the conformably overlying Table Point Formation. A similar succession is found in the Aguathuna Quarries, although the Aguathuna-Table Point boundary is demonstrably erosional (Knight and James, 1987).

Correlation throughout western Newfoundland

Biostratigraphic relationships between the Port au Port, Table Point, Port au Choix, and Hare Bay sections are illustrated in Figure 6. It can be seen that the Aguathuna Formation on Port au Port Peninsula terminates earlier than similar sediments of the Table Point-Daniel's Harbour region. All but the uppermost Aguathuna Formation at Back Arm Bay near Port au Choix is missing from the section; uppermost Aguathuna strata rest disconformably upon uppermost Catoche Formation strata in this area. At Hare Bay (Big Spring Inlet), a solution horizon approximately 13 m beneath the base of the Table Point Formation represents a stratigraphic break of two local assemblages. Deposition, however, resumed earlier at Hare Bay than near Table Point.

Assemblages within the Catoche Formation are generally recognizable, where sampled, throughout the Great Northern and Port au Port peninsulas. On the basis of lithological evidence (Knight and James, 1987), the succession of faunas in the Aguathuna Formation is probably disrupted within Assemblage IV (the ''Breccia Bed'' of Stouge, 1982), between Assemblages IV and V, and between Assemblages VI and VII. However, no significant break in the faunal succession is recorded at the level of the Breccia Bed. Stratigraphic level and conodont faunas of the transition from Assemblage IV to V are everywhere consistent, and the erosion surface separating these two faunas is not likely to represent a major hiatus. It is apparent that a regional regression affected all areas simultaneously (Barnes, 1984) prior to the deposition of Whiterockian sediments of Assemblage V. The Assemblage VI to VII transition is diachronous, and more long-lasting at Table Point than Port au Port or Big Spring Inlet.

CONCLUSIONS

In summary, the faunas of the upper lower Catoche and Aguathuna formations may be subdivided into seven assemblages. Their diagnostic species, stratigraphic level within the St. George Group, and probable correlations are summarized in Table 1.

The proposed biostratigraphic zonation, based upon the Table Point-Daniel's Harbour region, is recognized in other sections of the upper St. George Group throughout western Newfoundland. Comparison of the conodont succession throughout western Newfoundland indicates an omission surface in the middle part of the Aguathuna Formation, which is contemporaneous in inboard (Table Point) and outboard (Port au Port) settings. Subsequent strata contain faunas that do not correlate exactly across the region; an upper omission surface in the Aguathuna Formation is diachronous both beneath and above the omitted strata (Fig. 6); the base of the Table Point Formation is also diachronous. This interval also contains evidence of reworking of sediments.

The detailed conodont biostratigraphic framework for the upper St. George Group developed herein should enable further insight into tectonic development of the carbonate platform immediately prior to its foundering.

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Biogeography of Ordovician Chitinozoa

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Abstract

Biogeographic differentiation of chitinozoans may help explain some of the problems associated with correlating the microfaunas of Quebec with those of southwest Europe, North Africa, and the Baltic region of Scandinavia. This differentiation becomes evident when comparisons are made with the respective regional zonation, vertical stratigraphic distribution charts, and microfaunal assemblages.

An assessment of the number of species in common between chitinozoan assemblages of different regions, and the similarity coefficient thus obtained, indicates affinities among the microfaunas of North Africa, southwest Europe, Great Britain, and Bohemia; these regions were situated in high latitudes during the Ordovician. Similarly, microfaunas from eastern Canada, Australia, Spitsbergen, and the United States, which occupied subequatorial latitudes, have many feature in common. Faunas from the Baltic region of Scandinavia appear to have occupied an intermediate position.

Résumé

Les difficultés rencontrées lors des essais de corrélation entre les microfaunes de chitinozoaires ordoviciennes du Québec et celles de l'Europe du sud-ouest, d'Afrique du Nord et de la Baltoscandinavie s'expliquent par une différenciation biogéographique des chitinozoaires. Cette différenciation est mise en évidence en comparant les zonations régionales, les tableaux stratigraphiques et l'ensemble de la microfaune de ces régions.

La détermination du nombre d'espèces communes entre les assemblages de chitinozoaires de diverses régions et l'analyse des coefficients de similitude s'y rapportant révèlent des affinités entre les microfaunes d'Afrique du Nord, d'Europe du Sud-Ouest, de Grande-Bretagne et de Bohême qui étaient localisées sous les hautes latitudes. L'Est du Canada, l'Australie, le Spitsberg et les États-Unis, qui occupaient une position subéquatoriale, présentent également des liens entre leurs microfaunes. La Baltoscandinavie semble occuper une situation intermédiaire entre ces deux ensembles.

INTRODUCTION

The study of Ordovician strata of Quebec led to the recognition of 22 chitinozoan assemblages (Achab, 1989). The strata containing the microfossils on which the zonation is based may be correlated with the standard North American graptolite zones. This work shows that Quebec microfaunas were significantly different from those of similar age in Europe and the Baltic region of Scandinavia, hereafter referred to as Baltoscandia. These differences, which appear to be related to the geographic distribution, have been investigated to determine if, in fact, there is provincialism. Little previous work has been done on the provincialism of chitinozoans.

Rauscher (1973) compared the assemblages he had studied in France with the data then available for Ordovician chitinozoans. In the Arenig, he recognized a southern assemblage group, including assemblages from the Sahara, France, and Belgium, and a northern assemblage group, occurring in Great Britain, Baltoscandia, and the Russian platform. The analysis of microfaunas from the Llanvirn indicated a third intermediate group between the Sahara and southern Europe. Laufeld (1979), describing the geographical distribution of Ordovician chitinozoans, came to conclusions similar to those of Jenkins (1967). These authors favoured a fairly free communication, especially during the Caradoc, between the chitinozoans of Shropshire, and those of Baltoscandia. The relationship between the chitinozoans of northern

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Europe and those of North Africa is, in contrast, more limited. Laufeld (1979, p. 78) pointed out that, although it might be premature to introduce the concept of chitinozoan provinces, certain genera are found mainly in the "North Atlantic Area," whereas others occur in a southern province. Furthermore, he stressed the difficulty of correlating the strata between these provinces.

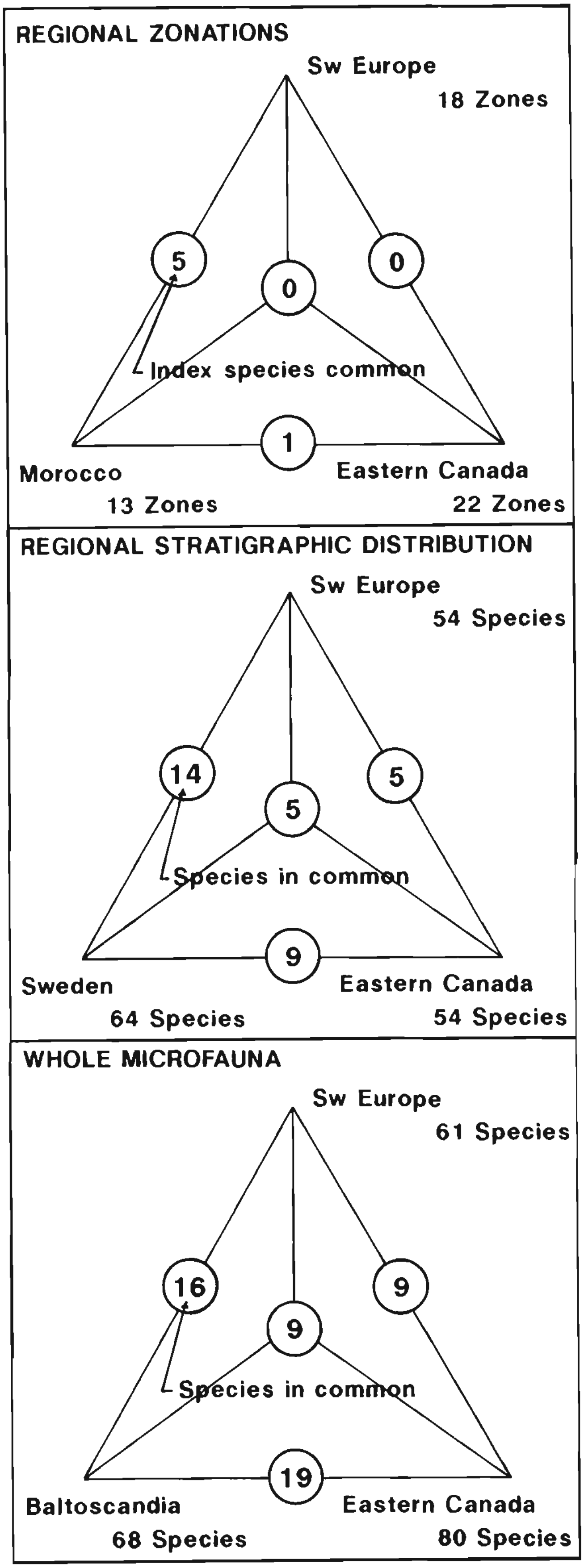
Paris (1981), on the other hand, approached the problem of provincialism firstly at the generic level, and secondly at the specific level. He believed that any selective geographic distribution of chitinozoan species, and, similarly, genera, could not be substantiated. Moreover, he considered that some of the differences recorded were the result of variations in sampling techniques, and concluded that it would not be valid to use chitinozoans to assess similarities, or differences, between various regions.

During the last decade, numerous chitinozoan assemblages from well dated sections in different parts of the world have been described. These new data have provided a better understanding of geographic distribution. Achab (1988), from a limited selection of taxa, showed that the geographic distribution of Ordovician chitinozoans was apparently related to paleolatitude. The aim of the present study is to make an analysis of the known global distribution of the microfaunas.

METHODS

When comparing microfaunas of a geological system, the approximate similarity between assemblages of two regions may be obtained by comparing the respective zonation. This comparison, because it is based on only one or two index species when the total assemblage generally contains seven to eight, is biased, but it does provide a first approximation of the degree of similarity or difference between microfaunas of two regions. This estimate needs to be documented precisely, because the choice of index species is subjective. It is possible that two zones based on similar microfaunas, but described by two different authors, may be designated by different zone index species names; conversely, the presence of an index species in two assemblages does not necessarily imply similarity in overall composition.

The first approximation can be subsequently refined, by comparing selected species from two regions that represent the stratigraphic distribution of chitinozoans in a given time interval. In addition to the index species, selected taxa, which characterize regional assemblages by their abundance and their stratigraphic qualities, should be taken into account. This comparison has the advantage of being based on well defined selected taxa, which are representative of evolutionary change occurring in the assemblages of two regions. This approach serves to assess the degree of similarity that exists between two, or several, regions. However, an even more precise evaluation may be obtained by taking into account the complete assemblage of species in the region, during a given time interval.



It is also possible to consider, in a more global sense, the assemblage of chitinozoan species known from different regions. All of these data can then be treated in order to determine the degree of similarity existing between regions, and to make conclusions concerning the geographic distribution of microfaunas.

Comparison of regional zonation

There are few publications proposing a regional chitinozoan zonation for the Ordovician System. The first was that of Paris (1981) for the Ordovician of southwest Europe. More recently, Achab (1989) established a regional zonation for the Ordovician of Quebec and western Newfoundland. The zonation of Elaouad-Debbaj (1987), although less complete, provides interesting data on the Ordovician of Morocco.

These zonations not only deal with the whole Ordovician, but have the advantage of being based on well dated sequences closely correlated with graptolite zones; this allows comparison of more or less synchronous zones.

From a comparison of these chitinozoan zones, it appears (Fig. 1) that of the 18 defined by Paris (1981), and the 13 recognized by Elaouad-Debbaj (1987), 5 are designated by the same index species. Five of the zones, which have different names, have been defined in Morocco from stratigraphic intervals that in Europe have either not been studied or recognized, or which lack chitinozoans. The three remaining zones are from intervals whose stratigraphic position is not certain; two of the zones are designated by species known to occur in southwest Europe.

If a comparison is made between the zonation of Paris (1981), and that proposed by Achab (1989) for Quebec, it becomes clear that none of the taxa designated as characterizing southwest European zones is the same as the Quebec index species. Also, it appears that the index species used in Europe, and in Morocco, are little known, or unknown, in eastern Canada.

At first sight, it thus appears that there is a marked difference between the Quebec assemblages and those of Morocco and Europe.

Comparison of the regional distribution of chitinozoans

Paris (1981), Grahn (1982), and Achab (1989) provide syntheses of the vertical stratigraphic distribution of the main chitinozoans for southwest Europe, Sweden, and Quebec, respectively. Comparison of the vertical stratigraphic distribution is facilitated by the fact that in each of the regions, an almost identical number of taxa (50 to 60) were selected to illustrate chitinozoan distribution during the Ordovician. Also, as mentioned above, there is good stratigraphic control for the chitinozoan-bearing strata, thus enabling comparison of synchronous time units.

Comparison of stratigraphically important species indicates that five species are common to all three regions (Fig. 1). When the regions are compared two at a time, it is evident that no additional species are common to southern Europe and Quebec, but nine are common to southern Europe and Sweden; this brings the total number of species common to these two regions to 14. There are a total of nine species common to Sweden and eastern Canada. The results suggest a clear distinction between the microfaunas of the three regions. To make the results meaningful, one can also take into account the greater contrast between Canada and southwest Europe, and the fact that faunally, Baltoscandia occupies an intermediate position. The difference between the European and Canadian microfaunas corresponds with that noted above, when only the zonal index fossils were considered.

Comparison of the total regional microfauna

The data summarized above can be supplemented by that obtained from the remaining palynological literature, and includes well dated, but more restricted, stratigraphic intervals in Europe (Rauscher, 1968; Rauscher and Doubinger, 1967a, b), Baltoscandia (Eisenack, 1931; Laufeld, 1967; Männil, 1971), and eastern Canada (Jansonius, 1964; Martin, 1975, 1983; Neville, 1974). These data, integrated with those previously discussed, include 209 records of 166 species that have been described from the three regions. Four additional species occur in Baltoscandia, southwest Europe, and eastern Canada, making a total of nine species in common among the three regions. When the regions are compared two at a time, the number of species in common (Fig. 1) are 16 in Baltoscandia and southwest Europe, and 19 in Baltoscandia and eastern Canada, while the nine species common to Canada and Europe also exist in Baltoscandia. These results compare well with those described in the previous section. It appears that southwest Europe and eastern Canada have distinct microfaunas, and that Baltoscandia has as many affinities with Europe as it does with North America.

At whatever level the comparisons are made, the results are consistent; namely, chitinozoans, like other fossils, show important biogeographic differentiation during the Ordovician. Paleogeographic reconstructions for continents during the Ordovician indicate that the eastern part of North America occupied a subequatorial position, whereas southern Europe was situated in low latitudes and Baltoscandia was located in an intermediate position. Biogeographic distribution of chitinozoans therefore seems to have been related to paleolatitude.

SIMILARITY BETWEEN ORDOVICIAN MICROFAUNAS

Although the approach outlined above might be used for Ordovician chitinozoan assemblages, problems may arise when the relative importance of stratigraphic levels, and thus

Figure 1 (opposite). Number of species common to different regions when taking into consideration — 1: regional zonation; 2: selected species from vertical stratigraphic distribution charts; 3: the whole microfauna.

the number of species described, differ in respective regions. For example, only the Lower Ordovician has been studied in Spitsbergen and Bohemia, whereas in England there are no data for this particular interval. For the comparisons to be reliable, successive time intervals should be studied. One should first consider the Lower Ordovician, then the Middle, followed by the Upper. From the species lists available for different regions, it is then possible to determine the number of species common to two regions, and to calculate the resultant similarity coefficient. The analysis of the number of species in common, and the similarity coefficients, allows a determination to be made regarding the degree of affinity between different regions. The similarity coefficients most frequently used are those of Simpson, of Jaccard, and of Dice (Gradstein et al., 1985). These authors pointed out that the coefficient of Simpson accentuates similarities, whereas that of Jaccard emphasizes differences. Intermediate is that of Dice, and this coefficient was used in the present study. It may be expressed by $I = 2Cij/Ni + Nj$ where Cij = number of species common to the two regions and Ni and Nj = number of species from regions i and j.

To support the contention that the similarity between certain microfaunas is related to Ordovician paleogeography, the spatial distributions of certain chitinozoan species were plotted on paleogeographic reconstructions for the Lower and Upper Ordovician (Figs. 2, 3).

The distribution of chitinozoan species, or genera, shown on each of the maps, takes into account their numerical and stratigraphic importance, thereby giving a typical character to the assemblages.

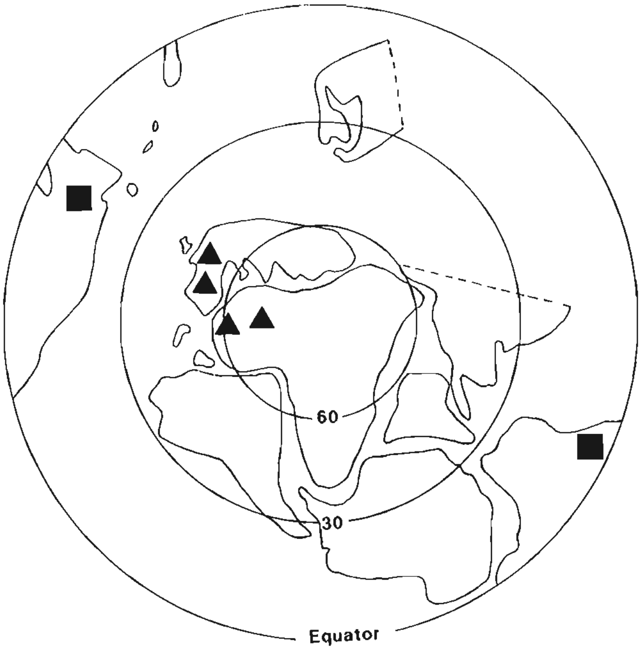


Figure 2. Lower Ordovician chitinozoan assemblages containing *Eremochitina baculata* (triangle) and *Fustichitina langei* (square); map from Whittington and Hughes (1972).

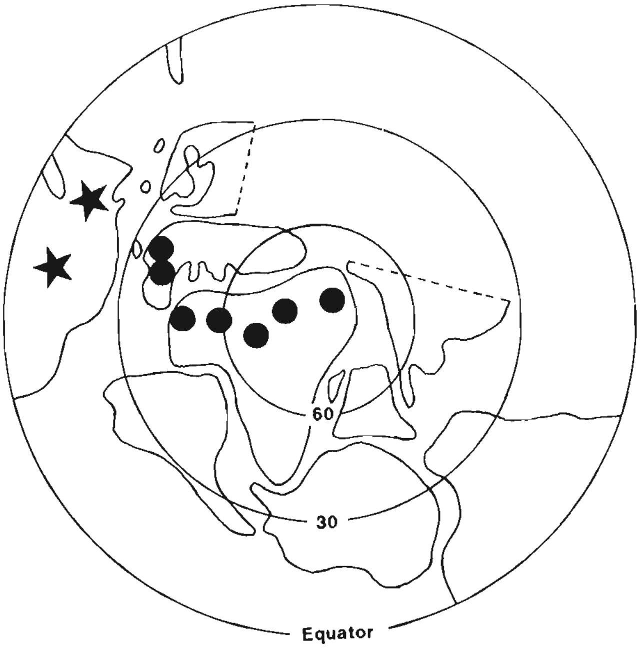


Figure 3. Distribution of Upper Ordovician chitinozoan assemblages containing *Hercochitina* sp. (star) and *Armoricochitina nigerica* (dot); map from Whittington and Hughes (1972).

Lower Ordovician

The data used above have been supplemented by those from relatively limited time intervals of the Lower Ordovician including: Spitsbergen (Bockelie, 1980), Australia (Combaz and Péniguel, 1972; Achab and Millepied, 1980), the United States (Hutter, 1987), Bohemia (Paris and Mergl, 1984), and North Africa (Taugourdeau and de Jekhowsky, 1960; Benoit and Taugourdeau, 1961; Molyneux and Paris, 1985; Elaouad-Debbaj, 1987). From a preliminary appraisal of the largest number of species common to the different regions (Table 1), it is clear that southwest Europe, North Africa, and Bohemia are part of a single entity. Nine of the 16 European species, and five of the six Bohemian species are, in fact, present in Africa. Included in the table are affinities between the microfaunas of eastern Canada and Australia, and those between the United States and Australia, as outlined by Achab (1982) and Hutter (1987). For Baltoscandia, only the data presented by Grahn (1982) concerning chitinozoans of Hunderum and Langevoja age (late Arenig), have been included.

The comparison of the similarity coefficients (Table 2) shows a consistent phenomenon. If one selects, in each of the regions, the two highest values of similarity coefficient, it is again clear that southwest Europe, North Africa, and Bohemia have much in common. On the other hand, Canada, Australia, the United States, and Spitsbergen make up a separate entity. Baltoscandia appears to bridge these two areas.

Detailed analysis of chitinozoan microfaunas shows that in the Lower Ordovician, there are a number of species with wide geographic distribution, reported from several locations. Of these, *Conochitina symmetrica* Taugourdeau and

Table 1. Number of chitinozoan species common to different regions of the Lower, Middle, and Upper Ordovician (regions showing the highest number of species in common are shown in grey)

LOWER ORDOVICIAN	SW EUROPE	AFRICA	BOHEMIA	BALTOSCANDIA	EASTERN CANADA	AUSTRALIA	UNITED STATES	SPITSBERGEN
SW EUROPE	16	9	3	0	1	0	0	0
AFRICA	9	23	5	3	5	1	0	1
BOHEMIA	3	5	6	2	3	1	0	1
BALTOSCANDIA	0	3	2	15	1	0	0	1
EASTERN CANADA	1	5	3	1	13	5	2	3
AUSTRALIA	0	1	1	0	5	6	2	1
UNITED STATES	0	0	0	0	2	2	2	1
SPITSBERGEN	0	1	1	1	3	1	1	5
MIDDLE ORDOVICIAN	SW EUROPE	AFRICA	GREAT BRITAIN	BALTOSCANDIA	EASTERN CANADA	AUSTRALIA	UNITED STATES	
SW EUROPE	30	10	7	8	1	2	2	
AFRICA	10	13	4	2	0	1	1	
GREAT BRITAIN	7	4	11	5	2	2	1	
BALTOSCANDIA	8	2	5	34	2	5	5	
EASTERN CANADA	1	0	2	2	16	2	5	
AUSTRALIA	2	1	2	5	2	11	2	
UNITED STATES	2	1	1	5	5	2	10	
UPPER ORDOVICIAN	SW EUROPE	AFRICA	GREAT BRITAIN	BALTOSCANDIA	EASTERN CANADA	UNITED STATES		
SW EUROPE	20	6	4	10	6	5		
AFRICA	6	23	6	6	11	6		
GREAT BRITAIN	4	6	11	7	6	4		
BALTOSCANDIA	10	6	7	45	13	15		
EASTERN CANADA	6	11	6	13	47	15		
UNITED STATES	5	6	4	15	15	23		

de Jekhowsky, 1960, *Conochitina decipiens* Taugourdeau and de Jekhowsky, 1960, and *Lagenochitina esthonica* Eisenack, 1955 are selected because they have a limited vertical stratigraphic range and they enable correlation to be made over large distances.

In addition, this analysis shows the existence of species of regional stratigraphic interest, with a more restricted geographic distribution that appears related to Ordovician paleolatitudes. In the regions marginal to Gondwanaland, which occupied high latitudes, one notes the constant presence of species characterized by differentiation of the periderm, such as *Eremochitina baculata* Taugourdeau and de Jekhowsky, 1960, *Velatachitina veligera* Poumot, 1968, and *Desmochitina* sp. (species possessing a velum). These taxa are associated with *Tanuchitina achabae* Paris, 1981, *Sagenachitina oblonga* Benoit and Taugourdeau, 1961, and *Laufeldochitina baculiformis* (Rauscher, 1968), species which seem equally restricted to high latitudes (Fig. 2).

In low latitudes, the microfauna is characterized by large chitinozoans with a simple shape and without periderm, such as *Fustichitina langei* (Combaz and Péniguel, 1972), *Conochitina poumoti* (Combaz and Péniguel, 1972), and *Conochitina pirum* (Achab, 1982).

Middle Ordovician

Species described by Combaz and Péniguel (1972) from Australia, by Jenkins (1967) from Great Britain, and by Grahn and Bergström (1984), Grahn and Miller (1986), and Hutter (1987) from the United States, have been integrated.

Table 1 shows that the differentiation outlined for the Lower Ordovician continues into Middle Ordovician times. Europe and Africa have the most species in common, and of the 11 species known in Great Britain, seven exist in Europe. Canada has more affinities with the United States, which shows, however, the same number of elements in common

Table 2. Similarity coefficients linking different regions of the Lower, Middle, and Upper Ordovician (regions with the highest similarity coefficient shown in grey)

LOWER ORDOVICIAN	SW EUROPE	AFRICA	BOHEMIA	BALTOSCANDIA	EASTERN CANADA	AUSTRALIA	UNITED STATES	SPITSBERGEN
SW EUROPE	1.000	0.462	0.273	0.000	0.069	0.000	0.000	0.000
AFRICA	0.462	1.000	0.345	0.158	0.278	0.069	0.000	0.071
BOHEMIA	0.273	0.345	1.000	0.190	0.316	0.167	0.000	0.182
BALTOSCANDIA	0.000	0.158	0.190	1.000	0.071	0.000	0.000	0.100
EASTERN CANADA	0.069	0.278	0.316	0.071	1.000	0.526	0.267	0.333
AUSTRALIA	0.000	0.069	0.167	0.000	0.526	1.000	0.500	0.182
UNITED STATES	0.000	0.000	0.000	0.000	0.267	0.500	1.000	0.286
SPITSBERGEN	0.000	0.071	0.182	0.100	0.333	0.182	0.286	1.000

MIDDLE ORDOVICIAN	SW EUROPE	AFRICA	GREAT BRITAIN	BALTOSCANDIA	EASTERN CANADA	AUSTRALIA	UNITED STATES
SW EUROPE	1.000	0.465	0.341	0.250	0.043	0.098	0.100
AFRICA	0.465	1.000	0.333	0.085	0.000	0.083	0.087
GREAT BRITAIN	0.341	0.333	1.000	0.222	0.148	0.182	0.095
BALTOSCANDIA	0.250	0.085	0.222	1.000	0.080	0.222	0.227
EASTERN CANADA	0.043	0.000	0.148	0.080	1.000	0.148	0.385
AUSTRALIA	0.098	0.083	0.182	0.222	0.148	1.000	0.190
UNITED STATES	0.100	0.087	0.095	0.227	0.385	0.190	1.000

UPPER ORDOVICIAN	SW EUROPE	AFRICA	GREAT BRITAIN	BALTOSCANDIA	EASTERN CANADA	UNITED STATES
SW EUROPE	1.000	0.279	0.258	0.308	0.179	0.233
AFRICA	0.279	1.000	0.353	0.176	0.314	0.261
GREAT BRITAIN	0.258	0.353	1.000	0.250	0.207	0.235
BALTOSCANDIA	0.308	0.176	0.250	1.000	0.283	0.441
EASTERN CANADA	0.179	0.314	0.207	0.283	1.000	0.429
UNITED STATES	0.233	0.261	0.235	0.441	0.429	1.000

with Baltoscandia. This last region seems to occupy an intermediate position, having nearly as many features in common with southwest Europe, Great Britain, and North America.

The similarity coefficients (Table 2) show the same differentiation, thus confirming on the one hand the affinities between Europe, Africa, and Great Britain, but on the other, affinities between Canada, the United States, and Australia. The intermediate position of Baltoscandia is again reflected by the distribution of the similarity coefficients.

In the Middle Ordovician, there occur species such as *Conochitina hirsuta* Laufeld, 1967 and *Cyathochitina calix* Eisenack, 1931, which, since they have been described in equivalent stratigraphic intervals in Europe, Baltoscandia, and North America, are useful for correlation over large distances. The microfaunas of these regions also contain taxa with long vertical stratigraphic ranges extending into the Upper Ordovician: e.g., *Desmochitina minor* Eisenack, 1931, *Cyathochitina campanulaeformis* Eisenack, 1931, *Cyathochitina kuckersiana* Eisenack, 1931, *Belonechitina micracantha* Eisenack, 1931, *Belonechitina robusta* (Eisenack, 1959), and *Rhabdochitina magna* Eisenack, 1931.

Species of regional stratigraphic value, restricted to the Gondwanaland margin, are: *Siphonochitina clavata* Jenkins, 1967, *Siphonochitina robusta* Jenkins, 1967, *Linochitina pissotensis* Paris, 1981, and *Laufeldochitina lardeuxi* Paris, 1981. The intertropical regions are typified by *Conochitina poumoti* Combaz and Péniguel, 1972, *Conochitina subcylindrica* Combaz and Péniguel, 1972 and *Lagenochitina* sp. A Achab, 1985.

Upper Ordovician

Palynological data for the Upper Ordovician are relatively abundant. References used for the present study include Männil (1971) for Europe; Jenkins (1967) for Shropshire (Great Britain); Stauffer (1933), Jenkins (1969, 1970), Miller (1976), and Hart (1986) for the United States. It should be pointed out, however, that few Upper Ordovician assemblages have been described from southwest Europe or North Africa. An evaluation of the number of species common to these regions (Table 1) seems to indicate that the differences previously observed are less marked. It is with Baltoscandia that southwest Europe has the most in common, Africa having more in common with Canada, and Baltoscandia more in common with North America and Europe.

The same conclusion is reached from analysis of the similarity coefficients. It is important to note that although southwest Europe, North Africa, and Great Britain have much in common, their microfaunas nevertheless contain a certain number of elements in common with Baltoscandia and Canada.

Species with wide geographic distribution include: *Desmochitina minor, Belonechitina micracantha, Belonechitina robusta, Belonechitina wesenbergensis* Eisenack, 1959, and *Cyathochitina kuckersiana* Eisenack, 1934. Most of these already existed in the Middle Ordovician. Certain taxa show a more limited stratigraphic distribution, and may be used for intercontinental correlation. Of particular importance are *Acanthochitina barbata* Eisenack, 1931, *Lagenochitina baltica* Eisenack, 1931, *Plectochitina sylvanica* (Jenkins, 1970), *Spinachitina bulmani* (Jansonius, 1964), and *Belonechitina taugourdeaui* Eisenack, 1968). The differentiation between microfaunas of low and high latitudes persists. Certain species, such as *Calpichitina lenticularis* (Bouché, 1965) and *Armoricochitina nigerica* (Bouché, 1965) are restricted to, and are characteristic of, microfaunas from the areas marginal to Gondwanaland; others, such as *Kalochitina multispinata* Jansonius, 1964 and species of the genus *Hercochitina* Jansonius, 1964 are, in contrast, only known in North American assemblages (Fig. 3).

CONCLUSIONS

Analysis of the species common to different zonation schemes, tables of vertical stratigraphic distribution, and complete assemblages of Ordovician chitinozoans from different regions, confirm a clear differentiation between the assemblages of the north Gondwana margin (North Africa, southwest Europe, Great Britain) and those of low latitude regions (eastern Canada, the United States, Australia, Spitsbergen). The differences help explain the difficulties palynologists experience when attempting to correlate assemblages from regions located at different paleolatitudes. The differentiation estimated by the simple determination of the number of species common to different regions, and the calculation of the similarity coefficient, appears to confirm work in progress, which utilizes more sophisticated methods of data analysis, such as correspondence analysis and ascending hierarchical classification.

The biogeographic differentiation of Ordovician chitinozoans is similar to that observed for other fossil groups, especially trilobites, conodonts, and graptolites. As with these groups, it would seem necessary to establish, and use, a separate zonation for each faunal province.

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Late Ordovician extinction in the Graptoloidea

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Abstract

During the Late Ordovician and Early Silurian, the composition of planktic graptolite faunas underwent a great change. Late Ordovician graptolites experienced nearly total extinction, with the exception of a few species with Pattern H astogeny (e.g., N. normalis, N. angustus, *and* N. extraordinarius). *None of the other middle Ashgill diplograptaceans survived into the* persculptus *Zone and on into the early Llandovery. Post-extinction morphological radiation developed entirely from these few species and was already well underway in the* persculptus *Zone.*

The Late Ordovician graptolite faunas do not consist only of holdovers from earlier, more vigorous lines; rather, they are dominated by rapidly evolving clades that experienced substantial increases in diversity during the Ashgill. These lineages are represented by the groups Climacograptus longispinus–C. venustus, Paraorthograptus pacificus–Amplexograptus latus, *and* "Glyptograptus" tenuissimus–"G." occidentalis, *as well as several dicranograptid and lasiograptid groups. However, all of these lineages (some 23 of 24 genera and three of the four extant families) became extinct within the middle Ashgill, prior to or during the interval occupied by the* Hirnantia *Fauna.*

The Ashgill graptolite extinctions coincide approximately with those among many other benthic to holoplanktic organisms. These extinctions are associated with increased oxygenation and possibly with cooling of basinal water masses, as well as regression and habitat restriction brought about by peak glaciation during Hirnantian time. It is suggested that the oxygenation of the basinal water masses resulted in disruption of the plankton communities and was a prime factor in causing graptolite extinction.

Revisions are made to the genera Normalograptus *and* Glyptograptus *on the basis of their astogenetic patterns.*

Résumé

L'Ordovicien supérieur et le Silurien inférieur ont été marqués par un changement important de la composition des faunes de graptolites planctoniques. Les graptolites de l'Ordovicien supérieur ont connu une extinction quasi totale à l'exception de quelques espèces de l'astogenèse de la configuration H (par ex. N. normalis, N. angustus *et* N. extraordinarius*). Aucun des autres diplograptacéens de l'Ashgillien moyen n'a survécu dans la zone à* persculptus *et jusque dans le Llandovérien inférieur. Le rayonnement morphologique postérieur à l'extinction s'est amorcé entièrement à partir de ces quelques espèces et était déjà bien en cours dans la zone à* persculptus.

Les faunes de graptolites de l'Ordovicien supérieur ne comprennent pas seulement des survivants de lignées anciennes plus vigoureuses; elles sont plutôt composées en grande partie de clades à évolution rapide qui se sont diversifiées substantiellement au cours de l'Ashgillien. Ces lignées sont représentées par les taxons Climacograptus longispinus - C. venustus, Paraorthograptus pacificus - Amplexograptus latus *et* "Glyptograptus" tenuissimus - "G." occidentalis *ainsi que par plusieurs groupes de dicranograptidés et de lasiograptidés. Cependant, toutes ces lignées (quelque 23 des 24 genres et trois des quatre familles conservés) ont disparu au sein de l'Ashgillien moyen, avant ou pendant l'intervalle occupé par la faune* Hirnantia.

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L'extinction des graptolites ashgilliens a coïncidé à peu près avec celle de nombreux autres organismes benthiques à holoplanctoniques. Ces extinctions sont associées à une augmentation de l'oxygénation et peut-être à un refroidissement des masses d'eau de bassin ainsi qu'à une régression et une limitation de leur habitat causées par une glaciation maximale au cours de l'Hirnantien. Il est proposé que l'oxygénation des masses d'eau de bassin a provoqué une perturbation des communautés planctoniques et que ce phénomène a été le principal facteur de l'extinction des graptolites.

Des révisions sont apportées aux genres Normalograptus *et* Glyptograptus *en fonction de leur configuration astogénétique.*

INTRODUCTION

It has long been known that the graptolites were affected by a major extinction event near the end of the Ordovician. Lack of stratigraphic precision from key Ordovician–Silurian boundary sections combined with a lack of understanding of graptolite lineages across this boundary has, however, prevented a full assessment of the nature and magnitude of this bioevent. The first problem was recently addressed with a concerted effort to assign a boundary stratotype and regional parastratotypes. Much information concerning graptolite distribution is now available from excellent sections around the world (Cocks and Rickards, 1988).

The second problem, the understanding of graptolite lineages, was also addressed recently through the proposal of a new, phylogenetic classification by Fortey and Cooper (1986) and Mitchell (1987) (see also Kearsley, 1985). This new scheme is based on early rhabdosomal astogeny (proximal development patterns) as a major criterion for supraspecific classification and has produced a refined understanding of graptolite phylogeny. The purpose of this paper is to examine upper Ashgill graptolite faunas and, in particular, the Ashgill extinction event, in light of this new view of graptolite phylogeny.

ASTOGENETIC PATTERNS IN ASHGILL GRAPTOLITES

It has long been observed that Ashgill faunas are dominated by dicranograptids, lasiograptines, and biserial diplograptaceans with spinose proximal ends, whereas the lowest Llandovery faunas are composed mostly of diplograptaceans with proximal ends bearing only a virgella. The significance of these observations has been masked by a classification that relied on thecal form rather than proximal end structure as its principal distinguishing criterion. The work of Mitchell (1987) on the astogeny of Middle and Upper Ordovician diplograptaceans has now been built upon by studies of exquisitely preserved Upper Ordovician (Stewart, 1987; Stewart and Mitchell, 1988) and Llandovery diplograptid faunas (Melchin, 1987a). These studies have led to the recognition of two new diplograptacean astogenies, Patterns J and K.

The proximal development types found in later Ashgill and early Llandovery faunas are schematically illustrated in Figures 1 and 2, together with typical species illustrating these patterns. Mitchell (1987) has provided some suggestions for recognition of these patterns in compressed specimens. The most obvious feature is that Patterns D–G and K (those forms that dominate pre-Hirnantian faunas) tend to have blunt proximal ends with a generally obscured sicula in reverse view, and proximal spines other than the virgella (i.e., paired antivirgellar spines, a mesial spine on th1^1 and rarely on th1^2). The Ashgill fauna, dominated by species with these primordial astogenies, will be referred to as the DDO (Dicranograptidae–Diplograptidae–Orthograptidae) fauna. Patterns H–J and M, which completely dominate post-Hirnantian faunas, possess a somewhat more acicular proximal end with a partially exposed sicula below th1^2 and usually no spines other than the virgella, which may be very elaborate. These are designated as the M (Monograptidae) Fauna.

These observations provide a framework for critical reexamination of the abundance of recently published literature with the aim of assessing the distribution and diversity of these newly assigned genera within the upper Ashgill.

CORRELATION OF MIDDLE AND LATE ASHGILL GRAPTOLITE ZONES

Any attempt to examine worldwide coincidence of the Late Ordovician extinction event relies heavily on the quality of the correlation of the various zonation schemes for Upper Ordovician rocks. Figure 3 shows a proposed correlation of the Ordovician–Silurian boundary zones from several key areas around the world. Examination of this and other published correlation charts shows that the end of the *pacificus* Zone in northwestern and arctic Canada, the *pacificus* Subzone of Scotland and northeastern Siberia, the *ornatus–latus* Zone of central Victoria, and the *anceps* Zone of southeastern Alaska are closely coincident. The termination of these zones corresponds to the nearly simultaneous disappearance of several key Ashgill genera including *Dicellograptus, Paraorthograptus*, species of the "*C.*" *latus* group, *Climacograptus* (*Diplacanthograptus*) (the *C. longispinus* group), *Orthograptus* s.s. (i.e., *O. amplexicaulis, O. truncatus*), and the Lasiograptinae (sensu Mitchell, 1987) (see Fig. 4 for generic ranges at selected boundary sections). The succeeding *extraordinarius* Zone yields a very low diversity fauna usually consisting of two to five species, all of which can almost certainly be assigned to *Normalograptus* as defined here (see Appendix). As pointed out by Williams (1982) and VandenBerg et al. (1984), the first occurrence of "*C*"? *extraordinarius* shows some overlap with the preceding fauna. In these areas, the base of the *extraordinarius* Zone is not recognized by the first occurrence of the nominate species but by the disappearance of the preceding fauna. Outside China, very few species are known for certain to have survived this extinction event: *Normalograptus normalis, N. angustus* (= *C. miserabilis*), and *N. extraordinarius* (e.g., Williams, 1983). The last species is restricted to the uppermost *pacificus* and the *extraordinarius* intervals, although it is preceded by the very similar *N. ojsuensis* in the U.S.S.R. (Koren' and Sobolevskaya, 1983).

Assessment of the boundary interval in China is somewhat complicated due to two main factors: the use of different, usually endemic species as zonal indices among the several regions of China; and the apparent diachroneity of the *Hirnantia* Fauna (Rong, 1984). Of particular interest is the position and range of the zone of "*Diplograptus*" *bohemicus*

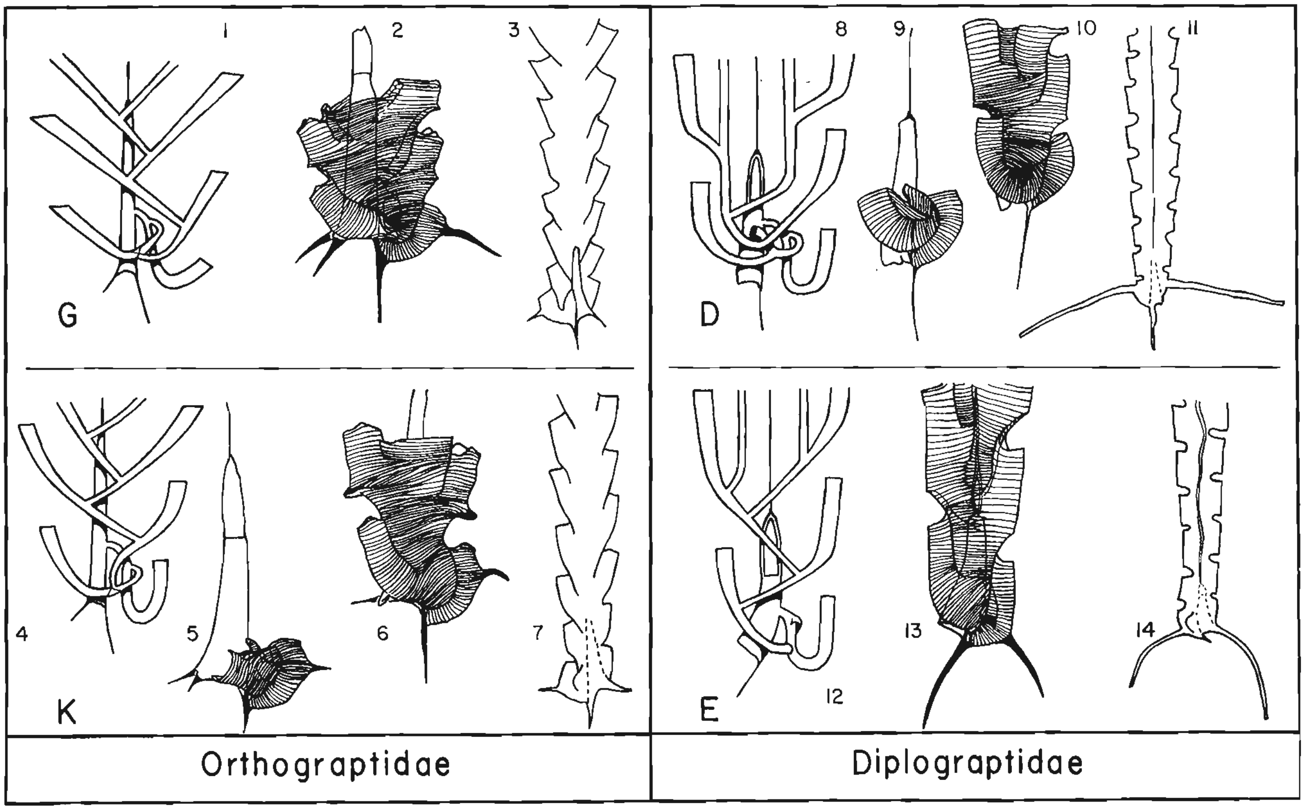


Figure 1. Comparison of the distinctive features of the astogenetic patterns exhibited by the exclusively Ordovician diplograptaceans. Shown for each are thecal diagrams illustrating the branching geometry of the early thecae; three-dimensionally preserved, isolated growth stages in reverse view; and compressed, nonisolated rhabdosomes in obverse view. 1.1–1.3, Pattern G; 1.1, thecal diagram showing prominent descending crossing canal of th1^2 and paired origin of th1^2 and th2^1 from two foramina produced by fusion of upward-growing flange and crossing canal; 1.2, postprimordial growth stage of *Amplexograptus maxwelli* (after Walker, 1953), note early origin of interthecal septum between th1^2 and th2^1 and the small initial size of the th2^1 protheca, x20; 1.3, *Orthograptus calcaratus basilicus* (236.75 m, "Utica Shale", Ohio Geological Survey Core 09-2082, Butler County, Ohio), showing upturned th1^1 with prominent gap between metatheca and protheca filled by th2^1, x6. 1.4–1.7, Pattern K; 1.4, thecal diagram illustrating dorsal origin of th2^1 and contribution of th1^1 metatheca to reduced th1^2 crossing canal; 1.5, early th1^2 stage of *Amplexograptus inuiti* (specimen newly isolated from type block, Cox, 1933), showing large dorsal foramen for th2^1 formed by upward-growing reverse lateral wall of th1^2 and hood produced from dorsal wall of th1^1 metatheca (see also Mitchell, 1987, Fig. 8k, m), x20; 1.6, postprimordial stage of same species as Fig. 1.5 showing relatively late origin of interthecal septum between th1^2 and th2^1 and large initial size of th2^1 (cf. Fig. 1.2), x20; 1.7, "*Glyptograptus*" *tenuissimus* (after Koren' and Sobolevskaya, 1983), showing tightly upturned th1^1 without gap, and narrow proximal end, x6. 1.8–1.11, Pattern D; 1.8, thecal diagram showing paired origin of th2^1 and th2^2; *Climacograptus (Climacograptus)* cf. *caudatus* (after Mitchell, 1987), 1.9, early th1^2 stage showing formation of this protheca via enclosure of reduced crossing canal by upward-growing flange, x22; 1.10, postprimordial stage of same species as Fig. 1.9, with th2^1 protheca ringed by semicircular th1^2, x22; 1.11, *C. (Climacograptus) trifidus spectabilis* (after Koren' and Sobolevskaya, 1983) with narrowly rounded, subsymmetrical proximal end, and proximal-facing sicular aperture, x5. 1.12–1.14, Pattern E; 1.12, thecal diagram showing strongly deflected sicula and dorsal origin of th2^1 from th1^2; 1.13; *Climacograptus (Diplacanthograptus) spiniferus* (after Mitchell, 1987), note simple pattern of differentiation of th2^1 from completed th1^2 protheca, x22; 1.14, *C. (Diplancanthograptus) longispinus supernus* (after Riva, 1974), showing laterally directed sicular aperture and virgella, th1^1 tucked well underneath proximal end and low thecal spine on th1^2, x9.

(a species here assigned to *Normalograptus*). Mu et al. (1984) and Chen and Lenz (1984) considered the *bohemicus* Zone to be approximately correlative with the *extraordinarius* Zone of Scotland. They also equated the *mirus* and *typicus* zones with the Scottish *pacificus* Subzone, and the succeeding *uniformis* Zone with the barren strata in Scotland. In part, Chen and Lenz (1984) based this correlation on the occurrence, in the Yukon *pacificus* Zone, of the distinctive and short-ranging Chinese genus *Diceratograptus.* Williams (1983) and VandenBerg et al. (1984) present a slightly different picture that equates the *typicus, mirus, uniformis,* and lower part of the *bohemicus* zones all with the *pacificus* interval. This is based on the co-occurrence of *Paraorthograptus* and other distinctive, cosmopolitan species in all of these Chinese zones and the termination of the DDO lineages within the *bohemicus* Zone as it is broadly applied in various parts of China.

In some of the Chinese studies where the *Hirnantia* Fauna occurs between the *bohemicus* and *persculptus* Zone, elements of the DDO faunas, such as *Paraorthograptus,*

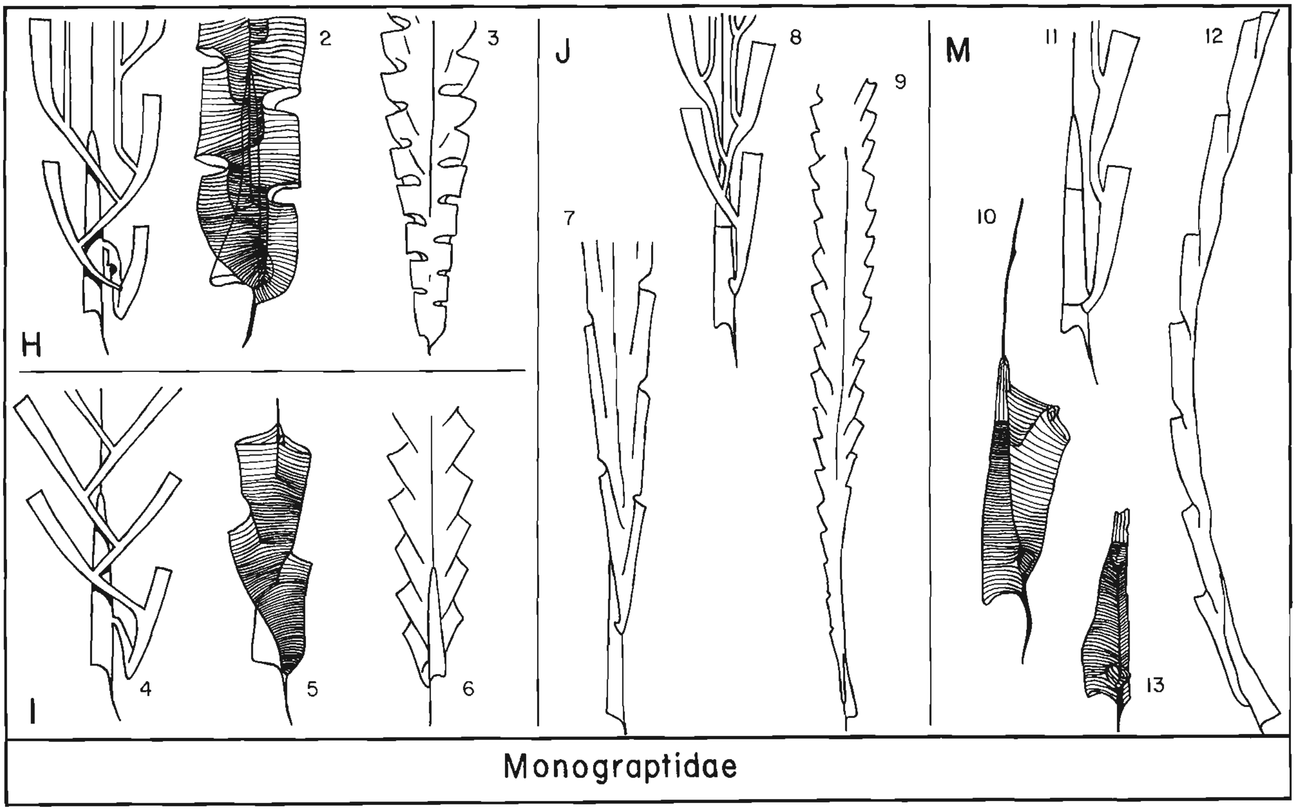
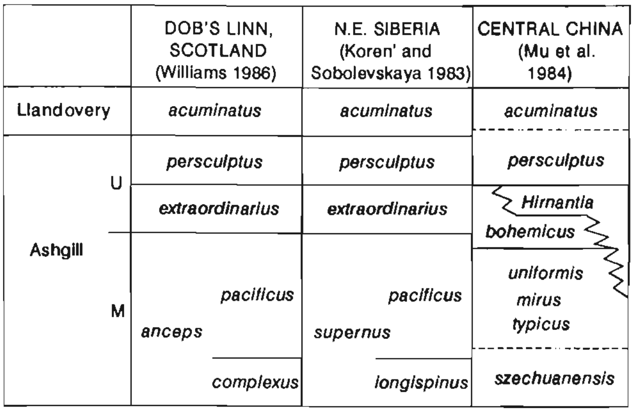


Figure 2. Comparison of the distinctive features of the astogenetic patterns exhibited by Upper Ordovician and Silurian Monograptidae. Shown are thecal diagrams illustrating the branching geometry of the early thecae; growth stages in reverse view (where available); and compressed nonisolated rhabdosomes. 2.1–2.3, Pattern H; 2.1, thecal diagram showing normal diplograptacean th1^1 protheca with th1^2 foramen and origin of th1^2 from free dorsal wall of th1^1; 2.2, *Normalograptus kuckersianus* (after (Mitchell, 1987), illustrating erect sicula and origin of dicalycal th2^1 from complete th1^2 protheca. x18; 2.3, reverse view, "*Diplograptus*" *tcherskyi* (after Melchin, 1987a), note fusiform proximal end with little exposure of the dorsal margin of the sicula, x7. 2.4–2.6, Pattern I; 2.4, thecal diagram illustrating th1^1 metatheca enclosing its reduced protheca, which lacks a th1^2 foramen; 2.5, *Glyptograptus tamariscus tamariscus* (after Melchin, 1987a), note exposure of sicula and acicular proximal end with origin of th1^2 from dorsal side of complete th1^1 protheca, x22; 2.6, obverse view, *Pseudorthograptus cyperoides* (after Sherwin, 1974), note short th1^1 protheca and acicular proximal end, x9. 2.7–2.9, Pattern J (known only from nonisolated specimens); 2.7, reverse view of *Akidograptus ascensus* (reconstructed from Williams, 1983), th1^1 with extremely short descending portion of protheca, th1^2 arises from completed protheca, x9; 2.8, thecal diagram illustrating inferred astogenetic pattern, th1^1 possesses no prothecal foramina; 2.9, obverse view of *Dimorphograptus confertus swanstoni* (after Lenz, 1982), x4.5. 2.10–2.13, Pattern M; 2.10, reverse view *Pristiograptus* sp. (Cape Phillips Formation, type section, northeastern Cornwallis Island, N.W.T., CP-77, 99.0–100.6 m, *M. griestoniensis* Zone) showing directly upward growth of th1 from porus, x18.5; 2.11, thecal diagram; 2.12, obverse view, *Atavograptus ceryx* (reconstructed after Rickards and Hutt, 1970), illustrating elongate proximal thecae and dominantly upward growth of th1, x14; 2.13, *Pristiograptus* sp., as above, showing origin of th1, x18.5.



		DOB'S LINN, SCOTLAND (Williams 1986)	N.E. SIBERIA (Koren' and Sobolevskaya 1983)	CENTRAL CHINA (Mu et al. 1984)
Llandovery		*acuminatus*	*acuminatus*	*acuminatus*
Ashgill	U	*persculptus*	*persculptus*	*persculptus*
		extraordinarius	*extraordinarius*	*Hirnantia* *bohemicus*
	M	*anceps* *pacificus*	*supernus* *pacificus*	*uniformis* *mirus* *typicus*
		complexus	*longispinus*	*szechuanensis*

Figure 3. Correlation of key Ordovician–Silurian graptolite zonations.

C. (*Diplacanthograptus*), and *Paraplegmatograptus* species continue through the *bohemicus* Zone and terminate at or below the base of the *Hirnantia* Fauna (e.g., Wang et al., 1983; Mu et al., 1984). In other areas of China, the *Hirnantia* Fauna occurs within a *bohemicus* Zone that consists entirely of Pattern H (and I?) species (Monograptidae) with no relics of the DDO lineages present (Mu and Ni, 1983; Li et al., 1984). In still other areas, the *Hirnantia* Fauna occurs between the *mirus-uniformis* Zone and the *bohemicus* Zone which, again, consists entirely of M taxa (e.g., Tongzi, Guizhou, Mu, 1983; Xixiang, Shaanxi, Yu et al., work in progress). This strongly suggests that the *bohemicus* Zone does not represent a uniform entity as it is used in different parts of China (VandenBerg et al., 1984) and that the pre-*Hirnantia* Fauna, lower *bohemicus* Zone with many DDO elements may be equivalent to the upper part of the *pacificus* interval, whereas the post-*Hirnantia* Fauna upper part is equivalent to the *extraordinarius* Zone. It is noteworthy that *N. bohemicus* is the only species consistently reported to span the entire *bohemicus* Zone in its broadest sense and is, therefore, another of the few species known with certainty to have survived the extinction event.

MIDDLE ASHGILL FAUNAS

The middle Ashgill, the interval occupied by the *anceps* Zone of Great Britain and the *supernus* Zone of the northeastern Siberia, has generally been viewed as a time of dwindling of previously more diverse lineages. Lineages comprising species with a Pattern D astogeny [*C.* (*Climacograptus*) and *Pseudoclimacograptus*], and some *Amplexograptus* species (Pattern G) indeed appear to be vestiges of previously more diverse groups (Fig. 5). *Climacograptus* (*Diplacanthograptus*) (Pattern E) and *Orthograptus* (Pattern G) maintain diversity through most of the mid-Ashgill, as some new taxa replace those lost. The lasiograpitinids (derived from a Pattern G-bearing ancestor) are still, for the most part, a poorly understood group. Although they form an important component of some mid-Ashgill faunas, their very high apparent diversity is most likely the result of oversplitting.

It is now clear, however, that some lineages were undergoing rapid evolution and that standing diversity declined only slightly until the latter half of the *pacificus* interval. A significant fraction of the diplograptacean fauna is occupied by a set of related species groups that show the Pattern K proximal development. This group may be restricted to the Ashgill (and possibly lowest Llandovery) and consists of several species groups, including ''*Glyptograptus*'' *tenuissimus* together with ''*G.*'' *hudsoni*, ''*G.*'' *occidentalis* and probably some forms assigned to ''*G.*'' sp. cf. ''*G.*'' *lorrainensis*; several climacograptid to amplexograptid species including the *A. latus* group; as well as *Paraorthograptus*, a genus in which more than 11 species and several subspecies have been named (many, but probably not all of which are synonymous — see Williams, 1983). Evolution within this group at both the generic and specific levels appears to have been very rapid and Pattern K taxa are dominant members of many *pacificus* interval assemblages. The Dicranograptidae also evolved some new colony designs during this time, including *Tangyagraptus* and *Diceratograptus*.

THE EXTINCTION EVENT

From the data reviewed here, the late Ashgill graptolite extinction event centres upon the *pacificus–extraordinarius* zonal boundary or within the lower part of the *bohemicus* Zone in China (Figs. 5, 6). It is an event of considerable magnitude, which resulted in the termination of some 21 genera, 3 families, and as many as 7 subfamilies between the base of the *pacificus* interval and the lowest part of the *acuminatus* Zone. The majority of the terminations take place at or very near the end of the *pacificus* interval; a few lineages survive, in much reduced diversity, into the succeeding zones.

The actual abruptness of the extinction event is still difficult to assess because it is difficult to establish the degree of synchroneity among localities from different cratons. In addition, the DDO graptolites exhibit different range patterns among these locales. Most of the important mid-Ashgill taxa at Dob's Linn (Williams, 1982) range up to the last *pacificus* Subzone sample but this abruptness is certainly heightened by the fact that the overlying 2.9 m of strata are barren, with the exception of a thin band of *extraordinarius* Zone graptolites (Berry, 1989). However, the fact that *Normalograptus extraordinarius* itself occurs at the top of the *pacificus* Subzone in this section indicates that it is probably equivalent to the lower (pre-*Hirnantia* Fauna) *bohemicus* Zone of China.

In some sections in the Soviet Union and China, where the middle to upper Ashgill are well represented, the passage does not show an abrupt termination of the DDO lineages (e.g., Koren' and Sobolevskaya, 1983; Wang et al., 1983; Mu et al., 1984). Furthermore, it is not always the same taxa that persist a little higher than the others (Fig. 3). This may indicate that different taxa persisted longer in different areas. On the other hand, the vagaries of preservation and sampling may account for some of the differences in the latest middle Ashgill species occurrences, especially since *uniformis* and *bohemicus* equivalent intervals are so thin, commonly 0.5 m or less (Wang et al., 1983; Mu et al., 1984). Signor and Lipps (1982) suggest that sampling and preservational biases are bound to obscure the true abruptness of a sudden extinction event. Nevertheless, the occurrence of several well preserved specimens of *Paraorthograptus* in *persculptus* Zone strata at two localities (Wang et al., 1983, Fig. 3; Lin and Chen, 1984) indicates that this genus did persist longer than the others in a few refuges. The same can be said of the occurrences, in the *persculptus* Zone, of specimens that probably represent a species of *Orthograptus* (e.g., Wang et al., 1983).

Occurrences of *Dicellograptus* have been reported from the *bohemicus* Zone by Ge (1984) and in the *Pristiograptus leei* Zone (middle Llandovery) by Wang (1984), in both cases with an otherwise entirely M Fauna and represented by single, robust, incomplete specimens, stratigraphically very close to *pacificus* Zone strata. It is possible that both of these stratigraphically isolated occurrences resulted from reworking, because in all other areas, *Dicellograptus* disappears with or before the other DDO genera. The mid-Llandovery genus, *Neodicellograptus*, which one of us (M.J.M.) collected recently in arctic Canada (Fig. 7), possesses a Pattern H astogeny and is almost certainly derived from a *Metaclimacograptus* ancestor rather than from *Dicellograptus* as suggested by Chen and Lin (1978) and Wang (1984).

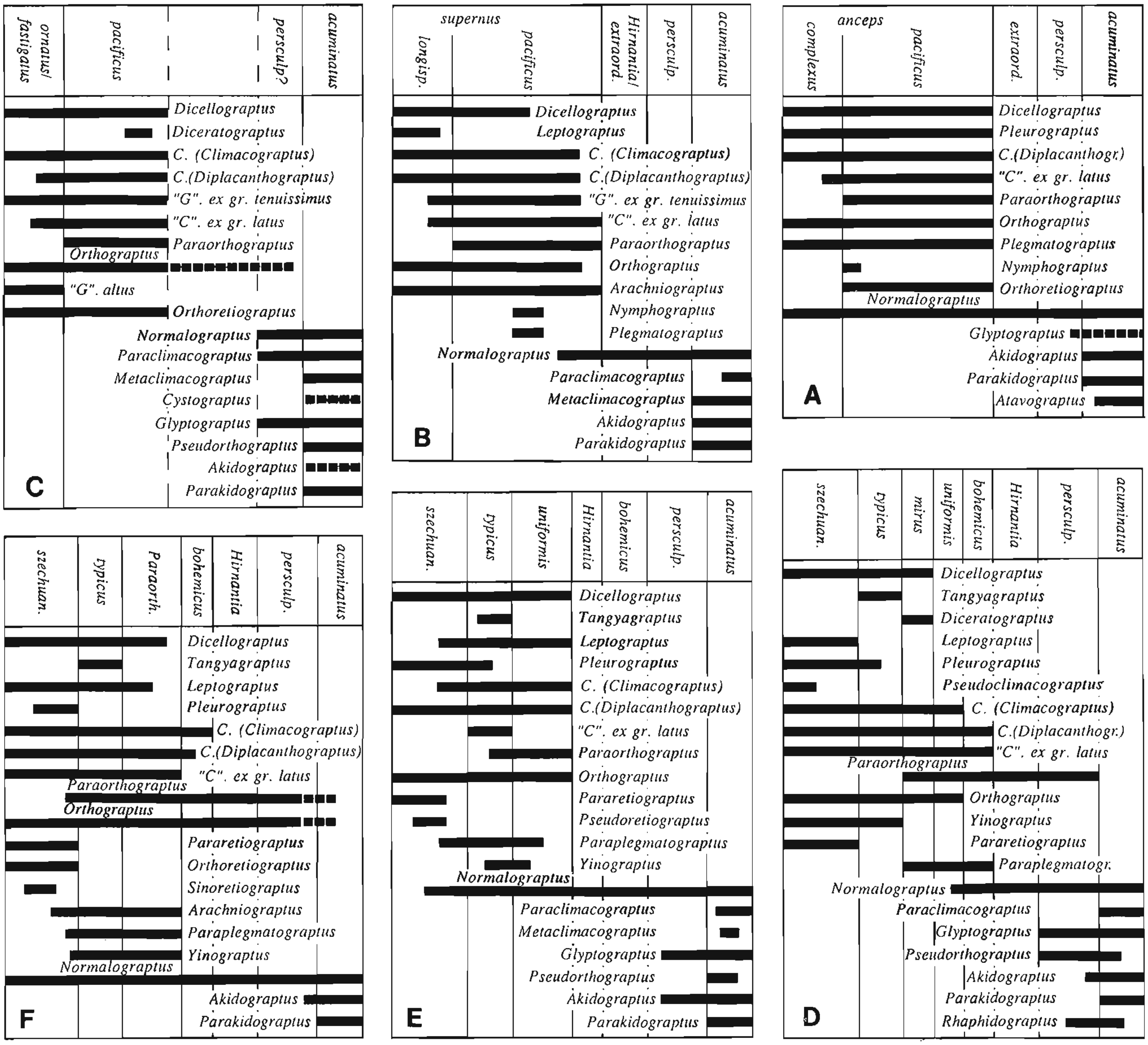


Figure 4. Distribution of graptolite genera across the Ordovician–Silurian boundary at several key sections of the world. Genera revised according to the phylogenetic classification of Mitchell (1987), Legrand (1987), and this paper. **A.** Dob's Linn, Scotland (Williams, 1982, 1983); **B.** Mirny Creek area, northeastern Siberia (Koren' and Sobolevskaya, 1983); **C.** northwestern and arctic Canada (Lenz and McCracken, 1982; Melchin, 1987b; **D.** Yichang, Hubei, China (Lin and Chen, 1984; Mu et al., 1984); **E.** Xixiang, Shaanxi, China (Yu et al., unpub. ms.); **F.** eastern Yangtze Gorges (Wang et al., 1983).

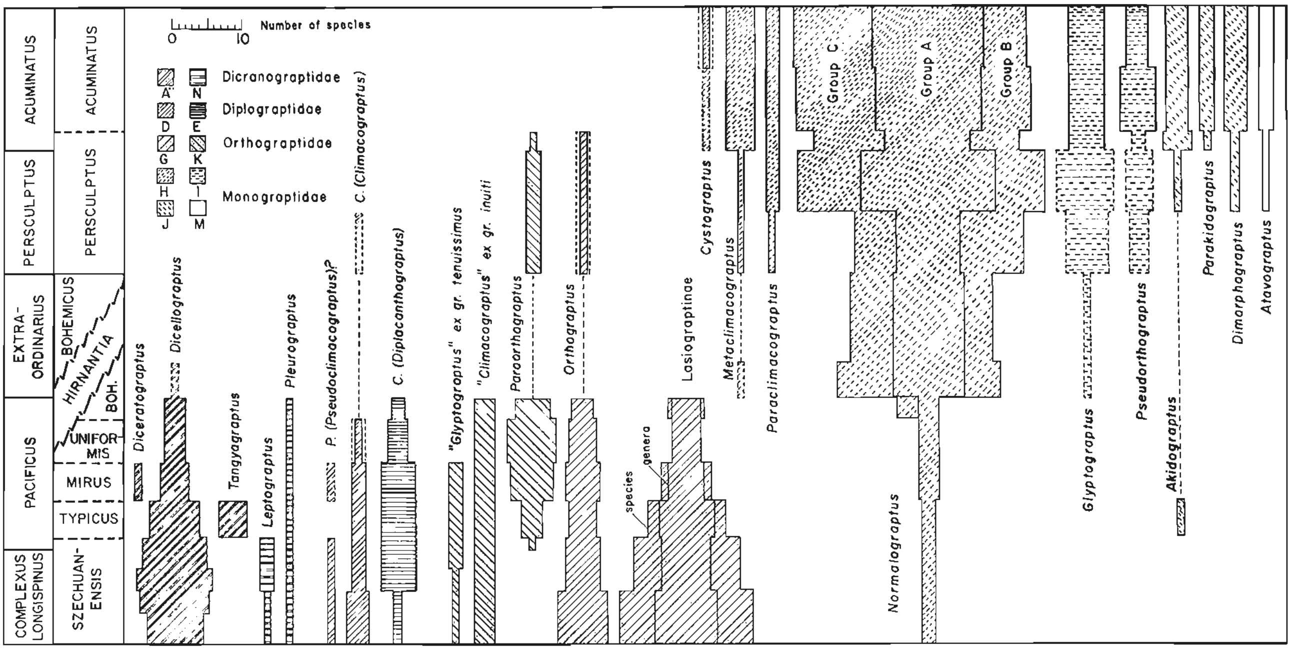


Figure 5. Composite range and diversity chart of Ordovician–Silurian graptolite genera worldwide. No attempt has been made to synonymize species from different areas. Data taken from Churkin et al., 1971; Rickards, 1976; Koren' et al., 1980; Li and Ge, 1981; Lenz and McCracken, 1982; Koren' and Sobolevskaya, 1983; Mu, 1983, 1988; Wang et al., 1983; Ge, 1984; Li, 1984; Li et al., 1984; Lin and Chen, 1984; Mu and Lin, 1984; Mu et al., 1984; VandenBerg et al., 1984; Berry, 1986; Legrand, 1986; Obut and Sennikov, 1986; Melchin, 1987b; Storch, 1988.

The extinction event itself, then, appears to have been an interval of accelerated extinction of both holdover lineages and rapidly radiating groups. It followed a long period of slow but not necessarily steady decline in graptolite diversity that began in the Caradoc and intensified during the relatively short *pacificus* interval (see also Koren', 1988).

The rediversification of the graptolites took place from only a very few lineages, all apparently derived from the Pattern H (*Normalograptus*) ancestral stock. In fact, all of the major Silurian graptoloid groups, the Pattern H (hedrograptines), Pattern I (petalograptines), Pattern J (dimorphograptines), and the uniserial monograptids were all established by the end of the *persculptus* Zone. The retiolitines and the cyrtograptids were the only main generic groups to evolve later in Silurian time. The pace of diversity increase was also rapid (Fig. 6). By early *extraordinarius* Zone time, species diversity had rebounded to more than 20 species and by mid-*acuminatus* Zone time it had nearly tripled. Thus, in an interval of probably less than 2 million years (Carter et al., 1980), the Monograptidae had increased in diversity from roughly six to approximately 60 species.

EXTINCTION MECHANISMS

Lithological change or a hiatus is a very common feature of Ordovician–Silurian boundary sections. As noted above, at Dob's Linn the graptolitic sequence is interrupted by barren, light-coloured strata. In many areas of China and Siberia, the graptolitic sequence is interrupted by beds bearing the *Hirnantia* shelly fauna at or very near the passage from DDO to M faunas. In some parts of northwestern and arctic Canada, the boundary interval is occupied by light-coloured carbonate strata (Lenz and McCracken, 1982; Melchin, 1987b). In other regions of northern Canada, as well as in other parts of the world, there are no strata for this interval. In Bohemia (Štorch, 1988), Algeria (Legrand, 1986), and parts of China (e.g., Mu and Ni, 1983) the graptolite-bearing sequence begins in strata equivalent to the *extraordinarius* or *persculptus* Zone that overlie nongraptolitic rocks of Hirnantian age or older.

That these interruptions of the normal graptolitic (anoxic or dysaerobic) sequence occur at virtually every well studied boundary section around the world, strongly suggests that the graptolite extinction event coincided with a change in global oceanic conditions. Oxygenation of deep water sediments, increased bottom water circulation (probably the cause of nondeposition or erosion observed in many areas), and major regression in shallow water sequences around the world (e.g., Lenz, 1976; Brenchley and Newall, 1980) may have been the result of the glaciation centred in North Africa. This Ashgill glaciation appears to have been at its peak during the Hirnantian (Spjeldnaes, 1981). The question is why a period of oceanic cooling, oxygenation, and regression should coincide with an almost complete extinction of a group of pelagic organisms.

Wilde and Berry (1984) have developed a model that attributes extinction events to oceanic turnover and the rise of inhospitable waters into the surface regions of maximum productivity, resulting from the onset of glaciation. Their model depicts warm, nonglacial periods as times of weak but stable water mass stratification. In tropical seas, there may have been two layers of anoxic (or at least strongly dysaerobic) water within the water column because of sluggish circulation and different salinity sources for water at different depths. During periods of transitional climate, the flood of cool, low salinity water from the polar regions would mix with the warmer, more saline water of the same density below the pycnocline (in the deep and intermediate layers of the oceans) causing a rise and merging of the anoxic layers. With the appropriate rates of climatic change, this could result in a temporary condition of density instability in the water column. Overturn could result, bringing toxic, oxygen-depleted, sulphide rich water to the surface. Periods of glacial maxima would be characterized by a high stability of oceanic stratification based on strong temperature gradients from very cold bottom waters of polar origin, and elimination of the anoxic layers as a consequence of movement of oxygenated waters into deep and intermediate depths.

Overturn and potential extinction in this model generally should coincide with the onset or decline of glacial activity, as this is the time when density instability would occur. However, the Late Ordovician graptolite extinction appears to coincide with the peak of glacial activity; ice caps had apparently been present over the south polar areas at least since Caradoc time.

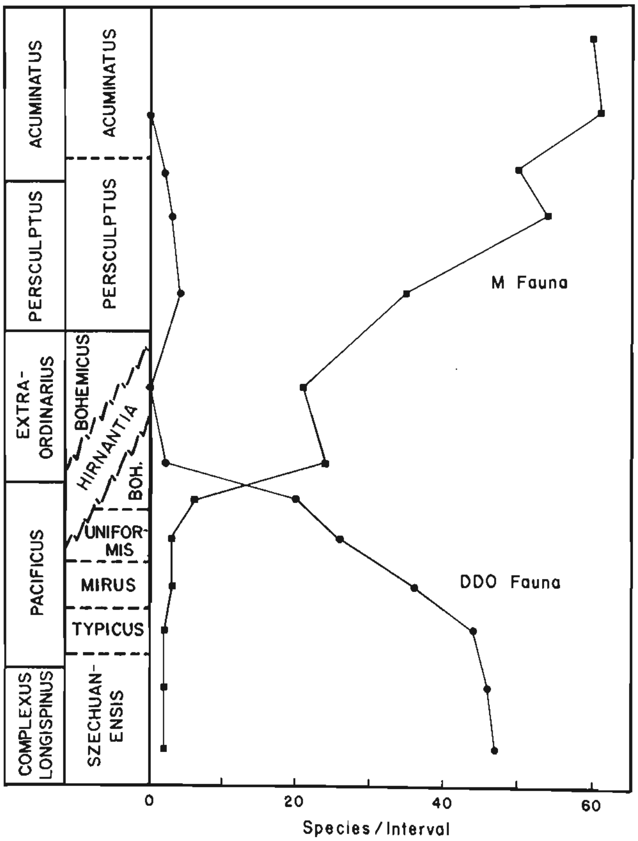


Figure 6. Numerical abundances of DDO and M Fauna graptolites across the Ordovician–Silurian boundary. See Figure 5 for data sources.

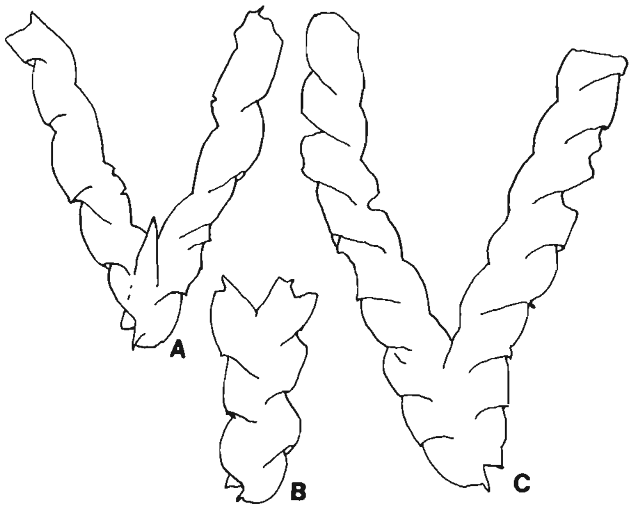


Figure 7. *Neodicellograptus* sp. cf. *N. siluricus* (Mu et al.) from small creek 10 km east of Cape Manning, Cornwallis Island, N.W.T., 75°27-N, 94°21-W. **A.** GSC 94728; **B.** GSC 94729; **C.** GSC 94730. All figures x12.

Brenchley (1984) reviewed the faunal extinctions of the Late Ordovician and found that the deep shelf faunas suffered a significant extinction coincident with the graptolite extinction at or near the Rawtheyan–Hirnantian boundary, whereas the shallower shelf faunas suffered fewer extinctions throughout Hirnantian time. Brenchley (1984) attributed the extinctions of the shelf faunas to regression and the associated reduction of shallow shelf area. The extinction of the pelagic faunas (especially graptolites) was attributed to contraction of the tropical plankton belts. Jablonski (1980, 1985), Wise and Schopf (1981), and Stanley (1984) argued that reduction in habitable area by itself is not sufficient to cause large-scale extinction. Stanley argued that overall oceanic cooling is a more likely cause of the extinction of faunas adapted to tropical conditions.

Other planktic or nektic organisms that appear to have suffered from significant extinction are the acritarchs (Duffield and Legault, 1981; Colbath, 1986; Martin, 1988) and cephalopods (McKinney, 1985). Neither seem to have declined as profoundly as the graptolites, but this may be attributed to the fact that both had a significant representation in neritic waters where extinctions were less pervasive than in the pelagic realm. Asaphinid trilobites had planktic larvae and appear to have been more severely affected by the extinction event than other Ashgill trilobite groups with benthic larvae (Fortey and Chatterton, 1988). Likewise, the Chitinozoa, which may have been a planktic dispersal stage of some metazoan, also suffered a significant faunal turnover (Grahn, 1988).

The nature of the graptolite habitat is of crucial importance in the consideration of their extinction events. Graptolite faunas may have been distributed according to depth (Berry, 1962; Berry and Boucot, 1972; Erdtmann, 1976; Kaljo, 1978; Cisne and Chandlee, 1982). Alternatively, regional faunal variations may reflect lateral differentiation of faunas and their association with specific water masses (Berry, 1974, 1977; Watkins and Berry, 1977; Finney, 1984). Analogy with modern zooplankton suggests that both of these factors were probably operative in graptolite ecology.

Koren' and Rickards (1979) indicated that although a few graptolite species are found in aerated shelf environments, there is a very strong association between diverse graptolite faunas and dark carbonaceous mudstone. Accordingly, they proposed a model that links graptolite diversity patterns and their final extinction to the rise and fall of the "black shale environment", which is an indicator of climate and oceanic productivity (see also Leggett et al., 1981). Berry et al. (1987) suggested that most graptolites may have inhabited oxygen poor, nitrate rich waters that existed between the oxygenated surface layers and the deeper anoxic waters of the global oxygen minimum zone. This hypothesis would explain the close association of graptolites and anaerobic and dysaerobic sediments. Applying this to the Wilde and Berry (1984) model for extinction in the Ordovician, equitable Early Ordovician climates could have resulted in several anoxic/dysaerobic layers within a tropical water column (see Wilde and Berry, 1984, Fig. 2). Such stratification should be especially well developed in areas adjacent to continental shelves and epicratonic areas. This would provide potential graptolite habitats at several levels within the water column. As the climate cooled in the Late Ordovician, the potential for low oxygen habitats would have been reduced as the influx of cooler, oxygenated water caused the anoxic layers to rise and merge. This, together with constriction of the tropical plankton belts and increased circulation, resulted in the decrease in diversity and provincialism in the latter half of the Ordovician. With the onset of peak glaciation, the anoxic/dysaerobic layer system could have been completely disrupted with the upward and lateral movement of cool, oxygenated waters toward the surface in tropical areas. Not only would this temporarily destroy the main graptolite habitat, but these cooler waters would flood onto the deeper shelf areas putting the tropical, deep shelf faunas in jeopardy. Even in the absence of the overturn of toxic, anaerobic waters, peak glaciation could have been responsible for the disruption of the tropical pelagic and deep shelf habitats, whereas the accompanying regression would have drastically reduced the epicontinental and shallow shelf seas, thus contributing to the more subdued levels of extinctions in those habitats.

Even if most graptolite species did not actually inhabit these dysaerobic zones, the widespread distribution of anoxia suggests the presence of large areas of upwelling of nutrient rich waters that would be favourable areas for rich plankton communities. In view of the fact that anaerobic and dysaerobic waters were widespread and persistent during the Ordovician and Silurian, it is reasonable to suggest that the structures of planktonic communities were probably adapted to the oceanic conditions associated with these layers, either directly (if they inhabited the anoxic dysaerobic waters during all or part of their life cycle) or indirectly (if they relied on nutrients upwelling from the anoxic layers). The rapid disruption of this oceanic layering should, therefore, have had a major impact on this community. The fact that the major Ashgill graptolite extinction event coincides with what the sedimentary record indicates was a major oceanic oxygenation event adds further weight to the suggestion that the graptolite-dark shale association is more than just preservational.

Two of the species that persist across the extinction event, *Normalograptus angustus* (= *N. miserabilis*) and *N. normalis* originated well before the end of the Ordovician and were cosmopolitan in latest Ashgill and early Llandovery times. They are commonly the only, or among the only, taxa found in the first Llandovery samples above the extinction, even in shelf areas such as Anticosti Island (Riva, 1988). These two species are among the longest lived graptolite species known; *N. normalis* persisted for about 7 m.y. (Rickards, 1977). These were almost certainly eurytopic species and may have been able to withstand disruption of the traditional graptolite habitat. It was this genus, *Normalograptus*, which was responsible for the rapid rediversification and recolonization of the graptolite habitats following the extinction interval.

Sheehan (1973, 1975, 1982) suggested that following the decline of the Upper Ordovician tropical shelf brachiopod communities, the recolonization process was largely accomplished by invaders from the cooler areas of deeper water and/or higher latitude. A similar process of invasion of shallow shelf areas by previously cooler water, deep shelf taxa was proposed by Chatterton and Perry (1983) for Lower Silurian trilobites in northwestern Canada.

The re-establishment of anoxia in the water column appears to have taken place rather rapidly at about the beginning of the *persculptus* Zone with the reincursion of dark, graptolitic sediments in many areas, including some areas that, prior to the extinction, were not sites of graptolitic sedimentation. It has been suggested that the spread of anoxia corresponding approximately to the end of the Hirnantian was one of the factors contributing to the extinctions at that time among benthic faunas (Brenchley, 1984; Nowlan et al., 1988).

NOTE ADDED IN PROOF

Since this paper was submitted, Riva (*in* Riva and Ketner, 1989) has made generic revisions to a number of the Ordovician "climacograptid" taxa discussed here. In particular, *Climacograptus* (*Climacograptus*) *caudatus* has been designated as the type species of the genus *Ensigraptus* Riva, 1989, the *C.* (*Diplacanthograptus*) *longispinus* group has been reassigned to *Appendispinograptus* Li and Li, 1985, the type species of which is *A. venustus*, and *C.* (*C.*) *trifidus* and many other spinose forms have been assigned to *Euclimacograptus* Riva, 1989 (type species *E. hastatus*). The reader is referred to Riva (*in* Riva and Ketner, 1989) for further discussion of these genera. Although these generic changes will slightly modify the patterns seen in the pre-extinction faunas as shown in Figures 4 and 5 of the present paper, the overall patterns of extinction and radiation as described here remain unchanged.

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APPENDIX

Systematic Notes

Order GRAPTOLOIDEA Lapworth, 1875

Suborder VIRGELLINA Fortey and Cooper 1986

Superfamily DIPLOGRAPTACEA Lapworth 1873, emend.

Family MONOGRAPTIDAE Lapworth 1873, emend.

Genus *Normalograptus* Legrand, 1987 emend.

Type species. *Climacograptus scalaris normalis* Lapworth, 1877.

Diagnosis (emended herein). Pattern H species with (usually) unornamented glyptograptid to climacograptid thecae. Proximal end relatively narrow, rounded, and asymmetrical with strongly alternating thecae. Median septum straight and usually complete, with th2^1 or some later theca dicalycal.

Remarks. Legrand (1987) has recognized the distinctness of the proximal development patterns exhibited by Llandovery and some Ordovician diplograptaceans; an astogenetic pattern that he called the "keroblastic" mode of development (equivalent to the patterns H and I employed here). He established a new genus, *Normalograptus*, to encompass the keroblastic species with climacograptid thecae.

Riva (1988) proposed the genus *Scalarigraptus*, also based on the type species "*C.*" *normalis*, which encompasses the same group of species as *Normalograptus* and is based on essentially the same criteria. *Scalarigraptus*, therefore, is a junior synonym of *Normalograptus*.

As Packham (1962), Bulman (1970), and others have observed, there exists a morphological continuum between species with climacograptid thecae and those with glyptograptid and biform thecae (climacograptid to glyptograptid) within many diplograptacean lineages. This feature of rhabdosomal evolution is particularly pervasive among those Monograptidae (sensu Mitchell, 1987) that possess a Pattern H astogeny as in the "*Glyptograptus*" *persculptus*–"*Climacograptus*" *extraordinarius* group from the Ordovician/Silurian boundary interval and the "*G.*" *euglyphus*–"*C.*" *brevis* lineage of Caradoc age (Finney, 1986). Until much more information is available on the phylogeny of species with a Pattern H astogeny and glyptograptid thecae, it is advisable to include both, as well as some of the biform "*Diplograptus*" species (e.g., "*D.*" *modestus*), within an expanded *Normalograptus*. Legrand's genus *Neodiplograptus* is based on the type species *Diplograptus magnus* (H. Lapworth). Uncompressed material from arctic Canada of the very similar species *D. tcherskyi*, indicate that members of the *magnus-tcherskyi-thuringiacus* group may differ at the generic level from members of the *D. modestus* group, which appear to resemble other *Normalograptus* species more closely.

In the composite generic range chart (Fig. 5) *Normalograptus* species are grouped into three informal species groups that may or may not have any phylogenetic significance: Group A, species with climacograptid thecae throughout; Group B, species with glyptograptid thecae throughout; Group C, those with biform to polyform thecae.

Glyptograptus Lapworth, 1873, emend.

Type species. *Diplograpsus tamariscus* Nicholson, 1868.

Diagnosis (emended herein). Pattern I species with glyptograptid to, less commonly, climacograptid thecae. Proximal end rather slender, unornamented, and usually tapering. Rhabdosome circular to ovate in cross-section, usually aseptate although a partial median septum may be present.

Remarks. Mitchell (1987) felt that all of the Llandovery species with glyptograptid and climacograptid thecae have the same proximal development pattern (Pattern H) and he decided to include both within an expanded *Glyptograptus* until further information was available to distinguish more phylogenetically consistent subgroups. One of us (M.J.M.) has now obtained uncompressed specimens of *G. tamariscus* (Fig. 2.5) and several other glyptograptids from arctic Canada, and these clearly show that the primordial astogeny of some of the Llandovery "glyptograptids" are of the Pattern I type, including *G. tamariscus*, the type species, whereas some (e.g., *G. nikolayevi*) posses the Pattern H proximal growth form and are here grouped with the Pattern H "climacograptids" in the genus *Normalograptus*

Evolutionary crisis of the Ashgill graptolites

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Abstract

Ashgill graptolites are widespread in many regions, where four to six zonal associations are differentiated in continuous sections. Of importance among successive bioevents are: 1) late Caradoc-early Ashgill extinction; 2) moderate radiation and appearance of an impoverished complanatus Biozone assemblage; 3) short-term radiation and minor extinctions at the complexus/pacificus boundary; 4) terminal mass extinction (pacificus event), affecting several graptolite families and coinciding with maximum glaciation in the late Ashgill; 5) slight radiation of morphologically simple diplograptids, followed by a final episode of Ashgill extinction (the extraordinarius Biozone); 6) radiation with moderate numbers of a few significant taxa (the persculptus Biozone); 7) radiation in the acuminatus Biozone comparable in importance with the persculptus event, but more expressive with regard to morphological variety.

Radiation events in the persculptus and acuminatus biozones did not restore the taxonomic diversity lost due to the crisis at the lower Hirnantian boundary. A new radiation peak was attained in subsequent evolutionary phases of the early Llandovery graptolites as a response to deglaciation and onset of black shale sedimentation.

Résumé

Les graptolites ashgilliens sont répandus dans de nombreuses régions où quatre des six associations zonales sont différenciées dans des profils continus. Les bioévénements successifs importants sont les suivants : 1) une extinction au Caradocien supérieur-Ashgillien inférieur; 2) un rayonnement modéré et l'apparition d'un assemblage appauvri de la biozone à complanatus; *3) un rayonnement à court terme et des extinctions accessoires à la limite* complexus/pacificus; *4) une extinction massive terminale (événement* pacificus) *affectant plusieurs familles de graptolites et coïncidant avec la glaciation maximale à l'Ashgillien supérieur; 5) un faible rayonnement des diplograptidés morphologiquement simples, suivi d'un épisode terminal de l'extinction ashgillienne (la biozone à* extraordinarius); *6) le rayonnement d'un nombre modéré de quelques taxons importants (la biozone à* persculptus); *7) le rayonnement dans la biozone à* acuminatus *comparable en importance à l'événement à* persculptus, *mais plus expressif en ce qui concerne la variété morphologique.*

Les événements de rayonnement dans les biozones à persculptus *et à* acuminatus *ne redonnent pas la diversité taxonomique perdue par suite de la crise observée à la limite de l'Hirnantien inférieur. Un nouveau sommet de rayonnement est atteint au cours des phases évolutives subséquentes des graptolites du Llandovérien précoce par réaction à la déglaciation et le début d'une sédimentation de schistes noirs.*

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INTRODUCTION

During Ashgill times, the rich diplograptacean fauna (Mitchell, 1987) became extinct. The fauna began in the early Caradoc and underwent its maximum taxonomic variation during the mid-Caradoc. In late Caradoc and early Ashgill the disappearance of some genera among lasiograptids, nemagraptids, and dicranograptids resulted in a noticeable reduction in morphological variation. In the early Ashgill these subfamilies as well as the retiolitids are represented by only a few genera. The diplograptids passed through the Caradoc-Ashgill boundary without remarkable change, and during the Ashgill consisted of only five to eight genera. The latest Ashgill and earliest Llandovery associations are composed entirely of glyptograpti, with the exception of the earliest monograptids, which are first found in the *persculptus* Biozone (Rickards and Hutt, 1970). A drastic impoverishment of the diplograptaceans (the orthograptid-dicellograptid subfauna of Bulman, 1970) at the end of the Ordovician is particularly evident at the species level.

Within this paper an attempt is made to summarize some general and global tendencies in Ashgill graptolite dynamics based on current knowledge of their stratigraphic and geographic distribution.

In continuous graptolite-bearing sequences known from regions of Eurasia, Australia, and North America, four to six zonal assemblages can be distinguished. Some diagnostic taxa are widely distributed within tropical and warm-water temperate belts. In central China and in eastern North America many taxa are known that are either endemic or only rarely occur elsewhere. This observation is even more pronounced in the late Ashgill. In both areas it is most probably related to regional paleogeographic peculiarities (Riva, 1969, 1974, 1988; Mu, 1984, 1988).

The Late Ordovician regression, which can be correlated with changes in oxygen content and climate, led to an incompleteness of the graptolite record within hemipelagic and pelagic sequences of many regions (Koren, 1987). It is postulated that two maxima of the Late Ordovician glaciation fell within the *complanatus* and *extraordinarius* to *persculptus* biozones (Barnes, 1986; Cocks and Rickards, 1988). In many Hirnantian sequences, black shales give way to light-coloured carbonates and clastics, and graptolites are rare or absent in most beds.

The evolution of the Ashgill graptolites can now be generalized on the basis of evidence from several continuous sections in different regions. Even the most representative, monotonous sequences differ from each other in thickness and completeness of the graptolite record. A quantitative and qualitative evaluation of the extinction and radiation events can be made only approximately, because of varying approaches to taxonomy at the generic and species level among graptolite workers. For instance, it is especially difficult to evaluate many taxa described from several sections of the Wufeng Formation in central China. Here, although the Ashgill graptolite associations seem to be most diverse, many of the new taxa are probably synonymous with previously described graptolites from other regions.

The present author has made an attempt to use generic and suprageneric classification, which is based on the main astogenetic patterns (Cooper and Fortey, 1983; Fortey and Cooper, 1986; Mitchell, 1987). However, there is still a lot to be done to refine and understand the morphology of many species among diplograptaceans, which still cannot be unambiguously placed within this new taxonomic system.

The sections chosen for the comparative analysis of extinction and radiation bioevents in the Ashgill evolution of graptolites are shown in Figure 1. Zonal correlation is based on data currently available; however, there are problems connected with the vertical ranges of some diagnostic graptolites within the *complanatus* and lower *anceps* biozones, as well as the composition of *extraordinarius* and *persculptus* assemblages.

SECTIONS THROUGH THE ORDOVICIAN-SILURIAN BOUNDARY

Scotland: Dob's Linn

Five zonal graptolite associations have been recognized in the upper Hartfell and lowermost Birkhill shales at Dob's Linn (Williams, 1980, 1982a,b, 1983, 1987, 1988). However, not all the zonal boundaries can be defined precisely, because of barren intervals separating graptolite-bearing beds. Thus, it is difficult to evaluate exact extinction levels. A mass extinction of dicellograptids, leptograptids, and orthograptids, which characterize the upper Caradoc, took place in the early Ashgill near the base of the upper Hartfell Shale (11 species among 17 known). The appearance of *Dicellograptus complanatus*, *D. minor*, *Orthograptus socialis*, and some retiolitids, constituting the impoverished assemblage of the *complanatus* Biozone, was the first radiation event in the Ashgill evolution of the graptolites. Many of the species occur only within this biozone, but their disappearance - successive or simultaneous - cannot be established in the section. The maximum taxonomic diversity, comparable with that of the late Caradoc, was achieved later within the *complexus* Subzone. At this level the last and most impressive radiation in several Ashgill lineages took place, and a new dicellograptid and diplograptid fauna appeared. The most important members of the newly formed association are *Dicellograptus complexus*, *D. anceps*, *Climacograptus supernus*, *Geniculograptus latus*, and *Normalograptus normalis*. Altogether, about 12 species have been identified at this level. Almost all members of the *anceps* assemblage disappear at the top of the biozone (the *pacificus* event). Among others, the following taxa die out: *Dicellograptus*, *Pleurograptus*, *Plegmatograptus*, and *Paraorthograptus* as well as some species of the *Climacograptus latus* and *C. longispinus* groups. The uppermost part of the upper Hartfell Shale contains impoverished climacograptid and glyptograptid associations (about five species) diagnostic of the *extraordinarius* Biozone, most of them known from the overlying and underlying strata. The first moderate radiation is found at the base of the *persculptus* Biozone. It is marked by the first appearance of *Glyptograptus persculptus*, *G. avitus*, and *Normalograptus medius*. Somewhere within the zone the first monograptids also make their entry. All taxa cross the *persculptus/acuminatus* zonal boundary. A new radiative event started at the base of the *acuminatus* Biozone. It gave rise to many new morphotypes, notably *Akidograptus ascensus*, *Parakidograptus acuminatus*, *Normalograptus trifilis*, and other forms.

Northeast U.S.S.R.: the Kolyma Basin, Mirny Creek

The *supernus* Biozone (the lower Tirekhtiakh Formation) at the Mirny Creek section is characterized by diverse facies development, composed mainly of organo-detrital carbonates with a benthic fauna (Koren et al., 1983, 1988). Shale and siltstone layers yielding numerous graptolites occur within this sequence. Strata of the *extraordinarius* and *persculptus* biozones consist of light-coloured dolomite, marl, and siltstone, representing the terminal phase of Late Ordovician regression. The black shale sedimentation beginning at the base of the *acuminatus* Biozone is the first evidence of the Llandovery transgression (the Maut Formation). Graptolites do not occur continuously within the succession of the Tirekhtiakh Formation, which makes it difficult to define precisely the stratigraphic ranges of some species. Nevertheless, within the formation four zonal graptolite assemblages have been recognized above the *quadrimucronatus* Biozone. They all become impoverished toward the top of the unit. The first extinction coincides with the basal part of member *N*, with the last occurrences of *Climacograptus tubuliferus*, *C. trifidus*, and *Arnheimograptus lorrainensis*, characteristic of the early Ashgill. At the *longispinus* level, graptolite diversity increases as the result of the last Ashgill radiation. Numerous species of the *Climacograptus longispinus*, *C. hastatus*, and *Geniculograptus latus* groups appear in association with *Dicellograptus complanatus* and *D. complexus*. Ten new taxa first occur at this level. The mass extinction of almost all Ashgill graptolites coincides with the top of member *P*. At this level *C. longispinus*, *C. pogrebovi*, *Paraorthograptus pacificus*, *Arachniograptus* sp., and other diagnostic diplograptids disappear (the *pacificus* event). Species belonging to *Dicellograptus* are rare in the section and do not reach the top of the *pacificus* Subzone. The upper Tirekhtiakh Formation (member *Q*) is characterized by numerous climacograptids (dominant), glyptograptids (comparatively rare), and diplograptids (rare). Species diversity is very low: it is about five to six in the *extraordinarius* Biozone and four to five in the *persculptus* Biozone. All graptolites have a very simple morphology. Both zonal assemblages include such long-ranging species as *Normalograptus angustus*, *N. mirnyensis*, and *N. normalis*. The beds between the last occurrences of *G. persculptus* contain numerous small climacograptid rhabdosomes, a feature which has also been noted at the Dob's Linn section. The disappearance of the abundant *G.? extraordinarius* (the middle part of member *Q*) marks the last extinction at the end of the Ashgill. The onset of black shale sedimentation at the base of the *acuminatus* Biozone is the first evidence of Llandovery transgression in the region. The first significant radiation is indicated by the appearance of new genera and species at the base of the Maut Formation.

Central China: Yangtze Gorges

The condensed sequence of graptolite-bearing shales of the Wufeng Formation north of Yichang, Western Hubei has been subdivided into six biozones (Mu and Ni, 1983; Mu, 1984, 1988; Mu and Lin, 1984; Mu et al., 1984; Wang et al., 1984). The zonal assemblages are very diverse, containing both cosmopolitan and endemic species. The lowermost part of the Wufeng Formation (W_1, *yangtzeensis* Biozone) contains no more than seven species, which also extend into the succeeding biozone. The radiation among the dicellograptids and climacograptids marks the lower boundary of the *szechuanensis* Biozone. The diverse assemblage contains about 20 species, most of which have a wide geographic distribution. The first extinction event in the Ashgill took place at the beginning of the *typicus* Biozone. At this level the last dicellograptids, *Climacograptus tubuliferus*, *Geniculograptus latus*, ''*Orthograptus*'' *denticulatus*, and *Pararetiograptus sinensis* became extinct. In the middle part of the Wufeng Formation (W_3-W_5), taxonomic diversity increased again as a result of a final radiation event in the dicellograptid stock, when some specialized graptolites like *Tangyagraptus* and *Diceratograptus* appeared (*typicus* and *mirus* biozones, Mu et al., 1984; Wang et al., 1984). At the same time, the genera *Paraorthograptus* and ''*Yinograptus*'' made their first appearance, and in great abundance. The graptolite dynamics of this time are unstable, because of the extinction and appearance of several short-lived species, though both composition and diversity of successive assemblages are comparable (about 16 to 20 species). The most complete known graptolite sequence of the topmost Ashgill is represented by the upper part of Wufeng Formation (W_6, *bohemicus* Biozone). The final extinction of the last typical Ashgill taxa takes place in two successive events within a very short time span. *Orthograptus amplexicaulis*, *Geniculograptus inuiti*, the last *Dicellograptus*, *Leptograptus*, and ''*Yinograptus*'' die out at the *uniformis/bohemicus* zonal boundary. The final mass extinction of almost all Ashgill lineages takes place in the lower *bohemicus* Biozone. Only the long-ranging *N. normalis* and *N. angustus* range into the overlying beds.

Graptolites of the Ordovician-Silurian boundary beds (*bohemicus*, *persculptus*, and *acuminatus* biozones) near Yichang, in the South Anhui and Tibet are represented by variable but dense populations (Mu and Ni, 1983; Li, 1984; Li et al., 1984; Lin and Chen, 1984; Mu et al., 1984). The *bohemicus* and *persculptus* zonal assemblages include graptolites that are morphologically simple compared with the preceeding fauna. They include some species of *Glyptograptus* (dominant), ''*Diplograptus*'' (numerous), and *Normalograptus* and ''*Amplexograptus*'' (rare). Successive radiation events within these form genera coincide with the beginning of the *bohemicus* Biozone, the *persculptus* Biozone, and the *acuminatus* Biozone, respectively. In this area the last event is not as impressive as in the other regions. It is marked by the appearance of only the index species in association with long-ranging diplograptids and glyptograptids.

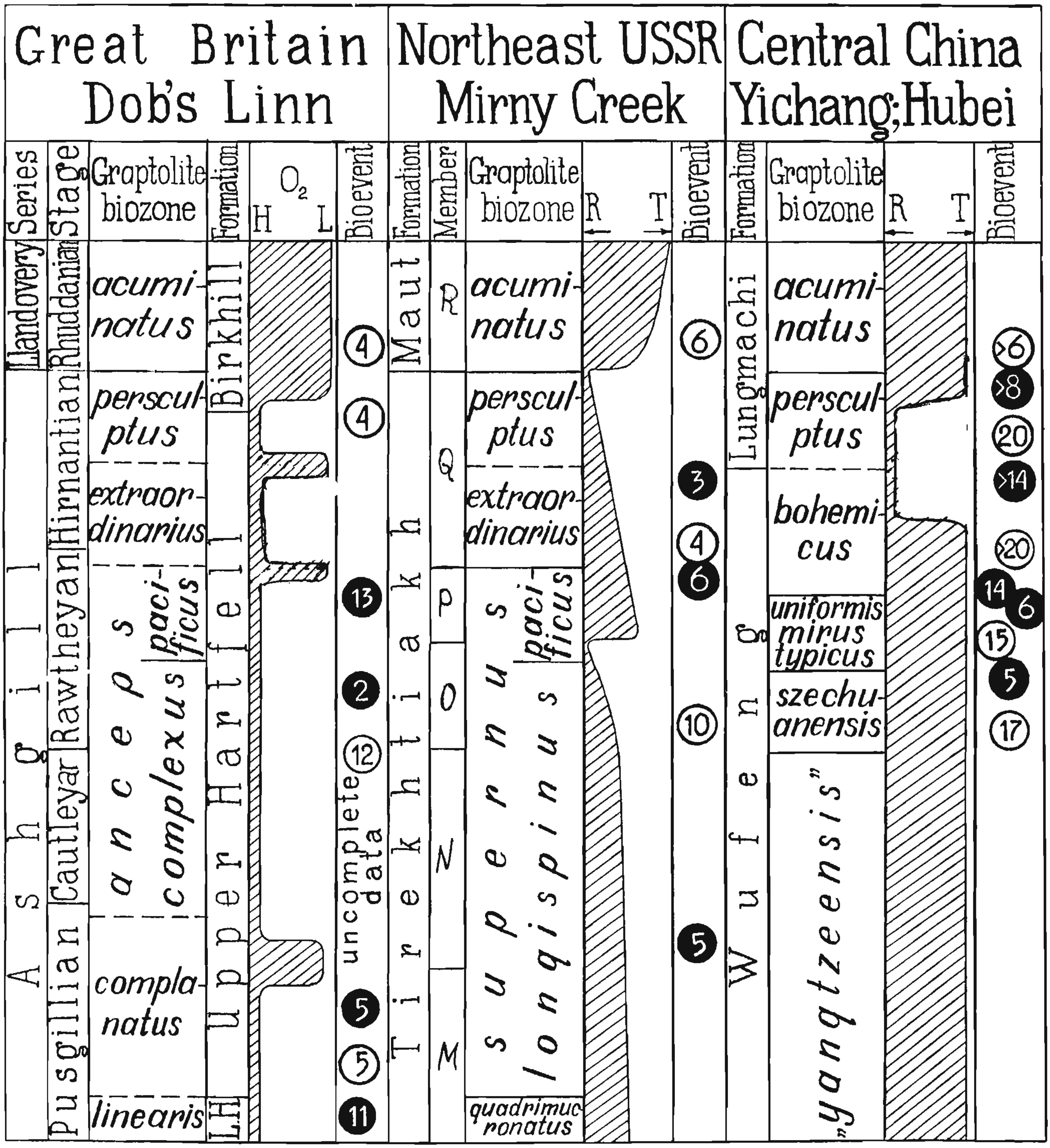

Figure 1. Ashgill-early Llandovery bioevents in graptolite evolution against interpreted changes in eustasy and oxygenation. Curves for Anticosti and Dob's Linn, and the curve showing global climatic changes are based on Barnes (1986, Fig. 1). Data on the transgressive-regressive cycles in Hubei and the Canadian Cordillera are taken from Geng Liang-Yu (in Mu [ed.], 1984) and from Lenz and McCracken (1982, 1988), respectively. Abbreviations: R, regression; T, transgression; Fm, Formation; H and L, high and low O_2 content in the bottom environment; L.H., Lower Hartfell; Mc, Macasly; E.B., Ellis Bay. Black and white circles show extinction and radiation, respectively. Arabic numerals give numbers of new and extinct species based on regional data, cited in the text. Roman numerals show successive extinction and radiation events.

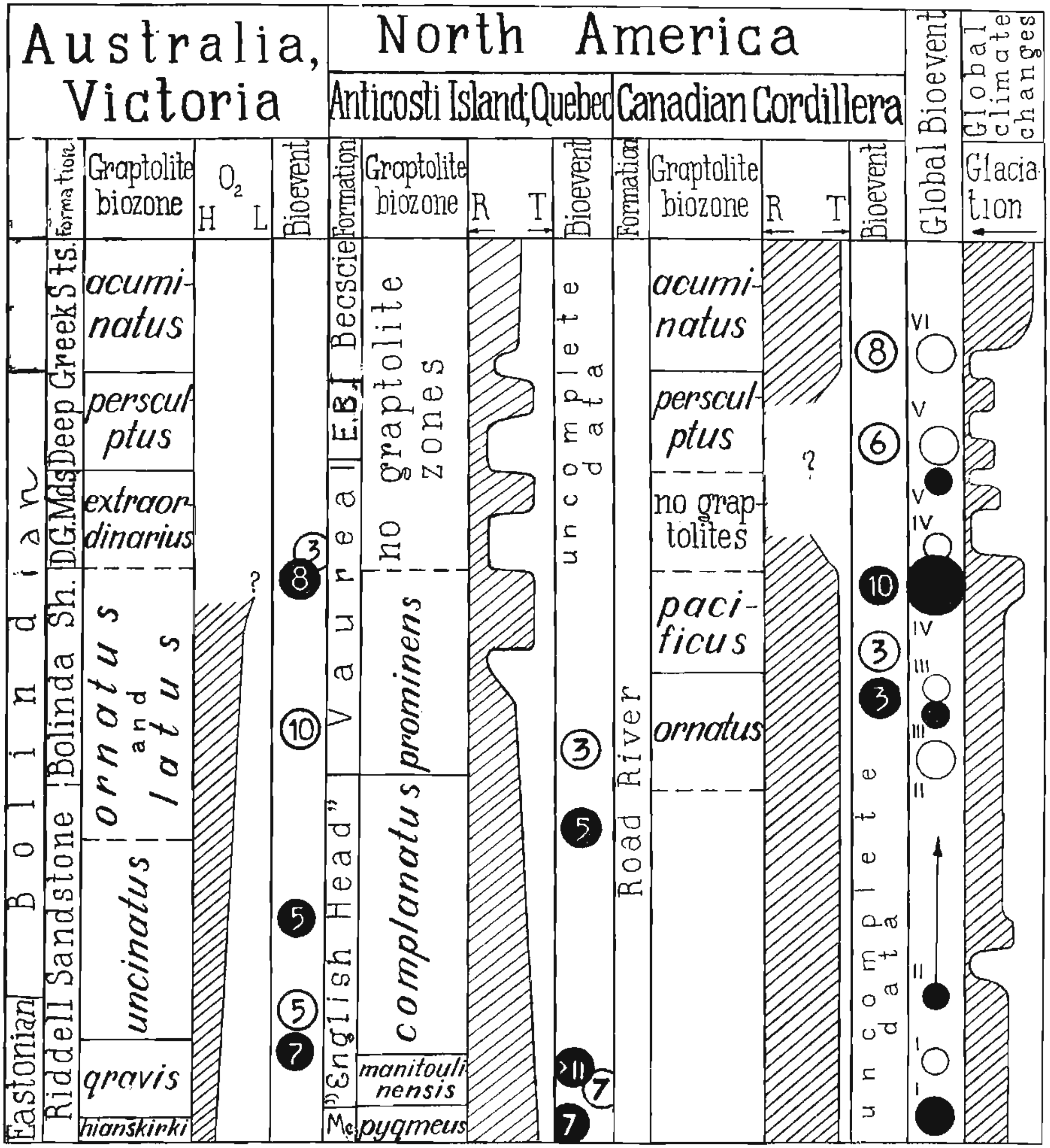

Figure 1. (cont'd.)

Australia: central Victoria

A continuous sequence of graptolite-bearing shales, indicative of the *gravis* and *uncinatus* biozones, is recognized in Eastonian and Bolindian boundary beds in the Mountain Creek and Wellington River (VandenBerg, 1981; VandenBerg et al., 1984; VandenBerg and Webby, 1988). Assemblages are represented mostly by cosmopolitan species with ranges similar to those known from other regions. Two extinction events are well expressed. The first coincides approximately with the *gravis/uncinatus* zonal boundary. It is characterized by the extinction of previous dicellograptids and orthoretiolitids. The radiation within the climacograptid stock follows this event, and the *C. longispinus*, *C. hastatus*, and "*C.*" *uncinatus* groups appear for the first time. Within the lower *uncinatus* Biozone, "*C.*" *uncinatus*, *Pleurograptus linearis simplex*, *Dicellograptus* sp. cf. *D. morrisi*, *Pararetiograptus pulcherrimus*, and others occur for the last time. The upper Bolindian graptolite sequence is known from the terrigenous rocks of Darraweit Guim. The *ornatus* and *latus* zonal associations have about the same diversity as that of the upper *gravis* Biozone (12 species). These biozones are characterized by cosmopolitan species that elsewhere are known from the *pacificus* Biozone. Almost all members of the *latus* zonal assemblage disappear at the top of the Bolinda Shale, that is, at a level coinciding with the main extinction event at the end of the Ashgill (the *pacificus* event). An impoverished diplograptid fauna characteristic of the *extraordinarius* and *persculptus* biozones (one to three species) is known from the Darraweit Guim Mudstone and Deep Creek Siltstone. In the same area the *acuminatus* Biozone is recognized by the occurrence of the zonal species. However, graptolite data from the Ordovician/Silurian boundary beds in central Victoria are insufficient to allow an evaluation of graptolite dynamics.

North America: Canadian Cordillera

Graptolite biozones of the Ashgill and the basal Llandovery have been studied in the 30 to 60 m thick shale sequence of the Road River Formation in the Richardson Mountains, northern Yukon (Lenz and McCracken, 1982, 1988). The assemblages of the *ornatus* and *pacificus* biozones are formed by cosmopolitan species. The Ashgill graptolite shales of the Descon and Maravillas formations of Alaska and Texas have similar zonal compositions (Churkin et al., 1970; Bergström, 1978). At the base of the *pacificus* Biozone the diversity of graptolites increased slightly. A mass extinction took place at the top of this biozone (the *pacificus* event). This sequence of latest Ashgill graptolites is not complete, most probably because of facies changes. Short breaks in sedimentation also cannot be excluded. The base of the *persculptus* Biozone is marked by a distinct radiation within the glyptograptid and climacograptid stocks. Among diagnostic taxa, the *Glyptograptus tamariscus* and *Normalograptus normalis-N. miserabilis* groups dominate. The base of the *acuminatus* Biozone is characterized by the appearance of a more diverse assemblage including the first *Akidograptus*, *Parakidograptus*, and *Cystograptus* as well as some new species of *Normalograptus*, *Paraclimacograptus*, and "*Diplograptus*".

North America: Anticosti Island and St. Lawrence Lowlands

The Ashgill graptolite sequence of eastern North America is known from the Utica Shale, and the English Head and Vaureal formations (Riva, 1969, 1974, 1988; Petryk, 1981). The peculiarities of their composition are especially well expressed at the Caradoc-Ashgill boundary (the *pygmeus* and *manitoulinensis* biozones) and close to the base of the Llandovery (*prominens* Biozone). However, the dynamics of graptolite associations during Ashgill times is not dissimilar to that revealed in other regions. The first extinction event is related to the end of black shale sedimentation at the top of the Utica or Macasty shales and the disappearance of the last typical graptolites of the *pygmaeus* Biozone. Here many dicranograptids, geniculograptids, and climacograptids disappear (about seven species). Radiation events within the *Climacograptus*, *Pseudoclimacograptus*, and *Diplograptus*? stocks are characterized by a discernible increase in taxonomic diversity in the *manitoulinensis* Biozone (to 18 species). The second extinction event of the older Ashgill fauna takes place at the top of this biozone. Here, "*Climacograptus*" sp. cf. "*C.*" *styloideus*, *Amplexograptus manitoulinensis*, *Pseudoclimacograptus* sp. cf. *P. clevensis*, and other taxa become extinct. The overlying *complanatus* Biozone, corresponding to most of the English Head Formation, is characterized by an impoverished assemblage of cosmopolitan dicellograptids and orthograptids (seven species). The third extinction event is documented at the base of the Vaureal Formation, where almost all typical Ashgill taxa become extinct. In the lower part of the formation only rare graptolites occur. They belong to *Arnheimograptus*, *Amplexograptus*, *Geniculograptus*, *Orthograptus*, and *Paraclimacograptus*?, constituting a peculiar assemblage of the *prominens* Biozone. The upper Vaureal Formation and Ellis Bay Formation consist of nine regressive series of shallow-water deposits and have no diagnostic graptolites. The occurrence of *Normalograptus normalis* and other long-ranging diplograptids allow a tentative correlation with the *persculptus* and, also possibly, with the *acuminatus* biozones.

CONCLUSIONS

In the evolution of Ashgill graptolites, several bioevents can be traced globally. Among them certain extinction and radiation events are recognized, even in those regions with different facies development. Thus, they reflect general tendencies in the evolution of graptolites and can, therefore, be used as good time-markers for global correlations. The events occur over relatively short, sometimes abrupt, intervals when one takes into consideration the average duration of the Ashgill graptolite biozone (Harland et al., 1982).

In addition, some bioevents may be more geographically restricted. For instance, in eastern North America there is an extinction event at the end of the *manitoulinensis* Biozone, while in central China there are two extinction episodes in the *bohemicus* and *persculptus* biozones, respectively. In addition, there seems to be a radiation at the base of the *typicus* Biozone in central China but this could well be an artifact of taxonomic oversplitting.

The succession of global bioevents in graptolite evolution can be summarized as follows.

1. The Caradoc-Ashgill boundary extinction event is characterized by the disappearance of many genera and species (Fig. 1, I). However, the taxonomic structure of the assemblages was not significantly affected.
2. The first radiation (I) is in the *complanatus* Biozone, although this is only moderate with respect to both morphological variety and the appearance of new species. The extinction pattern of these new taxa is not yet fully documented in continuous sections (II).
3. The most significant radiation event occurred within the early *complexus* Biozone, followed by a distinct acceleration of evolution during the *complexus* and *pacificus* phases (II, III). The taxonomic structure of the graptolite assemblages of this time reached a maximum diversity.
4. Insignificant extinction between radiations II and III, close to the *complexus*/*pacificus* boundary (III), can only be recognized in the most complete sections with continuous graptolite records (central China, Canadian Cordillera).
5. The major evolutionary crisis, which is expressed by the terminal mass extinction of Ashgill families, genera, and many species, takes place at the end of the *pacificus* Biozone (the *pacificus* event, IV). All morphologically specialized graptolites are eliminated; only generalists with simple morphology persist. The *pacificus* event can most probably be related to the maximum of the late Ashgill glaciation.
6. New morphological types appear together with the last representatives of the *pacificus* Biozone. These are the simple climacograptids and glyptograptids. They also become extinct but give rise to an *extraordinarius* assemblage (IV). This assemblage disappears at the end of that biozone (V).
7. Appearances of new species both at the lower boundary and within the *persculptus* Biozone are only moderate in number of taxa, but of great significance for the further evolution of graptoloids, as the Suborder Monograptina appears. Another evolutionary phenomenon, which occurs within both the *persculptus* and underlying *extraordinarius* Biozone, is a population burst in certain simple diplograptids.
8. The radiation at the beginning of the *acuminatus* Biozone is more or less comparable to that of the *persculptus* Biozone in the number of new taxa. However, it is more significant because of an appearance of several new diplograptid lineages (families, genera, and species).

The *persculptus* and *acuminatus* radiation events do not, however, restore the morphological diversity of graptolites lost at the Rawtheyan/Hirnantian boundary (the *pacificus* event).

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A chronology of North American Ordovician trilobite genera

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Abstract

Absolute duration of trilobite genera in North America is calculated for the Late Cambrian Pterocephalid and Ptychaspid biomeres and for the Ordovician. Generic duration for the two biomeres and each part of the Ordovician is distributed in a Poisson distribution. An absolute chronology for the Ordovician is calculated from Sweet's (1984) Composite Standard Section (CSS) and three high precision dates. The primary trilobite stratigraphic data are transposed into the CSS, which has been extrapolated to the beginning of the Atdabanian.

*There are 266 Ordovician non-agnostid trilobite genera that have an average duration of 12.2 m.y.; the minimum duration is about 0.1 m.y. and the maximum duration (*Shumardia*) 56.4 m.y. There are 64 genera with ranges longer than 20 m.y. The 95 Ibexian genera average 7.7 m.y., 164 Whiterockian genera average 16.4 m.y., 114 Mohawkian genera average 20.8 m.y., and 74 Cincinnatian genera average 22.1 m.y. For comparison, 32* Cedaria-Crepicephalus *Zone trilobite genera from Texas average 3.0 m.y., with a minimum of 0.2 m.y. and a maximum of 7.7 m.y. The average duration of 48 Pterocephalid biomere genera from the Great Basin is 0.9 m.y., with a minimum of 0.05 m.y. and a maximum of 3.7 m.y. The average duration of 41 Ptychaspid biomere genera from Minnesota is 1.6 m.y., with a minimum of 0.05 m.y., and a maximum of 5.3 m.y. The average duration of the Cambrian species is half that of the generic durations, and a similar or slightly smaller ratio appears to hold for the Ordovician.*

Résumé

On a déterminé la durée d'existence absolue de genres de trilobites de l'Amérique du Nord pour les biomères de Ptérocéphalidés et de Ptychaspidés du Cambrien supérieur et pour l'Ordovicien. La loi de Poisson illustre la durée d'existence des genres dans les deux biomères et chaque partie de l'Ordovicien. L'établissement d'une chronologie absolue de l'Ordovicien se fonde sur la coupe composée standard de Sweet (1984) et sur trois dates très précises. Les principales données stratigraphiques de trilobites sont transposées dans la coupe composée standard, qui est extrapolée jusqu'au début de l'Atdabanien.

*Il existe 266 genres de trilobites ordoviciens non-Agnostidés qui ont une durée d'existence moyenne de 12,2 ma; la durée d'existence minimale est d'environ 0,1 ma et la durée d'existence maximale (*Shumardia*), de 56,4 ma. Soixante-quatre genres ont existé pendant plus de 20 ma. Les 95 genres de l'Ibexien ont une durée d'existence moyenne de 7,7 ma, les 164 genres du Whiterockien, de 16,4 ma, les 114 genres du Mohawkien, de 20,8 ma, et les 74 genres du Cincinnatien, de 22,1 ma. Aux fins de comparaison, 32 genres de trilobites de la zone à* Cedaria-Crepicephalus, *au Texas, ont une durée d'existence moyenne de 3,0 ma, avec un minimum de 0,2 ma et un maximum de 7,7 ma. Quarante-huit genres du biomère de Ptérocéphalidés, dans le bassin Great, ont une durée d'existence moyenne de 0,9 ma, avec un minimum de 0,05 ma et un maximum de 3,7 ma. La durée d'existence moyenne de 41 genres du biomère de Ptychaspidés, au Minnesota, est de 1,6 ma, le minimum étant de 0,05 ma et le maximum, de 5,3 ma. La durée d'existence moyenne des espèces du Cambrien est la moitié moins de celle des genres; le rapport pour l'Ordovicien serait semblable ou légèrement moins élevé.*

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INTRODUCTION

This study was done for several reasons. There are few data on the durations of genera and species over long spans of time. There are theoretical reasons for suspecting a progressive increase in duration as adaptive radiation progresses. It is currently not possible to make a precise census of duration of taxa on a worldwide basis. The problems include those of correlation and time scale, as well as those of density of information. However, as a result of recent work on conodont biostratigraphy, it is possible to review the data for the Ordovician of North America. This study covers all the available literature on trilobite biostratigraphy of North America during the Ordovician, as well as Greenland, Spitzbergen, northwestern Scotland (especially the Girvan district), western Newfoundland, and northwestern Ireland. This study is primarily concerned with taxa at the generic level. A similar study at the species level is possible, but will require several more years. For comparison, the more detailed data for North America during the Cambrian is also described.

CAMBRIAN-ORDOVICIAN TIME SCALE

The major problem with the early Phanerozoic time scale is evaluation of data with varying precision, and interpolation between good dates. Snelling (1984) suggested that all early Paleozoic chronometric boundaries of periods, epochs, and ages should be rounded to the nearest 5 million years. I strongly disagree, as this discards all of the very large quantity of information that we have on the relative duration of these units. The time scale is too important to be left solely to those who measure ages.

The best time scales are the result of a synthesis by linear interpolation of time with a continuous independent variable such as the Cretaceous and Cenozoic magnetostratigraphic record derived from seafloor spreading data. Traditionally, developers of time scales have consulted the relative thickness of stratigraphic units as a way of interpolating between dates.

Shaw's (1964) method of graphic interpolation by the development of Composite Standard Sections is the most precise way known of partitioning the available time based on stratigraphic thickness. Sweet (1984) developed such a Composite Standard Section (CSS) for the Whiterockian to the end of the Ordovician by pooling data on conodont distribution from 81 measured and sampled sections spread throughout the United States.

Three good radioactive dates have been determined with high precision (two sigma errors of 2 m.y.) in or close to the Ordovician Period. There are also precise biostratigraphic data available. Two of these dates, Churkin's earliest Silurian date of 436 m.y. and Kunk and Sutter's (1984) date on the Deicke K-bentonite are included within or are very close to Sweet's CSS. Churkin's date provides a date of 438 m.y. for the end of the Ordovician. The Deicke date provides an age of 454 m.y. for the Black River-Trenton boundary.

Sloan (1987b) converted Sweet's (1984) CSS to a time scale for purposes of analysis of rates of evolution and sedimentation in the upper Mississippi Valley region. The equation for conversion of a CSS number to an absolute age is:

$$\text{Age} = 0.0544\ (1277 - \text{CSS}) + 438.$$

The chronology of the North American Ordovician from the base of the Whiterockian to the end of the Ordovician now rests on a very sound foundation. Unfortunately, the chronology of the type British series is less certain. It is based mainly on less precise methods, and is unlikely to improve except by correlation with North America.

The chronology can be extended to the Ibexian/Canadian with some ease. Sloan calculated the age of the end of the Ibexian to be 483.7 m.y. based on the value of CSS 387 that Sweet (1984) calculated for top of the series. An assumption of 504 m.y. for the Cambrian-Ordovician boundary (Ross et al., 1982) produces a duration for the Ibexian of 20.3 m.y. In turn, this determines the base of the Ordovician and the Ibexian as CSS 64. The temporal partitioning of the Ibexian in this study is based on the thickness of the Ross-Hintze trilobite zones at Ibex (Hintze, 1953, 1979, 1982). Unfortunately, the distribution of trilobites has only been reported to the zonal rather than the absolute stratigraphic level. Nonetheless, because of its rich fossil content, this is the best possible reference section for the entire Ibexian/Canadian.

The Cambrian time scale poses unique problems. There are few biostratigraphically controlled dates, and those that exist are of low precision. It is, however, possible to extrapolate the Ordovician time scale backward, by using the measured thickness of strata in the Great Basin. This seems reasonable because there are no major changes in tectonic style of sedimentation between the end of the Waucoban and the Whiterockian in the Great Basin. The thickness of the Ptychaspid biomere at the Ibex section (Miller et al., 1982) is 375 m compared to 805 m for the Ibexian (Hintze, 1982) at the same place. This suggests that the duration of the Ptychaspid biomere is 46.6 per cent of the Ibexian, or 9.5 million years.

Palmer (1971) presents data on the relative thicknesses of the biomeres and trilobite zones of the Eureka-House Range region (which includes the Ibex section) for the Middle and Late Cambrian, and the Inyo Mountains for the Early Cambrian. By simple ratios of thickness, the duration of the Ibexian and the Ptychaspid biomere can be extended (Table 1) to produce durations of the other Albertan-Croixan zones and biomeres. Naeser's (*in* Ross and Naeser, 1984) fission track date of 535 ± 12 m.y. for a tuff in the *Bolaspidella* Zone (estimated duration 527 to 540 m.y.) agrees closely with this extrapolation.

Unfortunately, Naeser's other fission track date of 563 ± 12 m.y. from the *Glossopleura* Zone is not quite as close to the extrapolated range of the *Glossopleura* Zone from 543.5 to 551 m.y., but the agreement is still satisfactory.

Table 1. Time scale of North American Cambrian and Ordovician zones, stages, and series as used in this paper

	Duration (m.y.)	Age at base (m.y.)
Silurian		438
Gamachian	0.5	438.5
Richmondian	3.9	442.5
Maysvillian	3.3	445.7
Cincinnatian, Edenian	3.8	449.5
Shermanian	3.7	453.2
Kirkfieldian	0.5	453.7
Trentonian, Rocklandian	0.5	454.2
Blackriveran	4.0	458.2
Mohawkian, Ashbyan	5.3	463.5
Chazyan	7.0	470.5
Whiterockian	13.2	483.7
Ibexian/Canadian	20.3	504
Ptychaspid biomere	9.5	513.5
Pterocephaliid biomere	4.8	518.3
Croixan, *Cedaria-Crepicephalus*	8.7	527.0
Bolaspidella	13.1	540.1
Bathyuriscus-Elrathina	3.4	543.5
Glossopleura	7.5	551.0
Albertella	1.8	552.8
Albertan, *Plagiura-Poliella*	1.3	554.1
Lenian, upper olenellid	14.1	568.2
Atdabanian, lower olenellid	8.5	576.7
Waucoban, Tommotian	9.4	586.1

A similar extension to the Waucoban poses more problems. The only long North American section including the entire Waucoban is that of the Inyo Mountains of southern California. This section, summarized in Palmer (1971), includes strata that can be recognized as of Tommotian, Atdabanian, and Lenian age as well as the Albertan and Croixan series. A simple extrapolation of the combined Albertan and Croixan duration of 50 m.y. back to the base of Tommotian rocks (shelly pre-trilobite Cambrian) produces an age of 616 m.y. for the base of the Cambrian. This seems excessive in view of the suggested ages for the base of the Phanerozoic listed in Cowie and Johnson (1984), in which the oldest possible age is about 600 million years. A solution to this dilemma lies in the position of the Inyo Mountain section. This section appears to be off the edge of the craton on the continental slope during the Waucoban, but on the craton in the Albertan-Croixan. A faster rate of rock accumulation during the Waucoban in the Inyo Mountains, about three times faster than during the Albertan and Croixan, would neatly resolve this conflict and make the base of the Phanerozoic 586 million years, in the middle range of the usual estimates. This ad hoc solution is adopted for the purposes of this paper. The age of the base of the Cambrian may well be even younger. In terms of evolutionary rates, this relatively young age seems reasonable because all other adaptive radiations involve rapid early evolution and then slow exponentially in rate. Failure to adopt this assumption would imply the reverse: slow evolution at first, followed by rapid evolution. This seems unlikely.

REVISED CLASSIFICATION

The classification of trilobites in this paper follows that of the Treatise (Moore, 1959) with modifications suggested by Fortey and Owens (1975, the order Proetida), Lane and Thomas (1983, the suborder Scutelluina), Fortey and Chatterton (1988, reclassification of the suborder Asaphina), and Thomas and Owens (1978). The classification of Bergström (1973) has not been followed.

BIOSTRATIGRAPHY

When Palmer proposed the biomere concept (1965a), he originally thought the upper and lower boundaries were slightly diachronous. More recently, Palmer (1984) and others have regarded them as unique extinction events, involving no break in deposition and substantially simultaneous over the craton. The upper Cambrian or Croixan trilobite biostratigraphy in North America has been studied in great detail, with metre by metre collections made and identified from many area. Only a few similar studies have been made on earlier Cambrian rocks, with that of Palmer and Halley (1979) a notable example. Most of the rest of the Cambrian studies have been at the level of the zone, with little or no detail on shorter durations. Figure 1 shows the number of North American Cambrian and Ordovician trilobite genera as a function of absolute age, based on the time scale adopted for this paper. Numbers of trilobite genera for Cambrian zones and subzones were taken from Lochman-Balk and Wilson (1958), the last complete review of Cambrian trilobite zonation, with some upgrade from Holland (1971), Palmer (1955, 1965b), and Stitt (1971, 1977). Numbers of Ordovician genera are from the current study.

The pattern of Cambrian trilobite history is notably different from that of the Ordovician. There are five cycles of radiation followed by extinction, the classic biomeres, during the Cambrian. These are 1) the Atdabanian-Lenian Olenellid biomere, terminated by an extinction of 72 per cent; 2) the early Middle Cambrian *Plagiura-Poliella* Zone to *Glossopleura* Zone biomere terminated by a 36 per cent extinction; 3) the late Middle and early Late Cambrian Marjumiid biomere, terminated by an 80 per cent extinction; and 4, 5) the Late Cambrian Pterocephaliid and Ptychaspid biomeres,

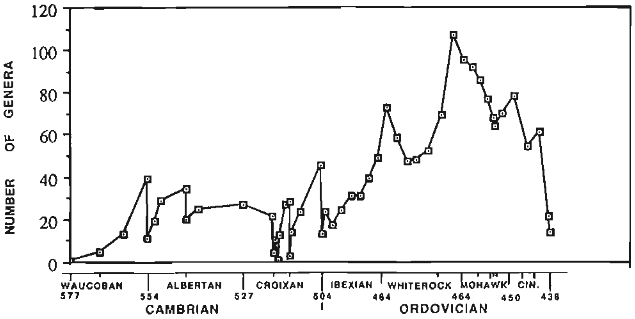


Figure 1. Number of North American trilobite genera plotted by zone or subzone (Cambrian) or 50 to 75 CSS units (Ordovician).

each terminated by extinctions of 95 per cent. The durations of Cambrian trilobite genera are short because of this pattern of multiple radiation, and because they have been reset by extinction.

In contrast, the diversity of Ordovician genera as a function of time is much simpler. The mean duration of trilobite genera increases from 7.7 m.y. for the Ibexian, to 16.4 m.y. for the Whiterockian, 20.8 m.y. for the Mohawkian, and 22.1 m.y. for the Cincinnatian. A plot of numbers of trilobite genera in North America during the Ordovician, at intervals of 2.5 to 5 m.y. shows a linear rise from 13 at the base of the Ibexian, to 73 at the base of the Whiterockian, and a gradual sag to a nadir of 47 during the middle third of the Whiterockian (a gradual extinction of 36 per cent), clearly due to the well known eustatic drop in sea level. It is followed by a linear rise to an absolute peak of 107 genera at the top of the Whiterockian. This is the greatest diversity of trilobites of all time. A linear decrease to 61 Richmond genera is interrupted only by the relatively minor catastrophic extinction in a tenth of the continent as a result of the Deicke eruption at the end of the Blackriveran. Minor increases in diversity are the result of recruitment from across the Iapetus, and evolution in situ. In the last 2 or 3 m.y. of the Ordovician, terminal extinction resulted in a reduction to 14 genera of trilobites, an extinction of 73 per cent.

Within each biomere or Ordovician series, the mean duration of trilobite genera almost exactly equalled the standard deviation of the durations. This property is typical of the highly skewed frequency distribution known as the Poisson Distribution, in which the majority of occurrences are less than the mean and there is a long tail of rare occurrences greater than the mean. Thus, although the mean of the 266 Ordovician non-agnostid trilobites for which there were at least two stratigraphically different occurrences was 12.2 m.y., the standard deviation was 11.8 m.y. and only 67 trilobite genera had a duration greater than 20 m.y. The genus with the longest distribution was *Shumardia* at 56.1 m.y.

One side effect of this analysis is that there is no sign, either in the Cambrian or the Ordovician, of the 26 to 30 m.y. cycle of major extinctions proposed by Raup and Sepkoski (1984). As a result of the reduction in frequency of extinctions in the Ordovician as opposed to the Cambrian, generic duration (Fig. 2) increased steadily with time.

CROIXAN BIOSTRATIGRAPHY

Shaw (1964) produced a composite section for the trilobites from the Late Cambrian part of the Marjumiid biomere from the Riley Formation of Texas, the classic *Cedaria* and *Crepicephalus* zones, which were described by Palmer (1955). From the earlier calculations, this part of the biomere is 8.7 m.y. in duration. The 21 genera have a mean duration of 3.0 m.y., standard deviation of 1.9 m.y., minimum duration of 0.2 m.y., and maximum duration of 7.7 m.y. The 32 species have a mean duration of 1.6 m.y. with a standard deviation of 1.6 m.y. The minimum duration is 0.2 m.y. and the maximum duration is 7.7 m.y.

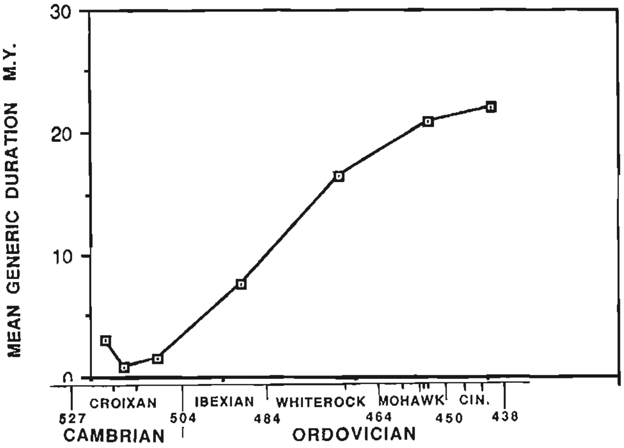


Figure 2. Mean duration of North American trilobite genera of the Croixan biomeres and Ordovician series.

The Pterocephaliid biomere was calculated above to be 4.8 m.y. in duration. Palmer's (1965b) Plate 21 summarizes the distribution of trilobites in this biomere throughout the Great Basin. On that basis, the 48 genera in this biomere average 0.9 m.y. in duration, the shortest generic duration is 0.05 m.y., and the longest is 3.7 m.y. Durations of 112 species are from 0.04 to 2.2 m.y. with the average and standard deviation both 0.4 m.y. The mean duration of species increased exponentially through the biomere.

Data for the 9.5 m.y.-long Ptychaspid biomere are available for Oklahoma (Stitt, 1971, 1977) and for the Llano region of Texas (Longacre, 1970). For the type area of the Croixan, the upper Mississippi Valley, data are presented in this report. Based on the original stratigraphic data for the collections summarized in Berg et al. (1956), the 41 genera have an average duration of 1.6 m.y., the minimum duration is 0.05 m.y., and the maximum duration is 5.3 m.y. As is the case in the other biomeres, durations were short at the beginning, and exponentially increased with time.

ANALYSIS OF THE ORDOVICIAN DATA

Analysis of the Ordovician data proceeded as follows. CSS numbers were calculated for the measured section of each paper, either directly from Sweet (1984) or from an analysis of conodont distribution, or less usually by direct physical correlation of beds, and in a few cases by matching of the first occurrences of trilobites. In each part of the Ordovician, analysis began on the trilobites from the same sections as those containing the conodonts that Sweet had studied. The data were robust, that is, after the first few rounds, changes in the distribution of common and abundant trilobite genera were small. The main outlines of trilobite distribution changed little from the stage where 185 trilobites were plotted, to the final stage where 266 genera were plotted. CSS numbers of first and last occurrences of trilobite genera were revised after each new paper was introduced into the system. The final version is given in Table 2.

IBEXIAN BIOSTRATIGRAPHY

The principle references consulted on the trilobite biostratigraphy of the Ibexian/Canadian are: Poulsen (1937), Ross (1952, 1953), Hintze (1953, 1979, 1982), Kindle and Whittington (1958), Robison and Pantojas-Alor (1968), Dean (1970, 1971, 1973), Fortey and Skevington (1970), Fortey and Bruton (1973), Fortey (1974a, b, 1975a, b, 1979, 1980, 1983), Whittington and Hughes (1974), Stitt (1977), Rushton and Tripp (1979), Fortey et al. (1982), Ludvigsen (1982a, b), and Miller et al. (1982). The standard section is the Ibex section of Hintze (1953, 1979, 1982), but as in Ross (1952), the detail is confined to the level of the zones, with no distribution data at the finer level. The Ibexian zones are: B = CSS 76-138; C = CSS 138-140; D = CSS 140-148; E = CSS 164-188; F = CSS 188-200; G1 = CSS 200-230; G2 = CSS 230-298; H = CSS 198-339; I = CSS 339-362; J = CSS 362-384; K = CSS 384-387. The continuing Whiterockian zones at Ibex are L = CSS 387-415; M = CSS 415-463; and N = CSS 463-515. There is much finer detail in the Cambrian-Ordovician boundary zones (CSS 64-79), and Fortey's Spitzbergen studies (CSS 140-495) provide very fine detail across the Ibexian-Whiterockian boundary. The average duration for agnostid Ibexian genera is 7.7 m.y.

WHITEROCKIAN BIOSTRATIGRAPHY

In addition to the references already listed for the Ibexian, the major Whiterockian references are: Whittington (1961, 1965b), Whittington and Kindle (1963), Shaw (1968, 1974), Ross (1970, 1972), and Ross and Shaw (1972). The major studies are those of Ross (op. cit.) for the Great Basin (CSS 350-750), Shaw (1968, 1974) for Oklahoma (CSS 350-825) and the Chazyan of New York (CSS 680-800), and Whittington (1961, 1965b) (CSS 387-680) for western Newfoundland. The average duration for 164 Whiterockian non-agnostid genera is 16.4 m.y.

MOHAWKIAN BIOSTRATIGRAPHY

The distribution of trilobites during the Mohawkian is based on data from Clarke (1894), Bradley (1925, 1930), Teichert (1937), Whittington (1941, 1950a, b, 1952a, b, 1953, 1954, 1956, 1959, 1963, 1965a, b, 1966, 1968), Evitt (1951, 1953), Frederickson and Pollack (1952), Cooper (1953), Weiss (1953), Whittington and Evitt (1953, 1954), Tripp (1954, 1962, 1965, 1967, 1976, 1979, 1980a, b), Williams (1962), DeMott (1963, 1987), Ross (1967), Shaw (1974), Chatterton and Ludvigsen (1976), Ludvigsen (1976, 1977, 1978a, b, 1979a, b), Evitt and Tripp (1977), Shaw and Fortey (1977), Chatterton (1980), Ludvigsen and Chatterton (1980, 1982), Tripp et al. (1981), and Tripp and Evitt (1981, 1983, 1986). The most extensive data come from the silicified faunas of Virginia studied by Whittington and Evitt (1953, 1954) (CSS 775-1000), from the very detailed work of Tripp (1954, 1962, 1965, 1967, 1976, 1979, 1980a, b), Williams (1962), and Paul (1981) on the Girvan faunas of Scotland (CSS 718-1277), and the silicified faunas of the Mackenzie Mountains of the Northwest Territories studied by Chatterton and Ludvigsen (1976) (CSS 750-1150). The work of Clarke (1894), Bradley (1925, 1930), Weiss (1953), and DeMott (1963, 1987) amply covers the upper Mississippi Valley region (CSS 950-1100). The average duration of 114 Mohawkian non-agnostid trilobite genera is 20.8 m.y.

CINCINNATIAN BIOSTRATIGRAPHY

The distribution of trilobites during the Cincinnatian is based on data from Slocum (1913, 1916), Troedsson (1928), Twenhofel (1928), Cooper (1930), Cooper and Kindle (1936), Dalve (1948), Bolton (1972), and Lespérance (1968). The only useful reference on the trilobite distribution of the type Cincinnatian is Dalve (1948), a summary and compilation of many minor observations. A major taxonomic and biostratigraphic review is in order. Slocum (1913, 1916) provides data on the distribution of Upper Ordovician trilobites in the upper Mississippi Valley. The other references are for Percé and Anticosti Island. In addition, there are data for the Maysvillian Whitehouse Group of the Girvan district of Scotland in Williams et al. (1972) and Thomas et al. (1984). The average duration of 74 Cincinnatian non-agnostid genera is 22.1 m.y.

EVOLUTIONARY ASPECTS

Durations of genera within the Ordovician and the three well sampled biomeres follow a distinctive pattern. Early in the interval concerned, durations are short, and they increase exponentially with time. The mean duration of genera is highly dependent on the elapsed time since the last major extinction (Fig. 3), fitting a second-order polynomial very well. Neither the small Deicke extinction nor the early Whiterockian extinction at 36 per cent was large enough to reset the evolutionary rates of trilobites. In many ways, the entire Ordovician corresponds to a typical Cambrian biomere.

Each Cambrian biomere and the entire set of Ordovician trilobites represents a normal adaptive radiation; each is exactly comparable to the latest Cretaceous and Paleocene adaptive radiation of ungulates (Sloan, 1987a) in progressive duration of taxa, and rates of evolutionary change.

There are only three significant extinctions in the record of trilobites in North America during the Ordovician. The first is the mid-Whiterockian extinction, which extends from CSS 450 to CSS 700, or from 483 to 469 m.y. This extinction involved a drop of 36 per cent from the peak number of trilobite genera in the earliest Whiterockian. It is most easily explained by the MacArthur and Wilson (1967) theory of island biogeography. There is a reduction of over 50 per cent in the total area of the continental shelf open to shallow water trilobites between the late Ibexian and the mid-Whiterockian. This reduction in area is due to the well known eustatic drop in sea level that produced the "Knox unconformity" and the end of the Sauk Sequence.

The second is the catastrophic extinction over the eastern half of the U.S.A. produced by the Deicke K-bentonite eruption (CSS 979) at the end of the Blackriveran. The extinction rate was about 17 per cent of the trilobite genera in the continent. The short duration of the Rocklandian and Kirkfieldian stages is a direct result of the rapid community

Table 2. Graphic representation of the ranges of North American trilobite genera by series, stages, and CSS units

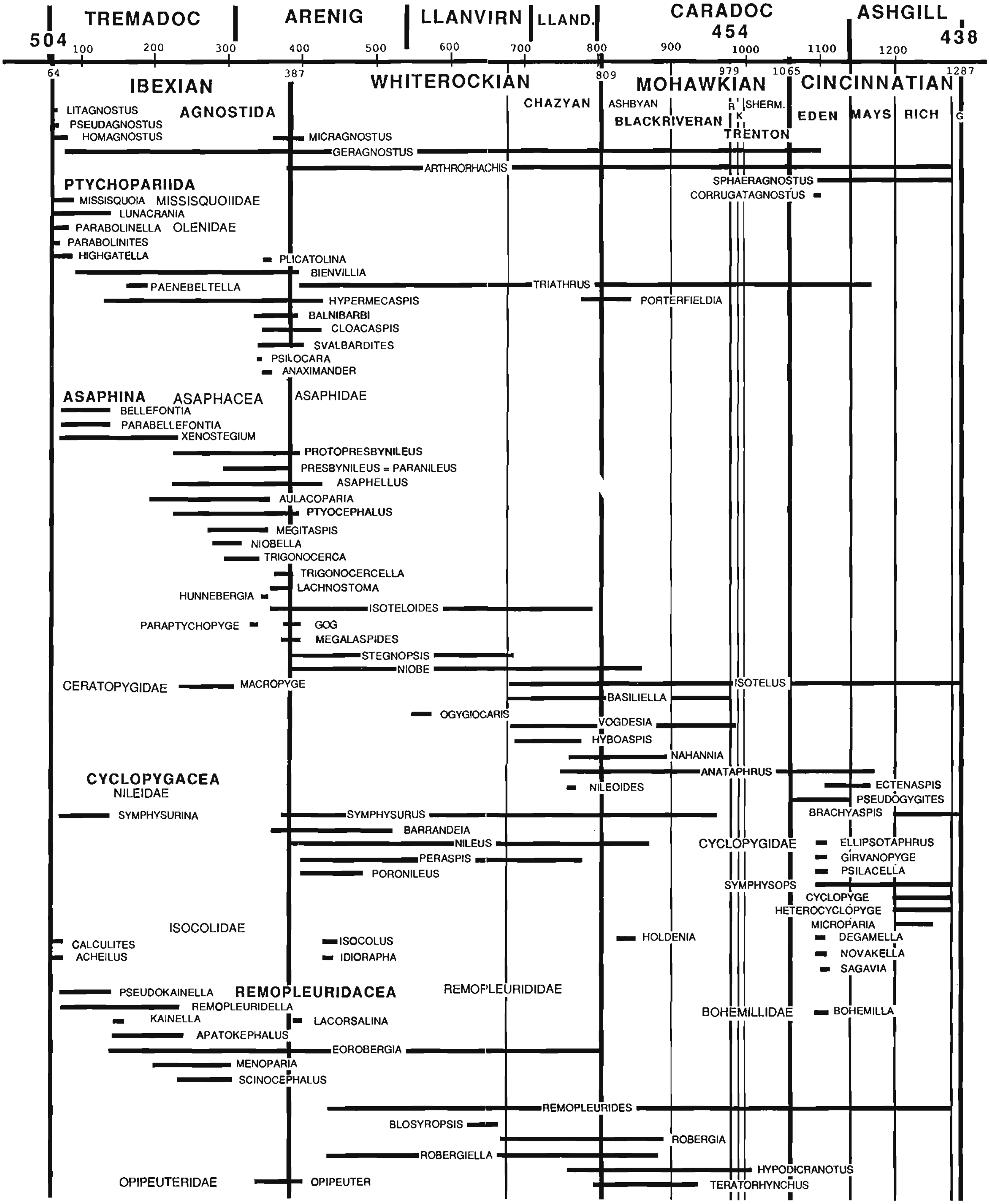
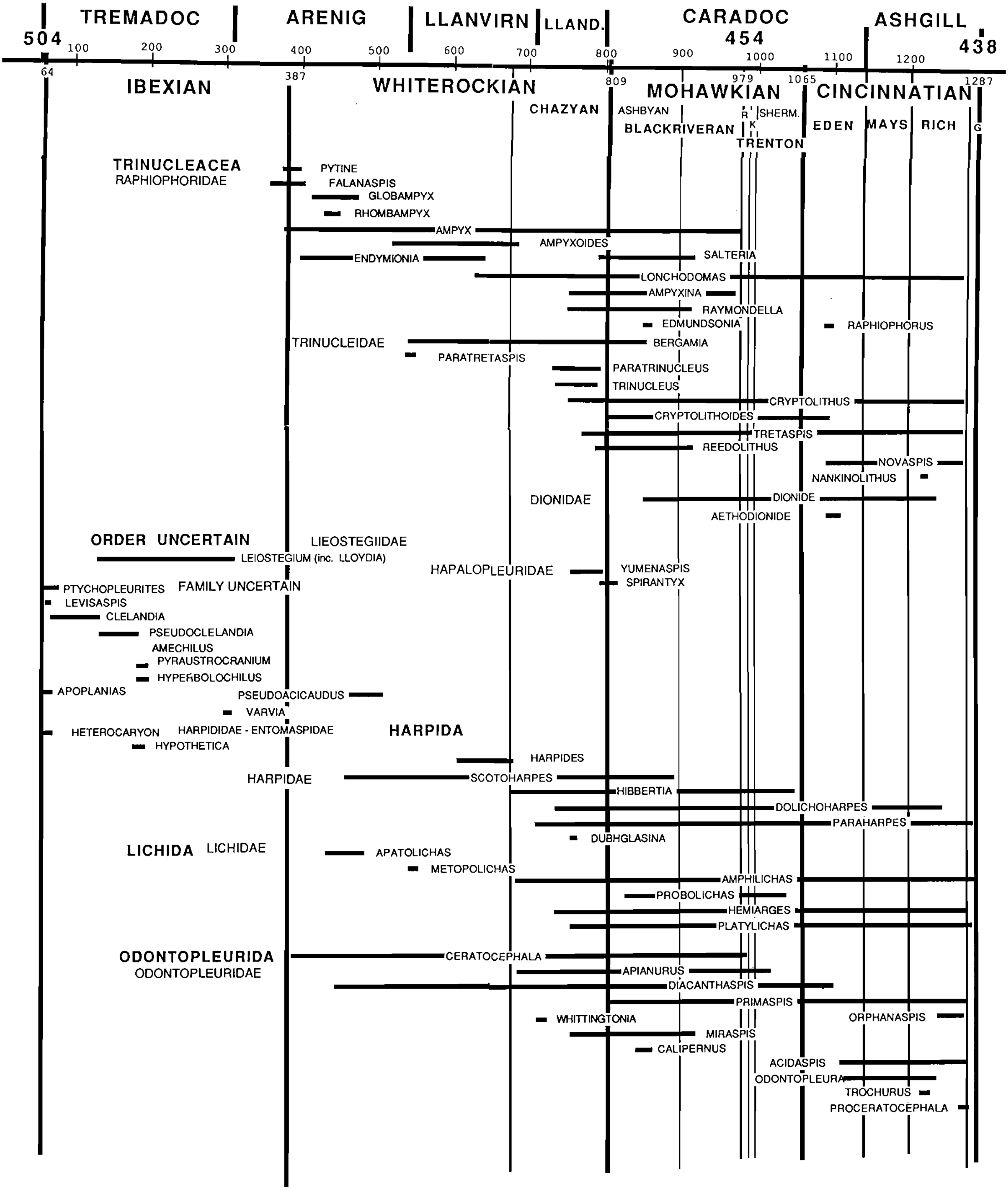


Table 2 (cont'd.)



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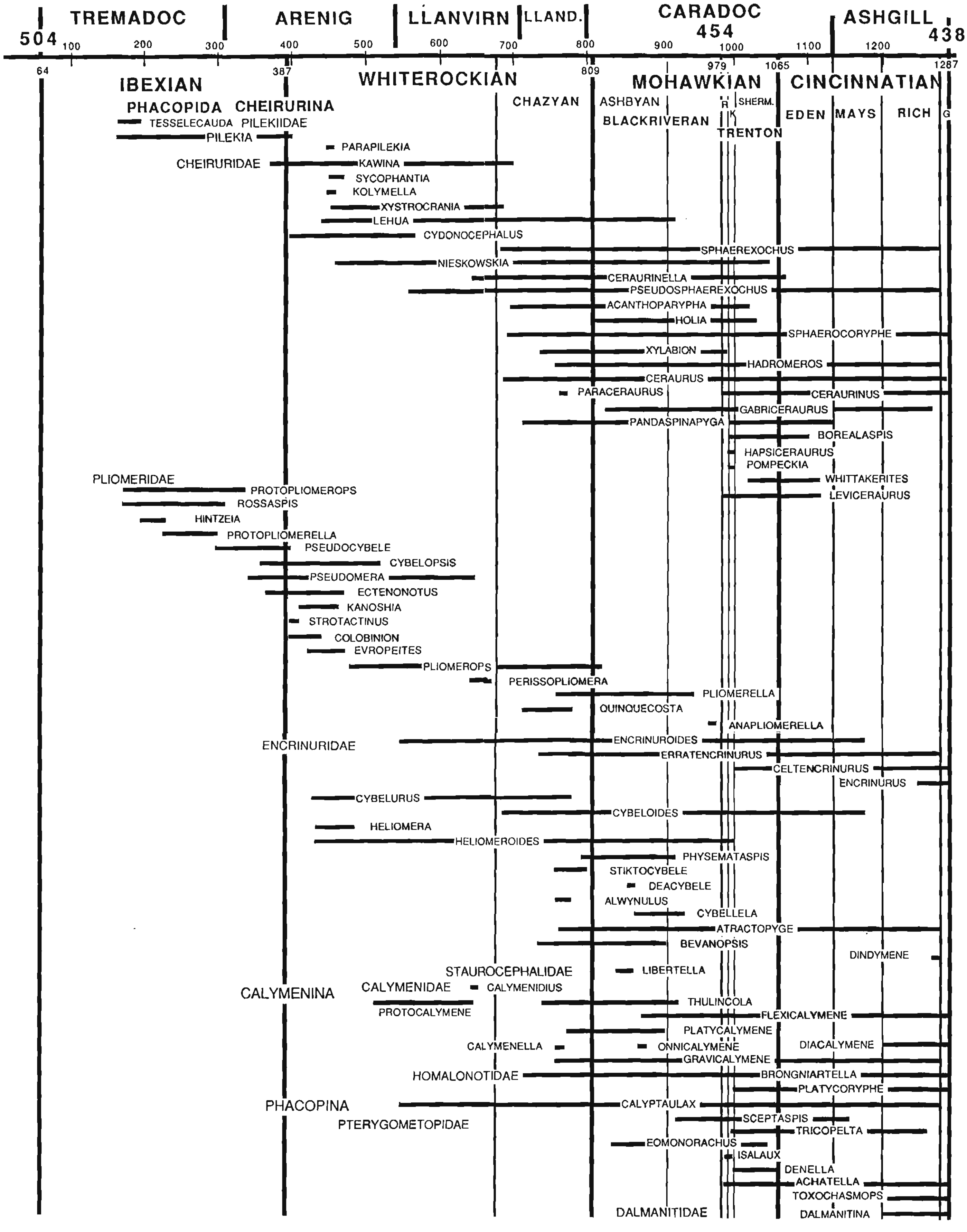

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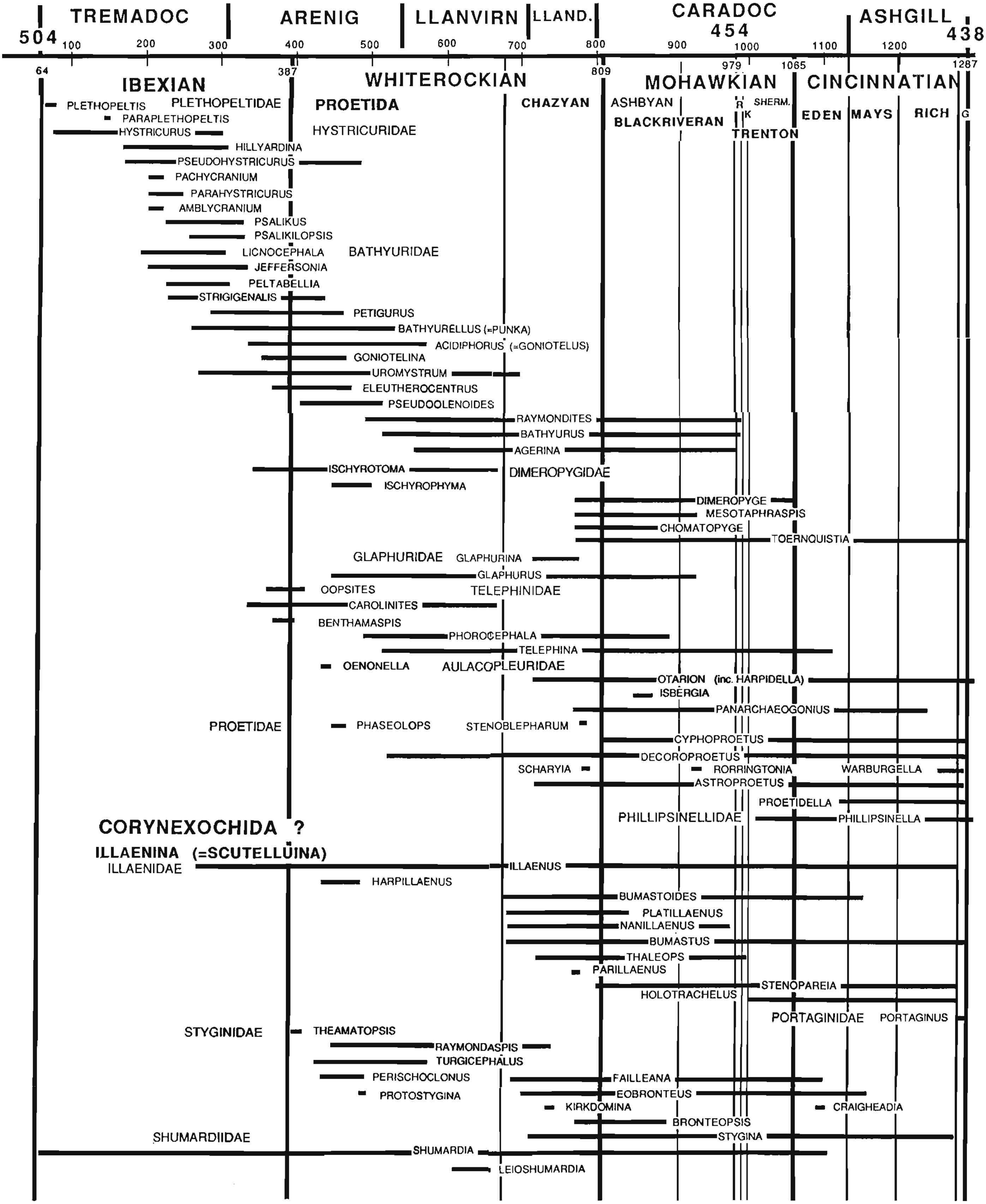

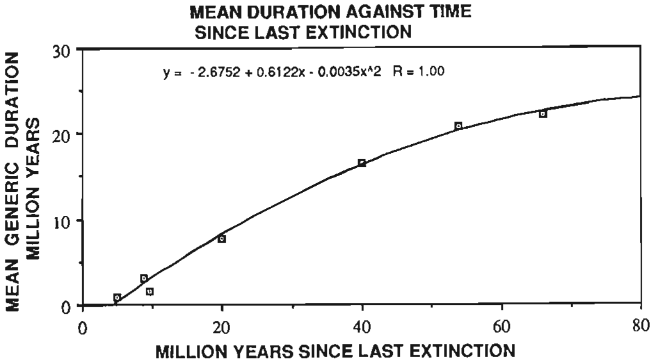

Figure 3. Plot of mean duration of North American Croixan and Ordovician trilobite genera against time in million years since the last major extinction.

changes made during the repopulation of the devastated area. Initial repopulation came first from deeper water faunas along the present eastern coast, then from shoal waters in Oklahoma, and finally, with about equal facility from the other side of the Transcontinental Arch or from across the Iapetus. Initial repopulation first occurred in trilobites with planktonic larvae and later in forms with nonplanktonic protaspis.

The third extinction is the well known terminal Ordovician extinction of eustatic origin. It resulted in a loss of 73 per cent of the Richmondian trilobites and is the biggest trilobite extinction after the Cambrian biomeres.

Most of the taxa in the Ibexian originated as immigrants from deeper water or from other cratons. Each immigrant species then underwent a radiation on this craton, with initial short duration of species, and rapid diversification of genera to produce distinctively North American families and subfamilies. Rates of immigration increased in the Mohawkian, and became more rapid and progressive in the Trentonian, and the Cincinnatian.

Chatterton and Ludvigsen (1976) documented a series of four trilobite shelf biofacies of increasing depth. In order, they are the *Bathyurus, Isotelus, Calyptaulax-Ceraurinella,* and *Dimeropyge* biofacies. Sloan (1987b) calibrated the first three as being from water shallower than 50 metres, on the basis of coexistence with Chlorophyta. Fortey (1983) added the olenid biofacies of deep water trilobites, and Fortey and Owens (1987) added two Ordovician deep water biofacies. These are the cyclopygid biofacies of large-eyed mesopelagic trilobites that occur deeper than 200 m and the atheloptic benthic biofacies of blind or small-eyed trilobites thought to occur at depths greater than 300 m. As previously documented by Whittington (1966), Whittington and Hughes (1972, 1974) and others, the pioneer immigrations across the Iapetus to North America first occurred in deep water, with later migrations onto the shelf. Three examples of Whiterockian or Blackriveran migrants that moved onto the shelf after the Deicke eruption at the end of the Blackriveran are *Flexicalymene, Brongniartella,* and *Cryptolithus.*

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Hirnantian trilobites and brachiopods in space and time

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Abstract

A global review of the range and distribution of Hirnantian shelly faunas reveals some marked contrasts. The brachiopods show a distinct latitudinal zonation, with the various associations of the typical Hirnantia Fauna widespread in subtropical and temperate latitudes, and the very different, diverse, Edgewood Fauna typifying equatorial regions. The Hirnantia Fauna includes some highly diverse assemblages and represents a mixture of some of the more eurytopic taxa from several earlier biotas as well as some genera whose provincial origins are not clear. The Hirnantia Fauna became extinct prior to the end of the Hirnantian. The trilobites suffered major extinction at the Rawtheyan-Hirnantian boundary and few Hirnantian assemblages contain more than 10 genera (most considerably less). Most genera are known from only a few sites and are simply relicts of earlier faunas in the area. The only commonly occurring association is that of a dalmanitine (Dalmanitina/Mucronaspis) and a homalonotid (Brongniartella/Platycoryphe), in some cases with the odontopleurid Leonaspis. This association had a greater depth and latitudinal range than that of the typical Hirnantia Fauna and persisted to the end of the Hirnantian. Analysis of the fate of Rawtheyan trilobite genera across a wide ecological spectrum shows that extinctions were disproportionately concentrated in deeper water environments. The same may apply to the brachiopods, with the extinction of most elements of the deep water Foliomena Fauna at the end of the Rawtheyan.

Résumé

Une étude de la répartition stratigraphique et géographique mondiale des faunes coquillères hirnantiennes fait ressortir certaines différences marquées. Les brachiopodes présentent une nette zonation latitudinale, les diverses associations de la faune à Hirnantia typique étant répandues à des latitudes subtropicales et tempérées tandis que la faune d'Edgewood, variée et très différente, caractérise les régions équatoriales. La faune à Hirnantia comprend quelques assemblages très variés et représente un mélange de certains des taxons plus eurytopiques de biotes plus anciennes et de certains genres dont l'origine provinciale n'est pas évidente. La faune à Hirnantia disparaît avant la fin de l'Hirnantien. Les trilobites connaissent une période d'anéantissement majeure à la limite du Rawtheyen et de l'Hirnantien, et très peu d'assemblages hirnantiens comptent plus de 10 genres (la plupart en contiennent beaucoup moins). La plupart des genres ne sont connus qu'à quelques sites; ce sont tout simplement des vestiges de faunes plus anciennes dans la région. La seule association que l'on rencontre fréquemment est celle d'un dalmanitidé (Dalmanitina/Mucronaspis) et d'un homalonotidé (Brongniartella/Platycoryphe), dans certains cas avec un odontopleuridé (Leonaspis). Cette association a une plus grande répartition en profondeur et en latitude que la Faune à Hirnantia typique, et elle persiste jusqu'à la fin de l'Hirnantien. L'étude des genres de trilobites du Rawtheyen sur un vaste spectre écologique révèle que les extinctions sont concentrées de façon disproportionnée dans les eaux plus profondes. Le même résultat pourrait s'appliquer aux brachiopodes, car la plupart des éléments de la faune benthique à Foliomena avaient disparu à la fin du Rawtheyen.

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INTRODUCTION

Throughout the world, the uppermost Ordovician trilobite and brachiopod faunas are distinctive and differ markedly from those of the underlying middle Ashgill in composition and (commonly) diversity. The latest Ordovician stage, the Hirnantian, is typified by these faunas but the base of the stage has yet to be formally defined. Its correlation with the graptolite zonal scheme is not firmly established and the marked provincialism of Upper Ordovician conodonts still presents major problems for international correlation (e.g., Barnes and Bergström, 1988). Both trilobites and brachiopods have been the focus of considerable attention in the context of the Late Ordovician extinction events (e.g., Sheehan, 1982; Brenchley and Newall, 1984; Briggs et al., 1988) and the brachiopods in particular have been influential in the positioning of the Ordovician-Silurian boundary in shelf facies (Cocks and Rickards, 1988). The timing and extent of the Upper Ordovician faunal changes differed from phylum to phylum and even province to province (cf. Eckert, 1988 and Donovan, 1988 on crinoids). In general though, the major extinctions occurred at the end of the Rawtheyan and during the early to mid-Hirnantian. By the end of the Hirnantian, many groups were beginning to diversify again (e.g., Barnes, 1986).

Global reviews of the detailed composition of Hirnantian brachiopod and trilobite faunas have recently been produced by Rong and Harper (1988) and Owen (1986) respectively. They need not be repeated here, but for the present paper form a substantial database, from which we can assess the diversity, spatial and temporal distribution, and the origins and fates of these faunas. Some marked contrasts between the two groups are highlighted. In particular, it is clear that in contrast to the trilobites, there is a more substantial group of brachiopods that had a wide distribution, and new brachiopod taxa, community structures, and provinces developed during the Hirnantian.

SPATIAL DISTRIBUTION

Diversity and geographic range

Figure 1 summarizes the available data on Hirnantian brachiopod and trilobite faunas at the generic level. In general, both groups show a marked decline in diversity and geographic distribution when compared to the underlying Rawtheyan Stage but there are some exceptions. This decline is most strongly marked in the trilobites.

Brachiopods

The majority of Hirnantian brachiopod faunas are of low diversity, commonly containing less than ten genera (Fig. 1A). Most genera are restricted to six or fewer sites but some are very widespread (Fig. 1B) and include the genera that Rong and Harper (1988) have shown to be the core of the *Hirnantia* Fauna. This fauna, formally named by Temple (1965), is critical for correlation of the Hirnantian Stage (Ingham and Wright, 1970) but problems exist in defining its precise temporal range. Moreover, although widespread, it is not cosmopolitan, and Elles, who first recognized the fauna (1922), emphasized the difficulties in using what may be a "facies fauna" for chronostratigraphic correlation (Elles, 1923).

The core of the *Hirnantia* Fauna comprises *Hirnantia* itself (known from 28 sites), *Dalmanella, Eostropheodonta* (and the closely related *Aphanomena*), *Hindella, Cliftonia, Kinnella, Plectothyrella*, and *Paromalomena*. The last three of these are restricted to the Hirnantian. In contrast to the very widespread elements of the core, the majority of genera known from the *Hirnantia* Fauna as a whole have a much narrower spatial range; most occur at fewer than five sites with many restricted to a single site. Virtually all of these elements are leftovers from the diverse middle Ashgill faunas.

Of the diverse Hirnantian brachiopod faunas (Fig. 1A), those of southwest Sichuan, west Hubei, and Bohemia are typical *Hirnantia* faunas. That at Garth, mid-Wales, contains a number of typical *Hirnantia* Fauna genera but these are numerically dominated by a large number of apparent relicts: genera unknown from other Hirnantian successions. The American Mid-continent Edgewood Fauna is very different from the *Hirnantia* Fauna (see below), and similar assemblages from the Oslo Region are also fairly diverse.

Trilobites

Almost all Hirnantian trilobite faunas are of very low diversity (Fig. 1A). Even in the two areas where the greatest number of taxa are found, the Oslo Region (see Owen, 1981, Table 1, 1986; Brenchley and Cocks, 1982) and Anticosti Island (see Owen, 1986; Lespérance, 1988b; R. Ludvigsen, pers. comm., 1988), different trilobites occur in low diversity assemblages at different levels. The detailed distribution of trilobites at the other sites of relatively high diversity is not known.

Most trilobite genera known from the Hirnantian are known from very few sites and these highly endemic taxa are commonly relicts of earlier faunas in the area (Owen, 1986). The notable exceptions are the odontopleurid, *Leonaspis*, the homalonotids *Brongniartella* and *Platycoryphe* and, especially, the dalmanitids variously ascribed to *Dalmanitina* and *Mucronaspis* (or *Songxites* now widely considered a subgenus of *Dalmanitina*). The diverse taxonomy affecting the dalmanitids was discussed by Lespérance (1988b, p. 363-365), whose synonymy of the Mucronaspidinae Holloway, 1981 in the Dalmanitininae Destombes, 1972 has provided some simplification. Lespérance also discussed the distinction between *Dalmanitina* and *Mucronaspis* but noted that many of the records of these taxa from the Hirnantian are based on insufficiently complete material or have not been adequately illustrated for generic differentiation. Consequently, they are treated together here pending further detailed taxonomic work. The same applies to the Hirnantian homalonotids. No trilobite genus is known to have appeared in the Hirnantian.

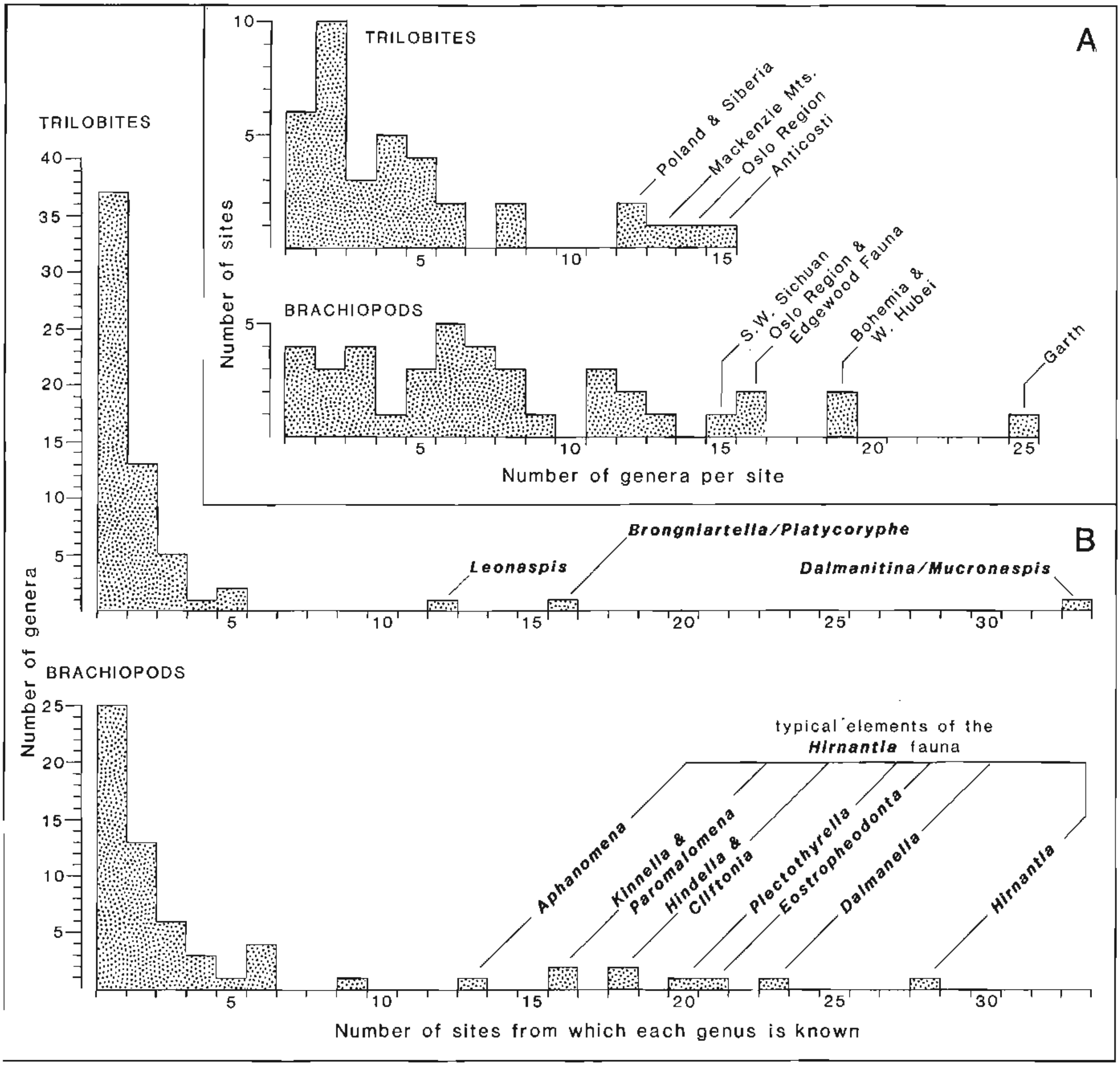

Figure 1. Summary histogram showing the diversity of Hirnantian brachiopod and trilobite faunas (A) and the degree of endemicity of each genus known from the Hirnantian (B). Brachiopod data from Rong and Harper (1988); trilobites based on Owen (1986, Table 2) but with some updating. Most notably: 1) the fauna from China is now considered under six sites and the entries for *Barrandia, Ptilaenus, Encrinuroides, Niuchangella,* and *Eoleonaspis* are removed; 2) *Tsunylichas* is a junior synonym of *Dicranopeltis* (see Thomas and Holloway, 1988, p. 193); 3) the entries for Anticosti and Argentina are amended in the light of Lespérance (1988a) and Benedetto (1986) and the fauna from Tasmania (Banks, 1988) is included; 4) *Panarchaeogonus* is present at Keisley, Bornholm, and Poland and indeterminate proetids at these three sites belong in *Decoroproetus.*

Biogeography

Brachiopods

In their global analysis of Hirnantian brachiopods, Rong and Harper (1988) demonstrated a marked latitudinal provincialism, which is summarized in Figure 2. This latitudinal differentiation reflected a temperature gradient that was probably steeper than earlier in the Ashgill. There is no evidence for significant cooling at equatorial latitudes (see Eckert, 1988, p. 159, 160). The typical *Hirnantia* Fauna (see above) colonized a variety of substrates in subtropical and temperate latitudes and characterizes the Kosov Province. Atypical *Hirnantia* faunas, commonly of low diversity, occurred at high latitudes marginal to the Gondwanan ice sheet and constitute the Bani Province. Although containing some of the genera of the typical *Hirnantia* Fauna — notably *Hirnantia, Eostropheodonta*, and *Plectothyrella* — their species are different from those of the typical faunas. The province also includes endemics such as *Arenorthis, Destombesium*, and *Undithyrella*. The North American Mid-continent fauna (see Amsden, 1974; Amsden and Barrick, 1986), the Edgewood province, known also from Kolyma in the U.S.S.R., was developed in equatorial latitudes. The province contains a substantial number of genera not known from the *Hirnantia* Fauna, including *Brevilamnulella, Thebesia*, and *Leptoskelidion*.

In addition to areas where the provincial attribution of the fauna is clear, there are areas at the interface of the Edgewood and Kosov provinces where the picture is more complex. Gorny Altai in northern Asia has a mixture of Edgewood and Kosov province elements whereas in Oslo, an Edgewood Fauna postdates a *Hirnantia* Fauna (see also Brenchley and Cocks, 1982). In eastern Canada, a typical Kosov Province fauna is present at Percé (Lespérance, 1988a) but the fauna on Anticosti may best be included in the Edgewood Province (Rong and Harper, 1988, p. 396). At Girvan, Scotland, the Hirnantian fauna is a mixture of relict (largely North American) taxa with *Hirnantia* Fauna immigrants.

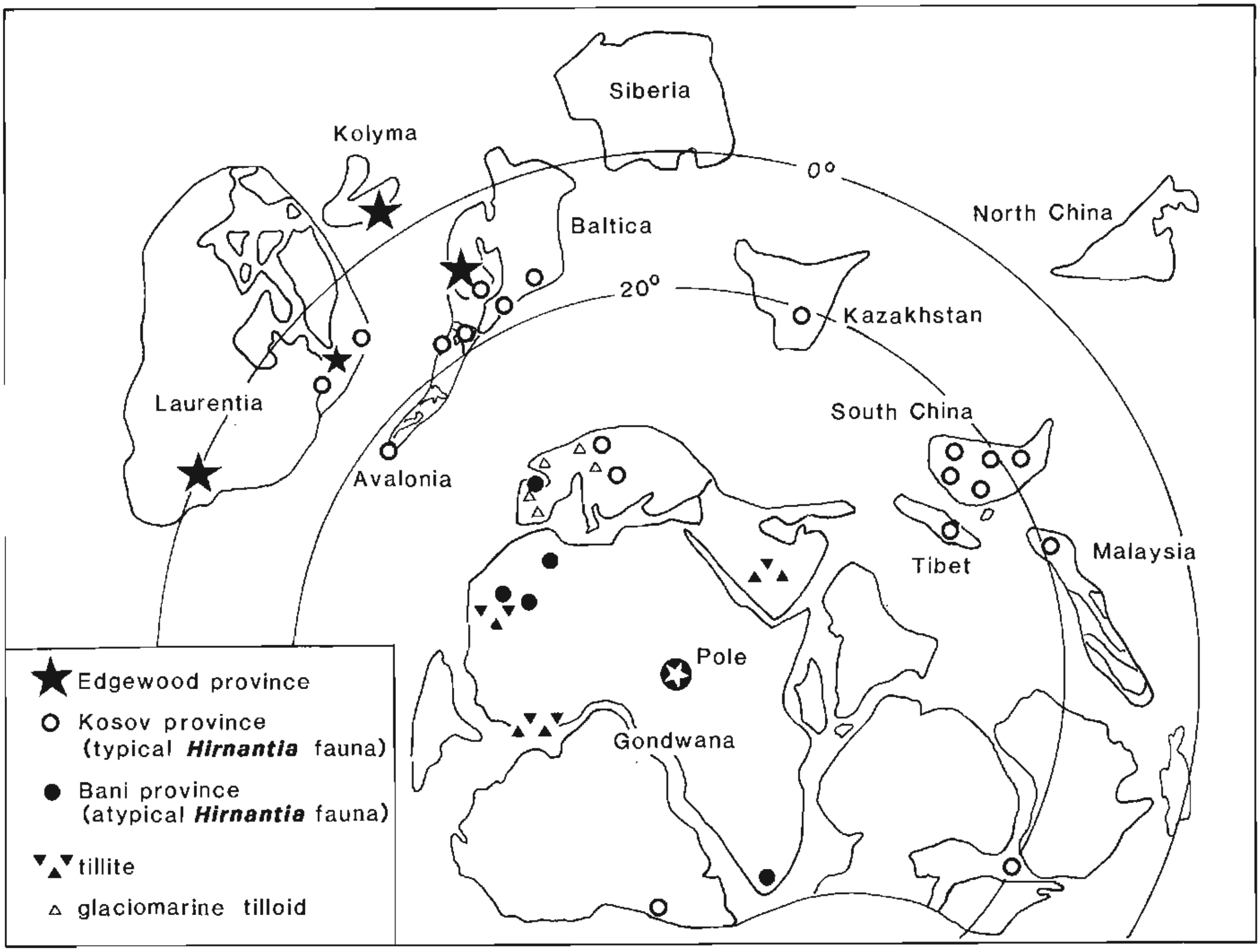

Figure 2. Hirnantian reconstruction showing the distribution of brachiopod faunas and provinces (from Rong and Harper, 1988, Fig. 1 and references cited therein; glaciomarine tilloid data from Brenchley and Newall, 1984; Brenchley, 1988; Robardet and Doré, 1988). Note that the most widespread trilobites — *Dalmanitina/Mucronaspis* and *Homalonotus/Brongniartella* — are known from sites from Laurentia to the margins of Gondwana.

Trilobites

Most Hirnantian trilobite genera were highly endemic local remnants of earlier faunas. The few widely distributed genera — *Dalmanitina/Mucronaspis, Brongniartella/Homalonotus* and, to a lesser extent, *Leonaspis* — were highly eurytopic and not latitudinally constrained. They are known from the high latitude Bani Province (e.g., Morocco), the midlatitude Kosov Province (e.g., Bohemia, Poland, China) and *Mucronaspis* is known from the Edgewood Province (see Lespérance, 1988b, p. 366-369). As with the brachiopods, the Oslo Region provides an interesting mixture of faunas at the interface between the Kosov and Edgewood provinces, which is also seen in the contrast between the Anticosti and Percé faunas in eastern Canada (see Lespérance, 1988b, p. 361,362). The Girvan Fauna is restricted to North American relicts (Owen, 1986).

Bathymetry

Brachiopods

A number of authors have recognized associations or communities within the *Hirnantia* Fauna (e.g., Rong, 1979; Harper, 1981; Brenchley and Cocks, 1982; Brenchley and Cullen, 1984). Rong and Harper (1988) have shown that the fauna is best developed in Benthic Assemblage Zone 3 and especially in the deeper part of the zone. In deeper water facies (BA4-5), lower diversity faunas contain *Dalmanella, Kinnella, Paromalomena, Fardenia/Coolinia*, and *Hindella* but are commonly dominated by *Skenidioides* and *Philhedra/Philhedrella* — forms otherwise rare in the typical *Hirnantia* faunas. Shallow water variants (BA2) are commonly of low diversity and are dominated by rhynchonellids and spire bearers (see Rong and Harper, 1988, Figs. 4, 5). The *Hirnantia* Fauna colonized a variety of substrates and is found in mudstone (e.g., Temple, 1965; Wright, 1968), siltstone (Temple, 1965; Bergström, 1968), sandstone (Harper, 1981), marl (Schönlaub, 1971), and limestone (Rong, 1979, 1984). The type Edgewood Province faunas thrived in a carbonate-forming environment in shallow, warm water with little or no siliciclastic input (Amsden and Barrick, 1986).

Trilobites

Dalmanitina/Mucronaspis with a homalonotid and, in some instances, *Leonaspis* have been described from deeper water settings than coeval *Hirnantia* faunas in North Wales (Brenchley and Cullen, 1984) and at Percé, Canada (Lespérance and Sheehan, 1976, p. 720). In the latter area, the so called *Mucronaspis* Community has been interpreted as occupying a BA6 position (Lespérance et al., 1987, p. 123). In Bohemia, Štorch (1986) recorded a low diversity "*Mucronaspis* Community" of trilobites and brachiopods from the base of the Kosov Formation (interpreted as the base of the Hirnantian) and a rich *Hirnantia* Fauna from near the top of the formation (*G. bohemicus* Zone). However, *Mucronaspis* and *Brongniartella* are also common in the *Hirnantia* Fauna (Štorch, 1984, p. 84). This co-occurrence is also seen elsewhere, most notably in parts of China where, in addition, Rong and Harper (1988) have also recorded *Dalmanitina* in shallow water sediments of the Nancheng Shale of southern Shaanxi. Similarly, Brenchley and Cullen (1984, Table 3) also recorded *Mucronaspis* in what have been interpreted as shallow water assemblages in the Hirnantian of the Oslo Region. Thus, as with the biogeographic distribution, the limited number of widespread Hirnantian trilobite genera were also bathymetrically more eurytopic than the typical *Hirnantia* faunas.

TEMPORAL DISTRIBUTION

Brachiopods

There are few sections available where the development of the *Hirnantia* Fauna may be monitored relative to graptolite biozones. Sections on the South China Plate display the relationships of the developing *Hirnantia* Fauna to the graptolite zonal scheme established within the Wufeng Shale (see Rong and Harper, 1988, for details). The first appearance of some elements of the fauna correspond to the *D. mirus* (W4) and *P. uniformis* (W5) zones. *Dalmanella, Paromalomena, Leptaena, Aphanomena, Fardenia,* and *Hindella* are absent; the fauna, which characterizes deeper water settings appears transitional. In Tongzi County, the typical *Hirnantia* Fauna first appears within the *P. uniformis* Zone. The fauna reaches its acme in the *D. bohemicus* Zone (W6) (= *C. extraordinarius* Zone). The *Hirnantia* Fauna disappeared suddenly, prior to the *G. persculptus* Zone of the Chinese zonal scheme, probably as a result of a severe reduction of the ventilation of the seabed. Excluding the Chinese sections, the *Hirnantia* Fauna of Kazakhstan and a variation on the *Hirnantia* Fauna documented from the English Lake District (Cocks, 1988) are recorded with *persculptus* Zone graptolites in terms of the British scheme. Both faunas thus represent occurrences at the younger end of the age range of the fauna. Significantly, both are from deeper water settings and the Kazakhstan fauna contains a number of mid-Ashgill relicts together with genera, but few species, of the typical *Hirnantia* Fauna (Koren et al., 1988). The correlation of horizons containing the Edgewood Province fauna with the graptolite zones is not known (see Amsden and Barrick, 1986, p. 47).

Trilobites

As noted by Owen (1986, p. 232, 233) *Mucronaspis* makes its first appearance in strata interpreted as uppermost Rawtheyan in Norway, Sweden, England, and Wales, although it had a longer previous history in peri-Gondwanan faunas. The low diversity Hirnantian trilobite associations centred on *Mucronaspis/Dalmanitina* are known from strata equated with the *D. mirus* to the *D. bohemicus* graptolite zones in China (Rong, 1984; Rong and Harper, 1988) and from the *G. persculptus* Zone in Kazakhstan (Apollonov et al., 1980) and Siberia (Chugaeva, 1983). An as yet unnamed dalmanitine presently ascribed to *Dalmanitina* (*Songxites*) (see Lespérance, 1988b, p. 365) is known from just below the *C. extraordinarius* Band at Dob's Linn. A broadly similar specimen originally ascribed to *Mucronaspis* (s.l.) by Siveter and

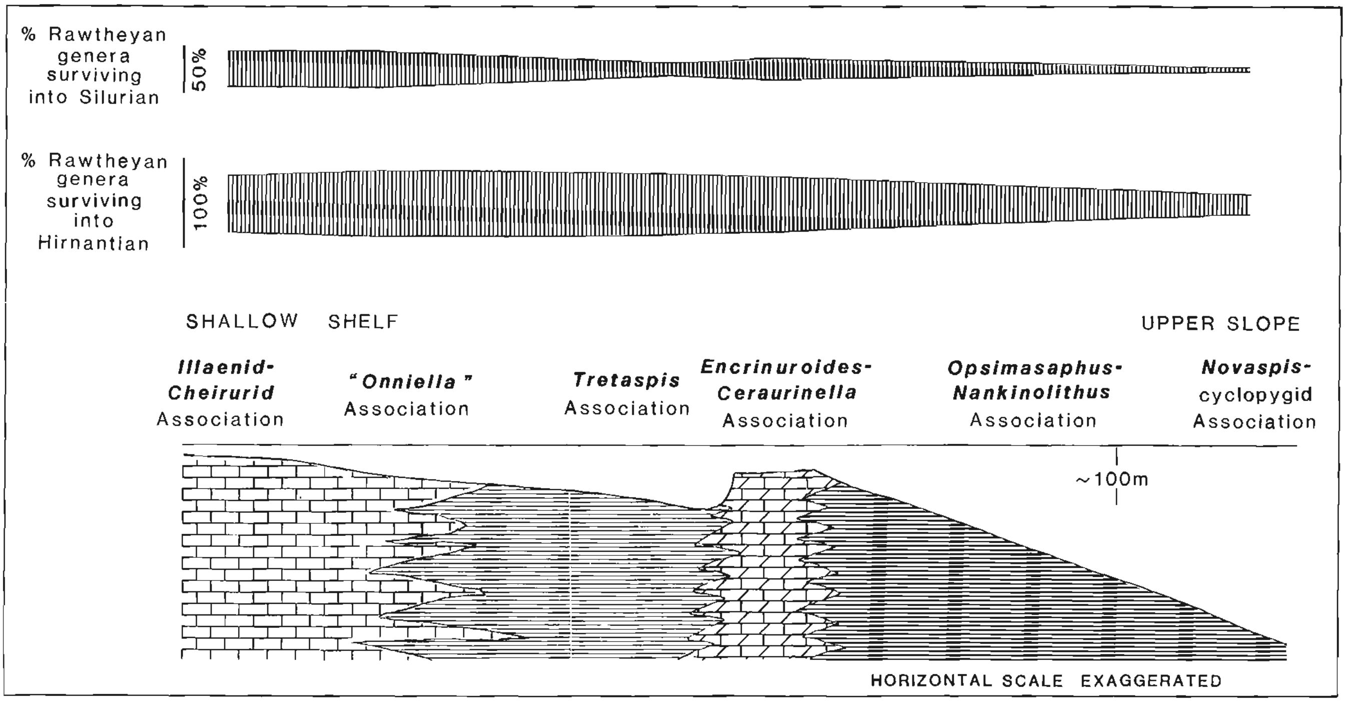

Figure 3. The survival of trilobite genera from a spectrum of Rawtheyan environments. The *Foliomena* brachiopod fauna is at the "deep end" of this spectrum and shows low survival rates into the Hirnantian and Silurian.

Ingham (*in* Siveter et al., 1980; see •najdr, 1987, p. 274; also Lespérance, 1988b, p. 365) is known from a level between the *D. anceps* and *G. persculptus* zones in Ireland. Interestingly, despite their very widespread occurrence throughout the Hirnantian, *Dalmanitina* and *Mucronaspis* did not survive into the Silurian (see Lespérance, 1988b, for discussion of putative Silurian species).

ORIGINS AND FATE OF THE HIRNANTIAN FAUNAS

The latest Ordovician extinction events shaped the composition and sealed the fate of the Hirnantian trilobite and brachiopod faunas. The extinctions were not concentrated in a single mass event at the end of the Ordovician Period (Brenchley and Newall, 1984) and their causes are still unclear, although the Late Ordovician glaciation and its several consequences are likely candidates (Brenchley, 1984, 1988; Brenchley and Newall, 1984; Barnes, 1986; Sheehan, 1988).

As a result of modern taxonomic studies, the composition of many Ashgill trilobite faunas, especially in Europe and China, is well known. It is possible therefore to assess the fate of Rawtheyan assemblages in a semiquantitative way and to do so in relation to a broadly bathymetric scheme. In contrast, unlike some of the earlier parts of the Ordovician, there are relatively few detailed monographic, or paleoecological works on Ashgill brachiopod faunas, and several important faunas remain to be documented, at least in modern terms. However, a few general points emerge even for the brachiopods.

Brachiopods

The end of the Rawtheyan marked the demise of the deep water *Foliomena* Fauna, which had flourished since the late Caradoc (see Cocks and Rong, 1988 for review) and the *Proboscisambon* "Community" of Havlíek (1982; see also Cocks and Rong, 1988), which also inhabited deep waters during the Rawtheyan. Only about 35 per cent of the genera known from the *Foliomena* Fauna survived into the Hirnantian and at least half of these also occurred in much shallower environments. The *Hirnantia* faunas comprised eurytopic taxa derived from several earlier provinces, together with newly evolved taxa and local relicts of earlier faunas. Further extinction toward the end of the Hirnantian marked the end of many of the widespread genera of the *Hirnantia* Fauna. As a result, the lowest Llandovery faunas are dominated by older relicts (including "Lazarus" taxa not known from the Hirnantian) together with some taxa from the Edgewood Province. Only later did the typical Silurian fauna become established (see Baarli and Harper, 1986; Cocks, 1988; Harper, 1988).

A relatively high proportion of the typical genera of the Edgewood Fauna, for example *Leptoskelidion, Biparetis, Thebesia*, and *Brevilamnulella*, appear initially at the base of the Hirnantian and were probably derived from ancestors within the North American province. Nevertheless, a number of genera arrived in the Edgewood Province at the beginning of the Hirnantian from pre-Hirnantian sites on the margins of the North American province — *Dolerorthis, Mendacella*?, *Dalmanella, Dicoelosia*, and *Leptaena*. All these genera are represented in the diverse faunas of the Drummuck Group of the Girvan district, southwest Scotland (Harper, 1982).

Table 1. The fate of the trilobite genera from a spectrum of Rawtheyan[1] environments

Locality	Unit	Faunal Type	Number of Rawtheyan Genera	Rawtheyan genera surviving into Hirnantian		Rawtheyan genera surviving into Silurian	
				No	%	No	%
Kildare, Ireland	Chair of Kildare Limestone	Illaenid-cheirurid Association	57	36	63	22	39
Keisley, Ireland	Keisley Limestone		29	19	66	12	41
Asker, Norway	Langåra Fm.	"*Onniella*" Association	31	24	77	13	42
Ringerike, Norway	Bønsnes Fm.	Intermediate Mid-shelf faunas	24	16	67	6	25
Hadeland, Norway	Kjørrven Fm.		18	13	72	5	28
Oslo, Norway	Husbergøya Fm.	*Tretaspis* Association	16	11	69	3	19
Girvan, Scotland	Upper Drummuck Group	Mixed shelf faunas	36	25	69	12	33
Bala, Wales	Rhiwlas Limestone	*Encrinuroides-Ceraurinella* Association	29	17	59	7	24
Västergötland, Sweden	Ulunda Mudstone	Intermediate deep shelf faunas	20	12	60	6	30
Holy Cross Mts., Poland	*Staurocephalus clavifrons* Beds		37	14	38	6	16
Bwlch Siglen, Wales	Abercwmeiddaw Group	*Opsimasaphus-Nankinolithus* Association	25	10	40	3	12
Abercwmeiddaw, Wales		*Novaspis*-cyclopygid Association	14	3	21	1	7
All genera in these 12 faunas			109	60	55	33	30

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[1] Sources of Rawtheyan data are as follows:

Kildare — Dean, 1978, p. 111, 112 (note: Otarion sp., Platylichas glenos, Trinodus sp. cf. T. tardus and proetid gen. et sp. inlet. of Dean reassigned to Panarchaeogonus, Arthrorhachis, and Cyphaspides, respectively).

Keisley — Thomas et al., 1984, Fig. 20 (see text).

Oslo and Asker units — Owen, 1981, Table 1; Brenchley and Cocks, 1982.

Ringerike and Hadeland units — Owen, 1981, Table 1 (note: species of *Platylichas* and *Paraharpes*? *wegelini* reassigned to *Autoloxolichas* and *Wegelinia*).

Bala — Whittington, 1968, Table 4 (note: *Astroproetus*, *Platylichas*, and *Trinodus* species reassigned to *Decoroproetus*, *Autoloxolichas*, and *Arthrorhachis* respectively; *Nankinolithus* present amongst the *Tretaspis* material).

Skultorp — Bergström, 1973, Table 1 (note: species of *Trinodus*, *Platylichas*, *Illaenus*, and *Zeliskella* reassigned to *Arthrorhachis*, *Autoloxolichas*, *Ectillaenus*? and *Calyptaulax*, respectively).

Poland — Kielan, 1960 (note: species of *Trinodus*, *Ogmocnemis*, and *Stubblefieldia* reassigned to *Arthrorhachis*, *Decoroproetus*, and *Pseudo-spaerexochus*, respectively).

Bwlch Siglen and Abercwmeiddaw — Price and Magor, 1984, Fig. 2 b, e. (note that *Opsimasaphus* may be a junior synonym of *Birmanites* — see Tripp et al., 1989, p. 36 and references therein).

Girvan — Thomas et al., 1984, Fig. 16 [note: this list only contains named species, the fauna is even more diverse but in need of revision (Owen, in prep.). *Nankinolithus* is included in the present analysis and the species of *Platylichas* reassigned to *Autoloxolichas*].

Note that Lazarus taxa have been accounted for in calculating the survival of Rawtheyan genera into the Hirnantian.

Trilobites

Brenchley and Newall (1984) estimated that only about 15 per cent of Rawtheyan trilobite genera survived into the Hirnantian. More recent reviews of Hirnantian faunas (Owen, 1986, and herein; Lespérance, 1988b) including the recognition of "Lazarus" taxa show that the figure is of the order of 50 to 60 per cent with only about 30 per cent of Rawtheyan genera surviving into the Silurian (Fig. 3). Most Rawtheyan families survived into the Hirnantian (Briggs et al., 1988, Fig. 9.5) but with very reduced generic diversity. By the end of the Hirnantian, half of the 29 Rawtheyan families were extinct. Thus, as Brenchley and Newall (1984) recognized, the Late Ordovician extinction of trilobites had two phases. Although the first, at the end of the Rawtheyan, was demonstrably rapid, the resulting low diversity and high endemicity (Fig. 1) make the speed of the latter phase of extinction impossible to assess accurately.

The low diversity of most Hirnantian trilobite faunas testifies to the marked disintegration of the Rawtheyan faunas, which represented well established biofacies. As is shown below, the extinction of taxa at the end of the Rawtheyan was not uniformly spread across the Rawtheyan ecological spectrum. Table 1 and Figure 3 show the fate of taxa from well documented faunas across this spectrum, from pure carbonate environments, through generalized shelf environments with a variety of lithofacies, to an integradational shelf to upper slope transect. Some comments on the faunas involved in this new synthesis are necessary before the synthesis results are analyzed.

Pure carbonate environments

Pure limestones of Ordovician to Devonian age commonly have a trilobite fauna dominated by illaenids, cheirurids, scutelluids, harpids, and lichids (Lane, 1972; Mikulic, 1981). Fortey (1975) named this association in the Llanvirn of Spitsbergen, the "illaenid-cheirurid community." All three authors cited above included the Ashgill Chair of Kildare Limestone in Ireland and the Keisley Limestone of Northern England in their analyses. The precise age of the former is not certain but a Rawtheyan age for at least part of it is likely (Dean, 1978, p. 110). If it is older, then the high generic survival rate is even more remarkable. The stratigraphy of the broadly coeval Keisley Limestone was summarized by Wright (1985), who demonstrated that the bedded limestone above the typical unbedded facies is Hirnantian in age (cf. Temple, 1968, 1969). The fauna of the higher strata, which includes *Mucronaspis* is not included in Table 1. Other Ashgill pure limestone faunas in need of modern documentation and thus not included here include those of the Pusgillian to early Hirnantian Boda Limestone of Sweden (Jaanusson, 1982a, Fig. 4) and a carbonate mound in Kazakhstan noted by Apollonov (1975, p. 379).

General shelf environments

The distribution of Caradoc to middle Ashgill shelly faunas was strongly substrate controlled (e.g., Lockley, 1983; McNamara and Fordham, 1981). Apart from the outer shelf-upper slope transect discussed below, the only detailed paleoecological study involving Rawtheyan trilobites is that of Brenchley and Cocks (1982) on the uppermost part of the stage in the Oslo-Asker district of the Oslo Region, Norway. They recognized two faunal associations, the *Tretaspis* Association in the mudstone of the east (Oslo) and the "*Onniella*" Association in nodular limestone and mudstone in the west (Asker). The naming of the latter is unfortunate, as the eponymous brachiopod is probably *Cryptothyris* not *Onniella*. The association is not the same as the offshore associations dominated by *Onniella* and small plectambonitaceans *Sericoidea* and/or *Chonetoidea* in the Middle and Upper Ordovician elsewhere (e.g., see Harper and Owen, 1986, p. 274,275). The trilobite fauna of the "*Onniella*" Association of Asker is dominated by proetids and the encrinurine *Erratencrinurus* (*Celtencrinurus*); the *Tretaspis* Association, by the eponymous trinucleid and, to a lesser extent, *Toxochasmops*, and *Lonchodomas*. Brenchley and Cocks argued that the *Tretaspis* Association inhabited the deeper water, which was still probably less than 100 m deep.

Two other faunas from the Oslo Region are included in this analysis, those of the limestone and shale of the Bønsnes Formation in Ringerike and the limestone, siltstone, and shale of the Kjørrven Formation in Hadeland. The Bønsnes Formation probably represents a shallow water, low energy environment (Hanken and Harper, 1985, p. 252, 253); the trilobite fauna is dominated by the scutelluid *Eobronteus*, illaenids, and *Remopleurides*. The Kjørrven Formation contains abundant *Tretaspis*, *Erratencrinurus*, *Toxochasmops*, and *Prionocheilus* in a fauna whose overall generic composition is close to that of the Cautleyan Stage in northern England (Owen, 1978, 1981, p. 8, 9). All but three of the genera in the formation are known from one or other of the three trilobite associations defined by McNamara and Fordham (1981) in the mid-Cautleyan of the English Lake District. The rarity of proetids and absence of *Gravicalymene*, important components of the English proetid and calymenid associations, indicates that the faunal list of the Kjørrven Formation does not simply represent a homogenization of the English associations.

Outer shelf-upper slope transect

Price (1980, 1981) and Price and Magor (1984) recognized a series of intergradational trilobite associations in the Ashgill of Wales and related them to a depth gradient. The shallowest of those, the *Encrinuroides-Ceraurinella* Association (Price and Magor, 1984, p. 195 = "intermediate faunas" of Price 1980, 1981) was considered to have occupied water depths greater than those of the "illaenid-cheirurid-lichid association" and the "shelf faunas" such as those of northern England (see above). The *Encrinuroides-Ceraurinella* Association occurs in the Rawtheyan Rhiwlas Limestone, interpreted as a small elevated site in the northern Welsh Basin. This limestone also includes the ten characteristic elements of the "*Phillipsinella parabola-Staurocephalus clavifrons* fauna" (Whittington, 1968, p. 116, 117). Price (1973) showed that this is an ecologically controlled assemblage, which appears at various levels in the Ashgill of Britain, Sweden, and Poland.

Price and Magor (1984) demonstrated a change downslope in the Rawtheyan of North Wales from the *Encrinuroides-Ceraurinella* Association through the *Opsimasaphus-Nankinolithus* Association to what they termed the "*Novaspis*-cyclopygid Association", which is an example of the cyclopygid biofacies of Fortey (1985, p. 222, 223) and Fortey and Owens (1987, p. 105-108). This biofacies existed during the Arenig and is numerically dominated by cyclopygids, which were probably mesopelagic, living between 200 and 700 m in the water column above the upper parts of the continental slope or deep marginal basins. The deep water benthos included trilobites with reduced, or no, eyes — *Novaspis, Nankinolithus, Dionide, Ectillaenus*, and raphiophorids in the faunas discussed by Price and Magor. In addition, some upper Caradoc and Ashgill cyclopygid biofacies occur with the deep water *Foliomena* brachiopod fauna (reviewed by Cocks and Rong, 1988). As with the trilobites, this fauna is integradational with shallower water assemblages and elements of the *Foliomena* fauna occur with the *Nankinolithus-Opsimasaphus* Association in North Wales (Price, 1981; Campbell, 1984). The data from North Wales analyzed here (Table 1) are based on the ends of the spectrum given by Price and Magor (1984) together with an example of the *Opsimasaphus-Nankinolithus* Association from the middle of this range.

Such a spectrum cannot be demonstrated in a single area elsewhere but other examples of parts of it are included in the present analysis (Table 1). The *Phillipsinella parabola-Staurocephalus clavifrons* fauna from Poland (Kielan, 1960) has many similarities to that of the Rhiwlas Limestone, as does the fauna of the Ulunda Mudstone in Västergötland, Sweden, described by Bergström (1973). The Swedish unit has a fauna similar to that of the upper Johnstorp Formation elsewhere in Sweden, which in turn has been allied to the *P. parabola-S. clavifrons* Fauna of Poland (Kielan, 1960; Jaanusson, 1982b, p. 173, 174). Both the Ulunda Mudstone and the Polish deposits have also yielded *Foliomena* faunas (Jaanusson, 1982b; Cocks and Rong, 1988), suggesting that they are from slightly deeper water than that of the Rhiwlas Limestone. Elements of the *P. parabola-S. clavifrons* fauna have recently been recorded as part of the diverse trilobite fauna of the lower Ashgill Tangtou Formation in South Jiangsu, Southern China (Tripp et al., 1989). This unit also contains a *Foliomena* Fauna (Cocks and Rong, 1988, p. 54, 55), thus demonstrating the longevity and wide distribution of this association.

A mixture of shelf elements

The diverse trilobite fauna of the Upper Drummuck Group at Girvan, Scotland, is included in Table 1, in addition to the well constrained faunas discussed above. Harper (1979, p. 442; 1982, p. 270) has shown that the brachiopod faunas of the group represent a mixture of assemblages accumulated during downslope transport. Although *Nankinolithus* and some elements of the *P. parabola-S. clavifrons* fauna are present in the Upper Drummuck Group in addition to elements of shallower assemblages, it lacks the diverse cyclopygid faunas of the deeper water assemblages. This provides a useful constraint on the depth of deposition of the unit. The subsequent survival rates of genera in the Upper Drummuck Group (Table 1) confirm the position on the palaeoslope suggested here.

The fate of the Rawtheyan trilobite assemblages

In Table 1 and Figure 3, the survival rates of genera from the various Rawtheyan faunas into the Hirnantian and the Silurian are shown. Those in the shelf assemblages fared much better than those in the deeper water regimes. About 60 per cent or more of the Rawtheyan genera survived into the Hirnantian from the *Encrinuroides-Ceraurinella* and various shelf associations. This figure decreased to only 20 per cent in the cyclopygid biofacies. The only record of a cyclopygid in the Hirnantian is from one locality in Pomeroy, Ireland (Owen, 1986, Table 2), whereas several genera were present in the Rawtheyan. Similarly, survival rates of the Rawtheyan shelf taxa into the Silurian were generally considerably higher and the only survivor from the cyclopygid biofacies was the raphiophorid *Raphiophorus*. Even within the shelf faunas, elements of the deeper water assemblages had a lower survival rate than those from shallower faunas. Of the faunas described by Brenchley and Cocks (1982) from Norway, only 19 per cent (3 genera) of the *Tretaspis* Association genera are known from the Silurian compared with 49 per cent (13 genera) from the shallower "*Onniella*" Association.

CONCLUSIONS

Evidence presented here indicates that deeper water trilobite assemblages suffered disproportionately greater levels of extinction than those from shallower environments. The extinction of most of the *Foliomena* Fauna at the end of the Rawtheyan suggests a similar pattern among the brachiopods. Moreover, although many genera did survive into the Hirnantian, the diverse recurring generic association of the earlier Ashgill cannot be recognized even among the shelf faunas, which at best contained very localized relicts of the earlier assemblages.

Faunas in all environments from the shallow shelf to the upper continental slope were severely disrupted at the end of the Rawtheyan. However, the concentration of generic extinctions in deeper water facies is counterintuitive to simple models of the spread of cold water onto low latitude shelves or simple species-area effects consequent on glacio-eustatic regression. Brenchley (1984) suggested that the extinction of deep shelf elements of the shelly fauna at the Rawtheyan-Hirnantian boundary may have been the result of the initial phase of sea level drop and consequent displacement of the outermost association "over the shelf edge." However, Owen (1986, p. 223) has noted the initiation of widespread faunal changes prior to the main regression. Moreover, although undoubtedly important at least locally on the shelf, it is difficult to see how not only benthonic faunas of the upper slope but also the mesopelagic cyclopygid trilobites could be directly affected by falling sea level. Similarly, these faunas are less likely to have been affected by the spread of cold water (either bottom waters or surface waters).

Oceanic overturn with consequent upwelling of anoxic and/or biologically unconditional (therefore toxic) waters at the slope-shelf boundary seem a more likely cause of the profound extinctions here. The factors leading to this at times of climatic change were discussed by Wilde and Berry (1984) who noted that overturn would occur both at the onset and at the end of a glacial episode. This could explain the extinctions both at the end of the Rawtheyan and the mid-Hirnantian. Fortey and Chatterton (1988, p. 216-218) noted that the groups of asaphine trilobites that survived to the Upper Ordovician but then became extinct probably had planktonic habits, and thus may have been particularly susceptible to major changes in oceanic circulation. Chatterton and Speyer (1989) have recently shown this same pattern of larval type and susceptibility to the end-Ordovician events in other trilobite stocks.

Reasons for the development of the *Hirnantia* faunas in high and mid-latitudes and the spread of *Dalmanitina*/*Mucronaspis* from Gondwana to an almost cosmopolitan distribution are not yet clear. Their eurytopy was considerable and thus their extinction before the Silurian even more remarkable.

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Lower to Middle Ordovician trace fossils from the Central Iberian Zone of Portugal and Spain

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Abstract

The Lower to Middle Ordovician strata of the Central Iberian Zone (C.I.Z.) in Portugal and Spain accumulated in a cold water, shallow shelf sea on the margin of Gondwanaland. The sequences are almost entirely siliciclastic and are dominated by thick (up to 1000 m) quartzose sandstone and mudrock (up to 350 m thick), the latter punctuated by widespread sheet sandstones of storm origin. Subordinate conglomerate and rare oolitic ironstone also occur.

Thirty-one ichnogenera and over 45 ichnospecies are now known from the Arenig to lower Caradoc of the C.I.Z. Range charts and diversity graphs illustrate the presence of trace fossils throughout the Lower to Middle Ordovician and their preponderance in the more heterolithic facies.

Five, possibly six, ichnofacies associations (A, B, C, C1, D, E) that are generally restricted to discrete sedimentological facies have been recognized. Heterolithic, shallow marine sequences contain examples of the Cruziana *(Association B) and* Skolithos *(Association A) ichnofacies, whereas shelf mudstones yield low-diversity assemblages with* Tomaculum *(Association D). Two associations (C, C1) that can be referred to the* Cruziana *ichnofacies are provisionally recorded from the storm interbeds; both are of low to moderate diversity and typically contain* Arthraria, Bifungites, Taenidium, *and* Phycodes. *Association E is dominated by U-shaped tubes with uncommon* Bergaueria. *It marks the recurrence of the* Skolithos *ichnofacies in a thin, but distinctive, oolitic ironstone sequence that is present throughout much of central Portugal and Spain.*

The composition of the assemblages and the general absence of typical deep water traces support the earlier suggestions of the existence of a relatively shallow shelf sea over much of central Iberia during Early to Middle Ordovician times.

Résumé

Les strates de l'Ordovicien inférieur et moyen de la zone Ibérique centrale, au Portugal et en Espagne, se sont accumulées dans une mer marginale froide et peu profonde, en bordure du Gondwana. Les séquences sont presque entièrement silicoclastiques et contiennent en prédominance des grès quartzeux épais (jusqu'à 1 000 m) et des pélites (jusqu'à 350 m d'épaisseur), ces derniers renfermant des nappes étendues de grès accumulées par suite de tempêtes. Elles contiennent aussi du conglomérat moins abondant et du fer oolitique rare.

On reconnaît actuellement 31 genres et plus de 45 espèces d'ichnofossiles dans l'intervalle de l'Arénigien-Caradocien inférieur de la zone Ibérique centrale. Les tableaux de répartition et les graphiques de diversité illustrent la présence d'ichnofossiles tout au long de l'Ordovicien inférieur et moyen, et leur prépondérance dans les faciès hétérolithiques.

On y reconnaît 5, et possiblement 6, associations d'ichnofaciès (A, B, C, C1, D, E) qui, en général, se limitent à des faciès sédimentologiques isolés. Des séquences hétérolithiques, de mer peu profonde, contiennent des exemples des ichnofaciès à Cruziana *(association B) et à* Skolithos *(association A), tandis*

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que des pélites de plate-forme renferment des assemblages de faible diversité qui comprennent Tomaculum *(association D). Deux associations (C, C1), qui pourraient faire partie de l'ichnofaciès à* Cruziana, *sont reconnues provisoirement dans les interstrates de tempête; leur diversité est faible ou moyenne et elles contiennent typiquement les genres* Arthraria, Bifungites, Taenidium *et* Phycodes. *L'association E contient en prédominance des tubes en U ainsi que des exemples peu fréquents de* Bergaueria. *Elle marque la réapparition de l'ichnofaciès à* Skolithos *dans une séquence mince mais distinctive de fer oolitique que l'on retrouve dans une grande partie du centre du Portugal et de l'Espagne.*

La composition des assemblages et l'absence générale d'ichnofossiles typiques d'eau profonde appuient l'hypothèse antérieure selon laquelle une mer marginale relativement peu profonde a recouvert une grande partie de la péninsule Ibérique centrale au cours de l'Ordovicien précoce et moyen.

INTRODUCTION

The diverse trace fossil assemblages in the Ordovician sequences of Portugal and Spain have long been known (Delgado, 1886, 1887; Hernández-Pacheco, 1908). However, the early studies, and indeed most of the later ones, have dealt almost exclusively with Lower Ordovician trace fossils (Costa, 1941; Hernández Sampelayo, 1950; Meléndez, 1954; Romano, 1974; Crimes and Marcos, 1976; Moreno et al., 1976; Baldwin, 1977a, b; Kolb and Wolf, 1979; Cooper and Romano, 1982; Pickerill et al., 1984b). Similarly in northwestern France, following the pioneer works of Rouault (1851), Tromelin (1878), and Lebesconte (1883), recent investigations (Durand, 1984, 1985a, b) have also tended to

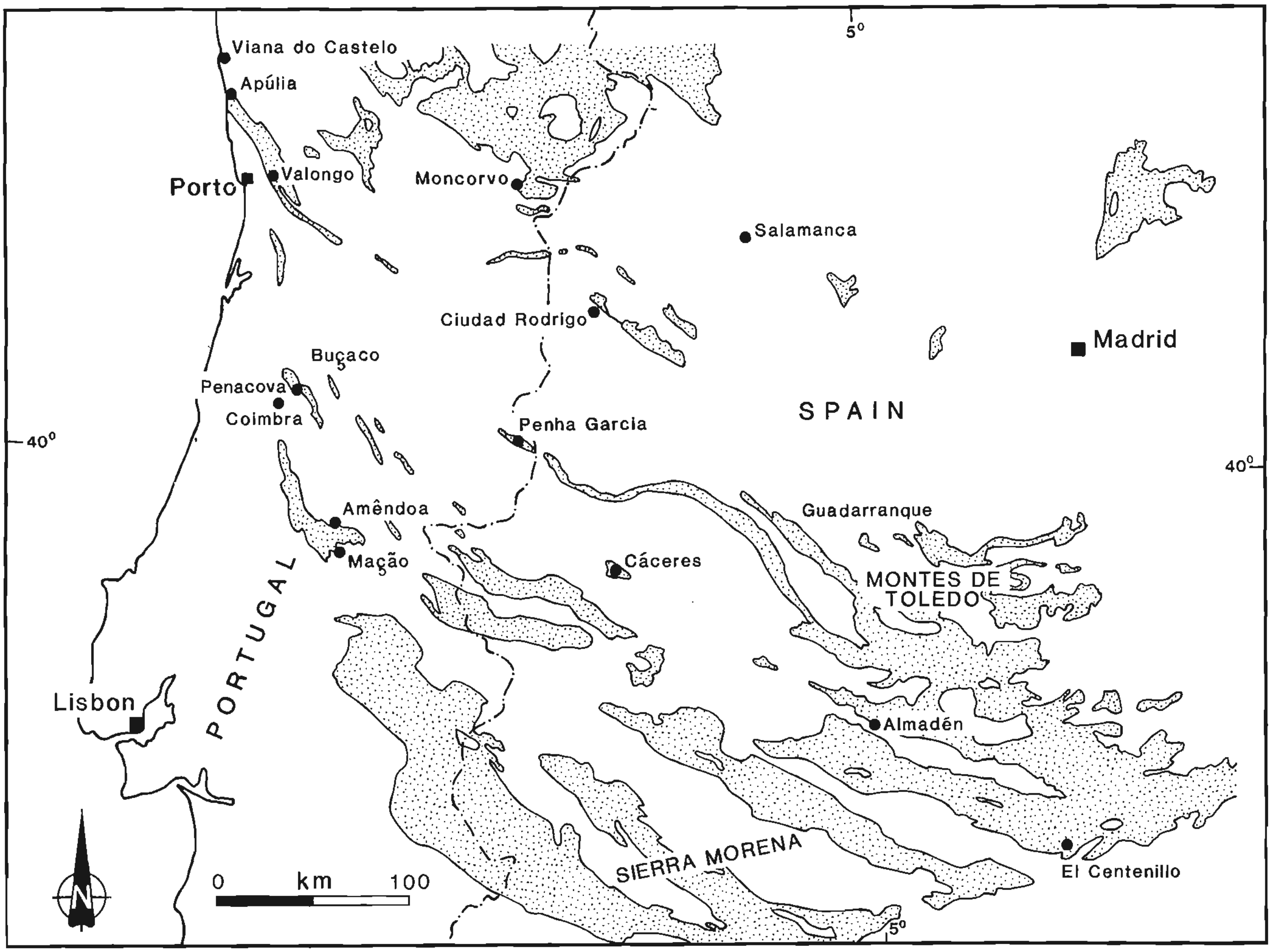


Figure 1. Map showing occurrence of Ordovician/Silurian rocks (stippled) in central Iberia (Central Iberian Zone) and localities mentioned in text.

concentrate on the Lower Ordovician. Rare exceptions include the works of Radig (1964) and Crimes et al. (1974) in Spain and Mélou and Plusquellec (1975) in Brittany.

This paper is concerned only with the trace fossils that occur in the Central Iberian Zone (C.I.Z.) (Hammann et al., 1982) (Figs. 1, 2) and are restricted to beds of Arenig to approximately middle Caradoc age. Delgado (1886, 1887) described and figured eight ichnogenera and 35 ichnospecies from the Lower Ordovician Amorican Quartzite in Portugal: 19 of these ichnospecies were included in *Cruziana*. A recent study of the trace fossils in the Armorican Quartzite from the Salamanca area, western Spain (Pickerill et al., 1984b) recorded 12 ichnogenera and 16 ichnospecies (four of which were assigned to *Cruziana*). Many of Delgado's *Cruziana* ichnospecies are now regarded as "behavioural and preservational variations" (Seilacher, 1970, p. 464). By comparison, similar assemblages are known from the Armorican Quartzite in Brittany; Durand (1985a) described 17 ichnogenera and 22 ichnospecies (four belonging to *Cruziana*) from this unit.

Despite the wealth of information on the variety of traces from the Armorican Quartzite, there has been little attempt to review and analyze their vertical distribution in central Iberia; this paper is partly concerned with this aspect of the Lower Ordovician ichnofossils. Radig (1964) described and figured *Tomaculum problematicum* from the Llandeilo of the Montes de Toledo (Fig. 1). Gutiérrez-Marco et al. (1984) listed seven ichnogenera from Llanvirn to lower Caradoc beds in central Spain (three were identified to ichnospecific level) but neither Delgado (1886, 1887, 1908) nor any subsequent author has studied Middle to Upper Ordovician traces from Iberia in any detail.

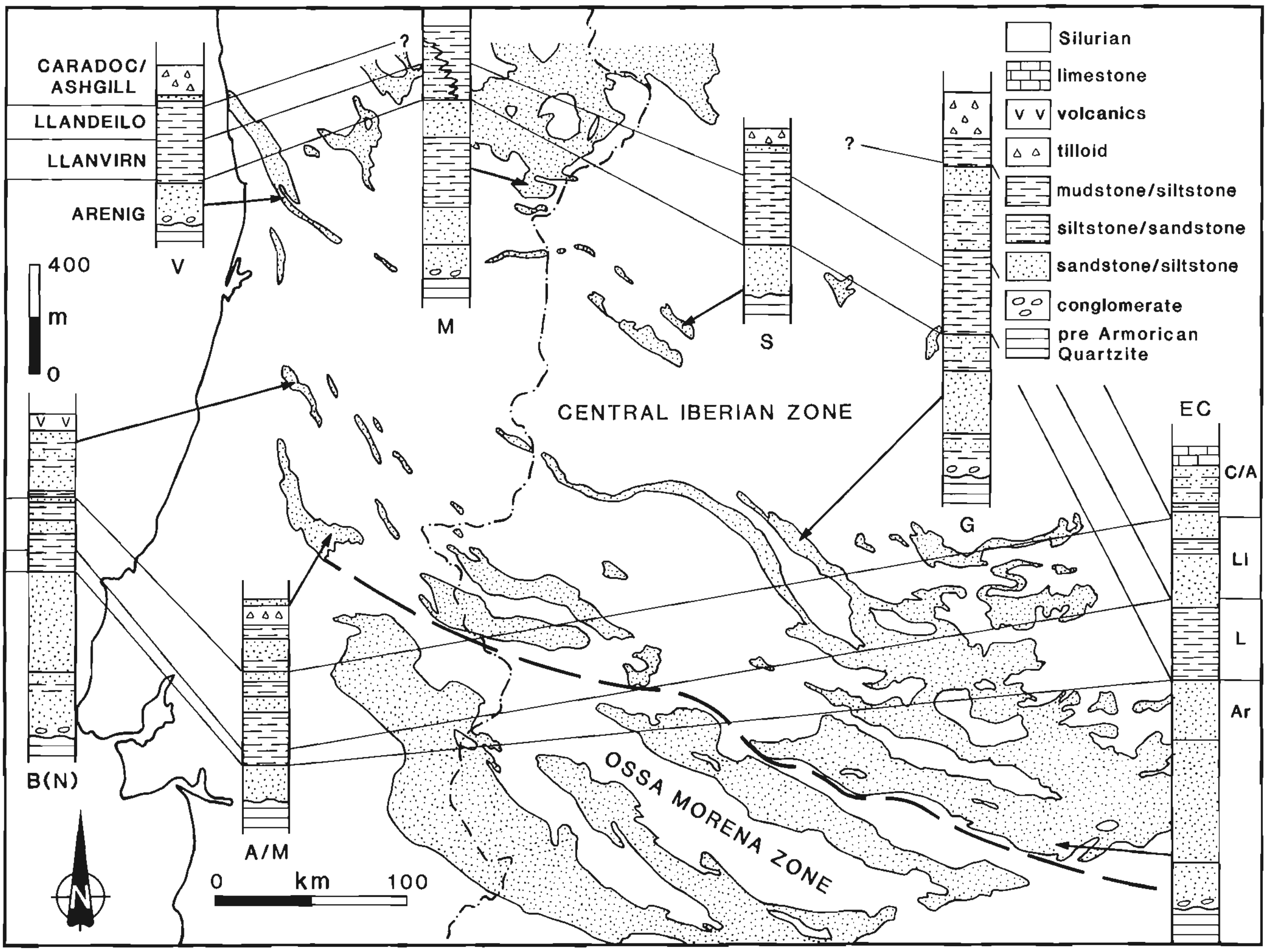


Figure 2. Selected lithostratigraphic sequences in the Central Iberian Zone. V, Valongo; M, Moncorvo, S, Salamanca; G, Guadarranque; EC, El Centenillo; A/M, Amêndoa/Mação; B(N), north Buçaco. Data from Valongo (Romano and Diggens, 1973–74), Moncorvo (Rebelo, 1980–81; Rebelo and Romano, 1988 for 1986), north Buçaco (Henry et al., 1973–74; Paris, 1981; Young, 1985, 1988), Amêndoa/Mação (Delgado, 1908; Young, 1985, 1988); Salamanca (Gutiérrez-Marco and Rábano, 1983), Guadarranque (Hammann et al., 1982; Hammann, 1983); El Centenillo (Hammann et al., 1982; Hammann, 1983).

The importance of trace fossils in sequences where body fossils are either rare or absent has been amply shown by Seilacher (1967, 1970), Crimes (1970), Baldwin (1977a) and others, and ichnofossils may provide the basis for age determinations and/or environmental interpretations. These aspects are briefly investigated below.

The data used to construct the range charts in Figure 3 were collected mainly by the author over a number of years during field work into other aspects of Ordovician biostratigraphy. Information in Spain is limited to a number of sections in the Montes de Toledo and Sierra Morena (Fig. 1). The following works were also used: Delgado (1886, 1887), Costa (1941), Bouyx (1966), Perdigão (1964–65), Seilacher (1970), Romano (1974, 1982), Romano and Diggens (1973–74), Cooper (1980), Cooper and Romano (1982), Pickerill et al. (1984b), Gutiérrez-Marco (1984), Gutiérrez-Marco et al. (1984), Rebelo and Romano (1988 for 1986). Further specimens were kindly made available by Drs. T.P. Young and A.H. Cooper.

STRATIGRAPHY

Seven simplified stratigraphic columns for selected regions in the C.I.Z. are shown in Figure 2. For each of the sequences the series boundaries for the Ordovician are indicated. The base of the Arenig is not shown due to the general absence of diagnostic faunas (body and trace fossils) although in the Montes de Toledo, Moreno et al. (1976) recognized the possible presence of beds of Tremadoc age below the Arenig Armorican Quartzite, on the basis of the range of *Cruziana* ichnospecies. The Arenig/Llanvirn boundary is drawn approximately at the top of the Armorican Quartzite and graptolitic faunas of early Llanvirn age occur immediately above the last quartzite bed in central Portugal (Romano et al., 1986). The Llanvirn/Llandeilo boundary normally lies within the mudrock sequence overlying the Armorican Quartzite; the exception is in the El Centenillo sequence (Fig. 1) where Hammann (1983) placed the boundary near the base of the El Caño Alternation. The evidence for this is inconclusive, however, as (presumably) Hammann used the presence of "*Neseuretus (N.) tristani* ssp. inc." (Hammann, 1983, p. 14) to indicate a post-Llanvirn age; this species ranges down into the upper Llanvirn. In central Portugal the base of the Caradoc was considered by Paris (1981) and Young (1985, 1988) to lie near the base of a mudstone unit (Carregueira Formation of Young); in north Portugal the Llandeilo/Caradoc junction probably lies toward the top of the Valongo Formation (Young and Romano, work in progress). Around Salamanca (Fig. 1) graptolitic shales of early Llandovery age overlie sparsely fossiliferous Llandeilo strata (Gutiérrez-Marco and Rábano, 1983), whereas in the Montes de Toledo (Guadarranque) and Sierra Morena (El Centenillo), Hammann (1983) located the base of the Caradoc near the base of a mudrock sequence, and toward the top of a quartzite sequence, respectively. The upper parts of the Caradoc and Ashgill are not considered in this paper.

The sequences illustrated in Figure 2 show a general similarity between the Arenig and Llanvirn, but there are differences in detail. McDougall et al. (1987) recently described the base of the Lower Ordovician Armorican Quartzite in the C.I.Z. and pointed out that, although it generally lies unconformably on the Complexo Xisto–Grauváquico (CXG) (a thick, upper Precambrian to Cambrian turbidite sequence), a conformable junction is seen in some places. The oldest proven, or assumed, Ordovocian rocks in the C.I.Z. are either the Armorican Quartzite or the "red beds" (Henry et al., 1973–74; Hammann, 1983) that locally occur below the quartzites. The latter (including the Sarnelha Formation at Buçaco) show a variety of fan–delta sediments (McDougall et al., 1987). The mudrocks overlying the Armorican Quartzite occur throughout the zone and generally form a thick monotonous sequence. They are locally interbedded with sandstone of Llanvirn and Llandeilo age that exhibit hummocky cross-stratification of storm origin (Brenchley et al., 1986). Neither of these sandstone incursions are known from the Valongo area (Romano and Diggens, 1973–74), nor have they definitely been recorded from Moncorvo (Fig. 1) or Salamanca, although in the former region the mudrocks of Llanvirn and possibly Llandeilo age pass laterally into a coarser facies (Rebelo and Romano, 1988 for 1986). In the Salamanca sequence, Gutiérrez-Marco and Rábano (1983) showed a more sandy facies at the top of the Middle Ordovician sequence. The Llandeilo sandstone strata are probably related to a regressive event (Fortey, 1984); they thicken to the south, toward the probable source area/shoreline. After the regressive phase, mud and silt deposition resumed in most regions until another short-lived, storm-generated event gave rise to hummocky sand at the top of the Llandeilo sequence (Young, 1988); this was then followed by mud of the early Caradoc transgression. In the southern regions, sedimentation was dominated during the Caradoc by thick sand accumulations interbedded with fossiliferous mud of considerable lateral extent. The sandstone frequently shows evidence of storm origin.

DIVERSITY AND VERTICAL RANGE OF TRACE FOSSILS IN THE CENTRAL IBERIAN ZONE

Figure 3 is a generalized lithostratigraphic log of the southern sections shown in Figure 2 (i.e., north Buçao, Amêndoa/Mação, Guadarranque, El Centenillo). For convenience only the lithostratigraphic terms of central Portugal are used (Henry et al., 1973–74; Cooper, 1980; Young, 1985, 1988). The graphs record the regional vertical ranges of trace fossils within these sections. In some cases the exact horizon is unknown (see explanation and key to Fig. 3), and in an attempt to represent all the ranges from different sections on a single log, some over-simplifications have inevitably been introduced. For instance, not all the ichnospecies shown in Figure 3 necessarily occur at any horizon or body of rock in one section.

Do the range charts give a true and accurate indication of the diversity and ranges of the taxa?

Of the ichnospecies listed in Figure 3, ten have not been recorded previously from the C.I.Z. Perhaps more importantly, of these ten ichnospecies, eight occur in units above the Armorican Quartzite, and two (*Zoophycos*? ichnosp. and *Lockeia*? ichnosp.) are so far only recorded from the Armorican Quartzite. The presence of these ichnospecies, previously unrecorded in post-Armorican Quartzite strata, indicates the general lack of ichnological work from this part of the sequence, and only three studies (Radig, 1964; Seilacher, 1970; Gutiérrez-Marco, 1984) have described post-Armorican Quartzite trace fossils.

The diversity and ranges of the taxa given in Figures 3 and 4 must then be open to question. Are the graphs faithful representations, or are they considerably biased by the very selective research done on these sequences? Although further work will undoubtedly add to the diversity of the vertical ranges, as well as expanding some of them, the general composition of the assemblages is considered to be essentially representative. For example, the only records of *Cruziana* occurring in beds overlying the Armorican Quartzite in the southern part of the C.I.Z. are of *Cruziana almadenensis* (Seilacher, 1970) from the Canteras (Botella) Quartzite, in the Sierra Morena (Almela et al., 1962) of Llandeilo to Caradoc age (Hammann, 1974, 1983) and of an unidentified specimen of *Cruziana* (possibly *Didymaulichnus*) from the Canaveral syncline, northeast of Cáceres (Carballeira et al., 1985), despite work on similar horizons and in similar facies elsewhere in the zone (Brenchley et al., 1986).

Diversity

Figure 4A depicts the diversity of the Ordovician trace fossils in the C.I.Z. Several notable features emerge from the graph. Diversities are highest in the coarser facies and, in particular, in the heterolithic parts of the sequence, such as at the top of the Armorican Quartzite and Monte da Sombadeira Formation. This latter feature may be due in part to the nature of the traces, which are more readily observable in heterolithic sequences (cf. Durand, 1985b). Diagenetic processes may enhance the preservation of trace fossils (Narbonne, 1984; Pickerill et al., 1984a), but these processes are clearly less effective in clean quartzose sandstone (Baldwin, 1977a), such as the Armorican Quartzite.

The diversity graph tends to mirror the nature of lithological boundaries. Thus, where formational boundaries are relatively sharp from sandstone/siltstone to mudstone (e.g., Monte da Sombadeira to Fonte da Horta Formation), there is a sudden drop in diversity (note that in Figure 4A this change does not quite coincide with the top of the Monte da Sombadeira Formation; this "effect" is due to combining the Portuguese lithological log with trace fossil information from the Los Rasos sandstone in Montes de Toledo, Spain). The greatest decline in diversity occurs where the Armorican Quartzite Formation grades into the overlying mudrocks.

Diversity is highest in the upper part of the Armorican Quartzite Formation. The lower diversity in the Monte de Sombadeira sandstone may be the result of environmental differences (see below). The difference in graph width between ichnogenera and ichnospecies is only really apparent in the Lower Ordovician. This is mainly due to the number of ichnospecies of *Cruziana*, and to a lesser extent, of *Daedalus, Didymaulichnus, Planolites*, and *Rusophycus*. The proportion of vertical to horizontal traces (Figure 4B) and its significance is considered later.

Trace fossils as age indicators in the Central Iberian Zone

Trace fossils generally have little use in age determination, although notable exceptions include ichnospecies of *Cruziana*. Cruzianids range from the Cambrian to Lower Triassic, but several authors (Seilacher, 1960, 1970; Crimes, 1968, 1969, 1970, 1975; Moreno et al., 1976; Baldwin, 1977a; Pickerill and Fillion, 1983; Rebelo and Romano, 1988 for 1986) have used some of the ichnospecies as age indicators in "unfossiliferous" or "poorly fossiliferous" strata. Of the cruzianid ichnospecies listed from the C.I.Z., *C. furcifera*, *C. rugosa*, and *C. goldfussi* are most frequently cited as indicating an Early Ordovician age; although, until the stratigraphic ranges of these ichnospecies are known in greater detail, age determination to series level should be regarded as tentative; particularly when only one or two of these ichnospecies are present. When all three ichnospecies co-occur, however, an Arenig age is more confidently indicated.

Rusophycus and *Phycodes* also have potential use as age indicators but, like cruzianids, it is doubtful whether they will ever achieve the age refinements of body fossils. Nevertheless, the presence of these trace fossils in "unfossiliferous" sequences is of considerable value in general age determinations. The value is lessened, however, in the early Ordovician of the C.I.Z. where a marked facies change occurs approximately at the Arenig/Llanvirn boundary and the full age range of these ichnospecies cannot be determined.

Vertical ranges and ichnofacies

Trace fossils occur throughout the Lower to Middle Ordovician sequences in the C.I.Z. (Fig. 3). Some, such as *Planolites* ichnosp., occur virtually continuously; others are restricted to part of the sequence (*Arthrophycus* ichnosp.); a number are only known from very restricted horizons (*Bergaueria hemispherica*).

Delgado (1908) noticed that the trace fossils in the Armorican Quartzite commonly showed a change from *Skolithos*-dominated assemblages in the lower part to *Cruziana*-dominated assemblages in the upper ("Quartzites à Scolithus" and "Quartzites à Bilobites"). Although this observation is probably true for many of the Armorican Quartzite (sensu lato) sequences in the C.I.Z., cruzianids do occur quite commonly low in the formation (Romano and Diggens, 1973–74) and *Skolithos* often ranges throughout.

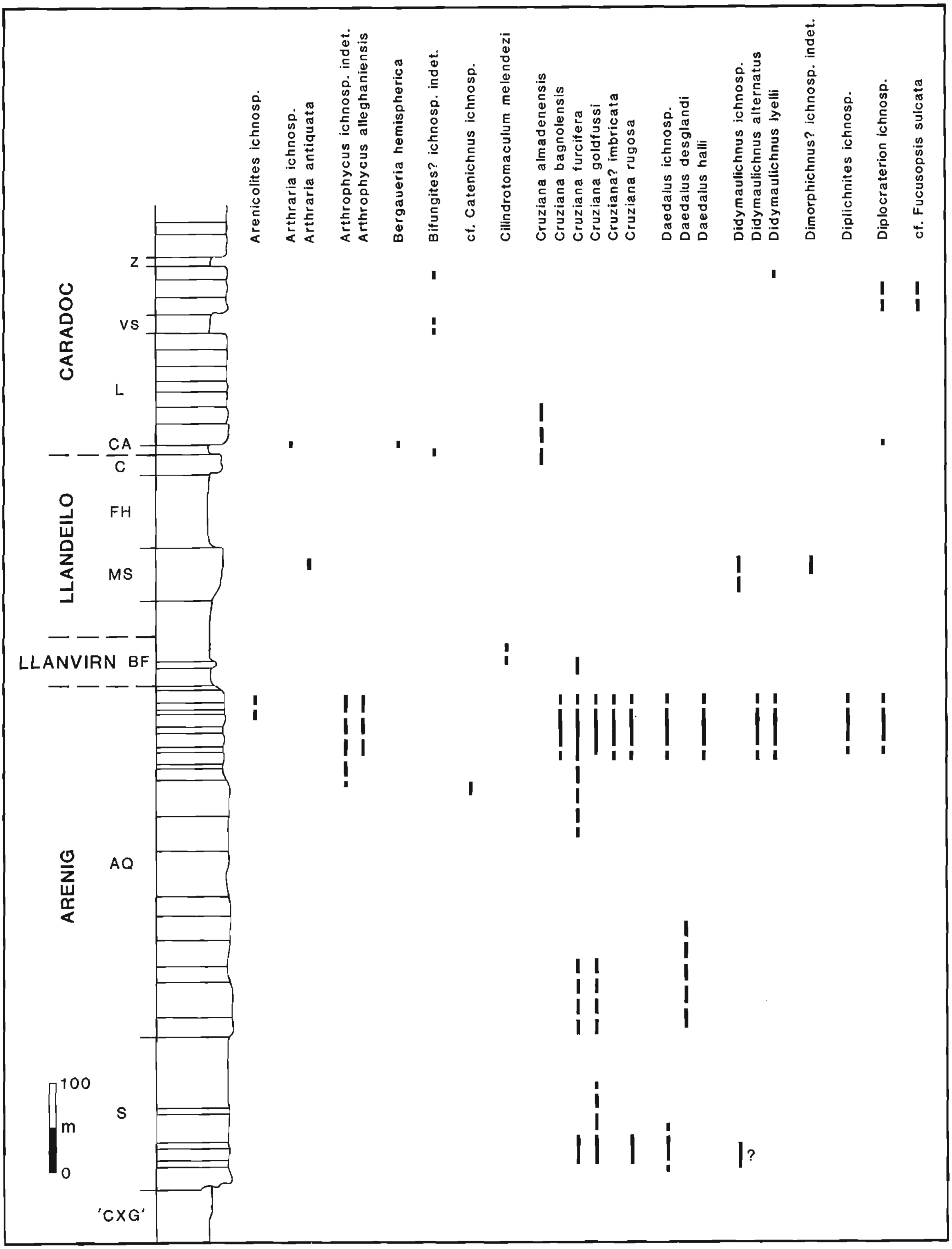
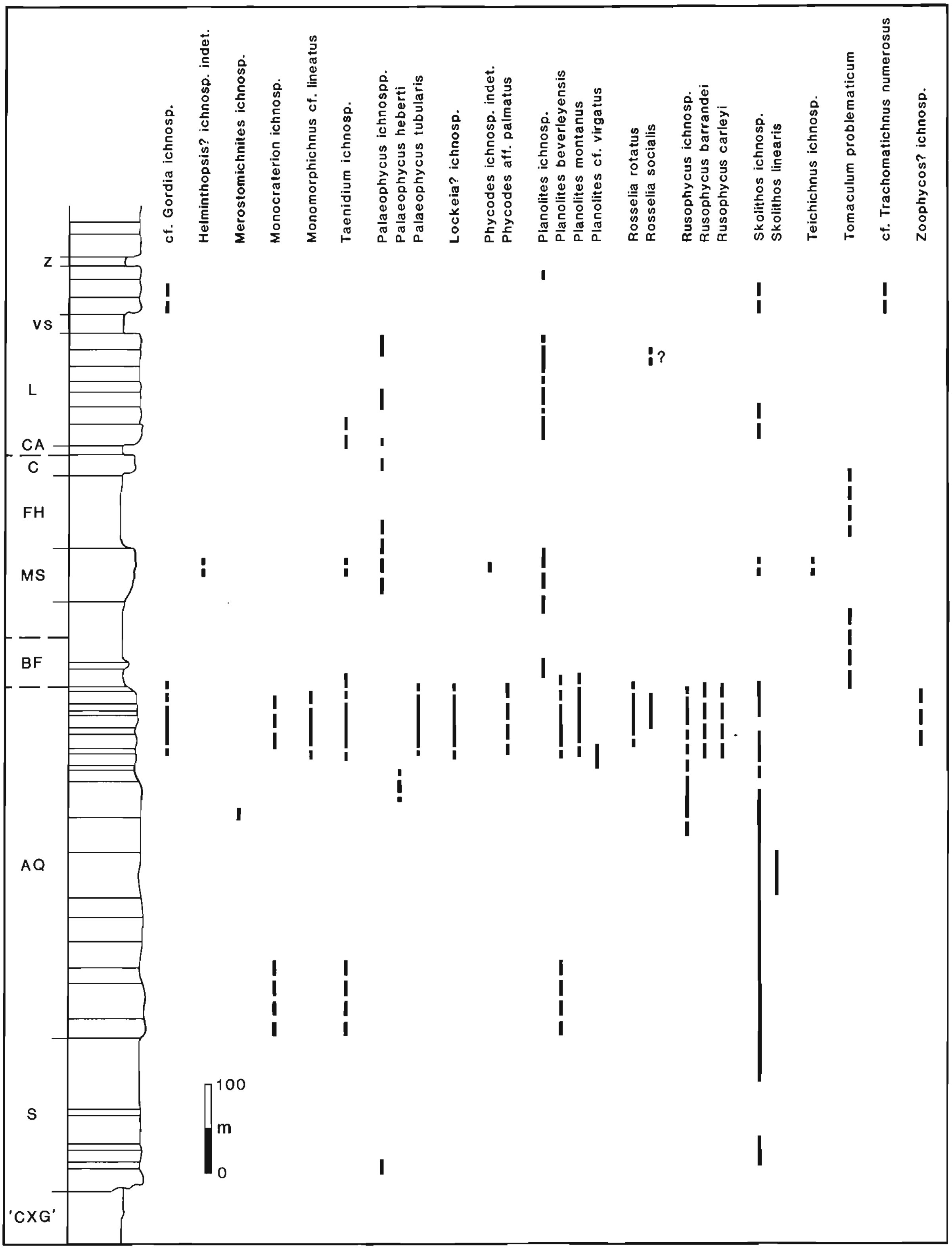


Figure 3. Generalized lithostratigraphy of the southern sections in the Central Iberian Zone with ranges of known trace fossils for the whole of the zone. Dashed lines indicate that exact horizon is not known. Abbreviations of lithostratigraphic units refer to those in Figure 4 or in caption. Note *Taenidium* and *Lockeia* are not in alphabetical order.



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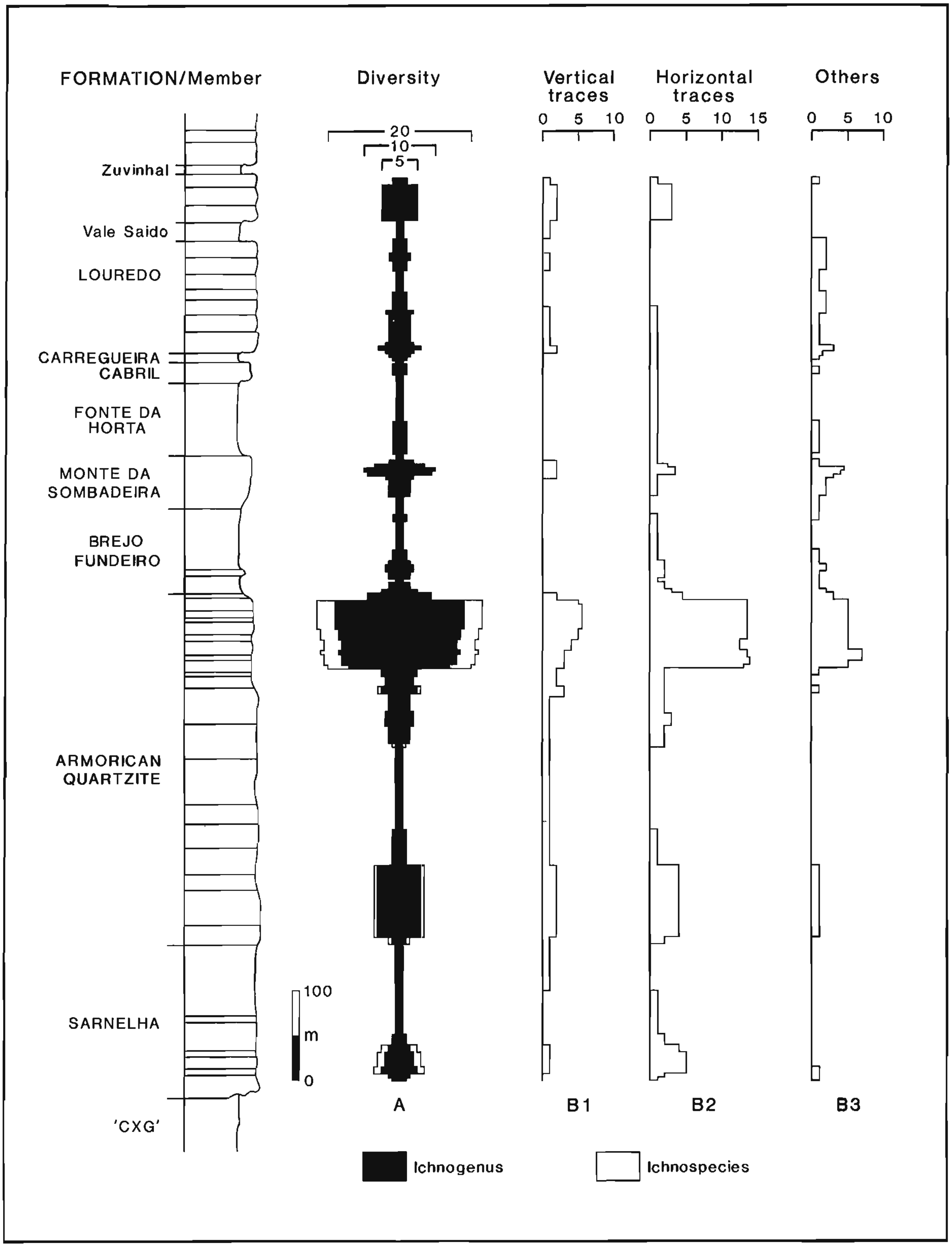
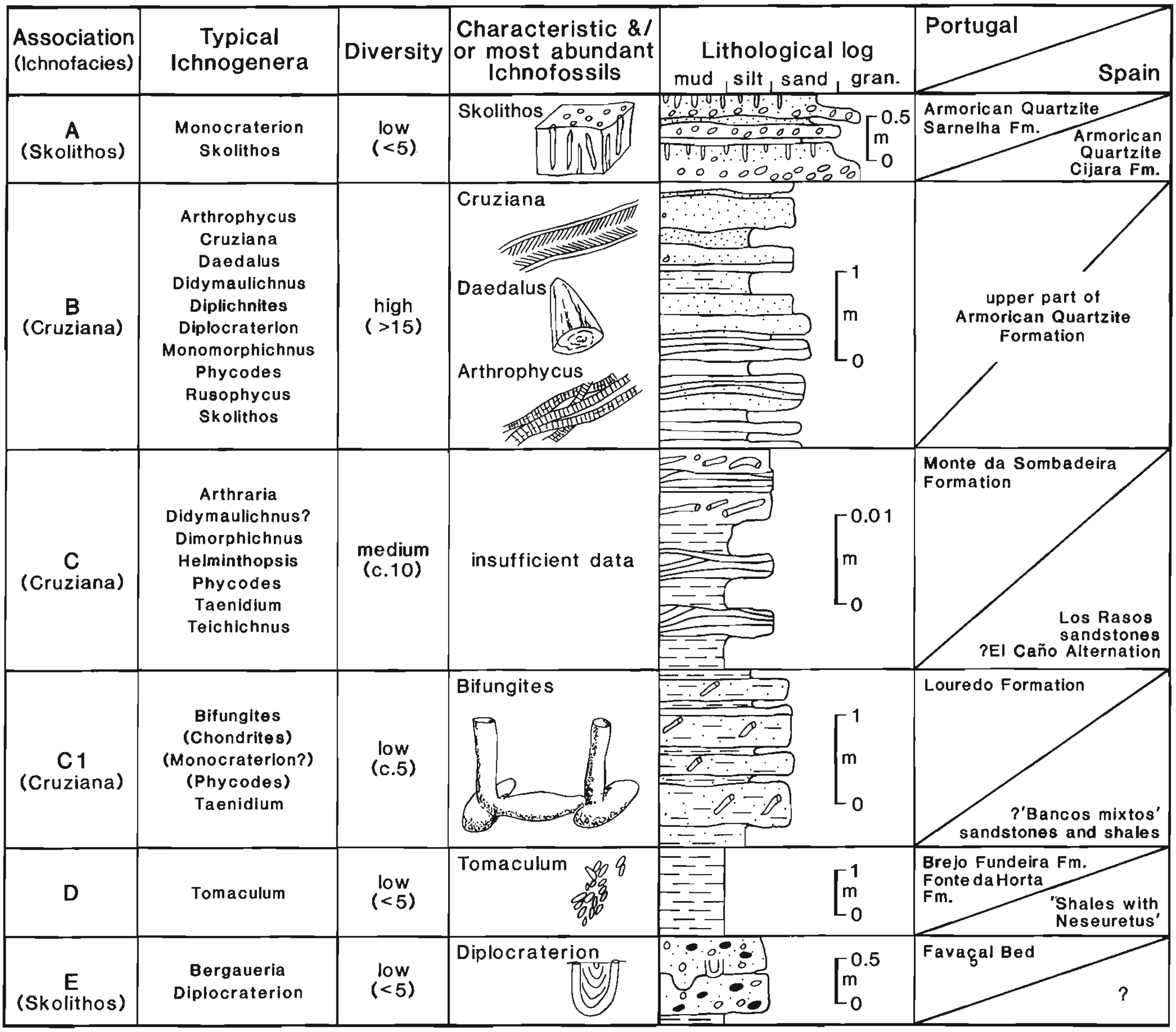


Figure 4. Diversity graph (A) of ichnogenera and ichnospecies from the Lower to Middle Ordovician sequence in the Central Iberian Zone. Also shown is distribution of vertical (B1), horizontal (B2), and other (B3) traces. See text for explanation. "CXG", Complexo Xisto-Grauváquico.

This "reversed" distribution was described from Apúlia, north Portugal (Romano, 1974) where cruzianids are most common some 40 m above the base and *Skolithos* is particularly abundant (up to 15,000 a square metre) between 85 and 110 m above the base. Unfortunately, the top of the Armorican Quartzite is not exposed at Apúlia and it cannot be determined whether there was a return to the *Cruziana*-dominated assemblage. Farther north still, at Viana do Castelo where the Armorican Quartzite is much attenuated, *Cruziana* and *Arthrophycus* also occur near the base of the sequence.

Of greater importance than the vertical ranges of individual ichnogenera (which may be modified following more detailed work on the sequences) is the presence of particular associations of trace fossils. These may be analyzed within the framework of ichnofacies developed by Seilacher (1964, 1967) and expanded by others (Bromley et al., 1984; Pemberton and Frey, 1984, 1985). At the present state of knowledge of Ordovician trace fossils from the C.I.Z., five or six associations have been recognized (Fig. 5). The generally ubiquitous ichnogenera *Planolites* and *Palaeophycus* are not considered in the definition of these associations. Association A (belonging to the *Skolithos* ichnofacies) and Association B (*Cruziana* ichnofacies) have previously been identified in other lower Paleozoic sequences. For example, Seilacher (1963) recognized the *Skolithos* (Pirispiki Red Beds) and *Cruziana* (Khabour Quartzite) ichnofacies from Iraq, and Crimes (1970) described virtually identical assemblages from the lower Paleozoic rocks of Wales. The trace fossils making up Association B in the C.I.Z. include *Rusophycus, Daedalus, Diplocraterion, Dimorphichnus, Phycodes*, and the ubiquitous



Association (Ichnofacies)	Typical Ichnogenera	Diversity	Characteristic &/ or most abundant Ichnofossils	Lithological log mud, silt, sand, gran.	Portugal	Spain
A (Skolithos)	Monocraterion Skolithos	low (<5)	Skolithos	0.5 m 0	Armorican Quartzite Sarnelha Fm.	Armorican Quartzite Cijara Fm.
B (Cruziana)	Arthrophycus Cruziana Daedalus Didymaulichnus Diplichnites Diplocraterion Monomorphichnus Phycodes Rusophycus Skolithos	high (>15)	Cruziana Daedalus Arthrophycus	1 m 0	upper part of Armorican Quartzite Formation	
C (Cruziana)	Arthraria Didymaulichnus? Dimorphichnus Helminthopsis Phycodes Taenidium Teichichnus	medium (c.10)	insufficient data	0.01 m 0	Monte da Sombadeira Formation	Los Rasos sandstones ?El Caño Alternation
C1 (Cruziana)	Bifungites (Chondrites) (Monocraterion?) (Phycodes) Taenidium	low (c.5)	Bifungites	1 m 0	Louredo Formation	?'Bancos mixtos' sandstones and shales
D	Tomaculum	low (<5)	Tomaculum	1 m 0	Brejo Fundeira Fm. Fonte da Horta Fm.	'Shales with Neseuretus'
E (Skolithos)	Bergaueria Diplocraterion	low (<5)	Diplocraterion	0.5 m 0	Favaçal Bed	?

Figure 5. Dominant ichnogenera in identified associations, characteristic ichnofossils, diversity (number of ichnospecies), generalized lithotype, and examples of lithostratigraphic units from Portugal and Spain.

Cruziana. In contrast, Association A is consistently of very low diversity (Frey and Seilacher, 1980) with *Skolithos* the dominant ichnogenus. One unusual aspect of the *Cruziana* ichnofacies (Association B) in the present study is the rare presence of *Zoophycos*? (also similar to *Spirophyton*) from the upper part of the Armorican Quartzite in central Portugal (see below).

The other four associations (C, C1, D, E) are less well defined, and are characterized in part by the absence of ichnogenera usually found in the *Skolithos* and/or *Cruziana* ichnofacies (Fig. 6). Associations C and C1 have a number of ichnogenera in common, namely, *Taenidium, Arthraria*, and/or *Bifungites* (the identification of these two latter ichnogenera is not without problems, see Fillion and Pickerill, 1984). The two associations are both of relatively low diversity; the former occurs in well bedded siltstone/mudstone sequences, the latter in more massively bedded, bioturbated, impure sandstone. The ichnogenera from Association C1 that are placed in parentheses in Figure 5 were recorded from the lower part of the Kermeur Formation in the Veryhac'h section, Crozon Peninsula, Brittany, which is equivalent to the lower part of the Louredo Formation of Buçaco, central Portugal. Associations C and C1 probably represent more distal expressions of the *Cruziana* ichnofacies (cf. Pemberton and Frey, 1984). They may not differ enough to be considered distinct, but the larger diversity in Association C, together with the presence of more common horizontal traces (*Didymaulichnus*?, *Dimorphichnus*, *Helminthopsis*) and complex feeding burrows (*Phycodes, Teichichnus*), suggest that the distinction is reasonable. Only further work will support or invalidate this suggestion.

Association D is of very low diversity. Apart from the description of *Cilindrotomaculum melendezi* (Gutiérrez-Marco, 1984) only *Tomaculum* has been definitely recorded from the generally homogeneous mud- to silt-grade sediments (Gutiérrez-Marco et al, 1984). Irregular strings of pellets reach up to 20 cm in length in the Llanvirn sequences and may be better referred to *Syncoprulus* (see Pickerill et al., 1987).

A distinctive lithofacies within the Ordovician sequences of the C.I.Z. comprises beds of oolitic ironstone, generally less than 1 m thick. The thickest and best documented of these are the Favaçal bed at the base of the Louredo Formation and Leira Ma Member at the base of the Porto de Santa Anna Formation (Paris, 1979, 1981; Young, 1988). The Leira Ma Member is not considered here. The Favaçal bed contains *Bergaueria hemispherica* (not recorded elsewhere in the Central Iberian Zone) as well as abundant U-shaped tubes referred to *Diplocraterion* ichnosp. This unique assemblage is recognized as a discrete association (Association E) referable to the *Skolithos* ichnofacies.

Ichnofacies and environments

Since the important early work by Seilacher (1967) on the relationship between trace fossil assemblages and bathymetry, various authors have noted the persistent recurrence of distinctive ichnofacies within particular lithofacies (Crimes, 1970; Baldwin, 1977a; Frey and Seilacher, 1980). The *Skolithos* and *Cruziana* ichnofacies are perhaps among the best documented and, although many of the published examples are from the lower Paleozoic, the two ichnofacies are also recorded from the Pliocene (Mayoral, 1986). The ichnological and lithological data from the Armorican Quartzite Formation (Figs. 3, 5) support previous environmental interpretations. Associations A and B, which represent typical expressions of these ichnofacies, are not further discussed here. Relevant works relating to these ichnofacies in the C.I.Z. can be found in Romano (1974), Moreno et al. (1976), Pickerill et al. (1984b) and McDougall et al. (1987), and in Durand (1984, 1985a, b) for Brittany. The presence of *Zoophycos*? in this ichnofacies is unusual and is perhaps more typically found in lithofacies that indicate a lower energy, offshore environment, although as noted by Eagar et al. (1985) and others, this trace fossil is now known to have a considerable bathymetric range.

The remaining associations (C, C1, D, E; Figs. 5, 6) occur in lithofacies that have not received much attention in the C.I.Z.; namely, shelf, storm sandstone (C) and bioturbated muddy sandstone (C1), shelf mudstone (D), and oolitic ironstone (E). Trace fossils representing Association C are present in the upper part of the Monte da Sombadeira Formation in central Portugal and Los Rasos sandstone in Montes de Toledo, central Spain, where they occur in laminated siltstone with interbeds showing low-angle crossbedding that resembles hummocky cross-stratification (Facies F3 of Brenchley et al., 1986, p. 244). The laminated siltstone strata are interpreted as having been formed during storms when silt was carried seaward and deposited from suspension. The facies is interbedded with thicker-bedded sandstone, which exhibits hummocky cross-stratification and rarely shows extensive bioturbation. The part more proximal to the Monte da Sombadeira/Los Rasos unit is represented by the El Caño Alternation of the Sierra Morena, central Spain. Recognizable trace fossils are generally rare in the latter, more nearshore sequences. The storm sands were laid down on a gently dipping shelf, above storm wave base, in probably no more than 80 m of water.

Association C1 occurs in generally massive, bioturbated and sparsely fossiliferous, muddy sandstone with thin interbeds of mudstone. The lower part of the Louredo Formation in central Portugal typifies the lithofacies, as does the lower part of the equivalent Kermeur Formation, Crozon Peninsula, Brittany. Most of the trace fossils recorded from this association were recovered from the French sequence. The poorly sorted sandstones have been heavily reworked by organisms, destroying the original mixed mud/sand sequence (Young, 1985) and may possibly represent a modified lithofacies such as that associated with Association C. The situation is similar to that of Suite 3 of Pemberton and Frey (1984).

Association D occurs in grey, weakly fissile mudstone, which represents the background sedimentation of the shelf (Facies F1 of Brenchley et al., 1986). Below the storm sandstone of the Monte da Sombadeira Formation, the mudstone commonly contains disarticulated and dispersed shelly faunas with infrequent bedding-plane assemblages; above this unit coquinas are most common, suggesting a slightly shallower depth, perhaps just below fair weather wave base. Trace fossil diversity is low and bioturbation is more common in the younger beds.

The trace fossils (Fig. 6) of Association E occur within a thin sequence of strongly bioturbated oolitic ironstone with phosphatic intraclasts. The oolitic ironstone rests disconformably on the underlying mudstone and generally grades up into a sandstone/ mudstone sequence. The ooids developed virtually in situ on a shallow, broad shelf and are interpreted as having been formed in a relatively stable environment on a transgressive surface (T.P. Young, pers. comm.). The lithofacies is widespread and probably represents a period of very slow sedimentation or almost nondeposition (Young, 1985, 1989). Association E is comparable to Suite 1B of Pemberton and Frey (1984) and represents an episodic occurrence of the *Skolithos* ichnofacies.

Early observations by Seilacher and others noted the preponderance of vertical traces in higher-energy (typically shallow water) environments and of essentially horizontal, more complicated traces in lower-energy (typically deeper water) environments. The trace fossils from the C.I.Z. were classified according to their vertical or horizontal attitude in an attempt to determine whether there was any correlation with the lithotype/energy level. An initial problem was deciding what constitutes a vertical or horizontal trace, as the majority of traces show vertical and horizontal components. For example, essentially bedding-plane cruzianid traces may quite frequently exhibit steeply sloping bases and deep vertical components (Pickerill et al., 1984b; Goldring, 1985), whereas *Zoophycos* shows excavations in both vertical and horizontal directions. Simple burrow systems (such as *Planolites* and *Palaeophycus*) may also include a variety of attitudes. However, in most of the ichnospecies recorded, the traces have been assigned to one type or the other on the basis of the dominant behavioural pattern. The results are plotted in Figure 4 (B1, B2, and B3; B3 includes mainly simple, oblique burrows) and are seemingly inconclusive. In the upper part of the Armorican Quartzite Formation (*Cruziana* ichnofacies) horizontal traces dominate, whereas in the deeper water Monte da Sombadeira Formation, and possibly lower part of the Louredo Formation, there is no clear dominance. It would be more meaningful to attempt to interpret the morphology of the traces in terms of the presumed behavioural type.

Figure 6 shows the proposed associations and inferred behavioural types of the ichnogenera listed in Figure 3 (see legend to Fig. 6 for explanation of symbols). In Association A and possibly Association E, all traces (excluding *Palaeophycus* and *Planolites*) of these low-diversity assemblages are represented by dwelling burrows typical of the *Skolithos* ichnofacies, although Crimes et al. (1977) and Pollard (1988) regarded *Bergaueria* as essentially a resting trace. The ichnogenera in Association D (again excluding *Palaeophycus* and *Planolites*) are also of low diversity but are diagnostic in that locomotion, burrowing, or resting traces are virtually absent. None of the remaining associations shows such low diversity or restricted behavioural types.

S	▨	C	C	C	S	Ichnofacies	Be-haviour type
E	D	C1	C	B	A	Association / Ichnogenus	
					•	cf. Catenichnus	DB
				•	●	Monocraterion	DB
		○	○	•	●	Skolithos	DB
			○	●		Didymaulichnus	FB/FT?
				•		Arenicolites	DB
				●		Arthrophycus	FB
				●		Cruziana	FT/FB
				●		Daedalus	FB/DB
				•		Diplichnites	LT
				•		cf. Gordia	FT
				•		Lockeia	RB/EB
				•		Merostomichnites	LT
				•		Monomorphichnus	LT
		○		•		Rosselia	DB
				●		Rusophycus	RB
				•		Zoophycos	FB/DB
			•	•		Dimorphichnus	LT
			•	•		Phycodes	FB
		•	•	•		Taenidium	FB
●				●		Diplocraterion	DB
			•			Helminthopsis	FT
			●			Teichichnus	FB
○			•			Arthraria	DB?
		•				Bifungites	DB
		•				cf. Trachomatichnus	LT
		•				cf. Fucusopsis	FT
	●					Tomaculum	FP
	•					Cilindrotomaculum	FP
•						Bergaueria	DB
	●	●	●	●	○	Palaeophycus	FB
	●	●	●	●	●	Planolites	FB

Figure 6. Distribution of ichnogenera in identified associations from the Central Iberian Zone, and inferred behaviour type. • present, abundant, • uncertain identification; DB, dwelling burrow; FB, feeding burrow; FP, faecal pellet; FT, feeding trail, LT, locomotion trail; RB, resting burrow; S, **Skolithos**; C, **Cruziana**.

It is reasonable to conclude that under relatively persistent high- or low-energy conditions, which resulted in essentially homolithic sequences (pure quartzose sandstone or mudrocks) (cf. Baldwin, 1977a; Durand, 1985a), several factors could restrict the diversity of trace fossil types. These factors include:

1. Reduction in the number of organisms (diversity and/or abundance) living in that particular environment. This is not the case in the mudrock sequences, in which rich trilobite and infaunal bivalve faunas occur sporadically.
2. Dominance of particular behavioural types to suit the sedimentary regime (Purdy, 1964).

These two factors would be controlled by substrate, food supply, oxygen level, etc.

3. Low fossilization potential of traces in a well sorted sediment (Pickerill, 1975).
4. Reworking of sediment, which destroys a high proportion of traces, particularly the surface or shallow-burrowing types (Fürsich, 1975).
5. Intense bioturbation, which may destroy otherwise discrete traces and homogenize the sediment (Fürsich, 1975).

Conversely, the heterolithic sequences, with their frequently abundant and varied ichnofaunas, undoubtedly supported a rich benthic fauna, or diverse life styles; the destruction of traces was countered by the original abundance of traces, and higher fossilization potential (Baldwin, 1977a). The upper part of the Armorican Quartzite typifies this type of sequence, yet body fossils are generally rare. Macrofossils recorded are limited to rare trilobites, inarticulate brachiopods, and three species of bivalves. An abundant soft-bodied fauna was probably responsible for many of the diverse ichnofossils present.

CONCLUSIONS

Shelf mudstones (Brejo Fundeiro and Fonte da Horta formations) with locally rich bivalve, trilobite, ostracode, and brachiopod faunas contain low-diversity trace fossil assemblages, and little direct evidence of the environmental conditions may be deduced from the ichnofauna. The lower Llandeilo (Monte da Sombadeira Formation and Los Rasos sandstone) and lower Caradoc (Louredo Formation) storm siltstones and sandstones interbedded within the shelf mudstones yield relatively abundant and diverse traces, but this diversity decreases in the more sandy, proximal, and higher energy parts of the units (El Caño Alternation (cf. Pemberton and Frey, 1984). Shallow marine, storm dominated heterolithic sequences (upper part of the Armorican Quartzite Formation) contain abundant and varied trace fossil assemblages, but are virtually devoid of body fossils. The lack of complex horizontal traces, which are typical of the *Nereites* ichnofacies (Seilacher, 1967; Frey and Seilacher, 1980), indicates that none of the sequences under study was deposited in deep water. Thus, the composition of the trace fossil assemblages currently known from this region tends to support the paleoenvironmental interpretation of these sequences inferred from sedimentological and body fossil studies (Hammann and Henry, 1978; Romano, 1982; Gutiérrez-Marco et al., 1984; Rábano, 1984; Brenchley et al., 1986).

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Environmental cycles and bioevents in the Upper Ordovician Red River-Stony Mountain solitary rugose coral province of North America

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Abstract

Stratigraphic sequences within the Red River-Stony Mountain Province record two major environmental cycles in a vast area of epicontinental seas. The Red River Cycle began with a transgressive phase in the middle Edenian that reached its maximum in the early Maysvillian, and was followed by a regressive phase during the later Maysvillian and early Richmondian. The transgressive phase of the Stony Mountain Cycle reached its peak in the middle Richmondian, and was followed by a regressive phase during the later Richmondian and Gamachian. In general, the greatest water depth and most open marine conditions were attained during the first cycle, and seas were shallowest and most restricted toward the end of the second cycle. Recognition of the transgressive maxima, when deposition was most widespread, permits chronostratigraphic correlation throughout the province (and likely beyond).

Bioevents involving evolution and dispersion of solitary rugose corals in the Red River-Stony Mountain Province, as well as fluctuations in the relative abundance of taxa, were closely related to changes in water depth and degree of environmental restriction during the two cycles. The first major evolutionary event took place during the transgressive maximum of the Red River Cycle, when water depth in some areas exceeded the limit for corals and they were confined to basin margins and structural highs. The second occurred at the end of the Red River regressive phase, when most areas became inhospitable and corals were restricted to the centres of cratonic interior basins and to cratonic margin areas of the province. Corals became widely dispersed during the subsequent transgressive phase of the Stony Mountain Cycle. Late in the regressive phase of that cycle, taxa were introduced from outside the Red River-Stony Mountain Province, and the endemic corals became extinct.

Résumé

Les séquences stratigraphiques au sein de la province de Red River-Stony Mountain révèlent deux importants cycles environnementaux dans une vaste région de mers épicontinentales. Le cycle de Red River a débuté avec une phase transgressive à l'Édenien moyen qui a atteint son maximum au Maysvillien inférieur et il a été suivi par une phase régressive au cours du Maysvillien supérieur et du Richmondien inférieur. La phase transgressive du cycle Stony Mountain a atteint son maximum au Richmondien moyen et a été suivie par une phase régressive durant le Richmondien supérieur et le Gamachien. En général, c'est au cours du premier cycle que les profondeurs d'eau les plus grandes et les conditions marines les plus grandes ont été atteintes; de plus, les mers ont atteint leur niveau le moins profond et leurs conditions les plus réduites vers la fin du second cycle. La détermination des maxima de transgression, lorsque la sédimentation a atteint sa plus grande superficie permet d'établir une corrélation chronostratigraphique à l'échelle de toute la province (et probablement au-delà).

Les bioévénements liés à l'évolution et à la dispersion des rugosa solitaires dans la province de Red River-Stony Mountain ainsi que les fluctuations de l'abondance relative de taxons ont été étroitement corrélés aux changements de profondeur d'eau et au degré de restriction environnementale au cours des

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deux cycles. Le premier événement important lié à l'évolution a eu lieu au cours du maximum de transgression du cycle de Red River lorsque la profondeur de l'eau dans certaines zones a dépassé la limite des coraux et que ces derniers ont été confinés aux marges de bassin et aux hauteurs structurales. Le second événement a eu lieu à la fin de la phase régressive de Red River lorsque la plupart des zones sont devenues inhospitalières et que la présence des coraux s'est limitée au centre des bassins intérieurs cratoniques et aux zones de marge cratonique de la province. Les coraux se sont dispersés considérablement au cours de la phase de transgression subséquente du cycle de Stony Mountain. À la fin de la phase de régression de ce cycle, des taxons provenant de zones extérieures de la province de Red River-Stony Mountain ont été introduits et les coraux indigènes ont disparu.

INTRODUCTION

The Late Ordovician Red River-Stony Mountain Solitary Rugose Coral Province of North America occupied a vast area of shallow epicontinental seas, in which carbonate sedimentation was predominant (Fig. 1). Characteristic taxa are species of *Grewingkia* including triangulate to trilobate corals (the normal cross-sectional shape is circular), and the genera *Lobocorallium* (derived from *Grewingkia*), *Deiracorallium*, *Salvadorea*, and *Bighornia*. Paleoecological and taphonomic studies indicate a close ecological relationship between *Grewingkia/Lobocorallium* and *Deiracorallium*, and between *Salvadorea* and *Bighornia*. The former group evidently preferred relatively deep water (but above storm wave base), high energy, open marine environments, whereas the latter favoured comparatively shallow water, low energy environments and tolerated somewhat restricted conditions (Elias, 1982b, p. 1586, 1587; Elias, 1985, p. 13-16; Buttler et al., 1988, p. 56; Elias, Zeilstra, and Bayer, 1988, p. 33; Elias, Nowlan, and Bolton, 1988, p. 347). Therefore, temporal and geographic patterns in the distribution and relative abundance of these genera should prove useful for paleoenvironmental reconstruction.

The purpose of this paper is to document and characterize major environmental cycles in the Red River-Stony Mountain Province, to evaluate their chronostratigraphic significance, and to determine the relationship of bioevents involving solitary rugosan evolution and dispersion to these cycles.

WILLISTON BASIN, SOUTHERN MANITOBA

The middle Edenian through Gamachian sequence of southern Manitoba and elsewhere within the Williston Cratonic Interior Basin represents virtually continuous, mainly carbonate deposition, and records two major environmental cycles (Figs. 1, 2; Cowan, 1971, p. 239, 240; Kendall, 1976, p. 46-58). The lower Red River Formation (Dog Head, Cat Head, and Selkirk members) was deposited in open marine conditions. The Dog Head and Selkirk members are litho- and biofacies equivalents. The Cat Head Member between them, which contains chert nodules, sponges (Rigby, 1971), comparatively small cephalopods (Flower, 1971), and remarkably well preserved megascopic algae (Fry, 1983), is here considered to record the first transgressive maximum. The lower Stony Mountain Formation (Gunn and Penitentiary members) is also an open marine deposit. The argillaceous, highly fossiliferous Gunn Member marks the second transgressive phase. The Penitentiary Member represents a transition to more restricted conditions (Elias, 1982b, p. 1585). The upper Red River Formation (Fort Garry Member) and the upper Stony Mountain Formation (Gunton and Williams members) plus Stonewall Formation consist of sedimentary cycles, some of which include evaporites. These units were deposited in restricted conditions during later stages of the first and second regressive phases.

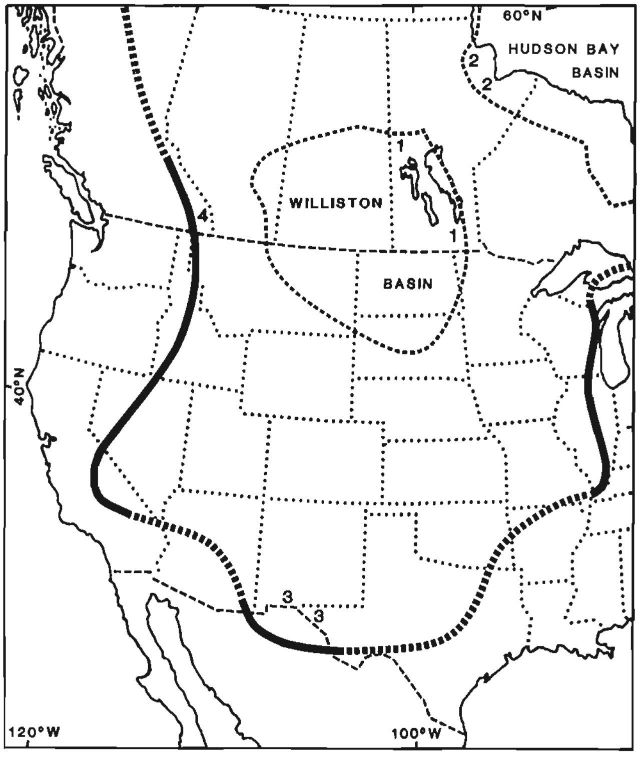

Figure 1. Late Ordovician Red River-Stony Mountain Solitary Rugose Coral Province in western and central North America (Elias, 1981, 1982a, 1983, 1985, new data; Buttler et al., 1988); thick solid and dashed lines show established, and uncertain boundaries, respectively. Medium dashed lines show outline of cratonic interior basins. Locations 1-4 are southern Manitoba, northern Manitoba, southern New Mexico and westernmost Texas, and Akutlak Creek in British Columbia, respectively.

Solitary rugose corals are diverse and common in the Selkirk Member of the Red River Formation, but Whiteaves (1897, p. 154) reported only a few comparatively small specimens from the Dog Head Member. The two small individuals documented by Caramanica (1973, p. 314, 351) are identified as *Salvadorea*? sp. and *Grewingkia* sp. herein. Caramanica (1973, p. 62) and Elias (1981, p. 10) observed single unidentifiable specimens in the Cat Head Member. It seems likely that the majority of species in the Selkirk Member were not present earlier. *Grewingkia robusta*, which is known from older strata elsewhere in the Red River-Stony Mountain Province, is considered to be the ancestor of *G. crassa crassa*, *G. lamellosa*, *G. dilata*, and *G. haysii selkirkensis* (a trilobate species) (see Elias, 1985, p. 17). The last three are endemic, suggesting evolution within the Williston Basin. The Selkirk assemblage is *Grewingkia*-dominated. *Salvadorea randi* is the only species known to range into the Fort Garry Member. This reflects the change from relatively deep and open environments to comparatively shallow and restricted conditions during the regressive phase of the first cycle.

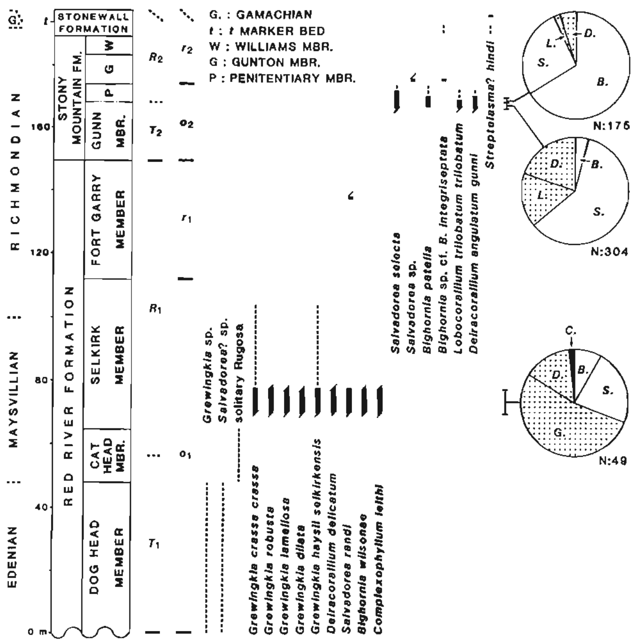

Figure 2. Composite section showing Upper Ordovician units and solitary rugose corals in southern Manitoba (see Fig. 1, loc. 1); T, transgressive phase; R, regressive phase; o, open marine stage; r, restricted marine stage; $_1$, first cycle (Red River Cycle); $_2$, second cycle (Stony Mountain Cycle) (see text). Section based on McCabe and Barchyn (1982, Fig. 10) and McCabe (pers. comm., 1987). Ages extrapolated from Sweet (1979, p. 54, Fig. 4). Ordovician-Silurian boundary (top of Gamachian) has been placed at top of Stonewall Formation (Stearn, 1956, p. 14-16) and at t marker bed within Stonewall (Brindle, 1960, p. 18, 19). Distribution of corals based on Caramanica (1973) for Dog Head and Cat Head members of Red River Formation, Elias (1981, 1982b, 1983, new data) and Elias, Nowlan, and Bolton (1988) for Cat Head to Fort Garry members of Red River Formation and Stony Mountain Formation, and Stearn (1956) for Stonewall Formation, with taxonomy updated; wide line indicates interval of occurrence, narrow dashed line indicates that exact position within an interval is uncertain, diagonal line indicates start or termination of range at base or top of measured section. Pie diagrams show relative abundance of genera in the intervals indicated, based on corals identified to species level by Elias (1981) for Red River Formation at Garson, and corals identified to genus level by Elias (1982b, Table 2; identified to species level by Elias, 1983) for Stony Mountain Formation at Stony Mountain; N, number of specimens.

New taxa appear in the Gunn Member of the Stony Mountain Formation. *Salvadorea selecta* arose from *S. randi*, probably in the Williston Basin. *Bighornia patella* descended from *B. wilsonae*, possibly in this basin. *Lobocorallium trilobatum* evolved from *G. haysii*. *Deiracorallium angulatum* arose from a Maysvillian species such as *D. delicatum*. All four species range into the Penitentiary Member. The Gunn assemblage is *Salvadorea*-dominated, suggesting that the sea was shallower and less open during the second transgressive phase than during Selkirk Member deposition. The *Bighornia*-dominated assemblage at the base of the Penitentiary Member reflects somewhat more restricted conditions. *Bighornia* sp. cf. *B. integriseptata* occurs at the base of the Gunton Member, where moulds of evaporite minerals mark the change to restricted environments during the regressive phase of the second cycle. This large form of *Bighornia*, which evolved from *B. patella*, is also present in the Stonewall Formation. *Lobocorallium* and *Deiracorallium* are not known from the restricted stage of the regressive phase. *Streptelasma*? *hindi*, introduced during Stonewall deposition, is not a Red River-Stony Mountain taxon.

HUDSON BAY BASIN, NORTHERN MANITOBA

The Upper Ordovician sequence of northern Manitoba and elsewhere within the Hudson Bay Cratonic Interior Basin represents predominantly carbonate deposition and records two major environmental cycles (Figs. 1, 3; Le Fèvre et al., 1976, p. 75-77, 79-81). It seems probable that sedimentation was virtually continuous from the Edenian through Gamachian, as in the Williston Basin. The Portage Chute Formation (lower Bad Cache Rapids Group) and Caution Creek Formation (lower Churchill River Group) were deposited in open marine conditions. Peaks in conodont frequency logs in the middle of the Portage Chute and lower part of the Caution Creek (Le Fèvre et al., 1976, Fig. 2) may mark the first and second transgressive maxima. Member 1 of the Caution Creek is the most fossiliferous unit in the entire sequence (Nelson, 1963, p. 14). The Surprise Creek Formation (upper Bad Cache Rapids Group) and the Chasm Creek Formation (upper Churchill River Group) plus the Red Head Rapids Formation accumulated in somewhat restricted to evaporitic environments during later stages of the first and second regressive phases.

The solitary rugosan assemblage in the Portage Chute and lower Surprise Creek formations is *Grewingkia*-dominated, indicating relatively deep, open marine conditions. Its disappearance higher in the Surprise Creek reflects the change to restricted environments during the regressive phase of the first cycle. New taxa appear in the Caution Creek Formation.

Salvadorea distincta distincta and *Bighornia patella* were introduced to the Hudson Bay Basin from elsewhere. The trilobate coral *Grewingkia haysii haysii* evolved from *G. robusta*, possibly in this basin. The representative of *Deiracorallium* arose from a Maysvillian species such as *D. harveyi*. The Caution Creek assemblage is *Salvadorea*/*Bighornia*-dominated, suggesting that the sea was not as deep and open during the second transgressive phase as during deposition of the Portage Chute and lower Surprise Creek formations. The Caution Creek taxa range into or through the Chasm Creek Formation.

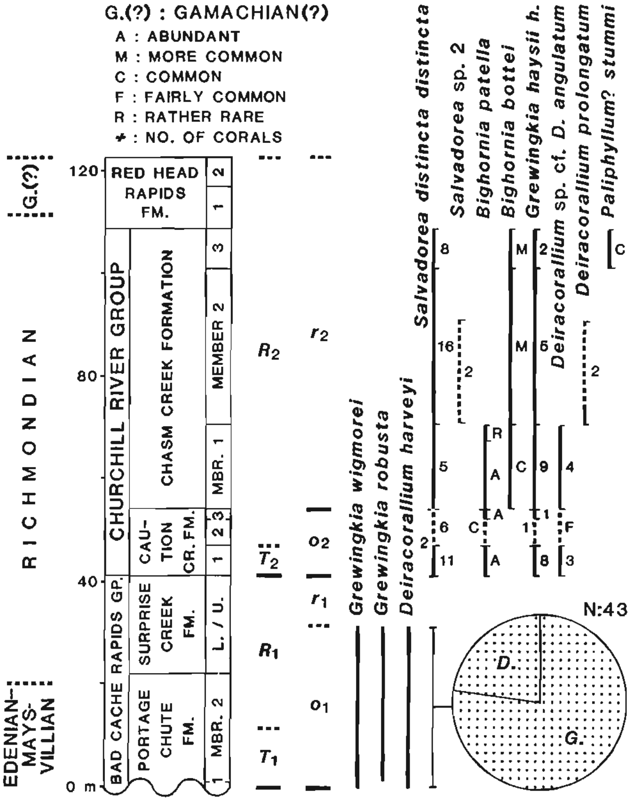

Figure 3. Composite section showing Upper Ordovician units and solitary rugose corals in northern Manitoba (see Fig. 1, loc. 2); T, transgressive phase; R, regressive phase; o, open marine stage; r, restricted marine stage; $_1$, first cycle (Red River Cycle); $_2$, second cycle (Stony Mountain Cycle) (see text). Section based on Nelson (1981, Fig. 2). Ages based on extrapolation from Williston Basin to Hudson Bay Basin using Sweet (1979, p. 54, Fig. 4) and Cumming (1975, Table 1). Distribution of corals based on Nelson (1963, 1981), with taxonomy updated; solid line indicates interval of occurrence, dashed line indicates that exact position within an interval is uncertain. Pie diagram shows relative abundance of genera in the interval indicated, based on corals identified to species level by Nelson (1981, Table 3); N, number of specimens. For Churchill River Group, relative abundance is indicated using number of corals identified to species level by Nelson (1981, Table 3) and statements by Nelson (1963) for the marked intervals; explanations of abbreviations are given above stratigraphic column.

Bighornia *bottei* (possibly a synonym of *B. integriseptata*), a large species that evolved from *B. patella*, appears at the base of the Chasm Creek. This coincides with the change from open to relatively restricted conditions during the second regressive phase. *Salvadorea* sp. 2 (of Nelson, 1981) and *Deiracorallium prolongatum* were later introduced to the Hudson Bay Basin. *Paliphyllum*? *stummi*, which appears in the upper Chasm Creek, is not a Red River-Stony Mountain taxon. Assemblages in the Chasm Creek are *Salvadorea*/*Bighornia*-dominated. The disappearance of *Deiracorallium* followed by other solitary corals at the top of the Chasm Creek mark changes to more restricted conditions near the end of the regressive phase.

SOUTHERN NEW MEXICO AND WESTERNMOST TEXAS

The middle Edenian to late Richmondian Montoya Group represents virtually continuous, predominantly carbonate deposition in a cratonic margin setting, several hundred kilometres from the Ouachita Geocline (Figs. 1, 4; Brimberry, 1984; Elias, 1985, p. 3-7 and references cited therein; Measures, 1985a, b). The Second Value Dolomite was deposited

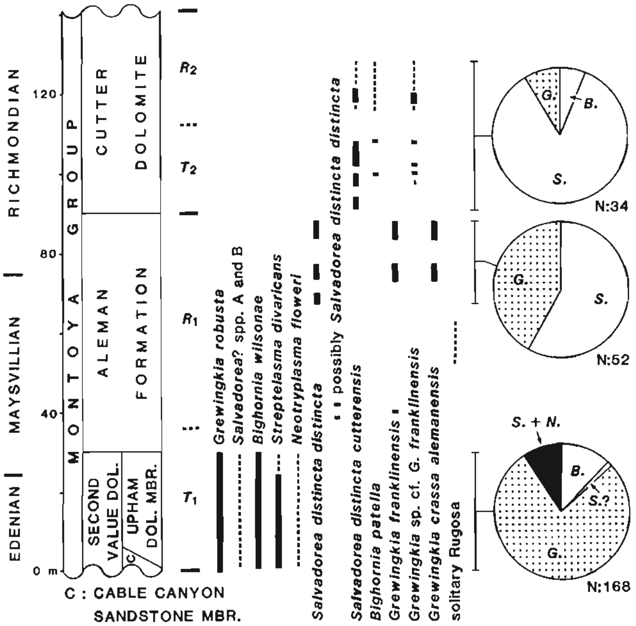

Figure 4. Composite section showing Upper Ordovician units and solitary rugose corals in southern New Mexico and westernmost Texas (see Fig. 1, loc. 3); T, transgressive phase; R, regressive phase; $_1$, first cycle (Red River Cycle); $_2$, second cycle (Stony Mountain Cycle) (see text). Ages and thickness of units based on Sweet (1979, Fig. 4). Distribution of corals based on Elias (1985, 1986), with taxonomy updated; wide line indicates interval of occurrence, narrow dashed line indicates that exact position within an interval is uncertain. Pie diagrams show relative abundance of genera in the intervals indicated, based on corals identified to species level (positively and probably) by Elias (1985, Table 1); N, number of specimens.

in open marine conditions during a transgressive phase. The basal portion of the Aleman Formation marks the first transgressive maximum. Relict anhydrite nodules scattered through the formation in one area suggest somewhat restricted conditions, at least locally, during the regressive phase. The Sierra Diablo area of westernmost Texas became emergent at the end of Aleman deposition, recording the regressive maximum. Argillaceous basal beds of the Cutter Dolomite mark the beginning of the second transgressive phase. From the composite conodont relative abundance log of Sweet (1979, Fig. 4), it is inferred that a transgressive maximum in the middle Cutter was followed by a regressive phase. Cyclic bedding has been reported in the Cutter.

The *Grewingkia*-dominated assemblage of the Second Value Dolomite reflects relatively deep, open marine conditions during the first transgression. These corals disappeared as water depth neared its maximum. *Grewingkia franklinensis* and possibly *Salvadorea distincta distincta* appear in the Aleman Formation above the transgressive peak. The first is an endemic, trilobate species that arose from *G. robusta*, likely in the vicinity of Montoya sedimentation. The second possibly evolved from a coral such as *Salvadorea*? sp. A or B (of Elias, 1985), which occurs in the Second Value. The endemic subspecies *G. crassa alemanensis*, which likely arose from *G. crassa crassa* of the Williston Basin and northern North America, appeared later during the regressive phase. The upper Aleman assemblage is *Salvadorea*-dominated, indicating that environments were shallower and more restricted during the regressive phase than during Second Value deposition. *Salvadorea distincta cutterensis*, an endemic subspecies, appears at the base of the Cutter Dolomite. It arose from *S. distincta distincta*, likely in the area of Montoya deposition. *Bighornia patella* was introduced during the second transgressive phase. *Grewingkia* sp. cf. *G. franklinensis* is the only representative of that genus in the Cutter. The high relative abundance of *Salvadorea* in the upper Montoya suggests that the sea was relatively shallow, and more restricted during the second environmental cycle than during the first. Solitary rugose corals disappeared from this region during the second regressive phase.

AKUTLAK CREEK, BRITISH COLUMBIA

The Beaverfoot Formation in the southern Rocky Mountains of British Columbia and Alberta represents dominantly carbonate deposition that began during the Late Ordovician and continued into the Early Silurian in a cratonic margin setting an unknown distance from the edge of the continent (Fig. 1; Buttler et al., 1988, p. 49, 50, 58-60). The *Bighornia-Thaerodonta* Zone of the lower Beaverfoot has been traced to 128 m above the base of the formation at Akutlak Creek (Fig. 5). All solitary rugosan taxa that are present (as well as *Salvadorea* sp. 2 [of Nelson, 1981] and *Bighornia* sp. cf. *B. bottei*, which have been found in this zone at other localities) also occur in the Churchill River Group of northern Manitoba. The entire *Bighornia-Thaerodonta* Zone is probably Richmondian in age, and strata of the lower Beaverfoot are here considered to have been deposited during the second major Late Ordovician environmental cycle.

Solitary rugose corals were introduced to this region during a transgressive phase, in which the dominantly clastic Whiskey Trail Member of the Beaverfoot Formation was deposited. *Deiracorallium prolongatum* may have evolved from *D. harveyi* of the Hudson Bay Basin. Norford (1969, p. 32) stated that "the horizon of widespread Beaverfoot transgression seems to be the top of the Whiskey Trail." The assemblage in the Whiskey Trail is *Salvadorea*-dominated, but *Grewingkia* is dominant in overlying carbonate beds of the Beaverfoot. This is considered to reflect deeper water and more open conditions at the peak of the transgression. Higher in the *Bighornia-Thaerodonta* Zone, the increasing relative abundance of *Salvadorea* and *Bighornia*, decreasing abundance of *Grewingkia*, and disappearance of *Deiracorallium* suggest a regressive phase and increasingly restricted conditions.

DISCUSSION AND CONCLUSIONS

Stratigraphic sequences within the Red River-Stony Mountain Province that represent virtually continuous deposition from the Edenian to Richmondian or Gamachian record two

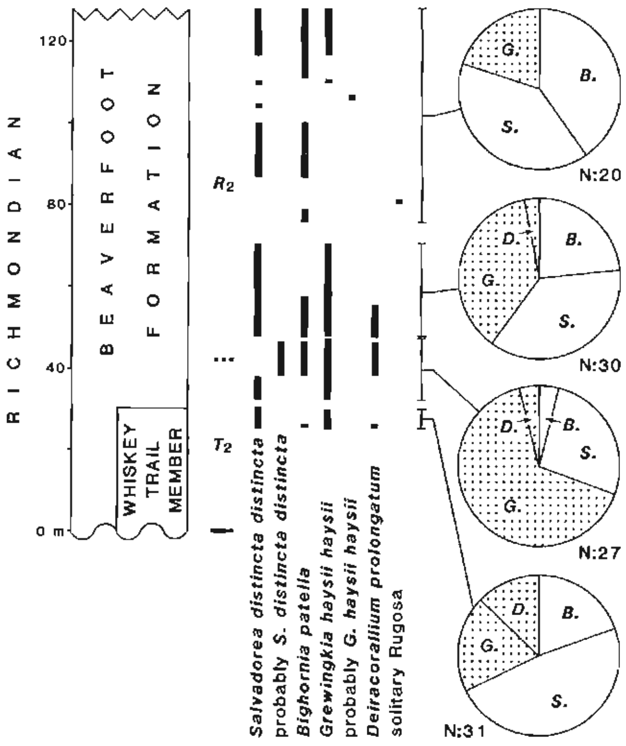

Figure 5. Section showing Beaverfoot Formation and distribution of solitary rugose corals at Akutlak Creek, British Columbia (see Fig. 1, loc. 4), based on Buttler et al. (1988); T, transgressive phase; R, regressive phase; $_2$, second cycle (Stony Mountain Cycle) (see text). Top of section marks highest level to which *Bighornia-Thaerodonta* Zone has been traced in the Beaverfoot of the southern Rocky Mountains, British Columbia and Alberta; entire zone is probably Richmondian. Pie diagrams show relative abundance of genera in the intervals indicated, based on corals identified to species level (positively and probably) by Buttler et al. (1988, Table 3.1, Fig. 3.4); N, number of specimens.

major environmental cycles in a vast area of epicontinental seas (Figs. 2-4). The first, here termed the Red River Cycle, began with a transgressive phase in the middle Edenian that reached its maximum in the early Maysvillian, and was followed by a regressive phase during the later Maysvillian and early Richmondian. The transgressive phase of the second cycle, here termed the Stony Mountain Cycle, reached its peak in the middle Richmondian, and was followed by a regressive phase during the later Richmondian and Gamachian. Le Fèvre et al. (1976, p. 76) noted that the latter regression corresponds to the latest Ordovician glacial maximum, and the first may reflect an earlier climax.

Greater thicknesses of sediment accumulated during regressive phases than during transgressive phases of the Red River and Stony Mountain cycles in cratonic interior basins and cratonic margin areas. Differences from place to place in the thickness of cycles and phases within them reflect variable rates of subsidence and sedimentation (e.g., Elias, Nowlan, and Bolton, 1988, p. 346). For both cycles in the Williston and Hudson Bay basins, relatively open marine conditions prevailed during the transgressive phase and early stage of the regressive phase, whereas environments were somewhat restricted to evaporitic during the later stage of the regressive phase. Cyclic sedimentation during the restricted phases in the Williston Basin (and probably elsewhere) was related to fluctuations in sea level (likely eustatic; Sweet, 1979, p. 54) and/or basin salinity (Kendall, 1976, p. 53-55, 57; Elias, Nowlan, and Bolton, 1988, p. 346).

In general, the greatest water depths and most open marine conditions were attained during the Red River Cycle, and seas were shallowest and most restricted toward the end of the Stony Mountain Cycle. In the western Midcontinent of the United States south of the Williston Basin, deposition was widespread during the early Maysvillian (maximum submergence) and middle Richmondian, and unconformities span the late Maysvillian-early Richmondian and the late Richmondian-Gamachian (maximum emergence) (see Sweet, 1979, Fig. 4). These unconformities suggest that seas in cratonic interior basins became isolated from ocean waters along the continental margin during regressive phases of the two cycles.

It seems probable that recognition of the transgressive maxima in the Red River and Stony Mountain cycles (when deposition was most widespread) would permit chronostratigraphic correlation among sections within the Red River-Stony Mountain Province (and likely beyond). Thus, the Cat Head Member of the Red River Formation, strata in the middle of the Portage Chute Formation, and basal beds of the Aleman Formation are correlable. Strata near the base of the Stony Mountain and Caution Creek formations, near the middle of the Cutter Dolomite, and in the *Bighornia-Thaerodonta* Zone of the Beaverfoot Formation a short distance above the Whiskey Trail Member, are considered correlative (Figs. 2-5). The above correlations between New Mexico-Texas and the Williston Basin in southern Manitoba agree precisely with those of Sweet (1979, Fig. 4) based on conodont biostratigraphy extrapolated to a subsurface section within the Williston Basin in Montana. Sweet equated the lower Cutter Dolomite in New Mexico-Texas with the Herald Formation (= Fort Garry Member of the Red River Formation) in the Williston Basin. If the boundary between the Red River and Stony Mountain cycles is considered synchronous, the lower Cutter correlates with basal beds of the Stony Mountain Formation. A problematic sequence containing Richmondian conodonts at Churchill in northern Manitoba (see Johnson et al., 1988, p. 96, 112) probably records the transgressive phase of the Stony Mountain Cycle, and likely represents the Churchill River Group (a possibility noted by Norford, 1971, p. 206).

Bioevents involving evolution and dispersion of solitary rugose corals in the Red River-Stony Mountain Province, as well as fluctuations in the relative abundance of taxa, were closely related to changes in water depth and degree of restriction during the Red River and Stony Mountain cycles. The first major evolutionary event evidently occurred at the time of the first transgressive maximum, in the early Maysvillian. It is suggested that as water depth in some places such as the Williston Basin and area of Montoya deposition exceeded the limit for most or all corals, their geographic ranges were reduced to basin margins and structural highs. Evolution apparently occurred rapidly in these populations, which were probably small, and the descendants and survivors dispersed as favourable conditions returned very early in the regressive phase. Trilobate species of *Grewingkia* and corals definitely representing *Salvadorea* appeared at that time. As conditions changed from relatively deep and open to shallower and more restricted during the first regressive phase, the *Grewingkia*-dominated assemblage disappeared from the Hudson Bay Basin, and *Grewingkia* was followed by *Salvadorea* as the dominant taxon in the Williston Basin and area of Montoya deposition.

The second major evolutionary event occurred at the end of the first regressive phase, in the early Richmondian. It is suggested that as most areas became inhospitable, geographic ranges of corals were reduced to the centres of cratonic interior basins and to cratonic margin areas in the Red River-Stony Mountain Province. Evolution apparently occurred rapidly, and the descendants and survivors dispersed widely within the province as favourable conditions spread during the second transgressive phase. *Bighornia patella*, and *Lobocorallium* or trilobate species of *Grewingkia* (but not normal to triangulate forms of *Grewingkia*), typically occur in these assemblages, which are generally *Salvadorea*-dominated. In the Williston and Hudson Bay basins, the appearance of a large species of *Bighornia* coincided with a change to more restricted conditions during the second regressive phase. Late in the regressive phase (late Richmondian-Gamachian), taxa were introduced from outside the Red River-Stony Mountain Province, and the endemic corals became extinct.

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Flexural interpretation of relationships between Ordovician tectonism and stratigraphic sequences, central and southern Appalacians, U.S.A.

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Abstract

Much of Ordovician deposition in the Appalachian Basin occurred in foreland basins developed in response to Taconian tectonism. The recent development of lithospheric flexure models suggests a scenario of foreland basin formation and "maturation" that can be followed in the stratigraphic record, thereby elucidating more clearly relationships between tectonism and sedimentation. The resulting scenario consists of a four-part, consecutive, process-response sequence: 1) process — collisional braking; response — development of an extensive unconformity; 2) process — active deformational loading; response — deposition of transgressive carbonates and overlying dark shale in a rapidly subsiding, underfilled foreland basin; 3) process — cessation of active tectonism and advent of "loading-type" relaxation accompanied by major erosion; response — infilling of foreland basin with flysch-like clastic sediment; and 4) process — rebound in orogen and advent of "unloading-type" relaxation; response — cratonward progradation of a marginal marine redbed sequence.

This dark shale through redbed sequence is repeated twice in the Ordovician of the central and southern Appalachians, indicating two distinct Taconian tectophases. Moreover, these sequences show evidence of migration in space and time both parallel and perpendicular to the strike of the orogen, suggesting that the sediments reflect the progress of orogeny along the orogen.

Although the general nature and timing of the Taconian Orogeny and resulting sedimentation are largely known, the parallelism between Ordovician stratigraphic sequences and predictions from lithospheric flexure models suggest that the above scenario may permit more exact interpretations of causal, temporal, and spatial relationships between tectonism and sedimentation. Clearly, the more fully preserved Ordovician sequences of the central and southern Appalachians may well serve as patterns for the application of this scenario elsewhere.

Résumé

Un grand nombre de dépôts ordoviciens dans le bassin appalachien se sont produits dans des avant-bassins qui se sont formés sous l'effet du tectonisme du Taconique. Le développement récent des modèles de flexure lithosphérique suggère un scénario de formation de l'avant-bassin et de "maturation" que l'on peut suivre dans les témoignages stratigraphiques, et qui éclaircissent les rapports entre le tectonisme et la sedimentation. Le scénario qui s'ensuit consiste en une séquence de quatre étapes consécutives de processus et effet: 1) processus — freinage après collision; effet — développement d'une discordance étendue; 2) processus — chargement déformationnel actif; effet — dépôt de carbonates transgressifs et de schistes argileux foncés surjacents dans un avant-bassin partiell ement rempli s'affaissant rapidement; 3) processus — cessation du tectonisme actif et apparition d'une relaxation du type "compression"; effet — remplissage de l'avant-bassin avec des sédiments clastiques du genre flysch; 4) processus — rebondissement de l'orogène et apparition d'une relaxation du type déchargement; effet — progradation vers le craton d'une séquence de couches rouges marines marginales.

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Cette séquence, qui va du schiste argileux foncé aux couches rouges, s'est répétée deux fois au cours de l'Ordovicien dans de bassin appalachien, indiquant deux phases taconiques distinctes. De plus, ces séquences portent l'évidence d'une migration spaciale et temporelle à la fois parallèle et perpendiculaire à la direction de l'orogène, ce qui suggère que les sédiments suivent l'evolution de l'orogénèse le long de l'orogène.

Bien que la nature et la chronologie générales de l'orogénèse du Taconique et de la sédimentation qui en résulte soient en grande partie connues, le parallélisme entre les séquences stratigraphiques orodoviciennes et les prédictions des modèles de flexure lithosphérique suggèrent que le scénario décrit ci-dessus puisse permettre d'interpréter de façon plus exacte les rapports causals, temporels et spatiaux entre le tectonisme et la sédimentation. Il est clair que les séquences ordoviciennes plus complètement développées dans les Appalaches centrales et méridionales pourraient bien servir de modèle pour l'application de ce scénario ailleurs.

INTRODUCTION

The Lower Ordovician rocks of eastern and central United States are a remarkably uniform sequence of dolomitic carbonates representing relatively shallow water deposition on the passive, southeastern platform margin of Laurentia. Abruptly, however, at about the Early-Middle Ordovician transition, the well defined patterns of Early Ordovician sedimentation were disrupted by uplift and erosion and succeeded by a complex of interbedded calcareous and detrital facies, which reflect the mobilization of the southeastern margin of Laurentia by the Taconian Orogeny. The term "Taconian" is used in the sense of Kay (1969) to include all Ordovician deformation along the Atlantic coast, although a growing trend exists to elevate what are herein called Taconian "tectophases" to the rank of orogenies (e.g., Drake et al., 1989).

In response to this Taconian mobilization of the Laurentian margin, simple patterns of platform-margin sedimentation were replaced by complex facies patterns largely restricted to resulting Appalachian foreland basins (Fig. 1). Although it was soon realized that Taconian tectonism was involved in the origin of these complex facies patterns, the exact mechanisms were uncertain. For example, Shanmugam and Walker (1978) and Benedict and Walker (1978), dealing largely with deep basinal facies, suggested that the necessary subsidence resulted from flexural bending or down-to-basin normal faulting. However, in the apparent absence of a major load, Shanmugam and Walker (1980) later argued that basement downwarping was a more likely explanation. Shanmugam and Lash (1982), Quinlan and Beaumont (1984), and Lash (1987a) were among the earliest to apply a more modern understanding of flexural mechanisms in explaining the Middle and Upper Ordovician sequence from foreland basins within the larger Appalachian Basin. These studies were also successful in relating some parts of the stratigraphic sequence in the Sevier and Martinsburg foreland basins to specific tectonic causes, but aside from the somewhat similar sequences in the two basins, the specific tectonic mechanisms and the relationships between the two sequences remained unclear. With the latest developments in the understanding of Taconian tectonism (e.g., Vick et al., 1987; Drake et al., 1989) and refinements in lithospheric flexure models (e.g., Beaumont et al., 1988; Jamieson and Beaumont, 1988), it may now be possible to integrate apparent stratigraphic complexities into coherent, regional patterns that will elucidate more clearly the relationships between tectonism and Ordovician sedimentation in central and southern parts of the Appalachian Basin and on adjacent parts of the craton; such is the goal of this paper. In addition, the vertical and areal distribution of lithologies in such a setting may hold supplementary information about the nature and timing of the orogeny itself. Hence, the use of lithospheric flexure models to interpret Ordovician stratigraphy may not only help to confirm and augment interpretations gained by other means, but may also serve as a model for use in other orogens and accompanying foreland basins.

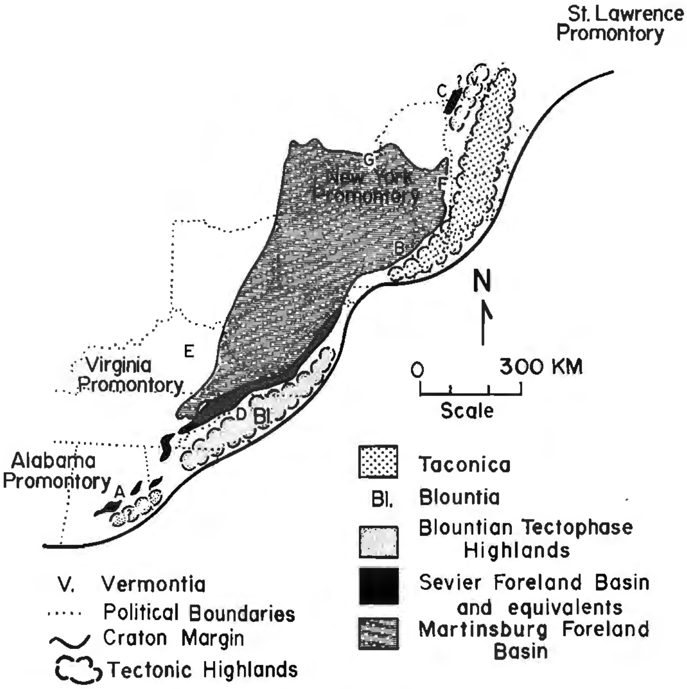

Figure 1. Study area in the central and southern Appalachians showing the positions of tectonic highlands and black shale foreland basins relative to continental promontories during the Middle and Late Ordovician. Note the northwestward and northeastward shift of the Martinsburg Basin relative to the Sevier Basin and equivalents (basin distribution from Dennison, 1986). Letters refer to mid- and end-points of sections illustrated in Figures 4, 5, and 6.

THE BASIC TECTONIC MODELS

Obviously, the advent of an orogeny along a cratonic margin influences sedimentation and the resulting stratigraphy not only by creating new sediment sources, but also by generating migrating foreland basins to receive that sediment. Possible mechanisms by which these basins form and migrate were not fully appreciated until the development of lithospheric flexure models (e.g., Price, 1973; Beaumont, 1981; Jacobi, 1981; Karner and Watts, 1983). These models basically suggest that subsurface and surface deformational loading in an orogen produces a downwarped flexural or retroarc foreland basin (sensu Dickinson, 1974) cratonward of the orogen and a peripheral bulge on the distal (cratonward) margin of the basin (Fig. 2) in response to regional isostatic compensation by the lithosphere. Although most of the loading is related to fold-thrust belt loading, a subordinate component of loading is produced by sediment loading (Beaumont, 1981; Tankard, 1986). As orogeny proceeds and the thrust loads continue to shift cratonward, or nearly normal to the trend of the orogenic belt, the foreland basin and peripheral bulge also migrate cratonward away from the load. Walker et al. (1983) have suggested a somewhat analogous basin-and-bulge migration in the Ordovician Sevier Basin, but their driving mechanisms were largely related to bimodal depositional rates and sediment-loading subsidence. If an orogeny is diachronous along its length, foreland basins and sediment sources will migrate parallel to the trend of the orogenic belt (Ettensohn, 1987).

A similar shifting of the peripheral bulge across distal parts of the foreland basin and the adjacent craton accompanies foreland basin migration (Quinlan and Beaumont, 1984). The result of bulge migration is widespread uplift of the foreland and formation of a regional unconformity (Fig. 2), especially if bulge uplift and migration represent the "initial docking of a substantial overthrust load" against a cratonic mass (Quinlan and Beaumont, 1984), although Dickinson's (1974, p. 22) more general mechanism of regional uplift as subduction is initially retarded by actual collision may be just as effective at generating such an unconformity.

Inasmuch as an orogeny probably occurs in one or more short pulses on the order of five million years or less in duration (Jamieson and Beaumont, 1988), pulses of tectonism and active deformational loading will cease, only to reoccur as later phases. Once active thrusting ceases, lithospheric relaxation related to a now static gravitational load — herein called "loading-type" relaxation (Fig. 3a) — causes the peripheral bulge to be uplifted and migrate toward the orogen, while the foreland basin becomes "deeper" and narrower (Fig. 3a). As the peripheral bulge migrates back toward the orogen, slight uplift and an accompanying erosional hiatus may develop in more cratonic parts of the sequence (Quinlan and Beaumont, 1984).

As tectonic highlands are lowered by erosion, another type of lithospheric relaxation — herein called "unloading-type" relaxation — subsequently occurs in response to erosional unloading. The response, however, is opposite (Fig. 3b) to that of loading-type relaxation in that major unloading results in rebound near the unloaded area and in what Beaumont et al. (1988) call an "antiperipheral bulge", which deepens and migrates toward the former load; a major unconformity may characterize the rebounded area (Beaumont et al., 1988).

SEDIMENTARY-STRATIGRAPHIC RESPONSES

The above tectonic scenario for foreland basins includes, in succession, four major processes: initial collision, abrupt subsidence due active deformational loading, loading-type relaxation following the cessation of deformation, and unloading-type relaxation due to rebound. If individual tectophases succeed each other rapidly enough, the final process may be shortened or completely eliminated, but in the typical foreland basin sequence each process can be expected to produce a virtually distinct stratigraphic and sedimentological response. The nature of such responses is the key to understanding basin history as well as the temporal and spatial relationships between orogeny and sedimentation. In the future, application of models like these may mean that just as much information about the nature of an orogeny can be gleaned from its accompanying foreland basin sequence as from the resulting structural, metamorphic, and regional tectonic relationships. The expected sequence of responses is briefly described below.

Unconformity development

Initial docking and subsequent peripheral bulge migration has the effect of buoying upward much of the foreland basin as well as most of the adjacent cratonic foreland, so that the

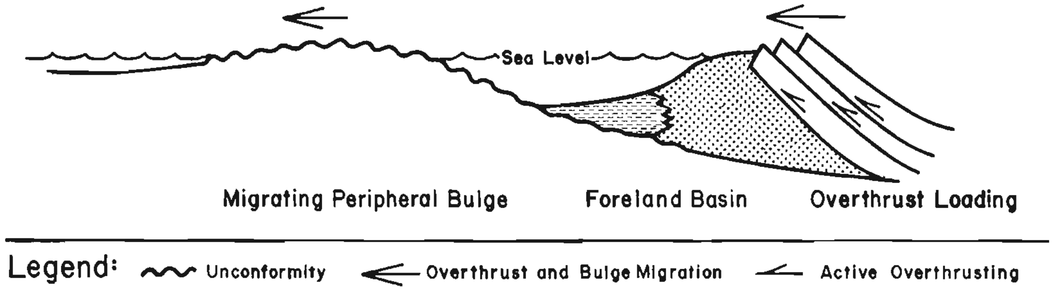

Figure 2. Schematic diagram showing development of a foreland basin and peripheral bulge due to deformational loading in the orogen (adapted from Quinlan and Beaumont, 1984).

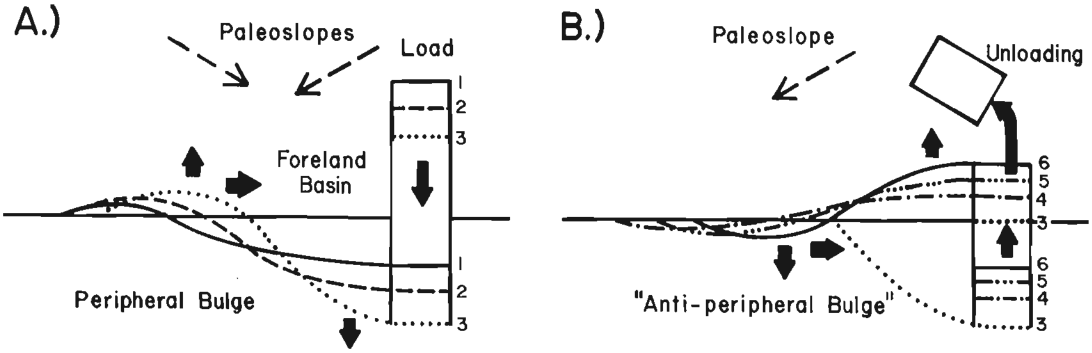

Figure 3. Two types of flexural response to lithospheric stress relaxation (redrawn from Beaumont et al., 1988). **a,** "loading-type" relaxation: response to a static gravitational load results in a deepening foreland basin and migration of peripheral bulge toward the load; **b,** "unloading-type" relaxation: response to erosional unloading results in rebound near the unloaded area and an "antiperipheral bulge" that deepens and migrates toward the former load.

response sequence begins with a major regional unconformity (Jacobi, 1981; Quinlan and Beaumont, 1984). However, in parts of the foreland basin just cratonward of the loci of most intense collision, unconformities should not be expected because subsidence due to residual deformational loading in these areas will generally be great enough to offset the effects of bulge uplift (Quinlan and Beaumont, 1984). For the Appalachian foreland basin during the Taconian (Hiscott et al., 1986; Lash, 1987a) and subsequent orogenies (Ettensohn, 1985a; Ettensohn and Chesnut, in press), these loci were generally the continental promontories (Thomas, 1977), which because of their protuberant nature apparently experienced a greater intensity of compression and deformation (Dewey and Burke, 1974; Dewey and Kidd, 1974). Because some loading would always have been present at these loci during an orogeny, continuous foreland basin subsidence would be expected to have occurred just cratonward of them. This is borne out by the fact that during the development of major North American cratonic unconformities (Early-Middle Ordovician, Early Devonian, and Early Pennsylvanian; see Sloss, 1963), parts of the Appalachian Basin just cratonward of the promontories exhibit gradational sequences (Ross et al., 1982; Patchen et al., 1985a, b) across intervals that are normally unconformable elsewhere in the basin.

Foreland basin subsidence

As active tectonism and deformational loading ensue, the foreland basin experiences rapid subsidence. Inasmuch as subsurface loading (buried, obducted blocks and flakes) is probably responsible for the initiation of foreland basin development and much of the early subsidence (Karner and Watts, 1983), during early phases of an orogeny no major source of externally derived sediment is available, and the basin becomes sediment starved. In fact, most major sediment redistribution occurs during subsequent periods of tectonic quiescence (Karner, 1987), and even then paleogeographic, paleoclimatic, eustatic, or structural mechanisms may act to enhance this sediment starvation (Ettensohn, 1985b, p. 71,72). In the absence of major clastic influx, organic matter and suspended clay and silt compose most of the sediment in these early basins, although on the cratonward margin of these basins away from the locus of major subsidence, a generally transgressive carbonate sequence may persist (Walker et al., 1983; Quinlan and Beaumont, 1984). Moreover, because of the rapid subsidence and slow sediment accumulation, the water columns in the basins quickly become stratified, especially if the basins are situated in tropical to subtropical regions where surface waters become so warm that the overturning of bottom waters becomes difficult (Ettensohn and Barron, 1981; Ettensohn, 1985b). As a result, this stage in the response sequence is characterized by dark, organic-rich shales, which accumulate in anaerobic to dysaerobic conditions. The faunas comprise largely planktic and nektic forms such as graptolites and cephalopods, although layers of broken and abraded, shallow water, benthic forms may also accumulate due to storm activity or mass-movement processes acting on adjacent shelf areas (Walker et al., 1983). Some of the classic, fissile, Paleozoic black shales such as those in parts of the Ordovician Athens, Blockhouse, Paperville, and Martinsburg shales, and in parts of the Devonian Ohio and Chattanooga shales, represent early deposition in rapidly subsiding foreland basins.

If the black shale stratigraphy of a foreland basin is well enough known to map the distribution of individual basinal units, these black shale units will typically track the progress of the orogeny. If an orogeny is largely contemporaneous along the length of the orogen, successive black shale basins will incrementally shift cratonward (Ettensohn, 1985a) in response to the cratonward migration of deformation. If the orogeny is diachronous along its length, successive basins will also shift parallel to the strike of the orogen in the direction of latest convergence (Ettensohn, 1987).

Though dark shale predominates during early foreland basin subsidence, on the distal or cratonward margin of the basin, carbonate deposition may persist for some time because of reduced rates of subsidence away from loading. Nonetheless, as deformation moves cratonward, even the distal margin will eventually subside and undergo inundation so that the overall sequence is transgressive.

Loading-type lithospheric relaxation

As active deformation and thrust migration cease, the lithosphere responds to the now static load by relaxing stress so that the basin deepens, while the peripheral bulge is uplifted and migrates toward the load (Fig. 3a). By this time, however, folding and thrusting have created an extensive surface load, which is eroded and transported into the foreland basin in the form of turbidites and debris flows, and this sediment load itself may be another important factor in basin subsidence (Walker et al., 1983) during this response. Despite continued basin subsidence, the influx of flysch-like sediment eventually exceeds subsidence to the point that sediments aggrade upward into dysaerobic and aerobic waters capable of supporting benthic faunas able to withstand clastic influx. Overall, coarser marine clastics predominate in the basin during this relaxation response, and two paleoslope directions arise (Fig. 3a).

Though marine clastic sediments dominate the centre and proximal margins of the foreland basin, carbonate deposition may continue on the distal margin (Walker et al., 1983). In contrast to the transgressive carbonate sequence that characterized the distal margin during the previous process-response, uplift and migration of the bulge during relaxation will give rise to a succeeding regressive carbonate sequence. If the amount of bulge uplift is large, the regressive sequence will be truncated by an unconformity.

Unloading-type lithospheric relaxation

As the foreland basin fills with clastic sediments and the orogenic highlands undergo substantial lowering by erosion, a brief period of elevational equilibrium ensues between the filled basin and eroded highlands. This is short-lived, however, for as isostatic adjustment continues, the orogen begins to rebound upward in response to the lost load, and a compensating "antiperipheral bulge" forms near the former distal margin, deepens and migrates toward the rebounding area (Fig. 3b) (Beaumont et al., 1988). Because the rebounding load now consists of formerly leveled highland and previously deposited foreland basin sediments, sediments eroded from the rebounded area will be mainly fine grained, consisting dominantly of siltstone, silty shale, shale, mudstone, or shaly carbonate, except in the most proximal areas where conglomerate may occur. Moreover, because the process begins from a state of approximate elevational equilibrium at or near sea level, a single, cratonward-dipping paleoslope will be established and the sediments will be deposited in mainly marginal marine or terrestrial environments. Hence, redbeds of molasse aspect will be common, and some of the classic red clastic wedges like the "Queenston Delta" probably record deposition as a result of unloading-type relaxation.

In the area of the former distal margin, on the other hand, enough deepening may occur due to "antiperipheral bulge" development so that a short-lived transgressive sequence of shallow open marine carbonate or shale may develop. Eventually, however, regression resumes as the wedge of marginal marine clastics progrades cratonward. Again, if uplift due to rebound is large, the resulting sequence may be truncated by an unconformity.

ORDOVICIAN STRATIGRAPHIC RESPONSES TO THE TACONIAN OROGENY

Tectonic framework

The Taconian Orogeny is now widely regarded as having resulted from a continental margin-island arc collison associated with an east-dipping subduction zone during the Middle and Late Ordovician (e.g., Malpas and Stevens, 1977; Hiscott, 1978; Rowley and Kidd, 1981; Rodgers, 1987). Although actual docking in the central and southern Appalachians probably did not occur until the Middle and Late Ordovician, evidence for Cambrian and Early Ordovician tectonic activity is recognized around the St. Lawrence promontory in New England and in some of the Canadian Atlantic provinces (Bird and Dewey, 1970; Rodgers, 1971; Hiscott et al., 1983; Hiscott, 1984; Boone and Boudette, 1989). Farther south, however, the subduction that would bring about later collision may have begun far east of the continental shelf as early as the Cambrian and continued throughout the Early Ordovician (e.g., Pavlides, 1981, 1989). These Cambrian and Early Ordovician phases of tectonism are sometimes called the Penobscottian Orogeny or the Penobscot phase of the Taconian Orogeny (e.g., see Neuman, 1967; Rodgers, 1967; Hatcher, 1987; Skehan, 1988; Rast, 1989).

The main part of the orogeny, however, is generally considered to have occurred from the Middle Ordovician to the early Late Ordovician (Llandeilian to Caradocian; Rast, 1989) beginning with the subduction of the Laurentian continental margin below island-arc complexes. These island-arc complexes may be represented by a series of terranes in the south — the Carolina terrane is probably one (Vick et al., 1987) — and by the Dunnage and Gander terranes in the north and central parts of the Appalachians (Williams and Hatcher, 1983). The occurrence of altered tuffs in upper Whiterockian and Mohawkian basinal and cratonic rocks from Tennessee and Kentucky (McFarlan, 1943; Conkin and Conkin, 1983; Shanmugam and Walker, 1983; Dennison, 1986) strongly supports the presence of these island arcs in the south. Continued eastward subduction below these arc complexes eventually resulted in the obduction of continental slope/rise sediments, continental and oceanic basement, as well as accretionary prism materials onto the Laurentian continental shelf; it was the accumulation of these materials on the former shelf in the form of folds, nappes, and thrusts that provided the deformational load necessary to initiate cratonward foreland basin formation.

Moreover, the temporal and spatial distribution of Ordovician foreland basin sequences in the central and southern Appalachians (Figs. 1, 4) suggests that the Taconian Orogeny occurred here mainly in two diachronous Middle and Late Ordovician tectophases related to enhanced convergence near the Virginia and New York promontories, respectively. Although major orogeny may have been concentrated at these

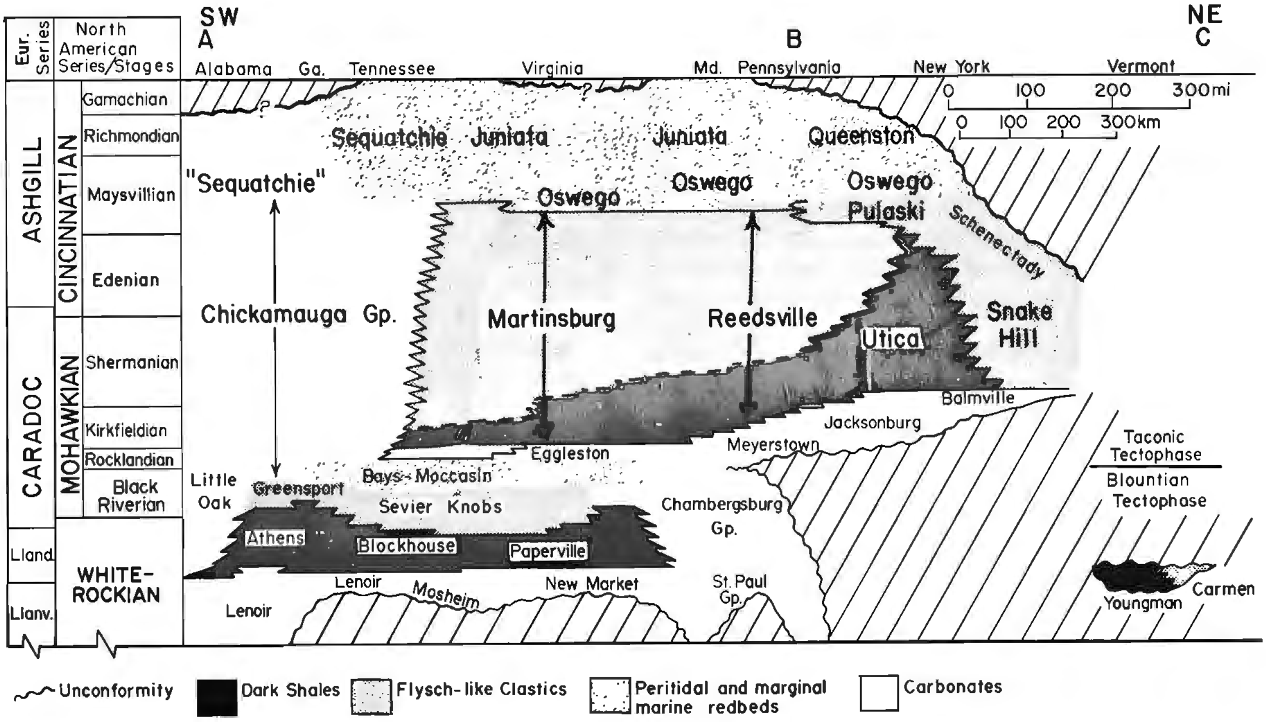

Figure 4. Highly schematic southwest-northeast section parallel to the strike of the Appalachian Basin showing the repetition and migration of basin sequences reflecting the different loci and timing of the Blountian and Taconic tectophases. The Martinsburg and Reedsville formations contain both flysch and dark shale. No vertical scale intended.

promontories during respective tectophases, both stratigraphic and structural data indicate that coeval tectonism occurred at other promontories as well (Hiscott et al., 1986).

The Ordovician chronostratigraphic framework used throughout this study is that of Ross et al. (1982).

Initial response to collision

Although subduction below the island arcs had probably been active for some time, perhaps since the Cambrian (Hatcher, 1987), not until the Early-Middle Ordovician transition is there any evidence in the study area for collision and the subsequent docking of a deformational load against the Laurentian platform margin. This evidence is in the form of a major unconformity throughout most of the Appalachian Basin (Fig. 4) and on adjacent parts of the stable craton called the post-Sauk Unconformity (Sloss, 1963), the Owl Creek Discontinuity (Wheeler, 1963), or unconformity C by Rodgers (1971). The unconformity probably represents the abrupt braking of subduction as the outer continental edge was initially drawn down into the subduction zone; in response, adjacent parts of the continent were apparently buckled up and eroded. Moreover, as the accretionary prism with newly incorporated thrusts of continental rise sediments moved onto the continental margin, the resultant deformational loading generated a migrating peripheral bulge that extended uplift and erosion cratonward (Jacobi, 1981; Quinlan and Beaumont, 1984). In fact, the unconformity becomes more pronounced to the west as erosional relief approaches 100 m in Virginia and Tennessee (Dennison, 1986). The resulting uplift and erosion occurred during the latest Canadian (latest Arenigian) and throughout much of the Whiterockian (Llanvirnian and Llandeilian) and was accompanied locally by the development of normal faults (Thompson, 1967; Zen, 1967; Williams and Stevens, 1974; Bradley and Kusky, 1986). Although this period of unconformity development has been called the Tinmouth phase of the Taconian Orogeny (e.g., Thompson, 1967; Rodgers, 1970, 1971), lithospheric flexure models suggest that it is really only the initial response in a stage of the Blountian tectophase.

Response to the Blountian tectophase

The major part of the Blountian tectophase was apparently the result of collision on and near the Virginia promontory (Fig. 1), and its effects are best illustrated cratonward of this promontory (Figs. 4, 5), although other promontories (e.g., eastward of the Youngman Basin; Figs. 1, 4) were also involved (Hiscott et al., 1986). By the middle Whiterockian (Llanvirnian), deformational loading on the promontory was great enough to have caused abrupt and drastic subsidence of the foreland basin in eastern Tennessee and western Virginia following post-Knox exposure and unconformity development (Walker, 1977; Shanmugam and Walker, 1978; Benedict and Walker 1978). However, farther south, cratonward of the Alabama promontory in Georgia and Alabama, foreland basin subsidence preceded or was simultaneous with unconformity development farther north (Roberson et al., 1988) so that basin deepening began there in the latest Canadian and earliest Whiterockian (latest Arenigian). Whether this was a separate event or just the initiation of a collision

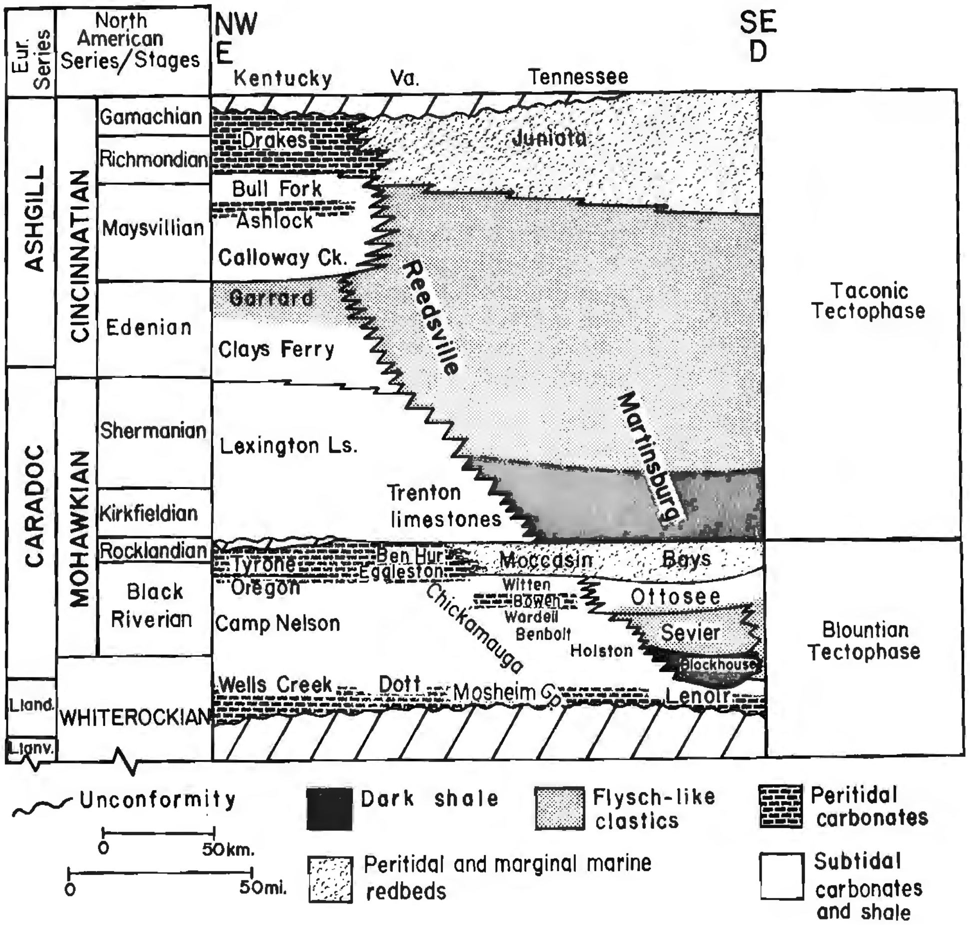

Figure 5. Schematic northwest-southeast section perpendicular to the strike of the southern Appalachian Basin, showing the repetition and relative cratonward migration of basinal sequences during the two tectophases. The Bowen-Witten and upper Ashlock-Bull Fork sequences may represent the passage of peripheral and "antiperipheral" bulges on the distal margins of the foreland basins. No vertical scale intended.

event that migrated northward toward the Virginia promontory with time is uncertain, but it is clear that some deformational loading was already present in the area as early as the latest Canadian.

In both areas, however, basin subsidence due to loading is indicated by the development of transgressive sequences beginning with peritidal units (New Market, Lenoir, and Mosheim formations) atop the unconformity (Figs. 4, 5) that pass successively upward into shallow, open marine, slope and basinal environments. These basinal environments are represented by the dark shales of the Athens, Blockhouse, and Paperville formations (Benedict, 1977; Shanmugam, 1977; Walker, 1977; Benedict and Walker, 1978; Shanmugam and Walker, 1978, 1980, 1983; Read, 1980) (Fig. 4), and developed first in Alabama and Georgia based on the ages of the Athens and equivalent shales (Ross et al., 1982; Patchen et al., 1985a). The distribution of the shale indicates that the basins migrated both northwestward, perpendicular to the orogen, and northeastward parallel to the orogen, as predicted by lithospheric flexure models for a northeastward-shifting orogeny. These patterns of basin migration are especially well illustrated in Read's (1980, Fig. 19) maps showing the distribution of Middle Ordovician environments. Moreover, the dearth of coarser clastic detritus in these organic-rich muds suggests that subaerial highlands had not yet formed to the east, and that most of the deformational loading at this time (latest Whiterockian-earliest Black Riverian; late Llandeilan-early Caradocian) was in the subsurface.

In contrast to my suggestion above that most of the rapid subsidence was related to loading in the orogen, K.R. Walker and his coworkers (e.g., Benedict and Walker, 1978; Shanmugam and Walker, 1978, 1980; Shanmugam and Lash, 1982; Walker, pers. comm., 1989) have suggested that this phase of subsidence was largely related to tectonic downwarping probably accomplished by down-to-basin faulting. However, no evidence of such faulting has been reported, and the origin of such faults is unclear, unless they could have formed during the previously described process of continental braking.

By the middle Black Riverian, tectonic highlands indicative of surface loading, called Blountia (Fig. 1), had developed to the east and were shedding coarser clastic debris, now preserved as conglomerate, graywacke, feldspathic sandstone, siltstone, and shale, westward into the subsiding foreland basin in eastern Tennessee and western Virginia. These rocks reflect debris flows, turbidites, and reworking by contour currents, and are represented in the Sevier, Tellico, and Knobs formations (Shanmugam and Walker, 1978, 1980, 1983; Read, 1980) (Figs. 4, 5). Locally in Alabama, similar lithologies in the Greensport Formation (Patchen et al., 1985b; Dennison, 1986) may reflect the filling of a foreland basin by sediments derived from a possible southern extension of Blountia (Figs. 1, 4).

Eventually, however, subsidence in eastern Tennessee and western Virginia declined at the same time as basin infilling and erosional reduction of the tectonic highlands to the east.

These events are reflected in the Sevier Basin by deposition of the shallow water shale, mudstone, and carbonate mudstone of the Ottosee Formation (Cantrell and Walker, 1985), which eventually spilled out of the basin northwestward onto basin margin, carbonate shelf facies of the middle Chickamauga Group (Walker, 1977; Walker et al., 1983; Ruppel and Walker, 1984) (Fig. 5).

If in fact the Sevier-Tellico-Ottosee clastics (Fig. 5) reflect the infilling of a basin that subsided in response to loading-type relaxation in the orogen, evidence for southeastward bulge migration toward the load should be expected. Probable evidence for this is present as a short-lived regressive sequence (upper Benbolt, Wardell, and Bowen formations; see Ruppel and Walker, 1984, and Wedekind, 1985) in the midst of the overall transgressive, basin margin, middle Chickamauga Group (Fig. 5). This regressive sequence from the upper Benbolt to the Bowen represents relatively rapid change from subtidal to peritidal facies; although the exact origin of these changes must remain uncertain, the timing and location of the abrupt shallowing are correct for the passage of a peripheral bulge.

The peritidal Bowen Formation in turn grades upward into deeper, subtidal carbonates of the Witten Formation. According to reconstructions by Walker (1977), and Ruppel and Walker (1984), the Witten grades eastward into the upper Ottosee, which by the late Black Riverian (Caradocian) represented shallow water, mixed clastic-carbonate deposition on a gentle cratonward-dipping paleoslope (Walker, 1980; Walker et al., 1983; Cantrell and Walker, 1985). The development of a single cratonward-dipping paleoslope during late Ottosee deposition signals the inception of unloading-type relaxation in the orogen. The time of this change in paleoslope direction may be represented by a widespread horizon of oolite shoals and reefs in the Ottosee (see Walker, 1980, and Cantrell and Walker, 1985). The subsequent deepening of the shelf reflected in the Witten (Fig. 5) and in coeval parts of the Ottosee (Ruppel and Walker 1977, 1984; Walker et al., 1983; Wedekind, 1985) probably reflects development of an "antiperipheral bulge", although Walker (1980) and Ruppel and Walker (1984) interpreted this to represent basin rejuvenation. The culmination of this phase of unloading, as well as of the Blountian tectophase, is represented by the westward progradation of the Bays-Moccasin-Eggleston tidal flat complex across much of the southern Appalachians (Figs. 4, 5). These units are composed of red and grey argillaceous limestone, red shale, and mudstone, which represent deposition in terrestrial, peritidal and very shallow marine environments (Read, 1980; Simonson, 1985).

While the Sevier foreland basin formed and filled with flysch-like sediment in response to the rise and erosion of Blountia in the southern Appalachians, areas roughly equivalent in position along strike in the central Appalachians experienced generally transgressive, shallow shelf deposition represented by carbonates equivalent to the Chazy Group in New York, as well as by carbonates of the Black River Group and its equivalents. Although a foreland basin would form here later in the Middle Ordovician (Rocklandian; Caradocian), no stratigraphic evidence supports the presence of a basin coeval with the Sevier in the central Appalachians. In northwestern Vermont and southeastern Quebec, however, an unconformity-bound basinal sequence represented by the dark Youngman Shale and Carman Quartzite (Kay, 1958) (Figs. 1, 4), as well as their nearby Chazy Group equivalents (Kay, 1958; Fisher, 1977, Fig. 3C), apparently represent a localized foreland basin sequence developed in response to the coeval uplift of tectonic uplands commonly called Vermontia (Fig. 1). The former extent of the basin is unknown because the units are cut out by an unconformity (Fig. 4) southwest of the Champlain Valley (Rodgers, 1971; Fisher, 1977).

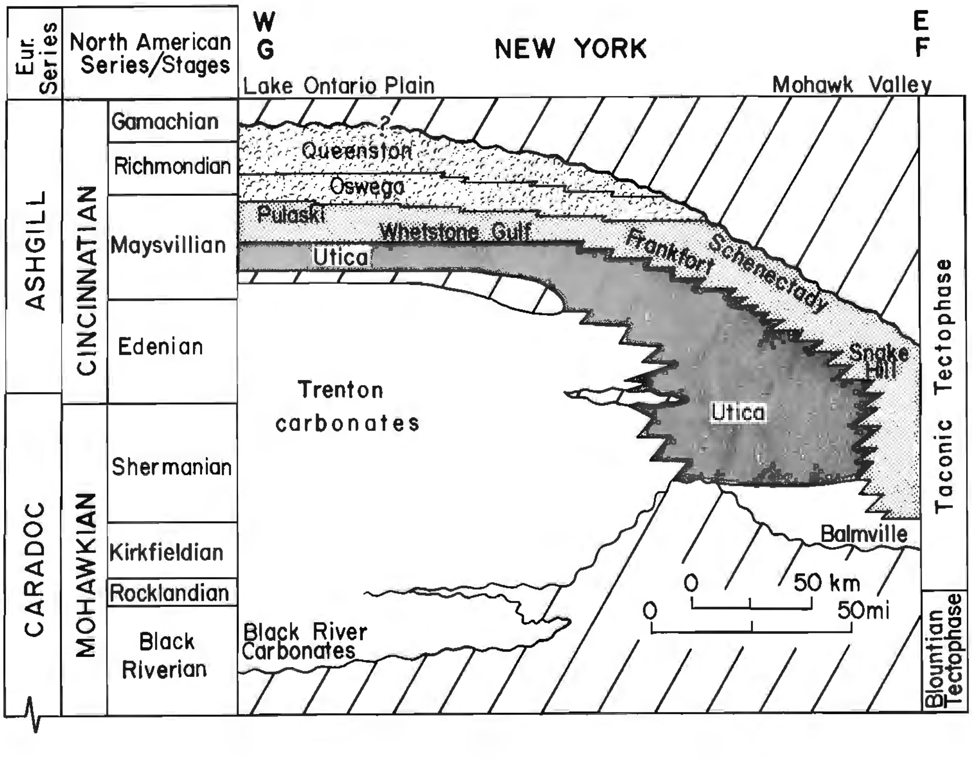

Figure 6. Schematic west-east section in New York showing the northward extension of the Martinsburg Basin reflected in the Utica Shale and the Utica's westward migration with time (after Fisher, 1977). No vertical scale intended; legend same as that in Figure 4.

While the Blountian tectophase proceeded in the south, and shallow water, carbonate platform deposition continued in the central Appalachians, eastward subduction of oçeanic and continental crust seaward of the New York promontory progressed to the point that the accretionary wedge at the advancing margin of the island arc began to shed sediments into a peripheral basin (sensu Dickinson, 1974) that impinged on the central Appalachian platform margin (e.g., Rowley and Kidd, 1981). The dark shale, siltstone, graywacke, and chert deposited in the basin are included in the allochthonous Normanskill Group (Fisher, 1977) and its equivalents, but are not part of the foreland-basin sequence.

Response to the Taconic tectophase

The Taconic tectophase (Kay, 1937) began in the late Middle Ordovician (Rocklandian; Caradocian) and apparently reflects the obduction of the aforementioned peripheral basin sediments and accretionary wedge material onto the platform margin, forming a tectonic highland called Taconica (Fig. 1). The obduction of these materials onto the platform generated the load necessary to form the Martinsburg (Utica) foreland basin (Figs. 4-6). Although what is called the Taconic tectophase here may be subdivided elsewhere into shorter named phases (e.g., Rodgers, 1971), the stratigraphic record in the central Appalachian foreland basin suggests that it is a single event.

The locus of the Taconic tectophase was largely centred near the New York promontory in the central Appalachians (e.g., Rodgers, 1971), but foreland basin subsidence, and hence deformational loading, apparently began earliest in northeastern Tennessee cratonward of the Virginia promontory, suggesting a substantial overlap in the areal extent of the tectophases (Fig. 4). In northeastern Tennessee, peritidal Bays-Moccasin mudstone from the last stage of the Blountian tectophase is sharply overlain by a thin sequence of transgressive carbonate, followed by the deep water black shale of the Martinsburg Formation (Walker and Diehl, 1985) (Figs. 4, 5), indicating abrupt Rocklandian (late Caradocian) subsidence. No unconformity, however, is apparent between the Bays-Moccasin and Martinsburg, suggesting that sufficient residual loading from the Blountian tectophase remained in the area of the Virginia promontory to offset the effects of any new bulge migration.

The New York promontory, in contrast, seems to have been very little affected by the Blountian tectophase, and residual loading was absent. Hence, while subsidence occurred to the south, uplift and erosion accompanying Taconic tectophase, peripheral bulge migration resulted in an unconformity in parts of the foreland basin just cratonward of the New York promontory in New Jersey, eastern Pennsylvania, and southern New York. In these areas, transgressive Rocklandian and Kirkfieldian (middle Caradocian) carbonate of the Jacksonburg, Myerstown, and Balmville formations unconformably overlie Canadian to Black Riverian (upper Arenigian to lower Caradocian) shallow water, platform carbonates of the Beekmantown Group and Chambersburg Limestone (Fig. 4) and grade upward into dark shale of the Martinsburg or its equivalents (Fisher, 1977; Patchin et al., 1985a; Lash, 1987a). A series of unconformities below the dark Utica Shale in central New York (Fig. 6) becomes progressively younger to the west (Fisher, 1977, Pl. 2, 4) and apparently represents the same bulge migration phenomenon. This stratigraphic response — unconformity, transgressive carbonate, and black shale — is analogous to that observed in the Sevier Basin during the inception of early Middle Ordovician Blountian tectophase (Shanmugam and Lash, 1982). Hence, the Martinsburg Basin (Martinsburg, Reedsville, and Utica shale) (Figs. 1, 4) reflects new basin subsidence accompanying the inception of the late Middle Ordovician Taconic tectophase. The much larger areal extent of the Martinsburg Basin (Fig. 1) may reflect the greater intensity of the Taconic tectophase combined with a possible middle Mohawkian to early Cincinnatian (middle Caradocian-early Ashgillian) rise in sea level (Leggett, 1978; McKerrow; 1979).

Inasmuch as deformational loading will migrate cratonward in time (e.g., Beaumont, 1981), the more westward position of the Martinsburg Basin relative to the Sevier Basin (Figs. 1, 5) reflects the extent of this shift between tectophases. The same pattern is also apparent within the Martinsburg Shale and its equivalents, for although the oldest parts of the Martinsburg in the eastern parts of its basin may be as old as Rocklandian, Kirkfieldian or Shermanian (middle to late Caradocian), at its maximum westward extent in eastern Kentucky, eastern Ohio, and West Virginia (Figs. 1, 5), it is Maysvillian (Ashgillian) in age (Patchen et al., 1985a,b). Similarly, the partially equivalent Utica Shale is Shermanian in the east, but in western New York, the unit is wholly Maysvillian in age (Fisher, 1977) (Fig. 6).

In an analogous fashion, the lower Martinsburg and its equivalents become younger northward along strike in the foreland basin (Fig. 4). In northeastern Tennessee, Martinsburg basinal black shale deposition began in the Rocklandian, but similar basinal environments only reached New York during deposition of the Utica ("Canajoharie") Shale in the Shermanian (e.g., Ross et al., 1982). This kind of shift in the inception of basinal deposition along strike probably reflects diachroneity in the collision of separate terranes along the orogen.

The cessation of active orogeny and the advent of loading-type relaxation is represented by shallowing during deposition of middle parts of the Martinsburg to the south (Walker and Diehl, 1985), by the influx of sandstone and siltstone turbidites and carbonate tempestites in middle parts of the Martinsburg of the central Appalachians (Shanmugam and Lash, 1982; Lash, 1987a,b), and by the sandstone, siltstone, graywacke, and shale of the Frankfort, Schenectady, Whetstone Gulf, and Pulaski formations in New York (Fisher, 1977; Zerrahn, 1978) from the Shermanian to the Maysvillian (latest Caradocian to Ashgillian). The concomitant uplift and eastward migration of the peripheral bulge on the western margin of the Martinsburg foreland basin (Fig. 3a) is difficult to document, because most of the rocks likely to reveal this event are poorly known in the subsurface of western West Virginia, eastern Ohio, and eastern Kentucky where they are commonly included in undifferentiated shale and limestone sequences (e.g., Patchen et al., 1985a). Nonetheless, in east-central Kentucky and south-central Ohio (Patchen et al., 1985b), abrupt shallowing from subtidal to peritidal conditions reflected in lower and middle parts of the Ashlock Formation (Weir et al., 1984) (Fig. 5) occurred at the correct time and place to represent the passage of a peripheral bulge.

Upper parts of the Ashlock and the Bull Fork Formation in the same area (Weir et al., 1984) reflect an equally sudden return to subtidal conditions and may indicate the subsequent development of an ''antiperipheral bulge'' accompanying the advent of unloading-type relaxation. The timing (latest Maysvillian; Ashgillian) is correct for such a shift, and these subtidal units are overlain by the Richmondian (Ashgillian) Drakes Formation, which is correlative with the Queenston, Juniata, Bald Eagle, Spitzbergen, and Sequatchie formations farther east in the foreland basin. The above foreland basin units are largely red in colour and consist of finer grained clastic sediments and argillaceous carbonates representing the cratonward progradation of fluviodeltaic, peritidal and very shallow marine facies on a unidirectional paleoslope developed in response to rebound accompanying unloading-type relaxation in the orogen. The Bald Eagle and Spitzbergen, however, are largely conglomerates (Thompson, 1970; Lash, 1987a), probably reflecting proximal source areas and the reactivation of basement structures during rebound (see Dennison, 1986, Fig. 26). In areas proximal to the New York promontory, moreover, a major unconformity overlies these redbed units. Although some workers have related this unconformity to thermal doming (Chapple, 1973; Coleman-Sadd, 1982), nothing more than unloading-type rebound compounded by the effects of a glacioeustatic drop in sea level near the Ordovician-Silurian boundary (Dennison, 1976; Hambrey, 1985) is probably necessary to explain it.

CONCLUSIONS

The Middle and Late Ordovician Taconian Orogeny resulted in a complete reorganization of stratigraphic patterns on the former eastern margin of Laurentia. Despite the complex nature of the resulting facies and of the nomenclature superimposed upon them, generalized lithostratigraphic patterns have emerged that confirm the predicted results of recently developed lithospheric flexure models. Consequently, these models have a high predictive value in interpreting the stratigraphic and sedimentological responses to orogenic events like the Taconian Orogeny. The generalized process-response sequence includes four stages: 1) collisional braking — development of a regional unconformity; 2) active deformational loading — thin transgressive carbonate followed by black shale in a rapidly subsiding foreland basin; 3) loading-type relaxation — flysch-like clastics infilling the foreland basin; and 4) unloading-type relaxation — cratonward progradation of a marginal marine redbed sequence. The stratigraphic response in each stage will generally track the progress of orogeny, but the black shale stage, because of its distinct lithology and ease of mapping, is most easily traced. The two Ordovician black shale sequences encountered in this study clearly reflect the initiation, migration, and foci of two Taconian tectophases in the southern and central Appalachians.

The process-response models discussed herein are mostly restricted to the interpretation of foreland basins, but the overall transgressive-regressive scenario inherent in the stratigraphic responses should be equally apparent in platform and ramp carbonates on adjacent parts of the stable craton. The migration of the peripheral bulge across these stable areas may also be apparent in the form of local unconformities or abrupt environmental changes. Although the stratigraphic and environmental fluctuations reflected in the process-response models for the Ordovician have been largely attributed to eustatic variations, their proximity to and contemporaneity with the Taconian Orogeny suggest that a large component of such variations must be related to concurrent tectonism.

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Taconian sedimentary basins of the Appalachians

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Abstract

Two categories of Taconian sedimentary basins can be recognized in northern Virginia and eastern West Virginia. Deep basins were topographic lows during deposition, and are recognized primarily on the basis of deep siliciclastic lithofacies. Shallow basins had little, if any, topographic expression during deposition, and are recognized primarily on the basis of isopach trends in strata that do not display very significant lithofacies changes. By comparison, similar Ordovician basins are recognized elsewhere in the Appalachians.

Deep basins were predominant during the Middle Ordovician and today are situated along the eastern margin of the Valley and Ridge Province of the Central and Southern Appalachians, and along the western margin of the Northern and Maritime Appalachians. The oldest deep basins, of early Middle Ordovician age, are found in Newfoundland. Deep basins of medial Middle Ordovician age are restricted to the Southern Appalachians. Late Middle Ordovician deep basins occur from the Central Appalachians to Vermont. Apparently, these deep basins were formed by thrust-loading of the eastern margin of North America.

Middle Ordovician shallow basins occurred in eastern Tennessee and southwestern Virginia. During Late Ordovician time, shallow basins occurred from southwestern Virginia to southern Pennsylvania. The origin of shallow basins is thought to be more directly related to subsidence due to sediment loading.

Subsequent to their initial development, many of the basins were structurally modified to become major synclinoria.

Résumé

On reconnaît deux catégories de bassins sédimentaires taconiques dans le nord de la Virginie et dans l'est de la Virginie occidentale. Les bassins profonds étaient des dépressions au cours de la sédimentation, et ils se reconnaissent principalement à leurs lithofaciès silicoclastiques profonds. Les bassins peu profonds avaient peu ou point d'expression topographique durant la sédimentation, et ils se reconnaissent principalement aux tendances des isopaques dans les strates où il n'y a aucun changement important de lithofaciès. Des bassins ordoviciens semblables se rencontrent ailleurs dans les Appalaches.

Les bassins profonds ont prédominé à l'Ordovicien moyen; de nos jours, on en trouve le long de la marge est de la province de Valley and Ridge dans les Appalaches du centre et du Sud, et le long de la marge ouest dans les Appalaches du Nord et des maritimes. Les bassins profonds les plus anciens, qui remontent au début de l'Ordovicien moyen, se rencontrent à Terre-Neuve. Des bassins profonds du milieu de l'Ordovicien moyen se trouvent uniquement dans les Appalaches du Sud. Des bassins profonds de la fin de l'Ordovicien moyen se rencontrent à partir des Appalaches centrales jusqu'au Vermont. Ils se sont vraisemblablement formés par suite de la poussée d'une charge sur la marge est de l'Amérique du Nord.

Des bassins peu profonds se trouvaient dans l'est du Tennessee et le sud-ouest de la Virginie à l'Ordovicien moyen, et à partir du sud-ouest de la Virginie jusque dans le sud de la Pennsylvanie à l'Ordovicien supérieur. Leur origine semble s'associer plus directement à la subsidence qu'à la charge sédimentaire.

Plusieurs des bassins ont été transformés structuralement en synclinoriums importants.

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INTRODUCTION

Taconian deformation, in the form of cleavage development, faulting, folding, and the development of unconformities has been recognized and described by numerous workers (see summary by Rodgers, 1971). The purpose of this paper is to emphasize another type of Taconian deformation, the development of sedimentary basins. This style of deformation is not recognizable in the form of structural features. Sedimentary basins may be recognized by stratigraphic variations that result from sediment accumulation in different parts of these basins.

The recognition of deep Taconian topographic basins is not new. These are flysch basins or foreland basins that have been described by previous workers, and this paper presents a summary of their interpretations.

This paper also presents evidence for what will be referred to as shallow basins. These were structural basins that had very little topographic expression during the Ordovician.

The basins discussed in this paper developed on continental crust. Excluded from the discussion are oceanic basins that developed during the opening of Iapetus, long before the Taconic event.

NORTHERN VIRGINIA TACONIAN CLASTIC SEQUENCE

The recognition of deep and shallow sedimentary basins in northern Virginia is based upon analysis of the Middle to Upper Ordovician Taconian clastic sequence. Taconian facies are markedly different on either side of the North Mountain front (Fig. 1). The following discussion presents a stratigraphic summary of each facies, as exposed in two stratigraphic sections located on the eastern and western sides of the Valley and Ridge Province in the northern Virginia area (Fig. 1).

Eastern facies

The eastern facies of the Taconian (Queenston) clastic sequence is well exposed in the Massanutten Synclinorium near Front Royal, Virginia. A stratigraphic summary of this eastern section is provided in Figure 2. A more detailed description of this section is provided by Diecchio (1985a, 1986) and by Fichter and Diecchio (1986).

The top of the Cambrian-Ordovician carbonate sequence is represented here by the Lincolnshire and Edinburg formations. The overlying Oranda and Martinsburg formations represent the Taconian clastic sequence. These clastic strata contain black shale of probable distal turbidite facies (lower Martinsburg), which are overlain by well developed, more proximal turbidites of the middle Martinsburg (Fig. 2). As emphasized by McBride (1962), these strata were probably deposited in a deep (flysch) basin. The presence of this postulated deep basin, in an area that shortly before was part of a shallow carbonate platform, suggests downwarping between deposition of the Edinburg and deposition of the Martinsburg. The transition from carbonate to clastic facies therefore coincides with a deepening event that represents the collapse of the carbonate platform. The base of the clastic sequence here is upper Blackriveran to lower Rocklandian (Ross et al., 1982). The uppermost Ordovician (Richmondian) is probably missing in the eastern section.

Western facies

The western facies of the Taconian clastic sequence is well exposed in the Wills Mountain Anticline in Pendleton County, West Virginia. A stratigraphic summary of this western section is provided in Figure 2. A more detailed description of this section is provided by Diecchio (1985a, 1986).

The limestones of the Trenton Group (Nealmont and Dolly Ridge formations) make up the upper part of the Cambrian-Ordovician carbonate sequence in this area. The Taconian clastic sequence (Reedsville, Oswego, Juniata formations) overlies the "Trenton" limestones. The clastic sequence here is upper Middle Ordovician (Shermanian, or upper Trentonian) and Upper Ordovician (Diecchio et al., 1985). The Dolly Ridge and Reedsville formations display the gradual transition from carbonate to clastic deposition in a shallow shelf environment. Throughout the full vertical extent of the Taconian clastic succession at this locality, there is no evidence that would indicate that water depth was ever much deeper.

One significant difference between the eastern and western sections is that the base of the clastic sequence is older in the eastern section than in the western section. This illustrates the progressive westward progradation of the easterly derived clastic wedge.

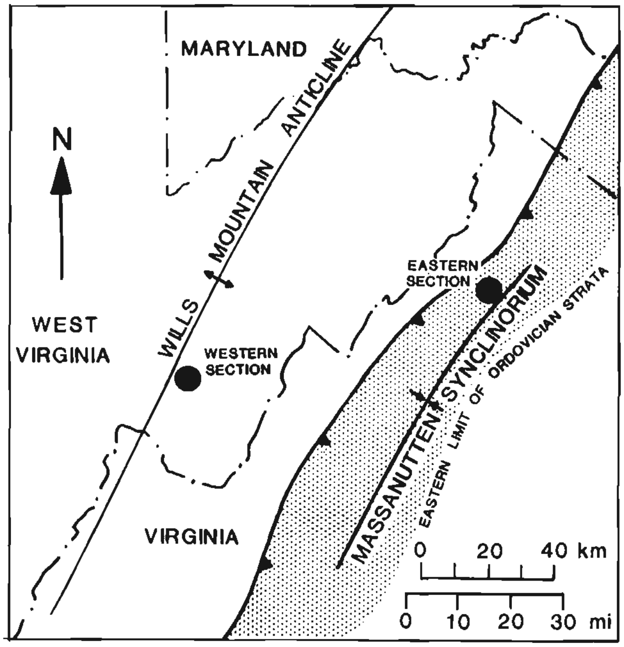

Figure 1. Location map of northern Virginia and eastern West Virginia showing eastern (stippled pattern) and western (no pattern) Taconian facies belts.

The eastern and western sections also exhibit a significant difference in their clastic depositional facies. In the eastern section the onset of Taconian clastic deposition is associated with a deepening event. Evidence of such deepening is not present in the shelf carbonate and clastic facies of the western section.

Sedimentary basins

As illustrated in the cross-section in Figure 3, the Taconian clastic sequence (Martinsburg/Reedsville, Oswego, Juniata formations) is thicker on either side of a positive area in the vicinity of the Wills Mountain Anticline. The actual position of this positive area cannot be determined with precision, but it defines two separate basins that will be referred to as the eastern and western basins.

Eastern basin

An eastern sedimentary basin existed to the east of the Wills Mountain Anticline (Figs. 1, 3). The eastern section is representative of strata in the eastern basin. Primary evidence for the existence of the eastern basin is the deep water clastic strata of the Martinsburg Formation in the eastern facies belt. Secondary evidence is the stratigraphic thickening of the Martinsburg Formation in this area (Fig. 3). The precise geometry of the deep eastern basin is not able to be determined with precision, due to lack of preserved information. However, Woodward (1951) and Chen (1977) provided isopach data that suggested the deep basin was generally coincident with the eastern facies belt (Fig. 1), with a northeasterly trending axis, subparallel to and roughly coincident with the axis of the Massanutten Synclinorium. A more regional expression of this basin is illustrated in Figure 4 as the extent of dark shale in the Massanutten Synclinorium.

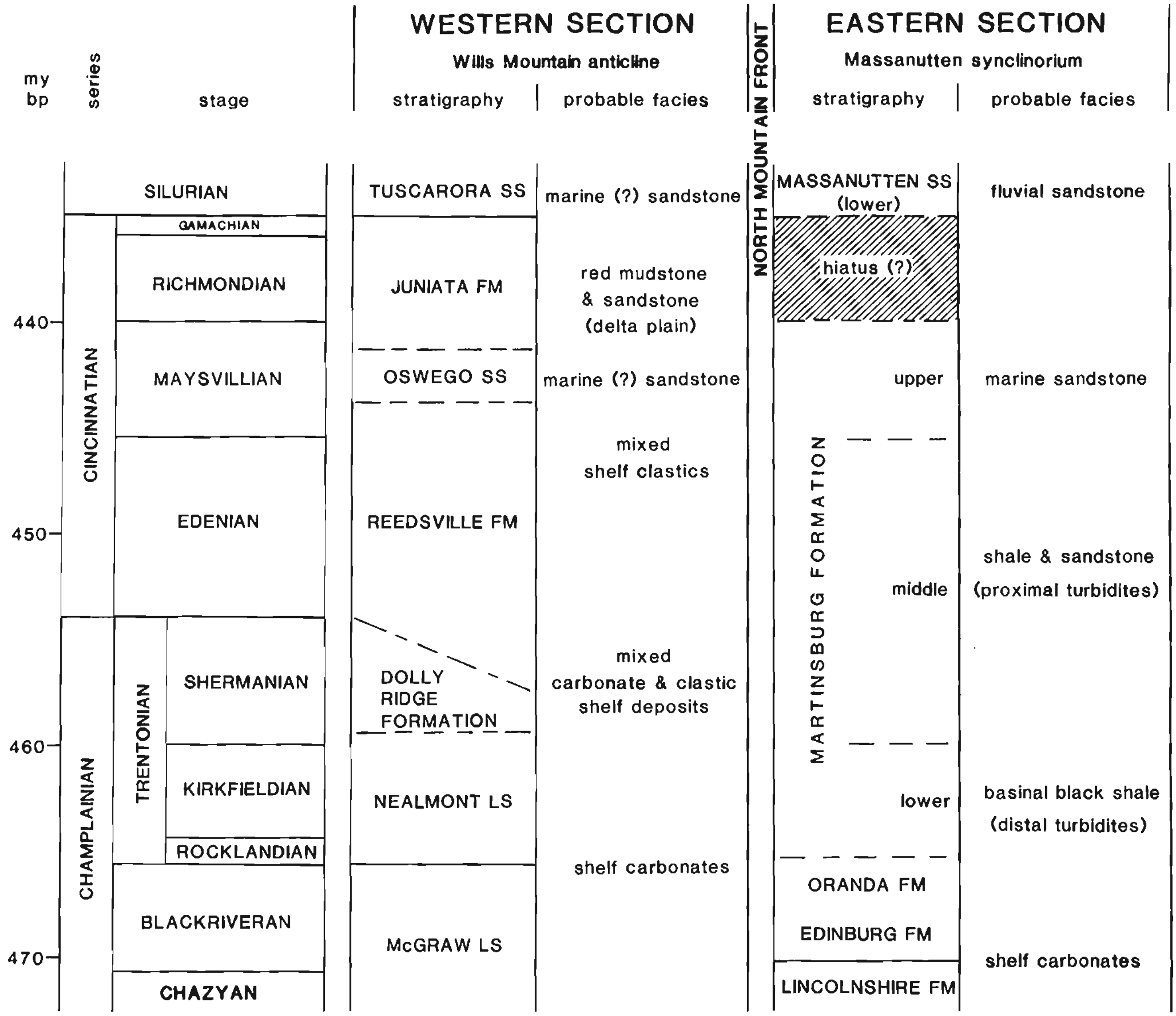

Figure 2. Stratigraphic summary of the eastern and western sections of northern Virginia and West Virginia. Chronostratigraphy of eastern section based on Ross et al. (1982).

Paleocurrent information provides support of the inferred basin geometry. McBride (1962) has shown that Martinsburg paleocurrents had two general components: toward the northwest, perpendicular to the basin axis; and toward the northeast, parallel to the basin axis. Northwesterly paleocurrents indicate that there was a positive area on the southeastern side of the basin. This is important, because it provides evidence that this was a trough-shaped basin, and not simply the edge of the continent. Paleocurrents flowing into the basin from the southeastern margin were rerouted to flow parallel to the basin axis toward the deeper part of the basin, which was probably near the Pennsylvania-New Jersey border.

Western basin

A western basin was situated west of the Wills Mountain Anticline. Primary evidence for this basin is the greater thickness of Taconian clastics in this area (Fig. 3). The western section (Fig. 2) is representative of the strata in the western basin. Facies changes in the subsurface across the western basin (Chen, 1977; Diecchio, 1985b) do not indicate the presence of an Ordovician topographic basin. The anomalously thick clastic section in the western basin apparently accumulated in shallow water, suggesting that, unlike the eastern basin, sedimentation was keeping pace with subsidence, and a topographic basin did not form. As shown in Figure 3, Middle and Upper Ordovician strata (Reedsville and Juniata formations) and Silurian strata (Tuscarora Formation) are thicker in this basin, indicating continued subsidence. A more regional expression of this western basin is illustrated in Figure 5c as the isopach anomaly of the Juniata Formation in eastern West Virginia.

DEVELOPMENTAL MODEL

Using the northern Virginia area as a model, the criteria for the recognition of deep and shallow basins can be characterized.

Deep basins

As defined in this paper, a deep basin is typically a structural basin that was originally a pronounced topographic basin. Using the eastern basin as a model, the characteristics of deep basins are 1) pronounced lateral and vertical facies variations that occur within the basin, indicating a paleotopographic depression; and 2) strata that thicken into the basin. A deep basin would result from a situation where subsidence occurred at a greater rate than sediment accumulation.

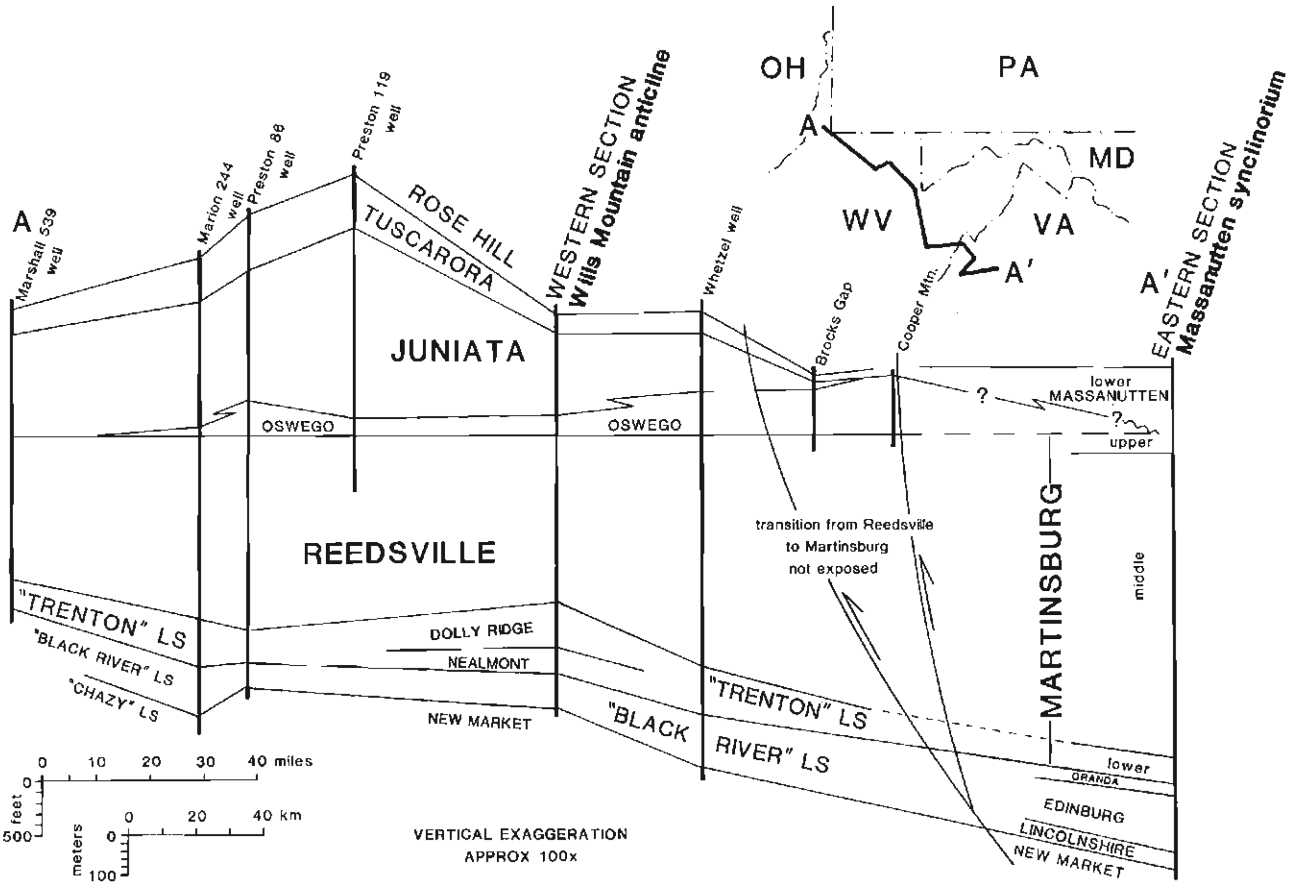

Figure 3. Stratigraphic cross-section across eastern West Virginia and northern Virginia. Spacing between sections is estimated palinspastic distances based on structural interpretations by Harris et al. (1982) and Kulander and Dean (1986). Refer to Figure 2 for a description of sedimentary facies.

Shallow basins

As defined in this paper, a shallow basin is typically a structural basin that was probably never a pronounced topographic basin. Using the western basin as a model, the characteristics of shallow basins are 1) strata that thicken into the basin; and 2) minimal facies changes across the basin, reflecting lack of pronounced topography. Shallow basins are the result of subsidence and sediment accumulation occurring at about the same rate.

MIDDLE ORDOVICIAN DEEP BASINS

Other areas, where deep marine siliclastic strata are present, are also proposed to have been the sites of Ordovician deep basins. Deep marine facies are the principal criteria for the recognition of these basins, whether or not they are associated with thickness anomalies. The areas that are thought to be Ordovician deep basins are shown in Figure 4. All Ordovician deep basins are Middle Ordovician (Champlainian) in age, as defined by Barnes et al. (1981). It is not clear whether each of the following areas was a separate basin, or whether each area is a structural remnant of one or more larger, more continuous basin(s).

Northern Virginia to Quebec

The late Middle Ordovician carbonate to basinal clastic transition of the Massanutten Synclinorium (eastern basin) can be traced northward to Quebec (Fig. 4). It is interrupted where it is overthrust by the Hamburg Klippe in Pennsylvania, the Taconic allochthon of New York and western New England, and numerous thrust slices just south of the international boundary. The stratigraphic similarity of the Middle Ordovician section throughout this area is the basis for postulating that it may have been part of one deep basin, or a few deep basins that developed at the same time, during the Trentonian (late Middle Ordovician).

Shanmugam and Lash (1982) described the collapse of a carbonate shelf and the development of a deep Ordovician basin, part of which is preserved in the Great Valley of Pennsylvania. Here the stratigraphic sequence is from shallow water carbonates of the Jacksonburg Limestone, to the pelagic sediments and turbidites of the Martinsburg Formation.

The Martinsburg outcrop belt extends from Pennsylvania to the Taconic region along the eastern border of New York. The significant aspects of the Middle to Upper Ordovician stratigraphic succession of this area have been summarized by Baldwin (1987) for strata of the Middlebury Synclinorium that are exposed on the southeast side of Lake Champlain. He described a vertical succession from shallow subtidal and open shelf limestone of the Black River Group and lower Trenton Group, to post-lower Trenton dark shales. These shales record a deepening event that occurred at about the same time as the deepening of the northern Virginia and Pennsylvania areas.

As described by Zen (1961), the Middlebury Synclinorium extends southward and continues into the autochthonous strata beneath the northern end of the Taconic Range in Vermont. These autochthonous strata include limestone of Chazy, Black River, and Trenton age, which are overlain by partly conglomeratic black slate and phyllite of the Hortonville Formation and the upper part of the Ira Formation. On the western side of the Taconic Range in New York, Fisher (1982) described an autochthonous section in the mid-Hudson Valley that contains a limestone to dark shale transition that is analogous to, but a little older (Kirkfieldian) than that of the Champlain Valley (Shermanian). Like the section on Lake Champlain, the autochthonous strata of the Taconic region also record a deepening event during the Trentonian.

The allochthonous rocks of the Taconic region represent an entirely different history than the autochthonous sequence. Allochthonous rocks of the Vermont Taconic region contain dark slate, and other intermixed lithologies, of Cambrian through Middle Ordovician age (Zen, 1961). The allochthonous rocks of the mid-Hudson Valley (Berry, 1962; Fisher, 1982) contain a similar sequence. These rocks represent the deep marine environments that existed east of the North American carbonate bank from the Cambrian until the Middle Ordovician. Apparently this deep marine area was uplifted during or prior to the emplacement of the Taconic klippen. This uplift formed the eastern margin of a deep basin or trough into which the klippen were emplaced. The western margin of this trough was the portion of the carbonate bank that had not foundered.

The stratigraphy of the area near the international boundary was reported by Shaw (1958) and Cady (1960). Of significance here is a lateral transition from the Cambro-Ordovician carbonate sequence to a predominantly shaly Cambro-Ordovician basinal sequence. Rodgers (1968; 1970, p. 73, 118) has interpreted this as a transition from the carbonate sequence to a deep basinal sequence that developed seaward of the carbonate bank. If this is the case, here we have a deep Cambro-Ordovician basin that, rather than having developed by Taconian downwarping of the edge of the carbonate bank, was coeval with the carbonate bank. This basinal sequence is similar to the allochthonous basinal deposits of the Taconic Range. Farther north, the Quebec basin contains a carbonate to shale transition deposited during the Blackriveran (Barnes et al., 1981) indicative of the destruction of a carbonate bank and the formation of a deep basin in that area.

Northeast of Quebec City, the Middle Ordovician dark shales are cut out by faulting along Logan's Line. The next exposure occurs along the northern coast of the Gaspé Peninsula, the stratigraphy of which was described by Enos (1969). In this area, the Middle Ordovician (Trentonian according to Barnes et al., 1981) Cloridorme Formation or Group is lithologically similar to the deep basinal strata under discussion. This area of northern Gaspé is therefore included in the list of probable Middle Ordovician deep basins. The Cloridorme Formation is allochthonous (Williams, 1978; Barnes et al., 1981), and not stratigraphically continuous with the underlying strata. It is therefore not possible to determine the age of this basin, or whether it developed by destruction of the carbonate bank, or seaward of the carbonate bank.

Southern Appalachians

South of the Massanutten Synclinorium are isolated exposures of Middle Ordovician dark shale that overlies carbonate. These strata are genetically similar to those deep basinal deposits already described. However, in the Southern

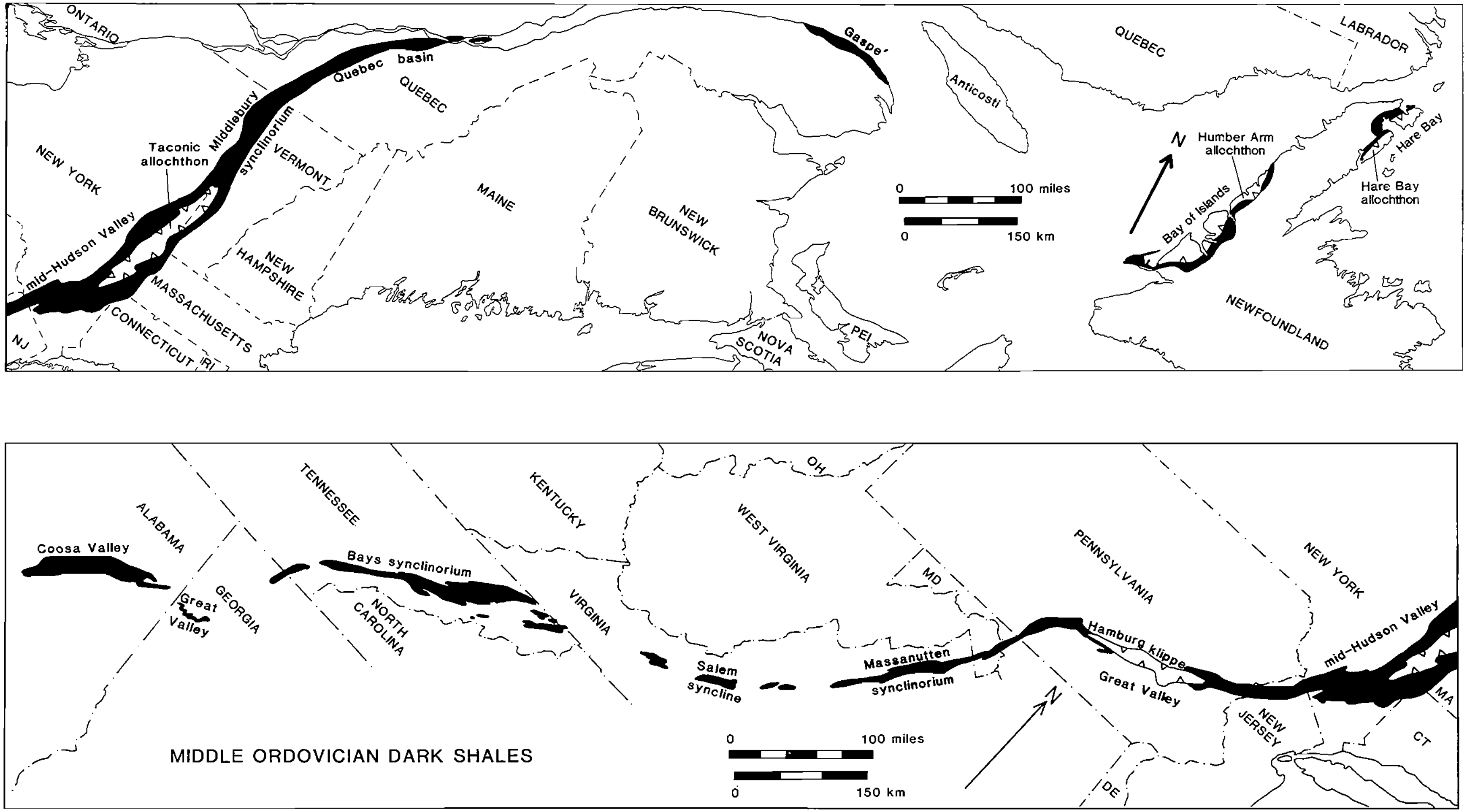

Figure 4. Extent of Middle Ordovician dark shales interpreted as having formed in deep basins. Modified from Williams (1978).

Appalachians, the basins are a little older, having developed during the Blackriveran or Chazyan (middle Middle Ordovician). Because of structural dislocations between each outcrop area, it is not possible to determine whether these were isolated basins, or a single basin.

The Ordovician carbonate section in the easternmost part of the Tennessee Valley and Ridge, between the Bays Synclinorium and the Blue Ridge, is terminated above by the thick black shale and turbidites of the middle Middle Ordovician Blockhouse and lower Sevier formations. It has been interpreted (Shanmugam and Walker, 1978, 1980; Shanmugam and Lash, 1982) that the carbonate shelf in this area underwent a sudden deepening.

For southwestern Virginia, Read (1980) documented a sequence of strata similar to that in eastern Tennessee. In this area the Lenoir Limestone is overlain by thick black shale of the Paperville Formation and coarser, submarine fan deposits of the Knobs Formation. This sequence implies the collapse of a carbonate shelf and the development of a deep basin (Read, 1980) during the Middle Ordovician. Leible (1975) documented the stratigraphic thinning of the Middle Ordovician limestone on the southeastern limb of this part of the outcrop belt. This further implies the presence of an eastern margin of this basinal structure during the Middle Ordovician. The fault-bounded Salem Syncline of southern Virginia contains a thinner and finer grained facies (Paperville Shale) of the deep basin clastic sequence (Paperville/Knobs) described above (Read, 1980). Subtle downwarping of the Salem area probably started prior to deposition of the Paperville Shale, as indicated by thickness variations in lowermost Middle Ordovician limestone (Lowry, 1960).

The Middle Ordovician stratigraphic succession east of the Clinchport Fault (Georgia) and the Helena Fault (Alabama) is predominantly clastic, unlike the predominantly carbonate section west of these faults (Drahovzal and Neathery, 1971; Chowns and McKinney, 1980). Within this area, the upward transition from the Lenoir Limestone to the dark Rockmart Slate (Georgia) or Athens Shale (Alabama) has been interpreted as the development of a deep siliciclastic basin during the Chazyan (Benson, 1986; Carter and Chowns, 1986). The dark shale/slate facies are best developed in the Great Valley of Georgia and in the Coosa Valley of Alabama, each of which is in the easternmost part of the Valley and Ridge Province in that area (Thomas and Drahovzal, 1974; Chowns and McKinney, 1980; Carter and Chowns, 1986). Based on measurements of oriented graptolites in the Athens Shale, Jones and Dennison (1970) have shown thatpaleocurrentsin this basin flowed from east to west, providing evidence for an eastern margin of this deep basin. Conglomerates in the Rockmart Slate contain clasts of Piedmont rocks, which would have been easterly derived (Cressler, 1970).

Western Newfoundland

Schuchert and Dunbar (1934) provided stratigraphic data for the Ordovician succession of the Bay of Islands Synclinorium, where Lower Ordovician limestone of the St. Georges Group is overlain by Middle Ordovician limestone and dark shale of the Table Head Group. This sequence is also interpreted as representing the development of a deep basin in an area where the carbonate bank had foundered (Jacobi, 1981; Cawood et al., 1988). Rodgers (1970, p. 150; 1971) pointed out that the dark shale strata in the western Newfoundland area are somewhat older (Whiterockian, or lower Middle Ordovician according to Barnes et al., 1981) than the lower black shale in areas farther south.

The stratigraphy of western Newfoundland has been described in such detail that the development of this deep basin and its contained allochthon is well documented (Cawood et al., 1988). Of significance is the observation that the deposition of the deep basinal strata (Table Cove and Black Cove formations) preceded emplacement of the Humber Arm Allochthon. This implies that the allochthon slid into a pre-existing depression. Further evidence for this interpretation is the conformable map relationships (Williams, 1978) between the allochthon and the outcrop trace of the basinal shale sequence. The post-allochthon cover sequence of Silurian carbonates indicates that the basin was filled almost to sea level after the allochthon was emplaced (Cawood et al., 1988). Also, the allochthon was derived from the southeast, providing evidence for a southeastern margin of the deep basin.

A sequence of strata similar to that in the Bay of Islands area (Rodgers, 1970, p. 152; Williams, 1978), and of similar age (Barnes et al., 1981) surrounds the Hare Bay Allochthon on the northern peninsula of Newfoundland near Hare Bay. It can be implied that this area was also part of an Ordovician deep basin that resulted from foundering of a carbonate bank, and into which the allochthon was emplaced.

ORDOVICIAN SHALLOW BASINS

Isopach data (Fig. 5) for shallow marine to nonmarine (not deep basinal) facies of the Taconian clastic sequence provide a basis for proposing that deposition of these strata was localized in shallow basins. Generalized regional thickness data (Fig. 5a) for the Middle Ordovician (Blackriveran) sandstone and redbed facies associated with the Blountian phase of the Taconian clastic sequence (Bays, Moccasin, and Bowen formations) are provided by Dennison and Wheeler (1975). These strata are anomalously thick in eastern Tennessee and southwestern Virginia. This implies that these areas, parts of which were deep basins during the middle Middle Ordovician, continued to develop, but as shallow sedimentary basins during the late part of the Middle Ordovician.

Isopach data for the Upper Ordovician sandstone and redbed facies of the Queenston phase of the Taconian clastic sequence, are shown in Figures 5b and 5c. The non-red sandstone and conglomerate facies (Oswego of Figs. 2, 3), although being facies-equivalent to the redbed facies (Juniata of Figs. 2, 3), are generally situated below the redbed facies. The sandstone and conglomerate facies are thicker in the western part of what in this paper is called the eastern basin in Virginia. This may indicate a westward shift in the axis of the eastern basin during the deposition of these coarser, and probably shallower, clastics. Isopach data in Figure 5b indicate that another shallow basin (or part of the same basin) extended northward into central Pennsylvania.

Thickness data for the redbed facies of the Queenston clastic sequence (Figure 5c) provides evidence of the shallow western basin discussed earlier in this paper. As illustrated in

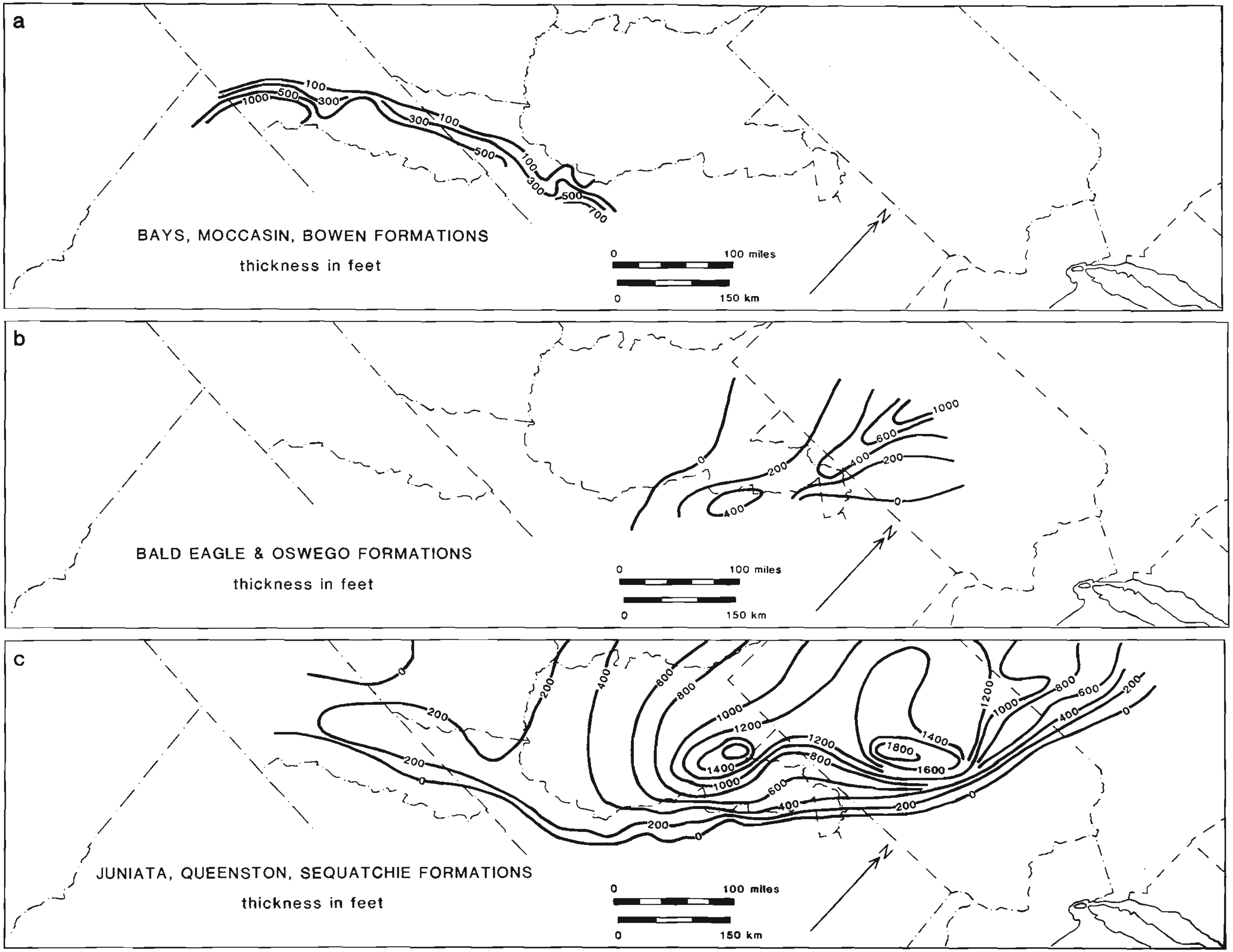

Figure 5. Isopach map of Taconian clastic strata exclusive of deep basinal facies. **a,** Blountian sandstone and redbeds; **b,** (lower) Queenston non-red sandstone and conglomerate; **c,** (upper) Queenston redbeds. After Dennison and Wheeler (1975); Dennison (1982); Diecchio (1985b).

Figure 3, the shelf deposits of the Reedsville Formation are thicker in the western basin, indicating that there was slow subsidence here, while the deep basin existed farther east. In addition, both the red and non-red Queenston clastics (Figs. 5b, 5c) delineate a shallow basin in central Pennsylvania. This is probably a continuation of subsidence of the same basin delineated by the Oswego isopachs (Fig. 5b).

CONTROL OF BASIN DEVELOPMENT

Deep basins

Most of the Middle Ordovician deep basins mentioned above had developed in association with the foundering of the pre-existing carbonate bank. The development of a deep basin requires that subsidence occur at a greater rate than sediment accumulation. These deep basins have been described as foreland basins associated with thrust-loading during subduction of the eastern margin of North America (Jacobi, 1981; Shanmugam and Lash, 1982; Quinlan and Beaumont, 1984; Lash, 1987). The isostatic effect of sediment loading, as well as other factors, is most likely of secondary importance in the development of deep basins.

Shallow basins

For shallow basins, there is not such a great disparity between the rate of downwarping and the rate of sedimentation, in fact the two rates should be about equal. There are two possible explanations for the development of a shallow basin. One reason might be coincidence, that the availability of sediment was sufficient to allow accumulation at the same rate as downwarping. An alternate, and more plausible explanation, is that sediment accumulation was the cause of the subsidence due to isostatic adjustment.

The question of why the downwarping and the sedimentation are localized to such specific areas as the basins themselves is another matter. Some structural control is probably responsible for this localization, but for the most part, these structures have not yet been recognized. One possible example has been described by Jacobeen and Kanes (1974, 1975) for the Broadtop Synclinorium that extends from northern Virginia into central Pennsylvania, the axis of which coincides with the axis of thickening of the Bald Eagle and Oswego formations (Fig. 5b). They have documented the presence of high angle normal faults that offset basement below the Broadtop Synclinorium. Differential rates of subsidence may have occurred on either side of these faults, either before or during accumulation of the Bald Eagle and Oswego formations, thus localizing the accumulation of these strata.

In at least two areas, Tennessee/southwestern Virginia and the eastern basin of northern Virginia, the development of a deep basin is followed by the existence of a shallow basin. In these areas, after the initial subsidence in response to the thrust load, it could be that sedimentation continued so as to fill these basins to shallower depths. Unequal rates of sedimentation, or localization of sediment accumulation within this area or part of this area, would have resulted in the development of one or more shallow basins.

Subsequent structural modifications

The sedimentary basins described above are features probably due to syndepositional conditions. Later compressional deformation (Acadian or Alleghenian) modified the original geometry of these basins by steepening their limbs or by redeforming these basins into more complex structural features. Also, many of these basins are allochthonous. However, for many of the basins, the original basinal structure is still recognizable in the form of a synclinorium. For example, the structural feature known today as the Massanutten Synclinorium had developed in the area that was an Ordovician deep basin (eastern basin of northern Virginia). Subsequent deformation (Acadian or Alleghenian) of the original basinal structure was responsible for the the present structural configuration (dips, fold axes, second-order folds) of this synclinorium. Also, the Massanutten Synclinorium is allochthonous, having been displaced westward along the North Mountain Fault (Fig. 1).

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Stratigraphy of the Middle Ordovician Long Point Group, western Newfoundland

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Abstract

The Long Point Group is the only fossiliferous Ordovician sequence in Newfoundland that postdates the deposition of synorogenic flysch and possibly also the emplacement of the allochthons. Thus, understanding the biostratigraphy and depositional history of this sequence is a critical element in any tectonic model for the Appalachian Orogen.

The Long Point Group consists of the lower, 75 m thick Lourdes Formation and the upper, 300 m thick Winterhouse Formation. Although the Lourdes Formation deepens up-sequence, it consists of three shallowing-upward sequences, each formally established as new members. The boundaries between the lower two members are unusual "planed-off" surfaces, whereas the top of the formation is transitional to the deeper Winterhouse Formation.

The lowest Shore Point Member begins with a green sandstone: probably reworked flysch or Humber Arm rocks. The remainder of the member is massive sandstone and interbedded sandstone and intertidal carbonate. The Black Duck Member is characterized by the presence of in situ coral heads. At the base, these are isolated heads, but in the upper six metres they form wall-like "reefs". These upper six metres contain a very diverse fauna, including trilobites, nautiloids, ostracodes, stromatoporoids, bryozoans, and gastropods. The uppermost Beach Point Member represents a deeper water environment and contains rolled and worn corals and other fossils in the upper subunit. The Winterhouse Formation is interpreted as a deeper water, possibly slope, deposit.

As the Long Point Group has been considered traditionally to immediately overlie the allochthonous Humber Arm Group, it provides a minimum age for the emplacement of the Taconic allochthons in western Newfoundland. Recent reinterpretations suggest that the present location of the allochthons was a result of the Acadian Orogeny. Resampling for conodonts has confirmed a predominantly Prioniodus gerdae Subzone age (Blackriveran to mid-Kirkfieldian) for most of the Lourdes Formation.

Résumé

Le groupe de Long Point est la seule séquence ordovicienne fossilifère à Terre-Neuve qui est postérieure à l'accumulation de flysch synorogénique et possiblement aussi à la mise en place des dépôts allochtones. Il est donc essentiel de connaître la biostratigraphie et l'histoire sédimentaire de cette séquence afin de pouvoir établir un modèle tectonique de l'orogène des Appalaches.

Le groupe de Long Point comprend une formation inférieure de 75 m d'épaisseur, la formation de Lourdes, et une formation supérieure de 300 m d'épaisseur, la formation de Winterhouse. Bien que la formation de Lourdes témoigne d'un accroissement de la profondeur marine vers le haut, elle se compose

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de trois séquences qui se caractérisent par une diminution de la profondeur marine vers le haut; on donne à chacune de ces séquences la désignation formelle de membre. Des surfaces «planées» peu communes séparent les deux membres inférieurs, tandis que le sommet de la formation passe progressivement à la formation de Winterhouse d'eau plus profonde.

Le membre inférieur, celui de Shore Point, se compose d'abord d'un grès vert, vraisemblablement un flysch remanié ou des roches de Humber Arm. Le reste du membre se compose de grès massif et de grès et de carbonate intertidal interstratifiés. Le membre de Black Duck se caractérise par la présence de têtes de corail en position de vie. Ces têtes sont des structures isolées à la base du membre et des «récifs» en forme de muraille dans les six mètres supérieurs, où l'on trouve une faune très variée composée de trilobites, de nautiloïdés, d'ostracodes, de stromatopores, de bryozoaires et de gastropodes. Le membre sommital, celui de Beach Point, représente un milieu d'eau plus profonde; sa sous-unité supérieure contient des coraux et d'autres fossiles roulés et usés. La formation de Winterhouse est un dépôt d'eau plus profonde, possiblement de talus.

Puisque le groupe de Long Point est considéré comme reposant directement sur le groupe allochtone de Humber Arm, il fournit l'âge minimal de la mise en place des allochtones taconiques dans l'ouest de Terre-Neuve. Selon une nouvelle interprétation récente, la mise en place des allochtones est attribuée à l'orogenèse acadienne. L'échantillonnage répété de conodontes confirme que la plus grande partie de la formation de Lourdes a l'âge de la sous-zone à Prioniodus gerdae (Blackriverien au Kirkfieldien moyen).

INTRODUCTION

The Long Point Group outcrops on the Long Point Peninsula of the Port au Port Peninsula (Fig. 1). Outcrops are predominantly to the southeastern side of "The Bar" and consist of nearly 16 km of strike section. The formations dip to the northwest at between 25° and 40°. The 75 m thick Lourdes Formation makes up the lower part of the Long Point Group. The upper part includes the poorly exposed Winterhouse Formation, which outcrops on the central and northwestern side of "The Bar" and is probably over 500 m thick. This study is primarily to clarify and define the stratigraphy of the Long Point Group, especially the Lourdes Formation (Fig. 2).

The Long Point Group unconformably overlies the highly deformed allochthonous rocks of the Humber Arm Group. This makes a determination of the age of its base of paramount importance in the understanding of the timing of the Taconic Orogeny in western Newfoundland, as it has traditionally been regarded as providing a minimum age of allochthon emplacement. This has assumed that the allochthons were transported during the early Middle Ordovician Taconic Orogeny. Distinguishing the tectonic and thermal processes between the Taconic and Acadian (Silurian–Devonian) orogenies has been difficult in western Newfoundland. Thermal effects in the region are generally modest (Nowlan and Barnes, 1987). New interpretations of the regional tectonics were developed after study of industry reflection seismic data, which were originally gathered offshore in 1973. From this data, Stockmal and Waldron (1989, p. A11) concluded that:

> "(i) the Long Point Group and the overlying Silurian–Devonian Clam Bank Formation are in a west-facing homocline flattening to the west; (ii) the underlying platform, possibly interrupted by short-throw steep faults, dips *eastward*, unaffected by the homocline above. The wedge-shaped region above the platform and below the Long Point is occupied by the Humber Arm allochthon and probably also by a transported slice of platform rocks. This geometry is interpreted as a structural "triangle zone", the significance of which is that the Long Point and Clam Bank, in the hanging wall of the east-vergent roof thrust of the triangle zone, must pre-date the development of the zone. Therefore, the Humber Arm allochthon was emplaced in its present position in the Acadian orogeny. The probable inclusion of a platform slice within the triangle zone implies significant displacement (>30 km) at this time."

If this interpretation is accepted, further studies would be useful to examine the stratigraphic relationships of the Long Point Group and the overlying Clam Bank Group. One might expect a much greater unconformable relationship than presently exists in the outcrop belt.

Previous studies have either only obtained an age for the upper part of the Lourdes Formation (Fåhraeus, 1973; Copeland and Bolton, 1977; Dean, 1977) or a vague correlation of the lower part (Cooper, 1956; Bergström et al., 1974). In an attempt to determine the age of the base of the Long Point Group, very large (up to 9 kg) samples from the basal limestone (SP2) were processed for conodonts. A summary of the results of this work is presented herein. In addition, the studies of the nautiloid faunas, coral reefs, diagenesis, and sedimentology of the Lourdes Formation are in progress. This detailed work has shown that the stratigraphy is in need of revision and formal definition. Bergström et al. (1974) established the Lourdes and Winterhouse formations and three informal members; these have been redefined and amplified herein.

The Long Point Group is the oldest unit in the Clam Bank belt of western Newfoundland. The Long Point Group unconformably overlies the chaotic Humber Arm allochthon of the Humber Belt on the southeastern side of "The Bar". However, toward the southwest, the sequence is overturned and seems to be overthrust by early Middle Ordovician Table Head Group rocks. The Long Point Group is thought to overlie the Mainland Sandstone, but the precise relationship cannot be observed.

Previous work

The stratigraphy of the Long Point Group was first described by Schuchert and Dunbar (1934, p. 70), who divided the "Long Point Series" into five informal units with a total thickness of about 500 m (Fig. 3). They considered the series to be in faulted contact with the underlying Humber Arm Group. Sullivan (1940, *in* Riley, 1962) divided the Long Point Group into seven units, the lower three of which were equivalent to the Lourdes Formation (Fig. 3). Riley (1962) mapped the Port au Port Peninsula, including Long Point. Corkin (1965) also studied the stratigraphy and mapped the Long Point Group as part of an assessment of its petroleum potential. Rodgers (1965) demonstrated that the contact with the Humber Arm Group was unconformable and only locally faulted. He (1965, p. 87) divided the lower part of the Long Point Formation (equivalent to the Lourdes Formation) into seven units with a total thickness of 90 m (Fig. 3) and estimated the remainder of the Long Point Formation to be 800 to 850 m thick. Kay (1969) divided the Long Point Formation into four units, the lower three of which are equivalent to the Lourdes Formation and the fourth to the Winterhouse Formation (Fig. 3). In theses on the stratigraphy and paleontology of the Long Point Group (Shaikh, 1971; Weerasinghe, 1970), the Lourdes Formation was divided into six units, with a seventh that encompassed the Winterhouse Formation. Fåhraeus (1973) also established a seven-fold subdivision for the Long Point Group and produced the first detailed measured sections and interpretation of the depositional environments (Fig. 3). Bergström et al. (1974, p. 1627) redefined the stratigraphy, formally establishing the Long Point Group, the constituent Lourdes Formation, and the Winterhouse Formation (Fig. 3). Dean (1977), in his description of the trilobite faunas, adopted the informal units of Bergström et al. (1974).

Surprisingly, only a limited amount of paleontological work has been undertaken on the fossiliferous Long Point Group. The first faunal list was provided by Schuchert and Dunbar (1934). More recent studies were by Flower (1952) on nautiloids, Cooper (1956) on brachiopods, Bolton (1965) on corals, Fritz (1966) on bryozoans, Copeland and Bolton

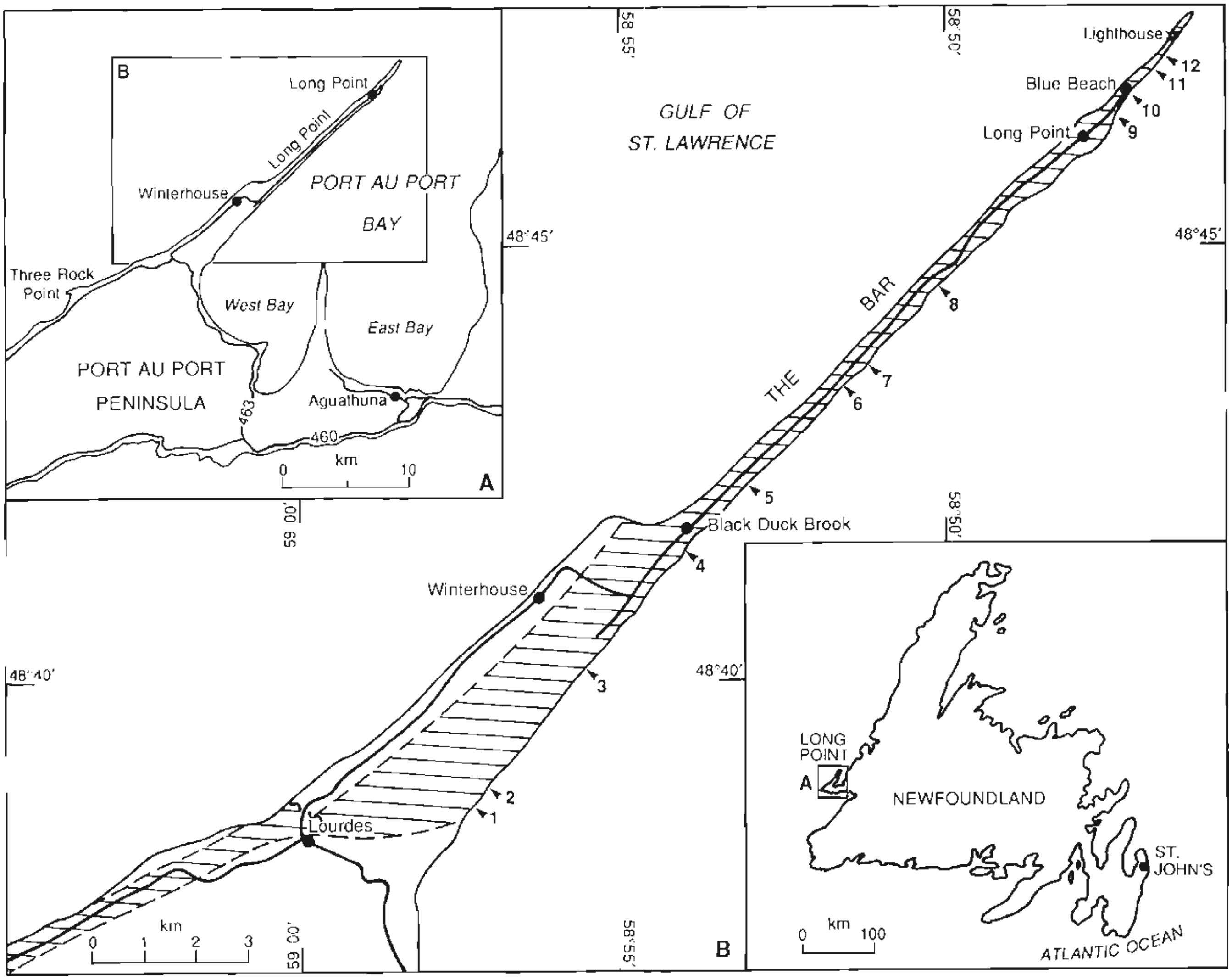

Figure 1. Long Point area of Newfoundland, showing localities mentioned in text.

(1977) on ostracodes and corals, Dean (1977) on trilobites, and Stait (1988) on nautiloids. Conodont data were provided by Fåhraeus (1973) and Bergström et al. (1974).

Depositional environmment

Although the Long Point Group was examined in some detail for this study, no attempt was made at the bed-by-bed analysis required to unravel the sequence of depositional environments represented. However, our preliminary observations allow generalizations to be made about the changes and pattern of sedimentation. This reconnaissance study has shown that an important story is to be obtained from these rocks. Preliminary studies of specific aspects of the sedimentology by R. Cuffy and N.P. James and by M. Coniglio have pointed to the importance of unraveling the entire complex of facies.

The following brief discussion is intended to place the stratigraphy developed herein into a sedimentological framework.

The Lourdes Formation consists of three members; each is a shallowing-upward sequence, with the lower two terminating in a "planed-off" bed or hardground, which is followed by deeper water sediments of the basal units of the overlying member. Each succeeding member, although a shallowing-upward sequence, represents a deeper water facies than the preceding member (Fig. 3).

The basal Shore Point Member has at its base green sandstone (unit SP1), which probably represents eroded and reworked Humber Arm facies in a subtidal setting. Unit SP2 represents the establishment of carbonate deposition over the area. The succeeding facies are very shallow water sequences, with unit SP4 probably representing a high intertidal setting.

The Black Duck Member has deeper water, silty carbonate and carbonate strata (BD1) that contain rare, in situ, isolated coral heads. The remainder of the member is from a shallower setting, showing considerable evidence of tidal and current activity. However, the uppermost unit (unit BD4) is still clearly subtidal.

The basal unit of the Beach Point Member was deposited in relatively deep water, with some beds having turbidite characteristics. The overlying unit (unit BP2) is also deep water, with rolled and worn fossils as well as evidence of currents.

In summary, the sequence of sediments probably reflects the subsidence that followed the emplacement of the allochthonous rocks of the Humber Arm Group. Carbonate deposition proceeded in a series of asymmetrical, shallowing-upward cycles, each deposited in deeper water until the water depth became too great for primary carbonate formation. Following this, sedimentation was dominated by the transported carbonates, clastics, and slumps typical of the lower Winterhouse Formation (Fig. 3).

Biostratigraphy

A major outstanding biostratigraphic problem in western Newfoundland has been the precise age of the base of the Long Point Group. Most of the fauna previously described was from the highly fossiliferous Black Duck Member, which is 40 to 50 m above the base of the Long Point Group. One of the aims of this study was to obtain biostratigraphic control on the basal rocks. This was achieved using conodonts, which were extracted from the two lowest units. Unfortunately, non-conodont faunas were found to be virtually nonexistent in the Shore Point Member and, when present, were of limited biostratigraphic use. The significance of the conodonts will be discussed further below.

Information for most fossil groups is restricted to the informal units BD4 and BP1, and no zonation is possible for any group (possibly with the exception of conodonts). The nautiloids are the only macrofossil group that has been collected systematically from the entire Lourdes Formation

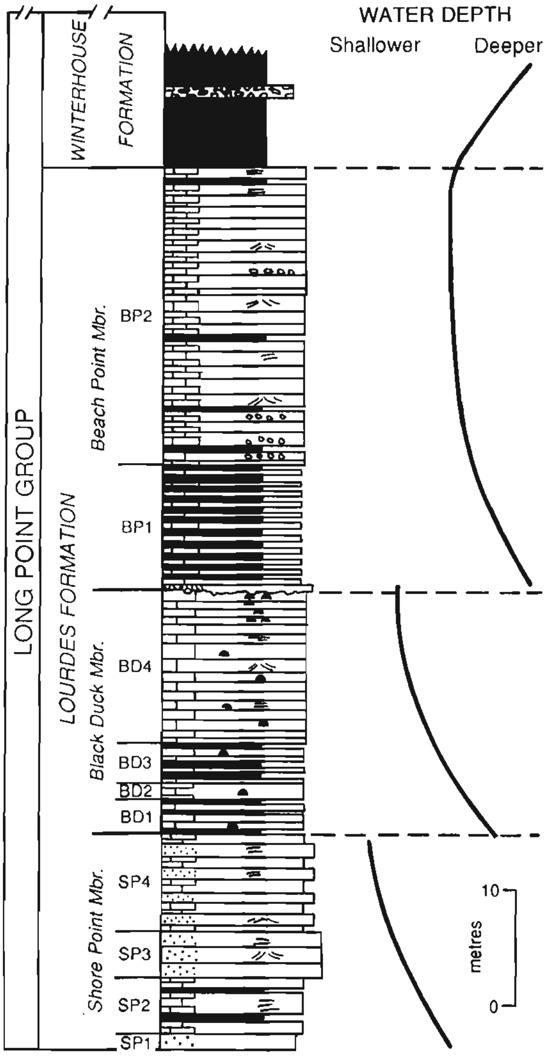

Figure 2. Stratigraphic column indicating the subdivisions and the relative water depths of the Long Point Group.

(Stait, 1988). Specimens occur from the nearly basal informal unit SP3 to the Winterhouse Formation but, again, they are most abundant in unit BP4 (Fig. 4).

Study of the nautiloids indicates a correlation with the Lowville to lower Rockland formations of the Ottawa Basin (Stait, 1988), with the species level correlations to the Lowville (late Blackriveran, Barnes et al., 1981). Copeland (*in* Copeland and Bolton, 1977) considered the ostracodes to be late Wildernessian to Rocklandian in age (late Blackriveran–Rocklandian, Barnes et al., 1981). Dean (1977) suggested a Blackriveran to Porterfield age, based on the study of trilobites from units BD4 and BP1. Thus, the macrofauna consistently indicate a late Blackriveran to Rocklandian age for the Black Duck Member and strata above. It should be noted, however, that none of the fauna provides a particularly precise correlation. Nautiloids are probably the most reliable indicators, due to the detailed study and species level correlations with the Ottawa Basin (Stait, 1988).

Fåhraeus (1973, p. 1822) considered that the "base of the Long Point Formation probably coincides with the base of the *Climacograptus peltifer* Zone of the British standard graptolite zonation and with that of the *Prioniodus gerdae* conodont subzone." This conclusion was based on the premise that the Lourdes Formation was deposited rapidly, and that the lowest conodonts in Fåhraeus' study came from the base of the Beach Point Member. Bergström et al. (1974, p. 1629) questioned this assumption, and stated that their collections from this portion of the Lourdes Formation, i.e., the lower two units, "do not contain stratigraphically very diagnostic forms but comparison of these collections with collections from the Southern Appalachians suggests that this part of the Lourdes may well be of pre-*gerdae* age." The only direct evidence on the age of the lower beds was based on the macrofossils in unit BD1 of the Black Duck Member, which Kay (*in* Bergström et al., 1974), considered to be Chazyan in age.

Collections made in the present study from the basal beds of the Shore Point Member (Fig. 4) have yielded a few conodonts. Abraded specimens of *Erismodus* sp. and *Polycaulodus* sp. occur together with a few well preserved elements of *Oulodus serratus* Stauffer. A latest Whiterockian (late Chazyan) to Blackriveran age is indicated.

Base of the Long Point Group

The precise relationship of the Long Point Group rocks to the underlying allochthonous rocks of the Humber Arm Group is extremely important in developing a framework for the tectonic development of western Newfoundland.

The first attempt to explain the relationships was by Rodgers and Neale (1963), who suggested that the contact was probably faulted. However, in the mid-1960s, Rodgers, Brueckner, Utting, and Cummings uncovered the contact at Tea Cove (locality 1 in Fig. 1). Rodgers (1965, p. 85) noted that "The contact is not a fault, though it is locally faulted; the lowest layers of the Long Point rest unconformably on red and green shale or friable claystone showing confused structure these rock types are characteristic of Humber Arm terrain."

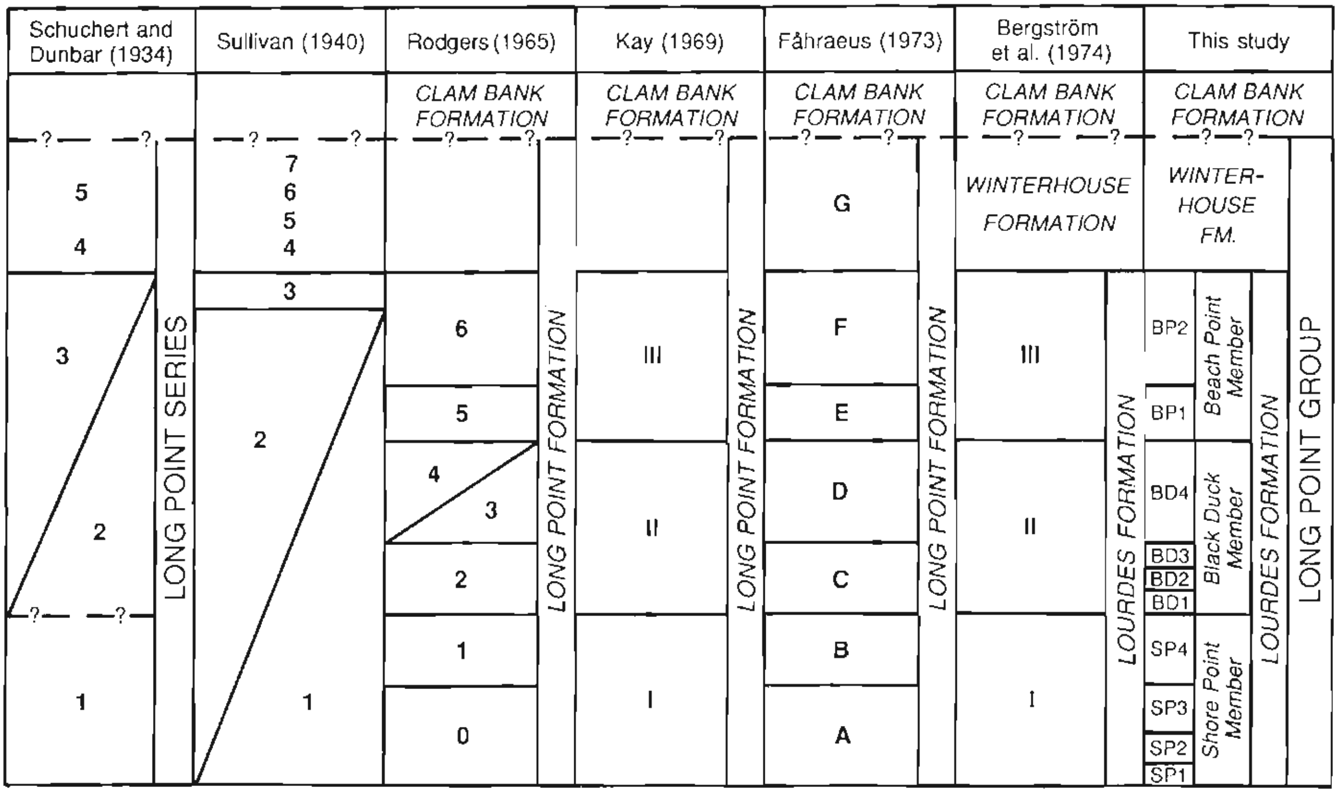

Figure 3. History of the stratigraphic classification of the Long Point Group.

The above description has been accepted by subsequent authors but has not since been confirmed. However, in 1984, B.A. Stait and N.P. James once again exposed the contact at the southeastern end of Long Point (locality 2 in Fig. 1). A detailed description of the contact is included below, but it is clear from our study that the contact is an unconformity and not a fault. The contact may now be locally faulted at some localities, but this is a later event that occurred preferentially along the contact, especially where the Long Point Group swings southwest across the peninsula, with the sequence overturned at Three Rock Point (Fig. 1,A).

A previously undescribed feature uncovered in our sections was the poorly consolidated green sandstone at the base of the Long Point Group. This green sandstone may have a

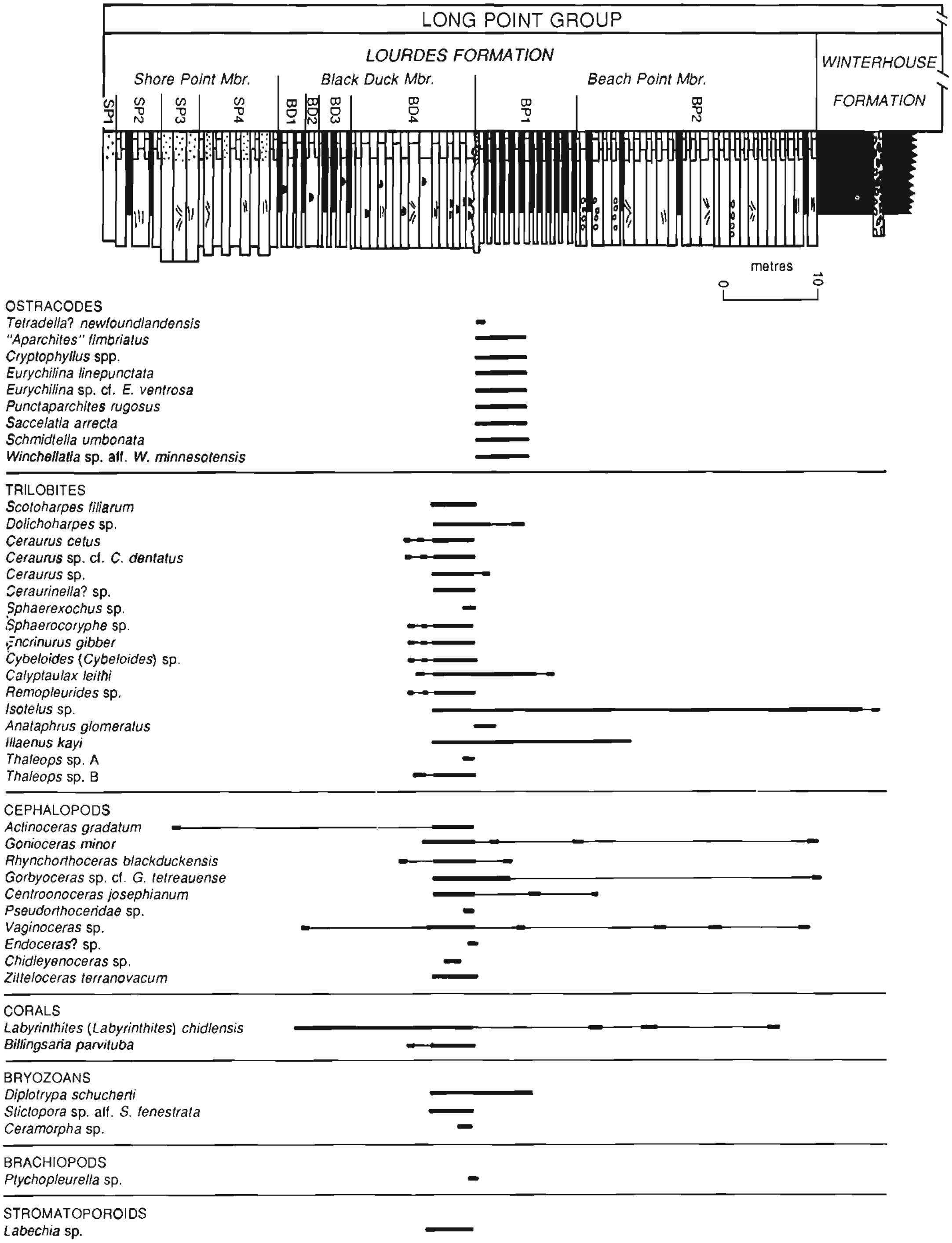

Figure 4. Ranges of fossils from the Long Point Group. Some are only general positions, as the precise stratigraphic position of some collections is not known.

localized distribution, filling lows in the eroded surface on the Humber Arm rocks, which would explain why it was not recorded by Rodgers and others.

The green sandstone is very similar to parts of the Mainland Sandstone but it is difficult to determine whether the green sandstone was formed as a late phase of flysch deposition, by reworking of the flysch during the erosional phase following emplacement of the thrust sheets, or by erosion of the underlying Humber Arm rocks. The last alternative is the most probable, as the upper Mainland Sandstone is carbonate-rich (Cow Rocks Member) rather than the typical green sandstone of the lower part. As the Cow Rocks Member appears to be older than the Long Point Group, deposition of the flysch would have ceased prior to the deposition of the lowest unit of the Long Point Group in this area.

STRATIGRAPHY

The general stratigraphy of the Long Point Group was described by Fåhraeus (1973) and Bergström et al. (1974). Recent work by the present authors enables the nature of the contact with the Humber Arm Group to be described and the members of the Lourdes Formation to be formally defined.

LONG POINT GROUP

The Long Point Group was originally defined by Bergström et al. (1974, p. 1627) as "a group consisting of two formations, the Lourdes Limestone and the Winterhouse Formation." They noted that the "Long Point rests unconformably on the Humber Arm Group." The Long Point Group is equivalent to the Long Point Series (Schuchert and Dunbar, 1934) and the Long Point Formation (Rodgers, 1965; Fåhraeus, 1973). The Long Point Group consists of the lower, dominantly calcareous Lourdes Formation and the poorly known, dominantly clastic Winterhouse Formation, which together underlie most of the Long Point Peninsula (Fig. 1).

The base of the Long Point Group is exposed at the southeastern end of "The Bar". It is most clearly exposed at Tea Cove (locality 2, Fig. 1). At this locality it consists of one metre of green, poorly consolidated sandstone (SP1), which grades up into the overlying carbonate unit (SP2). It overlies the highly deformed red and green shales of the Humber Arm Group with a relatively flat and erosional contact. Although the contact is not exposed elsewhere, the lowest unit of the Long Point Group overlies, with only a narrow gap, various rock types of the Humber Arm Group, including black graptolitic shale, greywacke, and igneous rocks.

Lourdes Formation

Bergström et al. (1974, p. 1628) defined the Lourdes Formation as consisting " of about 280 ft. (90 m) of dominantly limestone. It is fully exposed in an escarpment along the shore of the northwestern side of Port au Port Bay." This is a rather vague definition and, therefore, we herein define the base of the Lourdes Formation as the base of the green sandstone in the cliff two kilometres south of Portage on the southeastern side of Long Point (Stephenville 1:50 000) map sheet, grid ref. 563910; locality 2 in Fig. 1). The top of the formation is the base of the overlying Winterhouse Formation (see below). The Lourdes Formation is equivalent to units 1–3 of Schuchert and Dunbar (1934), units 0–6 of Rodgers (1965), and units A–F of Fåhraeus (1973) (Fig. 3). The Lourdes Formation is 75 m thick.

The subdivision of the Lourdes Formation has been a major point of difference between various studies (Fig. 3). Of the two latest studies (Fåhraeus, 1973; Bergström et al., 1974), we favour the subdivision of Bergström et al. (1974), as the three units are clearly separated by disconformable surfaces and have distinct lithological characteristics. This three-fold subdivision is sufficiently understood to allow formal definition of three new members of the Lourdes Formation. In ascending stratigraphic order, they are the Shore Point, Black Duck, and Beach Point members.

Shore Point Member

The Shore Point Member consists of 16 m of silty limestone, sandstone with calcareous cement, and minor argillaceous beds. The base of the member is defined at the base of the green sandstone in the cliff two kilometres south of Portage on the southeastern side of Long Point (locality 2 in Fig. 1; Stephenville 1:50 000 map sheet, grid ref. 563910), which disconformably overlies the highly deformed red and green shale of the Humber Arm Group. The member outcrops as cliffs along the southeast shore of Long Point Peninsula from Tea Cove to just north of Portage (between localities 1 and 4, Fig. 1). It is equivalent to units A and B of Fåhraeus (1973) and member I of Bergström et al. (1974), and consists of four informal units that we designate units SP1, SP2, SP3, and SP4, in ascending stratigraphic order.

Unit SP1

This basal unit of the Shore Point Member consists of green silty sandstone (1.0 m thick). It disconformably overlies the red and green shale of the Humber Arm Group. It outcrops rather poorly between localities 1 and 2 (Fig. 1).

Unit SP2

This 3.5 m thick unit is composed of silty limestone with minor interbeds of recessively weathering calcareous shale. It has a transitional boundary with the underlying green sandstone, with the transition occurring over a 30 cm thick interval. It outcrops sporadically between localities 1 and 2 (Fig. 1).

Unit SP3

This 5 m thick unit consists of massively bedded quartz sandstone with calcareous cement. The contact with the underlying unit SP2 is sharp but not erosional. Crossbedding is present within some beds, but most are planar laminated. Fossil fragments are concentrated in layers. The unit outcrops from Tea Cove (locality 1, Fig. 1) to just south of Portage (locality 3, Fig. 1). This unit contains a sparse, fragmented molluscan fauna.

Unit SP4

This 6.5 m thick unit includes quartz sandstone with calcareous cement and interbedded silty limestone, and minor shale beds. The quartz sandstone is commonly crossbedded and has disrupted and contorted bedding. The limestone has abundant bird's-eye structures, and typically a basal silty sandstone layer, which grades upward into fenestral limestone. The shale layers commonly contain mudcracks. The contact with unit SP3 is transitional over about one metre, with sandstone beds decreasing in thickness and silty limestone developing. This unit outcrops over a wide area from near Tea Cove (locality 1, Fig. 1) to north of Portage (locality 4, Fig. 1). The macrofauna is restricted to rare gastropods, which are transported and fragmentary.

Black Duck Member

The 21 m thick Black Duck Member consists of shale and nodular calcarenite near the base, and calcarenite and calcilutite in the upper two thirds. It is characterized by the presence of coral heads of ***Labyrinthites***, which form isolated patch reefs in the upper six metres. The base is marked by a sharp, probably erosional, contact with the underlying Shore Point Member at locality 3 (Fig. 1). The Black Duck Member is the most widely outcropping member of the Lourdes Formation. It occurs along the southeast shore of Long Point Peninsula from just north of Tea Cove (locality 1, Fig. 1) to north of Beach Point (locality 10, Fig. 1).

The wide area of outcrop and the diverse fauna of this member have made it the most studied part of the Lourdes Formation. The upper part of the member opposite the Black Duck Brook turnoff has been studied in great detail, particularly the morphology of the coral reefs (James and Cuffey, 1989).

The Black Duck Member is divided into four informal units. In ascending order, they are units BD1, BD2, BD3, and BD4.

Unit BD1

This unit consists of argillaceous beds (10 cm thick) with thin interbeds of nodular calcarenite (commonly surrounding isolated coral heads). The thickness varies along strike, with a maximum thickness of 2.7 m. The base is the same as that for the Black Duck Member. The outcrop extends from Tea Cove (locality 1, Fig. 1) to north of Portage (locality 5, Fig. 1).

Unit BD2

This unit is discontinuous along strike, but when present is a series of calcarenite beds with minor shale interbeds. It reaches a maximum thickness of 2.1 m at locality 4 (Fig. 1).

Unit BD3

This 3 m thick unit is lithologically similar to unit BD1, with thin beds of shale interbedded with nodular calcarenite. The lower boundary, where unit BD2 is present, is relatively sharp, but not erosional. Where unit BD2 is absent, unit BD3 merges with unit BD1.

Unit BD4

This unit consists of thin to moderately thick bedded calcarenite and calcilutite beds, with rare, thin argillaceous beds. Its thickness varies along strike, being 12 m at locality 2 and 15 m at locality 10 (Fig. 1). This unit forms most of the shoreline outcrop on the southeast side of the Long Point Peninsula from Shore Point (locality 2) to the north of Beach Point (locality 10). It contains abundant coral heads, especially in the upper six metres, where they form reefs up to six metres in height. The base is transitional with the underlying unit BD3; however, it is conveniently drawn at the point where calcarenite is dominant over shale. This unit is highly fossiliferous, and biocalcarenite and biocalcirudite flank the coral heads and reefs. The sedimentology and paleoecology of the coral heads and buildups at locality 5 were considered in detail by James and Cuffey (1989). They further subdivided unit BD4, although the subunits are only of local significance.

Important macrofauna are corals, nautiloids, trilobites, bryozoans, brachiopods, gastropods, echinoderms, sponges, ostracodes, and stromatoporoids. Conodonts are absent in, and close to, the reefal strata.

Beach Point Member

This is the uppermost member of the Lourdes Formation. It consists of a lower part of thin bedded shale with thin interbeds of nodular calcarenite, and an upper part of oncolitic biocalcarenite and calcilutite, with minor thin argillaceous beds. The member is variable in thickness; north of Beach Point it is 32 m thick. The base is marked by a conspicuous erosional surface on the top of the underlying Black Duck Member. This surface has been ''planed'' flat, dissecting corals and other fossils. The lowest bed in the Beach Point Member is often a thin sandstone, and the base of the member is defined as the base of this sandstone at the cliff on the southeastern shore of Long Point, opposite the turn off to Black Duck (Stephenville 1:50 000 map sheet, grid ref. 602953; locality 4, Fig. 1). This member is equivalent to units E and F of Fåhraeus (1973) and member III of Bergström et al. (1974). The Beach Point Member is divided into two informal units, BP1 and BP2, in ascending stratigraphic order. Fossils are abundant in this member, especially echinoderms, brachiopods, trilobites, conodonts, algae, nautiloids, sponges, and ostracodes, with minor corals.

Unit BP1

This unit is composed of interbedded shale and thin bedded nodular calcarenite and calcilutite, and thin, graded beds of silty limestone. The upper surfaces of the graded limestone beds are highly irregular, whereas their lower surfaces tend to be planar. The thickness of this unit at locality 5 is 10 m. The base is the same as for the Beach Point Member and is best exposed at locality 4, but is also exposed in the cliff section from south of this locality to north of Beach Point (locality 10).

Unit BP2

This 22 m thick unit is composed of thin to thick bedded biocalcarenite, calcilutite, and minor argillaceous beds. Oncolites are common throughout, but especially near the base, where they may constitute over 60% of the rock. The base is transitional, but is drawn at the first thick (40 cm) biocalcarenite; above this point, argillaceous beds become thin and rare. This unit is best exposed north of Beach Point out to the lighthouse on the southeast side of the Long Point Peninsula (localities 10 to 12, Fig. 1). However, it is well exposed all along the peninsula, south to locality 4. Crossbedding and trace fossils are particularly abundant in the unit. Rare, displaced, and eroded coral heads are also present.

Winterhouse Formation

This was defined by Bergström et al. (1974, p. 1628) as " a succession of thin-bedded silty shales, sandstones, redbeds, and occasional limestones with an estimated total thickness of 1000 ft (300 m)." Bergström et al. (1974) had a problem defining the base of the formation, commenting (p. 1628) that although "the lithological difference between the Lourdes and Winterhouse is striking, the contact between these formations appears to be gradational. " They defined it as occurring " beyond the light at the extreme northeast end of the peninsula." We agree with the location of the boundary and consider it to occur above the strike ridge of biocalcarenite that forms the projection of rock on the northeast side of the peninsula. Above this point, the sequence is dominated by shale and silty limestone.

The thickness of the Winterhouse Formation is uncertain due to the extremely poor outcrop. Earlier estimates were based on the dip and the outcrop width, e.g., 400 m by Schuchert and Dunbar (1934), 800 m by Rodgers (1965), 500 m by Fåhraeus (1973), and 300 m by Bergström et al. (1974). We are unable to refine these estimates and suggest that a stratigraphic drill hole would be of considerable benefit. The Winterhouse Formation is equivalent to units 4 and 5 of Schuchert and Dunbar (1934), units 4 to 7 of Sullivan (1940), and unit G of Fåhraeus (1973) (Fig. 3).

The Winterhouse Formation is fossiliferous, with several horizons containing receptaculitids, brachiopods, nautiloids, rugose corals, gastropods, conodonts, graptolites, and trilobites.

CLAM BANK FORMATION–LONG POINT GROUP BOUNDARY

We have not examined the boundary problem in this study and, therefore, agree with Fåhraeus (1973, p. 1824), who placed the discontinuity at the top of the crossbedded, pebbly gravel conglomerate.

ACKNOWLEDGMENTS

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Late Ordovician stratigraphy, paleoecology, and sea level changes in the Argentine Precordillera

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Sanchez, T.M., Benedetto, J.L., and Brussa, E., Late Ordovician stratigraphy, paleoecology, and sea level changes in the Argentine Precordillera; in Advances in Ordovician Geology, C.R. Barnes and S.H. Williams (ed.), Geological Survey of Canada, Paper 90-9, p. 245-258, 1991.

Abstract

In the Precordillera basin, Upper Ordovician rocks are exposed in the Sierra de Villicum, Cerro del Fuerte, and Talacasto areas. In the Sierra de Villicum, the facies sequence is interpreted in terms of a main lowering of the sea level during the Ashgill followed by a rise in the Hirnantian (persculptus *Zone). Lowstand deposits are represented by debris flow and channel fill conglomerates, whereas those of highstand include fine grained bioturbated facies. The base of the transgressive sequence contains a conglomeratic lag horizon, succeeded upward by mudstone containing a low diversity* Hirnantia *Fauna. A shallow water community characterized by the brachiopods* Hirnantia sagittifera *and* Dalmanella testudinaria *and by the bivalve* Modiolopsis cuyana *is recognized. The rise in sea level is documented in the Cerro del Fuerte section by deep shelf muddy deposits bearing* Anisopleurella *sp. cf.* A. gracilis *and in the Talacasto region by a transgressive sequence containing G.* persculptus *near the base.*

The fluctuations in sea level recorded in the Precordillera basin are synchronous with glacioeustatic sea level changes recognized worldwide.

The Ashgill debris flow deposits exposed in Subandean Ranges of northwestern Argentina are probably related to the same glacioeustatic event.

Résumé

Dans le bassin de la Précordillère, des roches de l'Ordovicien supérieur affleurent à proximité de la Sierra de Villicum, de Cerro del Fuerte et de Talacasto. Dans la Sierra de Villicum, la séquence de faciès est interprétée en fonction d'une chute importante du niveau marin à l'Ashgillien, suivie d'une montée à l'Hirnantien (zone à persculptus*). Les dépôts de bas niveau se composent de coulées de débris et de conglomérats de chenaux, tandis que les dépôts de haut niveau comprennent des faciès bioturbés à grain fin. La base de la séquence transgressive contient un horizon composé d'un résidu conglomératique de déflation, suivi vers le haut d'une pélite qui renferme une faune à* Hirnantia *peu variée. On y trouve une communauté d'eau peu profonde, caractérisée par la présence des brachiopodes* Hirnantia sagittifera *et* Dalmanella testudinaria *et par le bivalve* Modiolopsis cuyana. *La présence, dans la coupe de Cerro del Fuerte, de dépôts boueux de plate-forme profonde dans lesquels se trouve* Anisopleurella *sp. cf.* A. gracilis *et, près de Talasco, d'une séquence transgressive qui contient* G. persculptus *près de sa base, témoigne de la montée du niveau marin.*

Les fluctuations du niveau marin qui sont enregistrées dans le bassin de la Précordillère sont synchrones avec les fluctuations glacio-eustatiques qui se rencontrent partout au monde.

Les coulées de débris de l'Ashgillien qui affleurent dans les chaînons subandins, dans le nord-ouest de l'Argentine, sont vraisemblablement associées au même événement glacio-eustatique.

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INTRODUCTION

The Precordillera is a mountain chain about 500 km long and 80 to 100 km wide composed of a thick succession of Cambrian to Devonian marine rocks and Carboniferous to Triassic continental strata. The lower Paleozoic marine sequences were deposited on the western Gondwana continental margin and their study is essential for understanding the evolution of this margin (Aceñolaza and Baldis, 1986).

A major objective of this study is to examine the vertical and lateral facies change in Caradoc to Ashgill sequences of this region and to establish their relationship with the global sea level changes. The environmental interpretation is based both on the facies analysis and on the paleoecology of the benthic associations. Regional interpretation includes the biostratigraphic correlation of the Upper Ordovician units and the analysis of the basal stratigraphic gap.

Upper Ordovician rocks outcrop on the eastern flank of the Precordillera, between 29° and 33°S. For the purpose of this paper, three localities are discussed: the Sierra de Villicum, Cerro del Fuerte, and Talacasto (Fig. 1).

The Sierra de Villicum section is particularly important because the Ordovician-Silurian boundary beds are completely exposed and the Ashgill beds are very fossiliferous, containing brachiopods belonging to the *Hirnantia* Fauna (Levy and Nullo, 1974; Benedetto, 1986), trilobites (Baldis and Blasco, 1975), bivalves (Sánchez, in press), and graptolites. In the Cerro del Fuerte area, Ashgill beds are exposed in the Cerrito Puntudo section and consist of graptolitic mudstone containing a scattered brachiopod fauna dominated by the genus *Anisopleurella* (Benedetto, 1987), bivalves (Sánchez, in press), and graptolites (Brussa, 1987; Brussa et al., 1988). In the Talacasto section, Ashgill beds are represented by the lowermost part of the La Chilca Formation, which contains a graptolitic fauna belonging to the *G. persculptus* Zone, comparable in age to those of Villicum and El Fuerte (Cuerda et al., 1988) (Fig. 2).

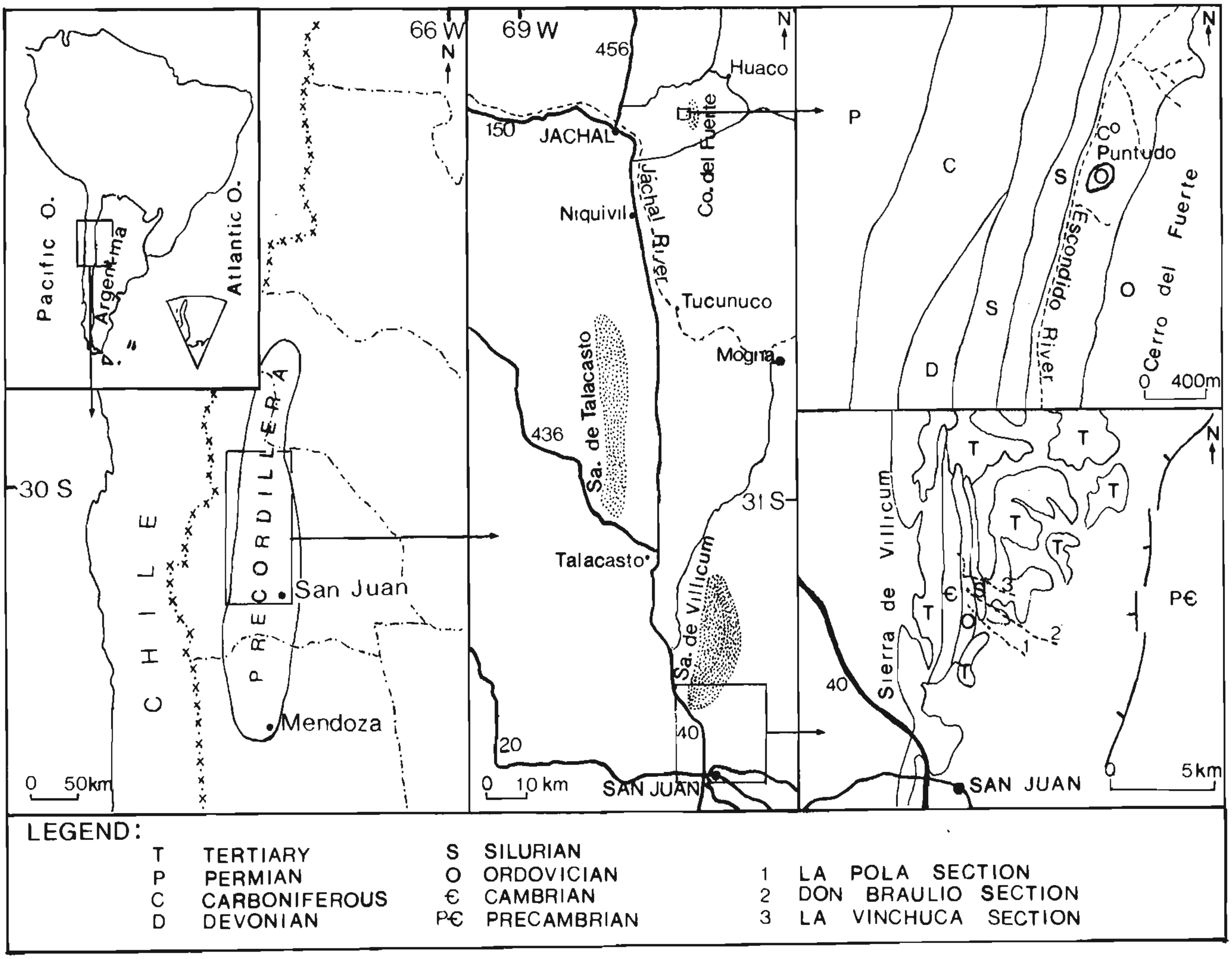

Figure 1. General locality maps and simplified geological maps of the Cerro del Fuerte and Sierra de Villicum areas.

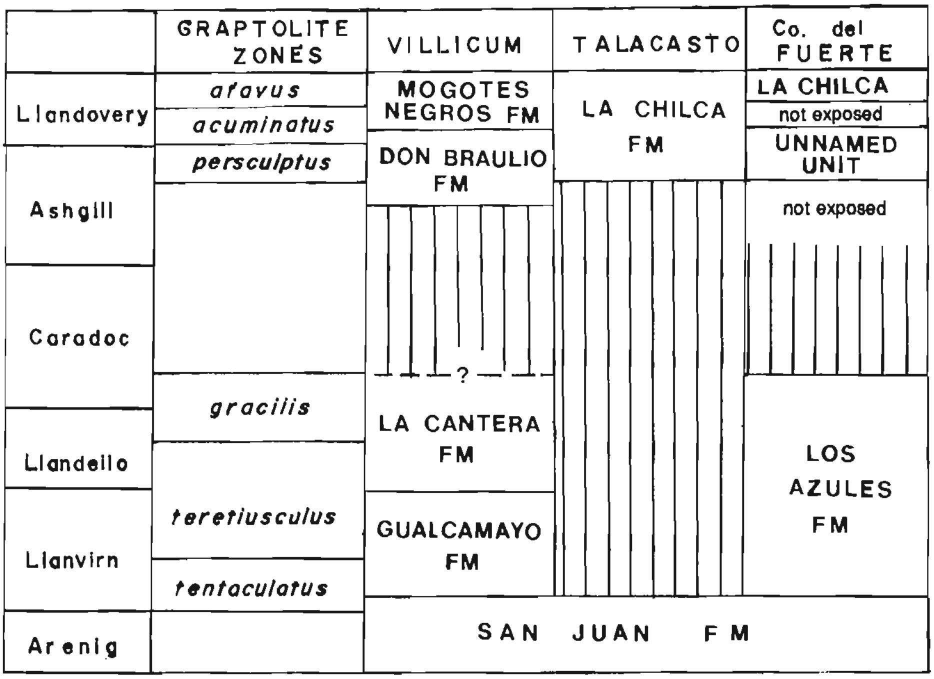

Figure 2. Correlation chart of the Ordovician units exposed at the three localities discussed in the text.

SIERRA DE VILLICUM AREA

Stratigraphy and facies analysis

In the Sierra de Villicum area, a nearly continuous succession of Lower Cambrian to Silurian rocks is exposed. Baldis et al. (1982) divided the Ordovician outcrop sequence into four units: San Juan, Gualcamayo, La Cantera, and Don Braulio formations, in ascending order. For the purpose of this study, only the last two named formations will be considered. Their type section is the Quebrada Don Braulio, on the eastern flank of the Sierra de Villicum (Fig. 1).

The La Cantera Formation consists of a 160 m thickness of turbidite sediments, including a variety of coarse to fine grained facies. The lower part of this unit is characterized by conglomeratic deposits, 20 to 30 m thick, composed of disorganized, clast-supported, coarse gravels. This facies has an erosional basal contact with the underlying graptolitic black shale of the Gualcamayo Formation (Llanvirnian). Channel-fill conglomerates are overlain by sandy facies represented by thick- to medium-bedded massive sandstone and cross-stratified, coarse grained sandstone. The upper 100 m of the sequence is dominated by fine grained turbidites (including facies C and D of Mutti and Ricci Lucchi, 1975).

Astini (work in progress) has recognized three main facies associations in La Cantera Formation: channel-fill deposits, channel-lobe transition deposits, and interchannel overbank deposits. He interpreted the facies sequence in terms of a rapid relative lowering of sea level, which can be related to the Guandacol orogenic phase of late Llanvirn-early Llandeilo age, followed by a cessation of the fan system during the late Llandeilo-early Caradoc, which could be produced by a slow rising of sea level in combination with a period of tectonic inactivity.

The Don Braulio Formation, which overlies the La Cantera Formation, consists of conglomeratic, sandy, and muddy facies. The thickness in the three sections measured (the Don Braulio, La Pola, and La Vinchuca sections) ranges from 50 to 60 m (Fig. 3). Four main facies have been recognized in this unit. Complete descriptions were given by Astini (1988).

A. *Pebbly mudstone facies.* This facies is characteristic of the lower third of the unit and consists of massive pebbly mudstone with clasts ranging from 2-50 cm in diameter, composed mainly of Lower Ordovician limestone, sandstone, and quartz. This facies contains up to 20% pebbles floating in silty or muddy matrix, which contains many reworked bryozoan colonies and shelly debris. This facies corresponds to A1.3 (disorganized gravelly muds) of Pickering et al. (1986).

B. *Channel-fill conglomerate and sandstone.* In the type section this facies consists of matrix supported conglomerate, 2 m thick, which passes upward into coarse grained sandstone. These conglomeratic beds have not been found laterally beyond distances of 100-200 m because of their channel geometry (Fig. 3). Basal pebbly sandstones rest on a sharp erosive contact on the debrites of underlying facies. They consist of well rounded clasts, mainly matrix supported, with 15-20% fine to coarse grained sandstone matrix. Most of these clasts are in the 5-8 cm size range and are composed of green to grey fine grained sandstone and rare quartz. The conglomerates lack internal organization. This coarse grained facies is well developed in the

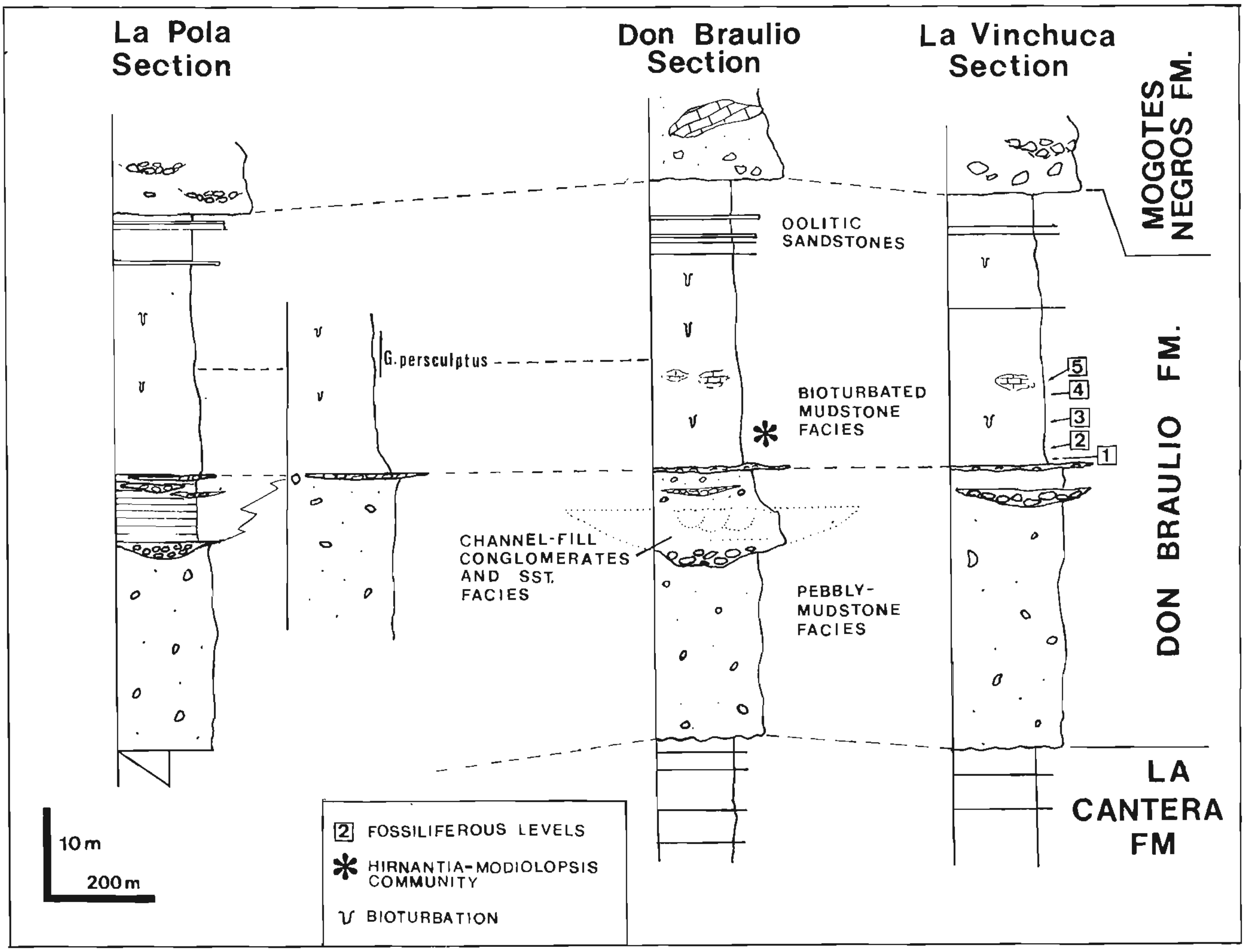

Figure 3. Facies sequence of the Don Braulio Formation and correlation of the measured sections.

axial part of the channel. Toward the top and laterally, the pebbly sandstones pass into amalgamated coarse grained sandstone. In these beds, internal sedimentary structures are diffuse, but parallel and medium-scale trough crossbedding may occur. Lenticular, clast supported conglomeratic beds are fairly common within this facies. These small-scale channels are commonly 2-10 m wide and 0.2-0.75 m deep and are composed of well sorted and rounded clasts of quartz (50%), quartzite (45%), and limestone (5%).

Another type of channel deposit, commonly associated with the pebbly mudstone facies, consists of small lenticular bodies of fine grained sandstone reworked by oscillation ripples that develop interference patterns near the top.

C. *Bioturbated fossiliferous mudstones.* This facies is characteristic of the middle part of the Don Braulio Formation. The thickness ranges from 5 to 8 m. It consists of extensively bioturbated green to grey muddy siltstone interbedded with horizons of calcareous mudstone containing a well preserved shelly fauna and graptolites. Brachiopods, trilobites, and bivalves occur both as mass concentrations or are uniformly distributed within the mudstone.

Near the boundary between this facies and the underlying pebbly mudstone, a discontinuous horizon formed by lenticular patches of conglomerate, 10-20 cm thick, is developed. It consists of mudstone rip-up clasts, quartz and sandstone pebbles averaging 2 cm in diameter, and many reworked shells of brachiopods (mainly *Hirnantia* and *Dalmanella*), pelecypods, and crinoid columnals.

D. Y*ellowish mudstone and interbedded oolitic sandstone.* The upper part of the Don Braulio Formation is characterized by massive, yellowish green mudstone that has been totally bioturbated (*Skolithos* is a common trace fossil), interbedded with sharp based, fine to medium grained hematitic sandstone beds up to 20 cm thick. Sandstone layers tend to be continuous at outcrop scale and are arranged in sequences about 1 m thick. These horizons are composed mainly of hematitic oolites and rounded clasts of chert (Peralta, 1986a). No internal structures are recognized. The interbedded dark grey shales contain inarticulate brachiopods and graptolites.

Interpretation

The mud supported conglomerates of the lower part of the Don Braulio Formation represent deposition from cohesive debris flows. The muddy matrix contains reworked bryozoan colonies and brachiopod shells that may have been derived from adjacent shelf areas. The limestone pebbles, on the other hand, could be derived from previously consolidated carbonate sediments equivalent in age to turbidite deposits of La Cantera Formation, although fossil evidence is not conclusive.

The channelized conglomerate and sandstone capping the first debris flow sequence are interpreted as mixed channel-fill deposits (Mutti and Normark, 1987). The erosional stage is indicated by their sharp basal contact cutting deeply into the underlying pebbly mudstones. The conglomerates that infill the axial part of the channel have a low percentage of sandy matrix. They are interpreted as residual conglomeratic deposits. Beds of trough crossbedded or diffuse planar-laminated sandstone overlie the channel-lag deposits or rest directly on the debris flow facies where the conglomerates pinch out toward the channel margins. These sandy deposits, Facies B2.2 of Pickering et al. (1986) accumulated by migration of sinuous crested bedforms within the channel. They may represent the depositional period of the channel-fill sequence. The presence of small-scale, shallow, lenticular bodies of clast supported conglomerate indicates that active channels developed within the main channel. In the La Pola section (Fig. 3), similar channel-fill deposits are associated with laminated green mudstone and rip-up clasts of the same lithotype. They probably represent channel related overbank deposits. The presence of interfering ripple systems at the top of some channelized bodies and capped wave ripples suggests very shallow water alternating with periods of subaerial submergence.

An abrupt change in the sedimentation regime occurs above the debris flow deposits and the interbedded channelized bodies: both facies pass upward into a dominantly fine grained succession with a sharp base marked by a conglomeratic lag horizon with mudstone intraclasts and fossil debris. It is laterally extensive and is useful for correlation between the measured sections (Fig. 3). This conglomeratic bed infills depressions eroded into underlying beds. The underlying mudstones must have been semilithified by the time this horizon had formed, because the conglomerates contain mudstone intraclasts and pebbles of clast supported conglomerate. Pebbles of quartzite and sandstone could have been supplied to the shoreline by rivers. These conglomerates can be interpreted as transgressive lag deposits, formed by intense wave action in a shallow water environment. The muddy facies developed above this horizon is interpreted as the product of the slow deposition of suspended muds, below the storm wave base. This environment was inhabited by a diverse benthic fauna, which produced a strong biological disturbance of the sediment. Brachiopod shells are especially abundant locally, occurring in nests, which suggest that the bottom was affected by weak traction flows that clustered the local fauna. The oolitic sandstone interbedded with the bioturbated mudstone near the top of the Don Braulio Formation indicate a fully aerated shallow marine environment with high turbulence.

This unit is overlain by the Mogotes Negros Formation of Silurian age (Peralta, 1986a). It consists of a sequence, over 900 m thick, of sandstone, clast supported resedimented conglomerate, and giant calcareous olistoliths derived from the Llanvirn San Juan Formation.

Fauna and age

La Cantera Formation

According to Peralta (1986b), the basal part of this unit contains graptolites that indicate the *Glyptograptus teretiusculus* Zone. Another graptolite association occurs approximately in the upper third of the formation and contains *Dicellograptus divaricatus salopiensis* Elles and Wood, *Nemagraptus* sp. cf. *N. gracilis* (Hall) and *Dicellograptus* sp. (Peralta, 1986b). In the fine grained turbidite beds exposed in La Pola section, about 30 m below the top, we have found the following association: *Pseudoclimacograptus sharenbergi* (Lapworth), *Reteograptus geinitzianus* (Hall), *Dicranograptus* sp., *Cryptograptus* sp., *Climacograptus* sp., *Glyptograptus* sp., *Dicellograptus* sp., and *Thamnograptus* sp. This fauna is indicative of a late Llandeilo-early Caradoc age. The abundance of dicellograptids, the first appearance of the genus *Dicranograptus* and the presence of *P. scharenbergi*, which is a common form in the *N. gracilis* and *D. multidens* zones (early Caradoc) in northwestern Europe and North America, indicate that this association belongs to the base of the *N. gracilis* Zone.

Don Braulio Formation

In the bioturbated mudstones, a rich shelly fauna dominated by brachiopods and trilobites has been found. These fossiliferous strata, about 5 m thick, contain both in situ faunas and transported ones. The brachiopods were described by Levy and Nullo (1974) who first recognized Ashgill faunas in this region. More recently, Benedetto (1986) reported the genus *Hirnantia* and other typical representatives of the cosmopolitan *Hirnantia* Fauna such as *Dalmanella testudinaria* (Dalman), *Hirnantia sagittifera* (McCoy), *Eostropheodonta* sp. aff. *E. hirnantensis* (McCoy), *Plectothyrella* sp., and *Coolinia* sp. New collections made recently in these levels have yielded well preserved specimens of *Paromalomena polonica* (Temple) and *Cliftonia oxoplecioides plicata* (Benedetto, in press). In order to clarify the composition of this Hirnantian brachiopod association, the genera *Villicundella* and *Bagnorthis* created by Levy and Nullo (1974) are being revised by one of us (J.L.B.). This review shows that the genus *Villicundella* was erected on dorsoventrally crushed shells of *Hirnantia* and the genus *Bagnorthis* probably includes immature specimens of *H. sagittifera*.

These beds also contain a trilobite fauna dominated by *Dalmanitina sudamericana* Baldis and Blasco and *Eohomalonotus villicunensis* Baldis and Blasco (1975). Of the bivalves, Sánchez (in press) described the species *Modiolopsis cuyana* Sánchez and *Palaeoneilo* sp.

The upper levels of the mudstone facies (Fig. 3) contain a low diversity graptolite fauna dominated by *Glyptograptus persculptus* and indeterminate dendroids.

The uppermost levels of the Don Braulio Formation have yielded only graptolites and microfossils. A diversified chitinozoan association was recognized by Volkheimer et al. (1980) from the dark grey shale interbedded with the oolitic sandstone. This association, of Llandovery age, includes: *Ancyrochitina* sp. cf. *A. ancyrea* (Eisenack), *Conochitina* sp. cf. *C. chydaea* Jenkins, *Cyathochitina* sp. cf. *C. campanulaeformis* (Eisenack), *Euconochitina* sp. cf. *E. filifera* (Eisenack), *Rhabdochitina* sp., *Spathachitina* sp. cf. *S. clarindoi* da Costa, and *Sphaerochitina* sp. From the same levels, Peralta (1986a) identified the following graptolites: *Climacograptus* sp. aff. *C. hughesi* (Nicholson), *Climacograptus* sp., *Monograptus* sp., and *Rastrites* sp. This association confirms an Early Silurian age.

On the basis of the fauna, the Don Braulio Formation can be regarded as partly Hirnantian and partly Llandovery. The available evidence indicates that the Ordovician-Silurian boundary falls within the upper part of this unit (yellowish mudstones). On the other hand, a stratigraphic break is evident between the La Cantera and Don Braulio formations, which encompasses almost the entire Caradoc and the lower middle Ashgill (Fig. 2).

Paleoecology

The calcareous lenses found in the lower part of the muddy siltstone at the Don Braulio section (Fig. 3) contain a well preserved *Hirnantia* Fauna, which is dominated by articulate brachiopods (over 60%). The modiomorphoid pelecypod *Modiolopsis cuyana* is a common element and is accompanied by the trilobite *Dalmanitina*. The specimens of *Hirnantia* are mainly articulated and well preserved. The high percentage of articulated shells of this species is significant, because the deltidiodont type of hinge facilitates the dissociation of the valves (Sheehan, 1978; Sánchez, 1986). The abundance of articulated shells indicates that they are untransported shells. The specimens of *Modiolopsis* are frequently found as articulated valves, some of which are in life position. Shells of various size are associated in the samples, which suggests the coexistence of adult and juvenile individuals of a normal population. Several articulated specimens of *Dalmanitina* provide additional evidence for an untransported and rapidly buried assemblage. We, therefore, regard this association as a fossil community. The term community refers to "the biological association of taxa usually related to particular environmental conditions" (Brenchley and Cocks, 1982, p. 786). The *Hirnantia-Modiolopsis* Community is characterized by the epifaunal suspension feeder brachiopods *Hirnantia sagittifera* and *Dalmanella testudinaria* and by the semi-infaunal, byssate, suspension feeder pelecypod *Modiolopsis cuyana*. This is a low diversity *Hirnantia* Association (Rong, 1985) with seven genera, three of which are dominant (Table 1; Fig. 4).

The *Trematis*-Bivalve Assemblage, described by Brenchley and Cocks (1982), and this community are similar in the abundance of modiolopsid pelecypods, but differ in the occurrence of the inarticulate brachiopod *Trematis norvegica* and the relative reduction of *Hirnantia* in the former. Brenchley and Cocks (1982) suggest a lower shoreface environment for the *Trematis*-Bivalve Association.

According to the environmental distribution of pelecypods (especially modiomorphoids) during the Ordovician (Jablonski et al., 1983), the relatively high percentage of *Modiolopsis* in the community suggests a shallow water environment. The associated brachiopods *Hirnantia* and

Table 1. Composition of the *Hirnantia-Modiolopsis* Community (Don Braulio Formation; Sierra de Villicum)

	Articulated valves	Isolated valves	Total	Life habits[1]
Brachiopods				
Hirnantia sagittifera	32	29	71	ep.s.
Dalmanella testudinaria	51	45	96	ep.s.
Paromalomena polonica	-	23	23	ep.s.
Lingulacea indet.	5	-	5	in.s.
Pelecypods				
Modiolopsis cuyana	45	36	81	se.s.
Trilobites				
Dalmanitina sudamericana	8	19[2]	27	ep.
Encrusting trepostome bryozoa scarce				

[1] ep: epifaunal; in.: infaunal; se.: semi-endobyssate; s.: suspension feeder.
[2] Includes isolated cephala and pygidia.

Dalmanella are not indicative because they probably were eurytopic genera (Brenchley and Cullen, 1984). In the Oslo region, for example, the *Dalmanella* Association described by Brenchley and Cocks (1982) was developed in a deep water environment. This genus, however, flourished in the Middle Ordovician shallow waters (Cocks and McKerrow, 1978).

The low diversity of this community could be the result of a fluctuating resource supply. According to the paleogeographic reconstruction proposed by Ziegler et al. (1977) and to the extension of the floating ice suggested by Brenchley and Newall (1984), this region has experienced a cold circumpolar climate. Under these climatic conditions there is a rich resource supply in summer, followed by a rapid decay of resources in winter. Consequently, only some eurytopic and opportunistic species can survive (Valentine, 1971).

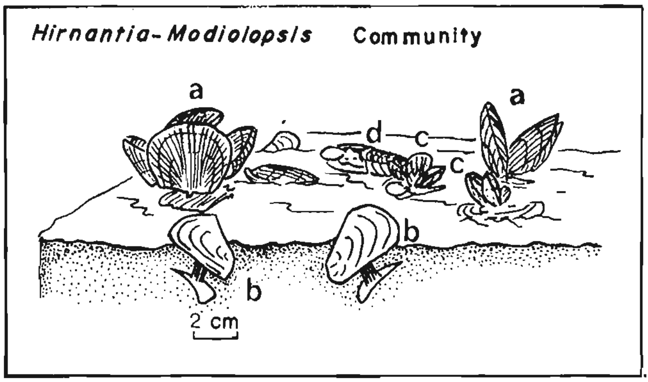

Figure 4. Reconstruction of the *Hirnantia-Modiolopsis* Community. a, *Hirnantia sagittifera*; b, *Modiolopsis cuyana*; c, *Dalmanella testudinaria*; d, *Dalmanitina sudamericana*.

In the same section, immediately above the *Hirnantia-Modiolopsis* Community, an assemblage of dispersed, locally clustered specimens of *Eostropheodonta* sp. cf. *E. hirnantensis, Plectothyrella* sp., *Eohomalonotus villicunensis, Palaeoneilo* sp., and conulariids are found (Benedetto, 1986).

In the La Vinchuca section, 500 m north, a 5 m thick sequence of fine grained sandstone and siltstone contains a *Hirnantia* Fauna represented by several assemblages. We have found five levels that appear to be distinguishable by relative proportions of genera (Table 2). Level 1 contains very transported specimens of *Dalmanella* and *Hirnantia*, which are the commonest genera, and rare *Paromalomena, Cliftonia,* and *Modiolopsis.* At level 2, we have collected a few specimens of *Dalmanella* and *Palaeoneilo* and sclerites of *Dalmanitina* and *Eohomalonotus.* Level 3 contains abundant sclerites of *Dalmanitina* and *Eohomalonotus* and a few valves of *Paromalomena.* At level 4, the assemblage includes isolated valves of *Paromalomena, Dalmanella,* and *Hirnantia.* At this horizon *Paromalomena* is the commonest genus, whereas *Dalmanella* and *Hirnantia* are rare. At the uppermost level (level 5), *Eostropheodonta* is the most abundant genus, whereas *Hirnantia* and *Paromalomena* are relatively scarce. We have not found at this locality the *Hirnantia-Modiolopsis* Community, which is apparently restricted to one horizon of the Don Braulio section.

There is undoubtedly considerable fluctuation within both a small area and a thin section in the relative proportions of the genera. This could be the result of normal fluctuations in the fauna belonging to the same association. However, it is important to note the replacement of *Hirnantia* by *Eostropheodonta* at the top in both sections and the progressive decrease both in number and diversity of the fauna (see Table 2), followed by the disappearance of the brachiopods at the graptolitic horizon. This fact is consistent with the decline and extinction of the fauna at the end of Hirnantian, recognized by Brenchley (1984), Brenchley and Newall (1984), and Sheehan (1988).

Table 2. Variation[1] of genera at the sections of Sierra de Villicum

	La Vinchuca Section					Don Braulio Section	
	Levels						
	1	2	3	4	5	*H.-M.* Comm.[2]	Levels above
Hirnantia	A	-	-	S	S	A	-
Dalmanella	A	C	-	S	-	A	A
Paromalomena	S	-	S	A	S	S	-
Eostropheodonta	-	-	-	-	A	-	A
Plectothyrella	-	-	-	-	-	-	C
Coolinia	-	-	-	-	-	-	S
Cliftonia	S	-	-	-	-	-	-
Modiolopsis	S	-	-	-	-	A	-
Palaeoneilo	-	S	-	-	-	-	S
Dalmanitina	-	S	C	-	-	C	S
Eohomalonotus	-	S	C	-	-	-	C

[1] A, abundant; C, common; S, scarce.
[2] *H.-M.* Comm. = *Hirnantia-Modiolopsis* Community.

This upper association resembles the *Hindella-Cliftonia* Association of Brenchley and Cocks (1982) in lacking *Hirnantia* and having *Eostropheodonta* and *Coolinia*, but differs in its lower diversity (lacking gastropods, bryozoans, corals, some brachiopods, and trilobites).

We can conclude that the lower horizons in both sections containing pelecypods (*Hirnantia-Modiolopsis* Community and Level 1) represent the shallowest part of the sequence, whereas the horizons dominated by brachiopod assemblages, found above, indicate a shelf environment below wave base.

CERRO DEL FUERTE AREA

Lithological description and environmental interpretation

The Cerro del Fuerte and Cerro Viejo form a north-south trending range of mountains located to the northeast of Jachal City (Fig. 1). On their western flank an unnamed Upper Ordovician stratigraphic unit is exposed in an isolated outcrop (Cerrito Puntudo) within the Rio Escondido valley (Benedetto et al., 1986; Brussa, 1987). Its basal and upper contacts are covered by recent alluvial sediments. This sequence is 119 m thick. The exposed basal beds consist of 3.5 m of grey, medium grained sandstone interbedded with lenticular, clast supported, polymictic conglomerate up to 20 cm thick, containing well rounded clasts of quartz and quartzite. They grade into a homogeneous succession of bioturbated, light green, silty mudstone, which locally contains brachiopods, molluscs, and graptolites. Sedimentary features include extensive bioturbation, with abundant oblique to subhorizontal cylindrical burrows and ferruginous concretions. The middle-upper part of the unnamed unit consists of bluish grey shale and sandy shale. These beds are poorly fossiliferous, containing only scattered remains of graptolites. They are overlain in apparent conformity by the La Chilca Formation, although the transition between the units is not exposed. The basal strata of the La Chilca Formation are composed of light blue, thin bedded, fine grained sandstone, which have yielded a well preserved graptolite fauna of early Llandovery age (Brussa, 1987; Brussa et al., 1988).

We interpret the fine grained facies of the unnamed unit as the deposits of a muddy shelf below the storm wave base. The lowermost sandstone-conglomerate facies can be interpreted as transgressive lag deposits, which originated in a high energy, shallow water environment.

Unfortunately, the stratigraphic relations at the base of the unnamed unit are unknown. However, black shale and marl containing a rich graptolite fauna of Llanvirn to early Caradoc age (Cuerda and Furque, 1975; Ortega, 1987), named the Los Azules Formation, outcrop about 4 km north of the Cerrito Puntudo section. Since no other Ordovician units are known in this region, the simplest interpretation of this boundary is that the Upper Ordovician sequence rests unconformably on the Los Azules Formation. Thus, the hiatus between these units would be the same as that in the Sierra de Villicum between the La Cantera and Don Braulio formations.

Fauna and age

The siltstone and mudstone that outcrop in the Cerrito Puntudo contain *Glyptograptus persculptus* (Salter). This graptolite zone is developed throughout the exposed sequence. The lower part of the unnamed unit contains a very sparse shelly fauna, containing *Anisopleurella* sp. cf. *A. gracilis* (Jones) and *Reuschella* sp. (Benedetto et al., 1986; Benedetto, 1987) and the bivalves *Costaledopsis fuertensis* Sánchez, *Whiteavesia* sp., and *Colpomya?* sp. (Sánchez, in press).

The *Atavograptus atavus* Zone is developed within the lowest 3.90 m of La Chilca Formation. *Atavograptus atavus* (Jones) appearance is placed at the first 0.50 m and *Climacograptus normalis* (Lapworth), *Paraclimacograptus innotatus brasiliensis* (Ruedemann), *Climacograptus*

Table 3. The composition of the assemblage at the unnamed unit (Cerro del Fuerte)

	Articulated valves	Isolated valves	Total	Measures[1]		Life habits[2]
				max.	min.	
Brachiopods[1]						
Anisopleurella sp. cf. *A. gracilis*	2	25	27	12	4	ep.s.
Reuschella sp.	1	14	15	17	4	ep.s.
Lingulacea indet.	-	3	3	3.5	2	in.s.
Pelecypods[1]						
Costaledopsis fuertensis	1	2	3	5	1.5	in.d.
Whiteavesia sp.	2	-	2	10	-	end.s.
Colpomya ? sp.	1	-	1	3.5	-	in.s.
Nautiloids (rare fragments)						
Conulariids: one specimen						

[1] Measurements in mm; in brachiopods, wide; in pelecypods, high.
[2] ep., epifaunal; in., infaunal; end., endobyssate; s., suspension-feeder; d., deposit-feeder.

sp. cf. *C. rectangularis* (McCoy), *Pseudoclimacograptus (Clinoclimacograptus)* n. sp., and *Diplograptus* sp. are found above (Brussa et al., 1988). It is supposed that the covered 200 m represent an interval between the unnamed unit and the base of the La Chilca Formation, and fall within the *acuminatus* Zone.

Paleoecology

The fossil assemblage described here occurs in the silty mudstone sequence. Fossils are disseminated throughout bioturbated rocks, and they occur individually rather than in clusters. The scarce fauna includes articulate and inarticulate brachiopods, pelecypods, graptolites, rare orthocone nautiloids, and one specimen of a conulariid. The orthocones occur as fragments, but brachiopods and pelecypods are represented by complete and often articulated shells. Disturbance of the shells prior to fossilization could have been the result of the activity of burrowing organisms, as is suggested by the abundant horizontal burrows. We regard this assemblage as essentially untransported, but because the fauna is very sparse we have considered it as an "assemblage" and not as a community. This fauna is presented in Table 3, which shows the commonest species, their relative size, and their life habit.

The assemblage is dominated by the epifaunal, suspension feeders *Anisopleurella* and *Reuschella.* In contrast, pelecypods are much less abundant. Both infaunal suspension feeders (*Whiteavesia* and *Colpomya?*) and infaunal deposit feeders (*Costaledopsis*) occur, but they are a minor component of the assemblage. Therefore, the suspension feeder type dominates. This suggests a firm substrate, clear water environment.

The pelecypods have a tendency toward small individuals and a low population density that might have been the result of decreased nutrient supply in deep waters. It is important to note the absence of the common Ordovician genus *Modiolopsis.* This can be the result of an inadequate environment (such as deep water).

The specimens of *Reuschella* are relatively smaller than the typical representatives of the genus (Benedetto, 1987), whereas those of *Anisopleurella* are large-sized individuals (cf. *A. gracilis* (Jones) in Cocks, 1970). *Reuschella* was reported as a common component in the Dolhir Formation of North Wales, which was deposited in a shallow shelf environment beyond the influence of wave action (Hiller, 1981). The smaller size of individuals of the Argentine species might indicate abnormal conditions for them, such as a deeper environment. *Anisopleurella* is a common genus of an Upper Ordovician association equivalent to Llandovery deep water *Clorinda* Community of Ziegler et al. (1968) (Cocks, 1970). This could be an explanation for their size not being affected or diminished. The evidence suggests a resource poor, deep shelf environment.

TALACASTO AREA

Lithological description and environmental interpretation

A fossiliferous sequence, 25 m thick, of Late Ordovician-Early Silurian age is exposed along the southern margin of the Talacasto River, approximately 10 km west of the Talacasto locality (Fig. 1). It corresponds to the lower part of the La Chilca Formation, a stratigraphic unit that usually reaches 80 to 100 m in thickness. The sequence rests unconformably on the Llanvirn limestones of the San Juan Formation (Fig. 2 and Fig. 5, column 1). The basal part consists of a monomictic, clast supported conglomerate up to 50 cm thick, composed of well rounded clasts of chert. It is overlain by 5 m of bioturbated, light green, graptolitic shale, which pass upward into fine grained, sharp based quartzitic sandstone, 10 to 20 cm thick, separated by shale partings. Traces on the top of sandstone beds are dominated by *Chondrites.* The upper 15 m exposed are composed predominantly of shale. Near the top there is a 90 cm thick, prominent bed of hematitic sandstone.

A model of deposition of the La Chilca Formation, based on facies analysis of a section outcropping 30 km north of Talacasto, has been proposed recently by Astini and Piovano (in press), who suggest an inner shelf to offshore transition, wave dominated and offshore environment. The facies succession exposed in the Talacasto area could have been formed in a muddy shelf below the mean wave base. The sandstone interbeds, however, indicate a periodic influence of low regime traction flows. The ferruginous beds were interpreted by Astini and Piovano (in press) as having been formed of resedimented oolites.

Fauna and age

The graptolites of this section were described by Cuerda et al. (1988). The fauna occurs in the lower 3.65 m of the La Chilca Formation. *Glyptograptus persculptus* (Salter) was only found in the lowermost 55 cm, associated with *Climacograptus angustus* Perner, *C. normalis* Lapworth, *C.* sp. cf. *C. medius* Tornquist, *Glyptograptus* sp., and *Pseudoclimacograptus (?Metaclimacograptus) robustus* Cuerda, Rickards, and Cingolani. The upper beds (from 0.6 to 1.75 m) are dominated by *Pseudoclimacograptus (?Metaclimacograptus) robustus* and also yield *C. acceptus* (Koren and Mikhailova), *C. angustus, C.* sp. cf. *C. medius, C. normalis,* and *Rhaphidograptus* sp., which would belong to a probable *acuminatus* Zone (Cuerda et al., 1988).

SEA LEVEL CHANGES IN THE PRECORDILLERA BASIN

The facies sequence present in the Upper Ordovician rocks of the Sierra de Villicum can be interpreted in terms of a relative lowering of sea level to account for the hiatus and facies change between the La Cantera and Don Braulio formations. The timing of this episode of low sea level cannot be accurately established because of the lack of fossils in the sediments immediately below the debris flow. On the basis of the graptolites recovered in the upper third of the La Cantera Formation, the hiatus spans at least the late Caradoc and the early Ashgill.

It seems likely that the eastern margin of the Precordillera basin was deeply notched and partially eroded at this time. The upper Ashgill strata were characterized by an influx of

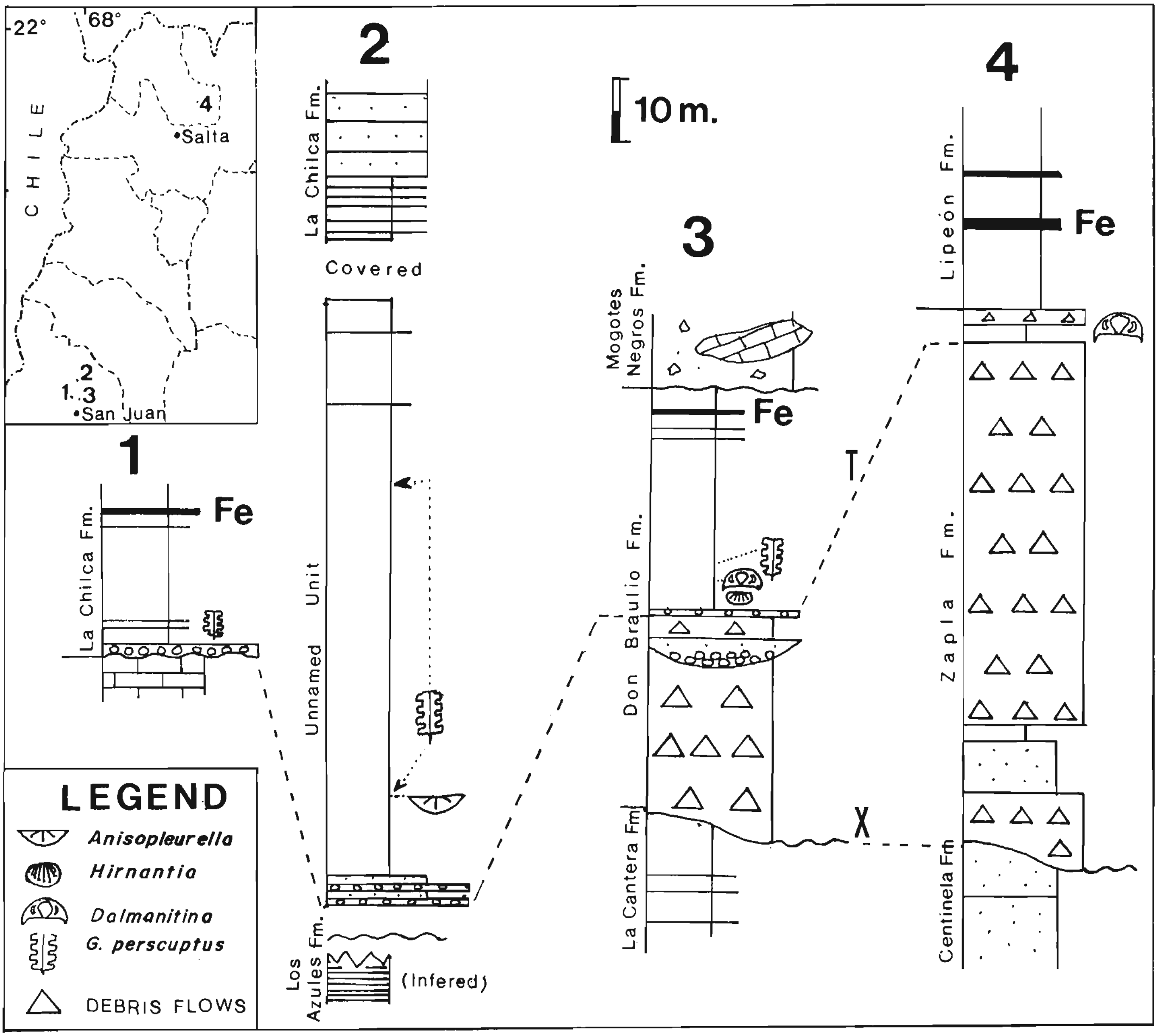

Figure 5. Correlation between the Precordillera sections (1, 2, 3) and Subandean Range section (4). 1, Talacasto section; 2, Cerro del Fuerte section; 3, Sierra de Villicum generalized section; 4, Sierra de Santa Bárbara section, Subandean Ranges (after Monaldi and Boso, 1987). X, stratigraphic break and erosion surface; T, base of the transgressive sequence; Fe, ferruginous horizons. The sections 2, 3, and 4 correspond to columns 15, 18, and 1, respectively, of the IUGS correlation chart (Aceñolaza and Baldis, 1986).

gravity flow sediments (base of the Don Braulio Formation), which might have been deposited in these large-scale erosional features. The presence of abundant fragments of bryozoans and brachiopods in the silty matrix of the debris flow indicates that these sediments were mainly derived from the contiguous shelf edge. Whether these deposits (facies A) were the product of a rapid lowering of the sea level or were primarily controlled by tectonic activity cannot be determined.

However, a short regressive episode during the Himantian is indicated by the change from the pebbly mudstone facies (A) to the shallower channel-fill facies (B). At this time, minor channels incised on the gravity flow sediments were infilled with shallow water sandstone and conglomerate (facies B). Although these coarse grained deposits (which were considered by Baldis et al., 1982, as the base of the Don Braulio Formation) were interpreted to have a tectonic origin (Villicum orogenic phase of Baldis et al., 1982), it appears likely that their deposition was primarily controlled by a sea level fluctuation. In our opinion, the stratigraphic relationships between these channel deposits and the underlying debris flows do not represent an unconformity of tectonic origin, but only the infill of the channel system at the beginning of the sea level rise after the main lowstand. At this time, a shallow water biota inhabited the muddy sediments, probably in the interchannel areas.

The rapid change from these shallow water sediments to the muddy, graptolitic facies is attributed to an increase in the rate of sea level rise. The resulting deposits are fine grained, low energy marine shelf facies which, according to the paleoecological evidence, progressively became deeper toward the top. In the *acuminatus* Zone, a short period of shallowing is suggested by the oolitic sandstone facies at the top (Fig. 3).

The silty mudstone containing a deep water mixed shelly and graptolitic fauna (*persculptus* Zone) exposed in the Cerro del Fuerte area can be correlated with the mudstone belonging to the same zone in the Sierra de Villicum sections (Fig. 5). Thus, they could represent the drowning phase at the end of the Hirnantian, and the lowermost coarse grained facies could correspond to the beginning of the transgressive phase.

In the Talacasto region, the stratigraphic gap at the base of the Ashgill sequence is more extensive than at other localities, spanning from lower Llanvirn to upper Ashgill (Fig. 2). It may be related to the existence of a tectonic high, uplifted during the Guandacol orogenic phase (upper Llanvirn-Llandeilo), which was not covered until the upper Hirnantian-Rhuddanian rise in sea level.

In summary, the evidence from the Upper Ordovician sequences of the Precordillera basin indicates a relative lowering of sea level followed by a transgressive event that started at the latest Hirnantian and reached a maximum in the *persculptus* Zone.

The timing of these events in the Precordillera basin shows a strong correlation with the glacioeustatic sea level changes recognized in many regions of the globe (Berry and Boucot, 1973; Sheehan, 1973; McKerrow, 1979; Brenchley and Cullen, 1984; Brenchley and Newall, 1980, 1984; Chen, 1984; Woodcock and Smallwood, 1987).

EVIDENCE OF LATE ORDOVICIAN SEA LEVEL CHANGES IN SOUTH AMERICA

Outside the Precordillera basin, Upper Ordovician rocks have been recently reported from the Sierras Subandinas in northwestern Argentina (Monaldi and Boso, 1987) (Fig. 6). Fossils were found in the upper part of the Zapla Formation. They include bivalves and trilobites of the genus *Dalmanitina (D. subandina* Monaldi and Boso). The Zapla Formation is principally composed of dark grey, massive pebbly mudstone, 80 to 100 m thick, with some intercalations of conglomeratic sandstone near the base. This unit paraconcordantly overlies the Caradoc(?) Centinela Formation and in turn passes gradationally into the Lipeón Formation. The latter starts with two prominent ferruginous beds, each up to 4 m in thickness, composed mainly of hematitic sandstone, which show crossbedding, lenticular bedding, and flaser bedding. These beds pass upward into a 650 m thick monotonous sequence of fine grained sandstone and siltstone.

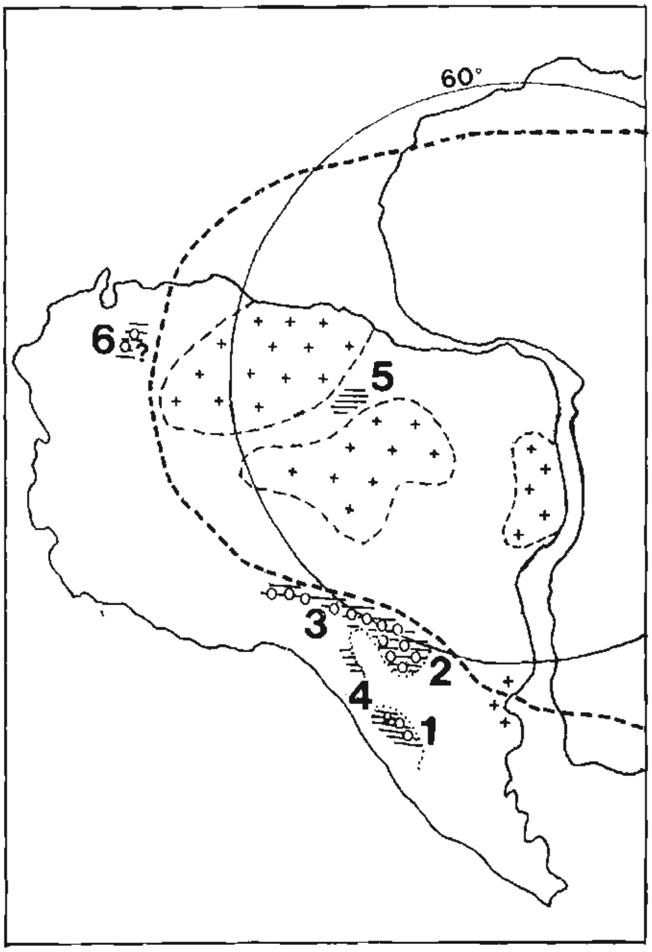

Figure 6. Geographic distribution of Ashgill lowstand deposits (open circles) and transgressive sequences of early Llandovery age (horizontal lines) on the western Gondwana landmass. Dashed line = estimated limit of Upper Ordovician ice cap (after Brenchley and Newall, 1984). 1, Precordillera basin; 2, Subandean Ranges; 3, Bolivia-Perú basin; 4, Puna region; 5, Amazonas basin; 6, Venezuelan Andes.

As Monaldi and Boso (1987) proposed, the *Dalmanitina* beds of the Zapla Formation are correlated with the ones that contain *Hirnantia* and *Dalmanitina* in the Precordillera basin (Don Braulio Formation). From this correlation we can infer that the diamictites of the Zapla Formation are equivalent in age to the debris flow deposits of the lower part of the Don Braulio Formation. The ferruginous beds of the lowest part of the Lipeón Formation in turn should be equivalent to the upper part of the Don Braulio Formation (Fig. 5).

The diamictites of the Zapla Formation have been classically interpreted as glacial deposits (Berry and Boucot, 1972). More recently, they were considered as gravity flow sediments that accumulated in an unstable basin (Antelo, 1978). On the other hand, a number of authors have regarded this coarse grained facies as being related to a Late Ordovician tectonic event named the Ocloyica Phase (Turner and Mendez, 1975; Moya and Salfity, 1982; Monaldi and Boso, 1987). According to the evidence presented in this paper relating to sea level fluctuations in the Precordillera basin, and the correlation between the northwestern and Precordilleran Upper Ordovician units, we infer that the Ashgill diamictites of the Zapla Formation may also be of glacioeustatic origin. Similarly, the transgressive nature of the Silurian sequence (Lipeón Formation) in the northwestern basin could reflect the sea level rise recognized in the Precordillera basin.

Deposits equivalent in facies and almost synchronous to those of the Zapla Formation have been recorded in Bolivia (Cancañiri and Sacta formations) and southern Perú, covering a 1500 km long and 200 km wide area along the western margin of Gondwana. Likewise, a stratigraphic lacuna just prior to the debris flow deposits is developed throughout this belt. Similar relationships are present in the Andes of Venezuela (Benedetto and Ramirez Puig, 1982). Here, the sequence consists of muddy limestone containing brachiopods, trilobites, and graptolites of Caradoc age (Hughes, 1980) capped by debris flow deposits. Lower Llandovery rocks outcrop in the same area (Boucot et al., 1972), but their stratigraphic relationships with the Ordovician unit are not clear.

As a basis for understanding the patterns of distribution of the Ashgill-lower Llandovery rocks, we have plotted on a paleogeographical map the data discussed in this paper. It shows that the effects of the global sea level fall would be recognized along the entire Paleo-Pacific margin of South America (Fig. 6).

The stratigraphic evidence of the subsequent rise in sea level related to the melting of the ice cap is particularly evident on the Brazilian Shield. In the Amazonas basin the first Phanerozoic transgression on the Precambrian basement is represented by the Trombetas Formation. The lowermost members of this formation (Autás-Mirim and Nhamundá members) consist of sandstone and shale, which pass upward to the graptolitiferous black shale of the Pitinga member containing *Climacograptus innotatus brasiliensis* and early Llandovery chitinozoa. According to Popp (1983), the lowermost horizons of this unit must be referred to the Upper Ordovician.

In the Andean orogenic belt this transgressive episode is well documented in the Subandean ranges of northwestern Argentina and Bolivia (Lipeón Formation and equivalents) and Cordillera Oriental of Bolivia (Uncia and Huanuni formations). Particularly significant is the presence of lower Llandovery strata near the border between Argentina and Chile (Isaacson et al., 1976), because this is the first record of marine Paleozoic strata covering the underlying Lower Ordovician folded rocks in the Puna region.

Sedimentary ferruginous deposits were unusually widespread during the lower Llandovery (Fig. 5). Based on the environmental parameters controlling the origin of these deposits, Beresi (1978) has suggested a tropical-humid climate for this region. This evidence appears to be contradictory to paleogeographic reconstructions based mainly on paleomagnetic data. We consider, however, that the origin of ferruginous beds was related to the more temperate regime initiated at the end of the Hirnantian when the ice cap melted (Brenchley and Cullen, 1984).

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Paleogeography and main features of volcanicity in the Ordovician of Kazakhstan and North Tien Shan

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Abstract

A general review is given of the Ordovician paleogeography and volcanicity in Kazakhstan and North Tien Shan. Paleogeographic environments that existed between the Early and the Late Ordovician have been studied. During the Ordovician, Kazakhstan and Tien Shan were considered to have been part of a Pacific-type continental margin including a system of volcanic and nonvolcanic (relict) island arcs, inter-arc basins with cherty-clastic sedimentation, and back-arc basins with successively developing spreading zones. The Kokchetav-North Tien Shan island-arc system, bordering the Dzhungariya-North Tien Shan back-arc basin, and the Chingiz-Tarbagatai system with the Zaisan back-arc basin are discussed in detail. The former comprises the Stepnyak-Betpakdala-North Tien Shan volcanic arc and the adjacent relict arcs: outer (Ishim-Naryn) and inner (Yermentau-Chu Ili). Within the relict arcs, volcanism was limited or absent.

Two types of volcanic activity have been established: island-arc volcanicity (basalt, andesitic basalt, rarely andesite, dacite, rhyolite) and volcanicity of spreading zones in back-arc basins (basalt predominating). The high alkalinity of Ordovician volcanics is noted and two maxima of volcanic activity are marked: Arenig-Llandeilo with associated maximum cherty sedimentation in the adjacent basins, and Ashgill, accompanied by accumulation of large volumes of volcaniclastic sediments.

Three stages in the history of Ordovician sedimentation of the Kazakhstan-North Tien Shan paleobasins are outlined:

1. *The Early Ordovician and the earliest Middle Ordovician were characterized by the predominance of deep sea environments, with pelagic and hemipelagic sedimentation in basins below the carbonate compensation depth, and in narrow island shelves.*
2. *The Middle Ordovician was a period of subsequent filling of the deep basins by clastic material, dominantly turbidites, which came mostly from regions of volcanic activity.*
3. *The Late Ordovician was a period when the inter-arc and back-arc basins were filled by clastic sediments, and a depositional shelf was distributed over the entire area.*

Résumé

La paléogéographie ordovicienne et la volcanicité dans le Kazakhstan et dans le nord de la chaîne Tien Shan sont présentées dans les grandes lignes. Les milieux paléogéographiques qui y existaient à l'Ordovicien inférieur et au début de l'Ordovicien moyen (Llanvirnien), à l'Ordovicien moyen et à l'Ordovicien supérieur ont été étudiés. À l'Ordovicien, le Kazakhstan et la chaîne Tien Shan sont considérés comme des parties composantes de la marge continentale de type pacifique comportant un

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tion cherteuse-clastique et de bassins d'arrière-arc incluant des zones d'expansion successives. Le système d'arc insulaire de Kokchetav-Tien Shan du Nord, bordant le bassin d'arrière-arc de Dzhungariya-Tien Shan du Nord, et le système de Chingiz-Tarbagatai avec le bassin d'arrière-arc de Zaisan servent d'exemples et sont traités de façon détaillée. Le premier comprend l'arc volcanique de Stepnyak-Betpakdala-Tien Shan du Nord et les arcs residuels adjacents : l'arc externe Ishim Naryn et l'arc interne Yermentau-Chu Ili. Au sein des arcs residuels, le volcanisme était limité ou absent.

Deux types d'activités volcaniques ont eu cours : une volcanicité d'arc insulaire (basaltes, basaltes andésitiques, rarement andésites, dacites, rhyolites) et une volcanicité de zone en expansion dans les bassins d'arrière-arc (basaltes prédominants). La forte alcalinité des roches volcaniques ordoviciennes est une caractéristique dont il est fait mention et deux maxima d'activités volcaniques sont relevés : l'Arénigien-Llandeilien associé à une sédimentation cherteuse maximale dans les bassins adjacents et l'Ashgillien, accompagné d'une accumulation de grandes quantités de sédiments volcanoclastiques.

Trois étapes de sédimentation ordovicienne dans les paléobassins de Kazakhstan-Tien Shan du Nord ont été circonscrites :

1. *L'Ordovicien inférieur et le tout début de l'Ordovicien moyen sont caractérisés par la prédominance de milieux océaniques profonds, à sédimentation pélagique et hémipélagique dans des bassins situés au-dessous de la profondeur de compensation des carbonates et dans les plates-formes insulaires étroites.*
2. *L'Ordovicien moyen a été une période de remblayage de bassins profonds par des roches clastiques, surtout des turbidites, qui provenaient en grande partie des régions volcaniques.*
3. *L'Ordovicien supérieur a été une période de remblayage des bassins d'inter-arc et d'arrière-arc avec des sédiments clastiques et une plate-forme de sédimentation s'est étalée dans toute la région.*

INTRODUCTION

In recent decades, because of a better understanding of modern marine geology, the search for ancient oceanic sediments and structures has developed within continental areas such as the Paleozoic deformed belts of Kazakhstan and Tien Shan. Here, the Paleozoic deposits contain ophiolite belts, and argillaceous, calcareous, and siliceous rocks that can be tentatively referred to pelagic and hemipelagic sediments are widely distributed. There is also a wide distribution of thick, green, clastic, rhythmically bedded sediments, which have all the features of turbidite sedimentation (Lisitsyn, 1985; Nikitin, 1973, p. 50, 51). Thus the early Paleozoic strata of Kazakhstan comprise relatively deep sea deposits, which belong both to oceanic structures with oceanic crust, and to continental margins with transitional-type crust — island arcs and back-arc basin systems (Lyapichev, 1977; Peive and Mossakovsky, 1982).

There is some controversy about the nature of these Paleozoic deformed belts in Kazakhstan during the early stages of development. Several possible interpretations have been proposed:

1. Kazakhstan is a mosaic of microcontinents formerly separated by micro-oceans that have completely disappeared (Avdeev, 1984; Seitov, 1988).
2. The Paleozoic strata are closely analogous to modern continental margins.
3. Relict parts of ancient oceanic structures similar to modern ones have been included (Sonenshain, 1976).
4. The lower Paleozoic structures are significantly different from structures of modern oceans (Abdulin and Shlygin, 1982).

An obvious approach to settling the debate is to compare the Ordovician stratigraphy and paleogeography of Kazakhstan and North Tien Shan with modern examples. There is much information about Ordovician deposits in Kazakhstan, including new data about the age and structure of cherty and volcanic formations (Nikitin, 1984; Zaicev, 1985; Nikitin et al., 1987).

Ordovician deposits are widely distributed in Kazakhstan and North Tien Shan (Fig. 1) as constituents of compound fold structures that are complicated by numerous faults. Most of these structures are asymmetric, with relatively simple, gently inclined western and southwestern flanks of synclinoria, and more steep, sometimes overturned, dislocated northeastern slopes. Such structures indicate a general vergence of Paleozoic belts of Kazakhstan and North Tien Shan, and assume tectonic transport from northeast to southwest.

The following tectonofacies zones are distinguished by sediment type, completeness of section, and volcanicity (Fig. 1) (Nikitin et al., 1986): 1) Ishim-Naryn (Ulutau-Karatau-Naryn); 2) Stepnyak-Betpakdala-North Tien Shan; 3) Yermentau-Chu Ili; 4) Chingiz-Tarbagatay; 5) Dzhungariya-Balkhash; 6) Zaisan; 7) Gorny Altai.

MAIN PALEOGEOGRAPHIC ENVIRONMENTS

Most of the area being discussed was covered by the sea during the Ordovician and was freely connected to the ocean covering the rest of the world (Vinogradov, 1968). In this sea was an island-arc system — the Kazakhstan Islands. Some of these islands probably existed in the Cambrian, but others appeared in the Ordovician and grew so that by the end of the Ordovician, they had formed a large region and had surrounded the inner part of Central Kazakhstan.

During the Tremadoc, Arenig, and Llanvirn, the Ishim-Naryn Zone (Figs. 1, 2) was a system of relatively deep marine basins with thin pelagic and hemipelagic argillaceous and cherty sediments up to 380 m thick (Nikitin, 1973; Azerbaev, 1978, 1988). Conodonts from near surface waters are found together with graptolites. Benthic faunas include thin shelled inarticulate brachiopods and rare gastropods. Graptolites frequently show parallel orientation, giving evidence of ocean currents that were probably cold and moved along the basin from south to north.

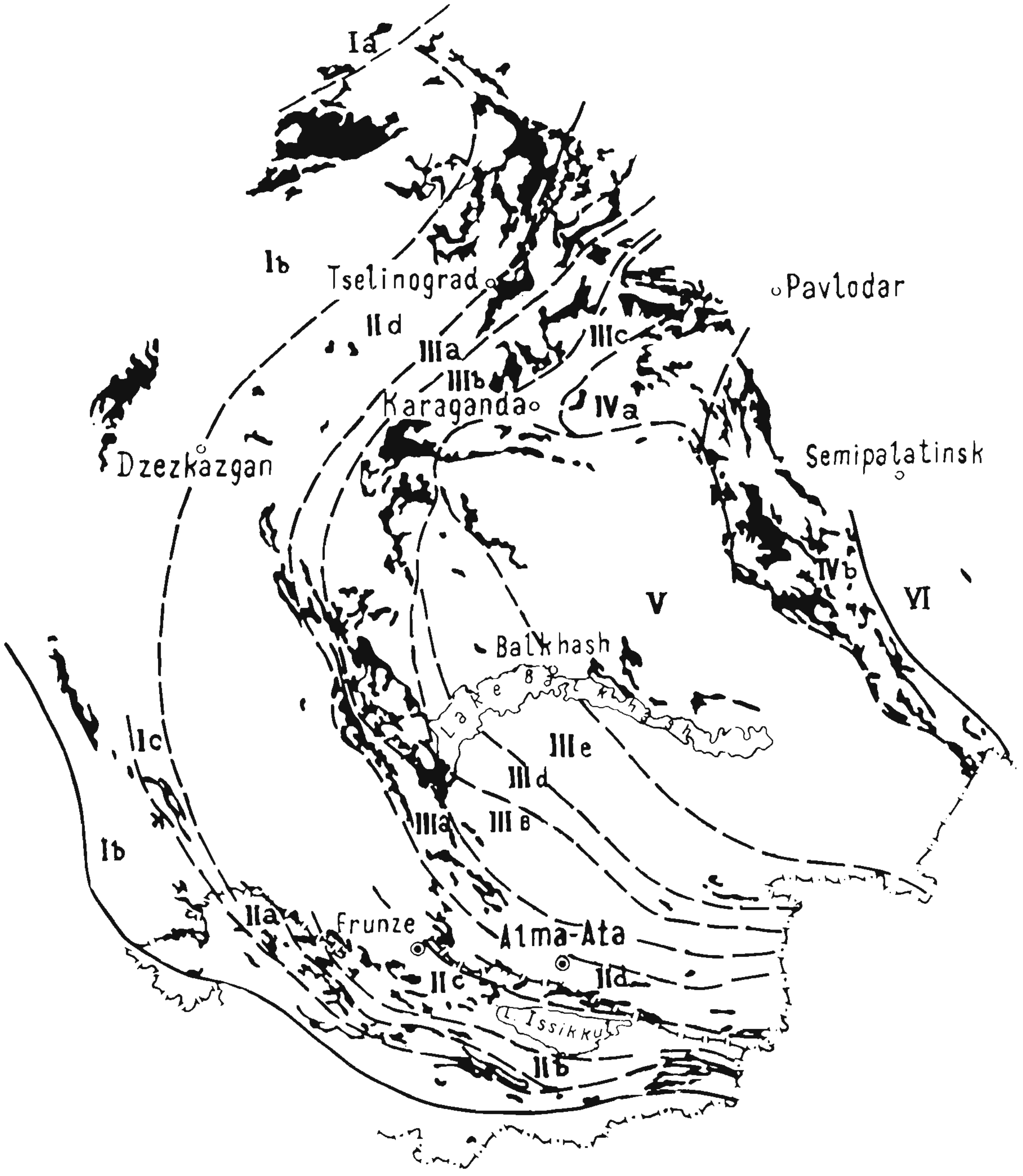

Figure 1. Ordovician structural zonation in Kazakhstan and North Tien Shan (according to I.F. Nikitin, V.G. Koroljov, M.K. Apollonov, and others). Shaded areas = outcrops of Ordovician deposits. I-VI: tectono-facies zones: I. Ishim-Naryn (Ia. Maryevka; Ib. Ishim-Sardzhas; Ic. Maly Karatau-Talas subzones); II. Stepnyak-Betpakdala-North Tien Shan (IIa. Susamyr-Dolon; IIb. Makbal-Burkhan; IIc. Keptash-Dzhumgal; IId. Stepnyak-Zailiyski Alatau subzones); III. Ermentau-Chu Ili (IIIa. Selety-Sugaty; IIIb. Ermentau-Burultas; IIIc. Shiderty; IIId. Mynaral-Southern Dzhungaria; IIIe. Mointy-Tekeli subzones); IV. Chingiz-Tarbagai (IVa. Boshchekul-Maikain; IVb. Alkamergen-Tarbagatai subzones); V. Dzhungaria-Balkhash; VI. Zaisan.

In the inner part of the zone, to the southwest (Maly Karatau-Talas Subzone), limestone was deposited. This region was probably an elevated submarine platform that had originated in the Cambrian. To the southwest and northeast of this seamount are calcareous rocks with features characteristic of slope facies. (Zhemchuzhnikov, 1986). Volcanic activity was very weak in some areas of the Ishim-Naryn Zone. It resulted in the appearance of minor coarse basic tuff and lava in some sections.

There is a suggestion of a landmass within the Ishim-Naryn Zone only in the extreme northwest, near the Kokchetav Massif. Lower to Middle Ordovician strata along the periphery of this massif include cherty as well as clastic sandy strata.

In the Stepnyak-Betpakdala-North Tien Shan Zone during the same time (Tremadoc-Arenig-Llanvirn), there are rhythmic, sandy, arkosic strata up to 2000 m thick. When clastic material was very limited, carbonaceous siliceous strata, and less commonly limestone, was deposited. Siliceous strata were most common in the late Arenig and early Llanvirn.

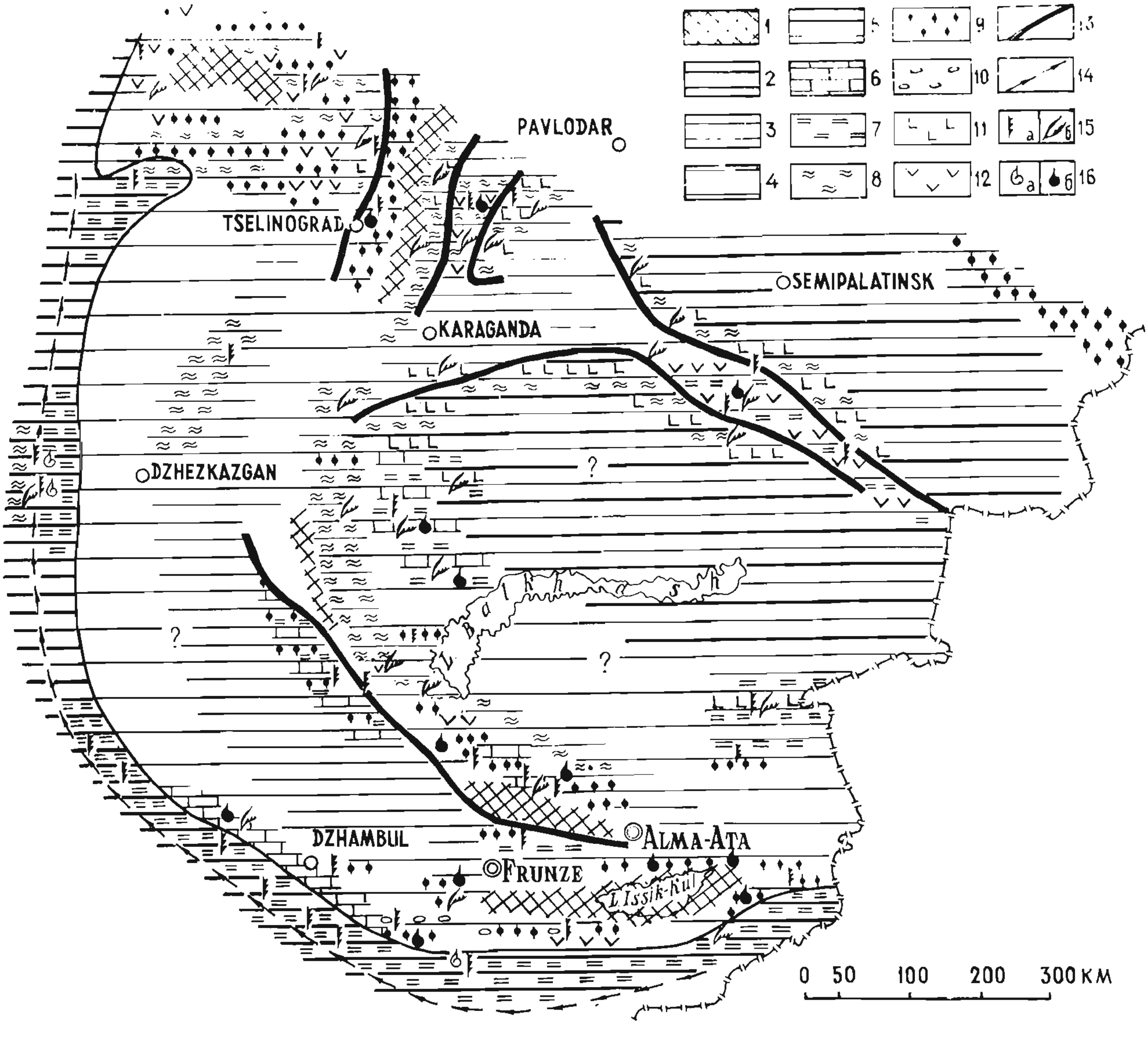

Figure 2. Paleogeographic reconstruction of Kazakhstan and North Tien Shan in the Early Ordovician and Llanvirn. 1. islands; 2. outer deep water basins (continental slope?); 3-4. island-arc basins (3. deep water basins; 4. shallow sea with submarine uplifts and island shelves); 5. back-arc basins; 6. limestone; 7. dark carbonate and siliceous sediment; 8. red siliceous sediment; 9. sandy sediment; 10. conglomeratic sediment; 11. deep water basaltic volcanism; 12. island-arc volcanism: basalt, andesite-basalt, rhyolite, and pyroclastic sediment; 13. spreading zones; 14. currents; 15. plankton (15a. graptolites; 15b. conodonts); 16. bottom communities (16a. deep-water inarticulate brachiopods; 16b. shallow-water bottom communities: brachiopods, trilobites, and others).

In the north they are replaced in the upper part by red siliceous strata, from which jasper formed. Local volcanism is known from the presence of basalt, rhyolite, and tuff.

In the dark siliceous strata, graptolites and conodonts are common and inarticulate brachiopods are rare. In the red siliceous strata, mainly radiolarians and conodonts occur. A typical benthic brachiopod community (*Clarkella, Nanorthis*; late Tremadoc) with trilobites (*Bathyuriscops granulatus*; Llanvirn) characterizes the shallow water limestone.

Along the western and southwestern margins of the Yermentau-Chu Ili Zone (Selety-Sugaty Subzone) there is an isolated part of the basin with an accumulation of apparently shallow, nearshore arkosic, and calcareous strata up to 500 m thick. There are many places where these strata appeared during the Arenig or Llanvirn, and they transgressively overlap Cambrian strata or Precambrian metamorphic rocks. The upper part of the clastic deposits is rhythmic and commonly contains lenses and bands of conglomerate. There are usually graptolites in the thin shale beds; in limestone, various trilobites, brachiopods, echinoderms, coelenterates, and other organisms produce typical Llanvirn benthic communities (*Aportophyla, Martellia, Yangtzella, Bathyuriscops granulatus*, and others). The composition, structural peculiarities and fauna of these Ordovician strata leaves no doubt that there was a system of islands, each surrounded by a narrow

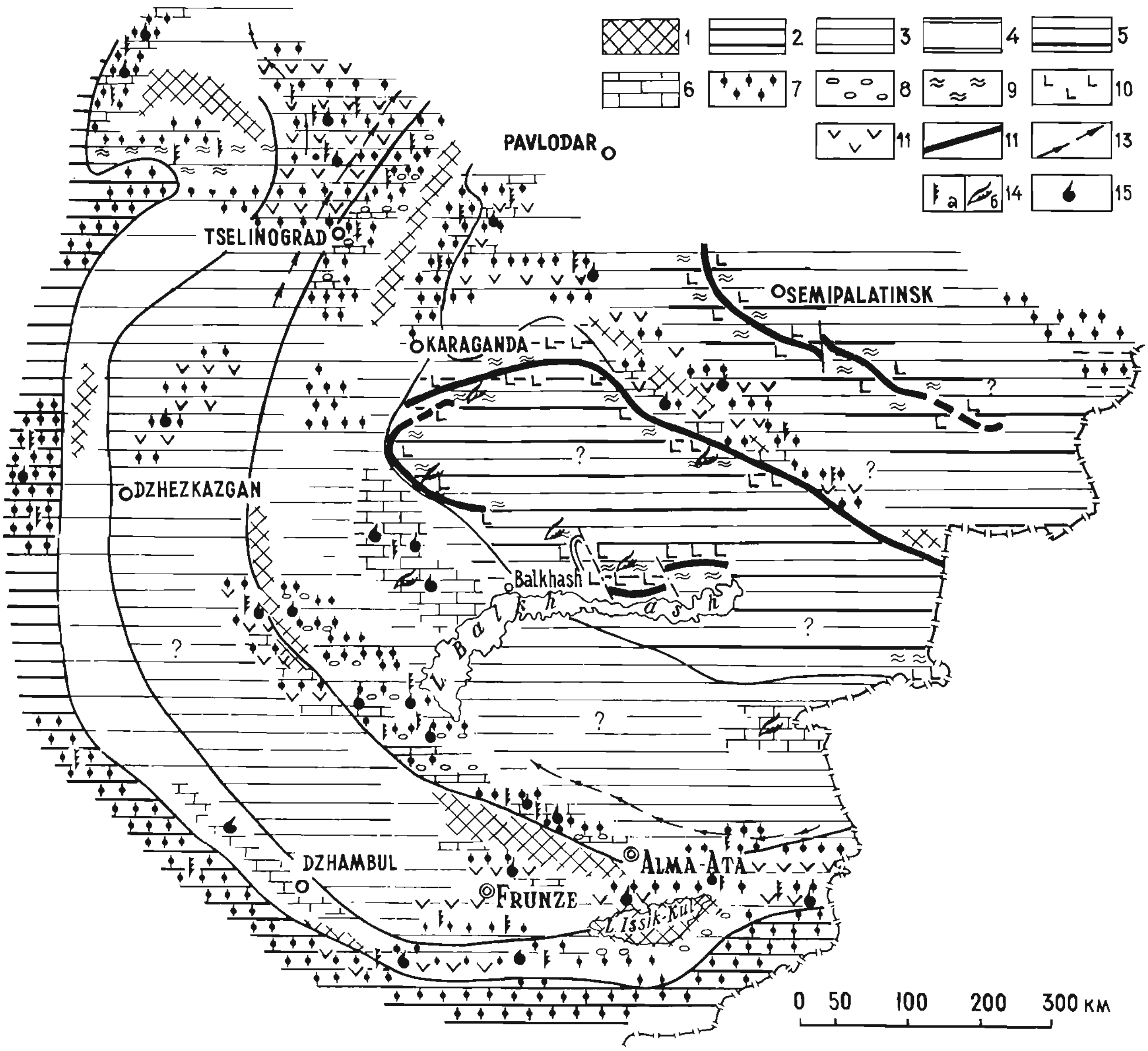

Figure 3. Paleogeographic reconstruction of Kazakhstan and North Tien Shan in the Middle Ordovician. 1. islands; 2. outer deep water basins (turbidite); 3-4. island-arc basins: 3. turbidite deep water basins and volcanic basins with volcanic islands; 4. shallow sea with submarine uplifts and island shelves; 5. back-arc basins; 6. limestone; 7. sandy sediment; 8. conglomeratic sediment; 9. red siliceous sediment; 10. deep water basaltic volcanism; 11. island-arc volcanism (basalt, andesite-basalt, andesite, and pyroclastic sediment); 12. spreading zones; 13. currents; 14. plankton (14a. graptolites; 14b. conodonts); 15. bottom communities: trilobites, brachiopods, and others.

depositional shelf. To the east these islands existed close to deep sea basins (Yermentau-Burultass Subzone), in which mostly red siliceous sediments accumulated; these strata have some iron and manganese and contain radiolarians and conodonts. Such deposits alternate with dark carbonaceous and cherty sediments, siliciclastic breccias, olistostromes, and volcaniclastics. In the northern part of the zone, within these deposits there are basalts and, in other places, the products of subaerial acid volcanism.

The islands of the Selety-Sugaty Subzone were probably small and supplied the adjacent basins with modest amounts of clastics; larger amounts of clastics and volcaniclastics were transported to these depressions from the east. In the eastern part of the depressions, adjacent to the Selety-Sugaty Islands, red siliceous sediment contains abundant argillaceous material that alternates with thick clastics (thicker than in the west). The same pattern can be observed in the southeast of the zone (Mynaral-South Dzhungariya Subzone) but clastics may have been delivered here by longitudinal currents.

In the southeastern part of this zone, dark carbonaceous and siliceous deposits or limestone are widespread. Limestone may have accumulated in the same way as in the Maly Karatau Range—on an underwater uplift or seamount, which had formed in the Cambrian. In the same region, Tremadoc volcanic activity is known from pillow basalts formed by underwater fissure eruptions.

The Chingiz-Tarbagatay Zone was an actively functioning volcanic island arc by the Early Ordovician. Cambrian volcanism continued during the Early Ordovician as subaerial andesitic basalt and andesitic dacite eruptions, which were accompanied by thick wedges of volcaniclastics up to 2000 m thick. On some narrow insular shelves, there were isolated quiet areas in which calcareous sediments accumulated, populated by the Tremadoc brachiopod community with *Clarkella* and *Nanorthis*.

At the end of the Early Ordovician there were mainly volcaniclastic sediments in this zone, which grade upward from coarse to fine grained. In the early Llanvirn there were carbonaceous, siliceous, and deep sea carbonate sediments with graptolites and conodonts. The Chingiz-Tarbagatay island arc was bordered by rift zones on the southwest and northeast. Thick sequences of ophiolitic basalt and chert were formed on the margins of this area, and in many places there are ultramafic intrusions. Textural peculiarities and content of these rocks indicate relatively deep sea conditions, and most of these strata formed below the critical level for pressure compensation in lavas and probably below the carbonate compensation level.

In the Middle Ordovician, at the end of the Llanvirn and the beginning of the Llandeilo (Fig. 3), there were important changes in all of the Ordovician marine basins. The appearance of clastic sediments in the Ishim-Naryn Zone suggests that islands had appeared nearby. Possibly the source was from numerous small volcanic islands that appeared in the adjacent zone.

Volcanic activity was also generated at the end of the Llanvirn in the Stepnyak-Betpakdala-North Tien Shan Zone, and the second active volcanic arc appeared. This volcanic arc resulted from the previous rifting zone, which is known from sutures with exposures of Precambrian ophiolites on the eastern boundary of the volcanic belt.

Volcanic activity in the Stepnyak-Betpakdala-North Tien Shan and Chingiz-Tarbagatay tectonic zones was generated by submarine basalt, andesite, and rarely dacite that poured out into depressions with high volcanic relief. Numerous, quickly eroded, and ephemeral islands appeared and supplied a large amount of volcaniclastic material. Volcanogenic deposits are widespread in the zone. They form an arcuate belt that can be traced from North Kazakhstan to North Tien Shan and farther to China as far as North Tsilyanshan (Zang Zhijin, 1984). The same peculiarities are observed all over the Chingiz-Tarbagatay Zone.

Caradoc volcanicity was less active. In most of the volcanic regions, basins were still deep enough to allow the accumulation of up to 3000 m of rhythmically bedded volcaniclastics with olistostromes. These strata were mainly the result of turbidite sedimentation during destruction of an older volcanic island.

Volcanicity became active again by the end of the Caradoc, but mostly in subaerial environments. Most of these volcanic rocks are andesitic basalt, andesite, and tuff. Dacite and dacitic pyroclastics occur rarely. On the east and west sides of the Stepnyak-Betpakdala-North Tien Shan volcanic arc there were deep water basins. By the end of the Llanvirn, siliceous sediments had been replaced by thick, rhythmically bedded clastics. These strata formed from volcaniclastics that came from the active volcanic island arcs and from another land area. Sedimentary basins were filled with turbidites and by the progradation of shelf platforms around islands. They surrounded the volcanic arc and are traced along the Ishim-Naryn and Yermentau-Chu Ili zones, from the northern margin of Kazakhstan to the eastern end of Tien Shan. In some basins the thickness of these complexes is 3000-3500 m and the deposition rate was about 450 to 500 m in one million years. In some regions (i.e., Kokdzhota area in the Ishim-Naryn Zone and the Yermentau-Burultas Subzone of Yermentau-Chu Ili Zone) the rhythmically bedded strata are more intensively deformed than in other parts of these zones. Deformation includes isoclinal folds, faults, and increased metamorphism (Sargaskaev, 1988). In the Yermentau-Burultas Subzone, among these strata are blocks of older chert and even ultramafic rocks. Based on these relationships, these deposits were referred, prior to faunal discoveries, to the late Precambrian or to the Cambrian. After chitinozoans, inarticulate brachiopods, and other fossils were found in Karatau (Ergaliev et al., 1987) and graptolites were discovered in the Burultas region (Nikitin et al., 1986), these deposits were assigned to the Ordovician and compared to deposits in other parts of the Ishim-Naryn and Yermentau-Chu Ili zones, which have a similar content but are less dislocated and metamorphosed. These deformed deposits may be the accretionary prisms that formed as a result of compression along the periphery of the volcanic belt of the Stepnyak-Betpakdala-North Tien Shan Zone.

During the Middle Ordovician, in the Dzhungariya-Balkhash Zone and probably in the Zaisan Zone, there continued to exist a relatively deep back-arc basin bounded by spreading zones. These spreading zones gradually migrated inside the basin. A series of highly faulted basalt, pillow lava, and chert was transported to these zones and is still preserved.

In large, relatively deep sea basins in which turbidites accumulated, the cherty and volcanic sediments were practically devoid of bottom communities. In siliceous rocks, numerous pelagic conodonts and radiolarians occur. Rare, disoriented fragmentary rhabdosomes of graptolites are found in rhythmically bedded clastics. More diverse graptolite assemblages are usually restricted to thin bedded siltstone and mudstone. Graptolites in these beds are frequently oriented along the structures, suggesting contour currents moving possibly from southeast to northwest. The faunas of narrow insular shelves of the Stepnyak-Betpakdala-North Tien Shan, Yermentau-Chu Ili, and Chingiz-Tarbagatai zones were more diverse. The most shallow water and nearshore community (e.g., the *Macrocoelia-Strophomena-Acculina* Community) is known in the North Tien Shan, where it occurs at the boundary of the red and green clastics (Nikitina, 1985), and the *Ancistrorhyncha* Community, which is common in the Yermentau-Chu Ili Zone. Deeper shelf communities were more varied, and inhabited calcareous sediments. One of

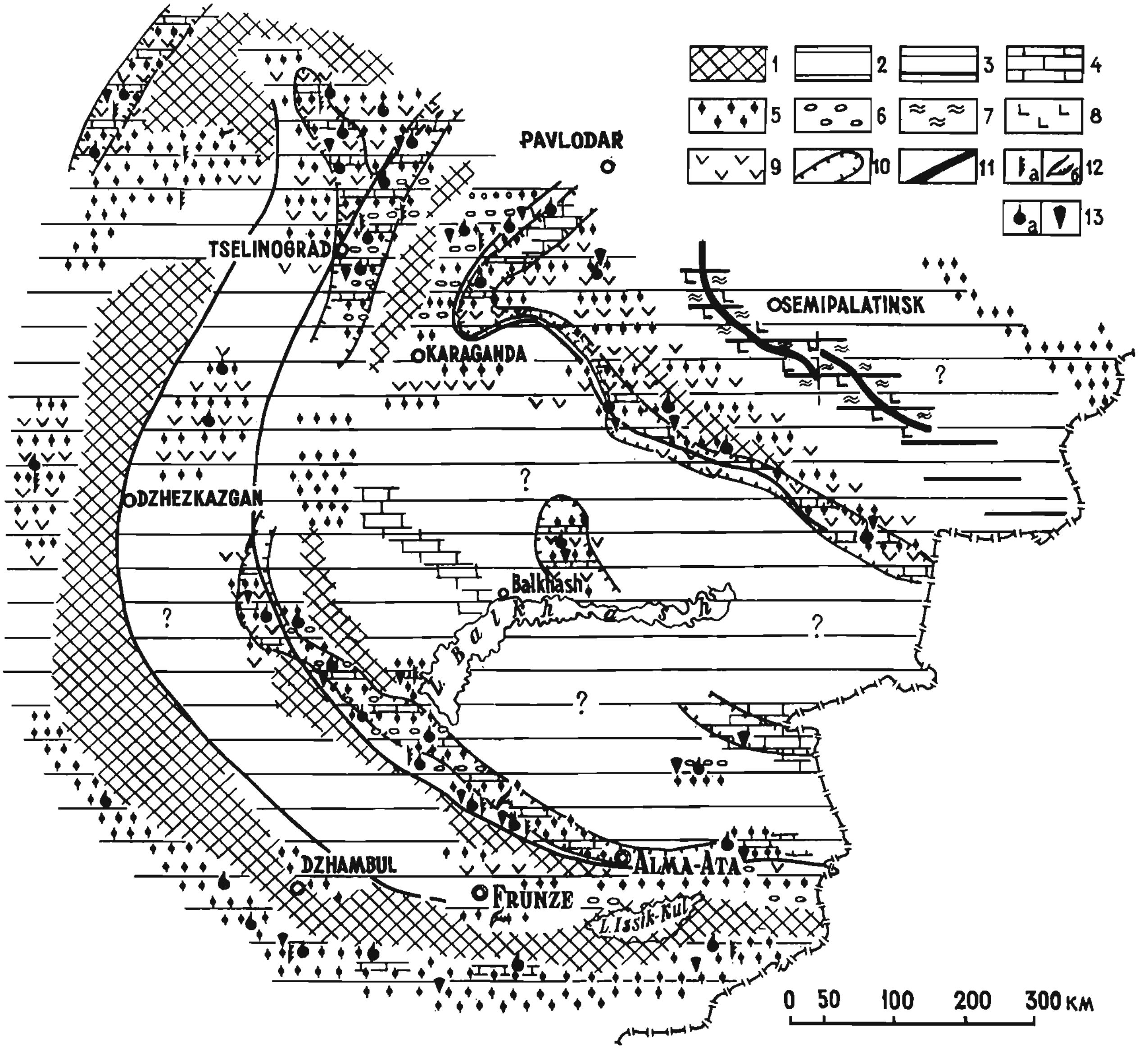

Figure 4. Paleogeographic reconstruction of Kazakhstan and North Tien Shan in the Late Ordovician. 1. islands; 2. shelf; 3. deep water basins; 4. limestone; 5. sandy sediment; 6. conglomerate; 7. red siliceous sediment (jasper); 8. deep water basaltic volcanism; 9. island-arc volcanism: basalt, andesite-basalt, andesite, rhyolite, and pyroclastic sediment; 10. boundary of organic reef distribution; 11. spreading zones; 12. plankton (12a. graptolites. 12b. conodonts); 13. bottom communities (13a. brachiopods and trilobites; 13b. corals).

them is the *Plectorthis-Ishimia-Strophomena* Community, which occurs in the north and south of the Ishim-Naryn Zone (Nikitin, 1974; Nikitin and Popov, 1983, 1985; Misjus, 1986), and in the Chingiz range (Klenina et al., 1984). The most diverse was the *Anoptambonites-Camerella-Cyclospira* Community where trilobites were found associated with organic reefs in the south of the Yermentau-Chu Ili Zone (Nikitin et al., 1974; Apollonov, 1976). The deepest community on the shelf was composed of trilobites (*Robergia-Ampyxinella-Dionidae* Community).

In the Late Ordovician (Fig. 4), there was a general shallowing of the Kazakhstan Sea and a further expansion of island areas. A vast highland massif formed in Ishim-Naryn, Yermentau-Chu Ili, and other zones and this marked the future contours of the later Silurian basin.

All the depressions that were adjacent to the island at the beginning of the Late Ordovician were filled with volcaniclastic sediments. The depositional shelf generally expanded during Late Ordovician time. Strata here consist mainly of green conglomerate, sandstone, and siltstone approximately 2000-3000 m thick. Along the periphery of the Dzhungariya-Balkhash back-arc basin, deep marine siliceous and volcanic formations of the spreading zones were overlapped by rhythmically bedded clastics. This was the result of the shelf progradation from the Yermentau-Chu Ili and Chingiz-Tarbagatai zones. In the lower part of these clastics there are locally thick mixtites that consist of large blocks of Middle Ordovician cherts. Only the back-arc basin of the Zaisan Zone remained as a deep sea trough, and here basic volcanics and cherty deposits accumulated.

In the main volcanic arcs (Stepnyak-Betpakdala-North Tien Shan and Chingiz-Tarbagatai zones), where volcanic activity had begun as early as the end of the Middle Ordovician, volcanism noticeably increased and became more widespread. Volcanic activity during the Late Ordovician was the same as at the end of Caradocian, manifested by eruptions and explosions of andesitic basalt, more commonly andesite, locally dacite and, in places, trachyandesite and trachyte, with brown and red rocks predominating. At the beginning of the Late Ordovician in the volcanic zones, there were separate islands, so that volcanic products accumulated mainly in shallow sea environments. As evidence of this, volcanogenic rocks are interbedded here with calcareous and clastic sediments containing marine faunas. Later, large volcanic islands emerged in the middle part of the Chingiz-Tarbagatai Zone and in the south of the Stepnyak Betpakdala-North Tien Shan Zone.

In the Upper Ordovician basins of Kazakhstan, reefs developed widely (Nikitin, 1973). These bordered some of the volcanic islands or formed extensive belts along the shelf margins of the island systems. Reef tracts extended for hundreds of kilometres in the Stepnyak-Betpakdala-North Tien Shan and Yermentau-Chu Ili zones and along the periphery of the Chingiz-Tarbagatai Zone. Those reefs in which calcareous algae played the dominant role were inhabited by various brachiopods, trilobites, corals, and other organisms. They formed the "dulankara" community, with *Anoptambonites*, *Camerella*, *Plectatrypa*, *Stenopareia*, *Glaphurina*, *Holotrachellus*, and others, inheriting the Middle Ordovician "anderken" community. Other communities, widely distributed in the shallow water shelf seas in sandy and argillaceous areas, were also mainly represented by brachiopods, trilobites, and corals. The most typical and common community for this environment is the *Plaesioms-Dinorthis-Otarorhyncha* Community. The *Parabasilicus-Dulanaspis-Pliomerina* Community and some others (Nikitin, 1972; Apollonov, 1976; Klenina et al., 1984; Misjus, 1986) are also found. At the end of the Ordovician, the *Holorhynchus giganteus* bank Community and the *Hirnantia-Dalmanitina* Community were typical for muddy depressions (Apollonov et al., 1980). Deeper-water graptolite associations were distributed locally in basins in the southern parts of the Yermentau-Chu Ili and Chingiz-Tarbagatai zones.

According to this brief survey of Kazakhstan and North Tien Shan paleogeography in Ordovician time, there existed systems of successively developing island arcs and back-arc basins (Figs. 5, 6). The en echelon Kokchetav-North Tien Shan and Chingiz-Tarbagatai island-arc systems are examples. The former system consisted of the Stepnyak-Betpakdala-North Tien Shan volcanic arc and the adjacent relicts of outer Ishim-Naryn and inner Yermentau-Chu Ili arcs. Within the

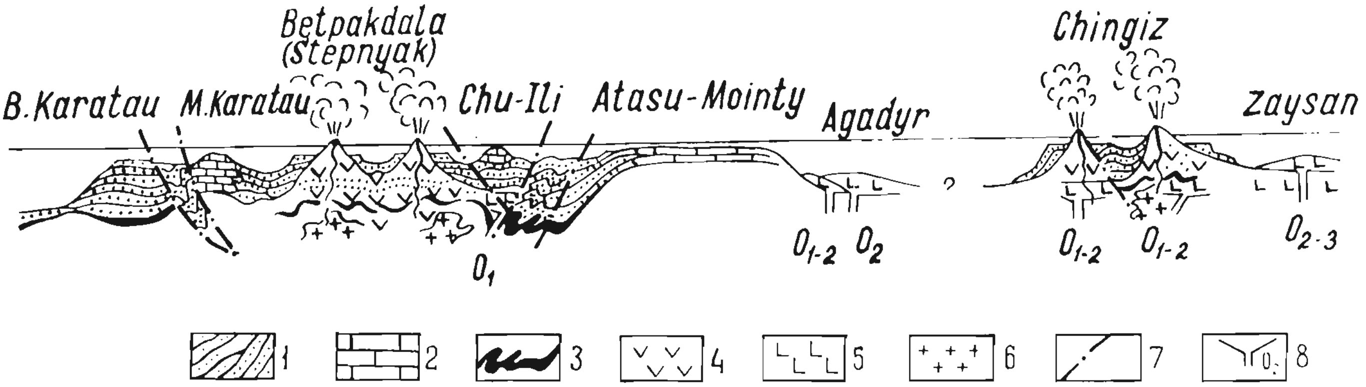

Figure 5. Schematic cross-section of Kazakhstan and North Tien Shan marginal basin in the Middle Ordovician. 1. clastic sediment; 2. limestone; 3. siliceous sediment; 4. island-arc volcanism; 5. volcanism of spreading zones; 6. granitoid intrusions; 7. faults; 8. spreading zones.

relict arcs, volcanism was weak or almost missing. The Chingiz-Tarbagatai island-arc system was likely to be traced to the West Siberia Lowland. Some island-arc systems bordered the back-arc basins: Kokchetav-North Tien Shan system surrounded Dzhungaria-Balkhash basin; the Chingiz-Tarbagatai system surrounded the Zaysan basin.

THE PECULIARITIES OF ORDOVICIAN VOLCANISM

Ordovician volcanicity in Kazakhstan was of long duration, it was significant and widespread, and was mostly in marine areas, with only rare nonmarine volcanism. The volcanic deposits are extremely thick, and diverse in composition. This diverse Ordovician volcanism may be subdivided into two types based on the character of the volcanic eruptions, by the content of the accompanying sediments.

The first type of volcanism unites volcanic eruptions of the main type, which were controlled by transregional magma sources. These volcanoes were connected at their bases, and they formed ranges of submarine and subaerial mountains, archipelagos, and volcanic island chains, which formed island-arc systems in Stepnyak-Betpakdala-North Tien Shan and Chingiz-Tarbagatai zones (Fig. 5).

The Ordovician volcanic deposits that took part in the building of these island systems have the same features as volcanics of these island arcs, such as: porphyritic and megaporphyritic structures, amygdaloidal structures of lavas,

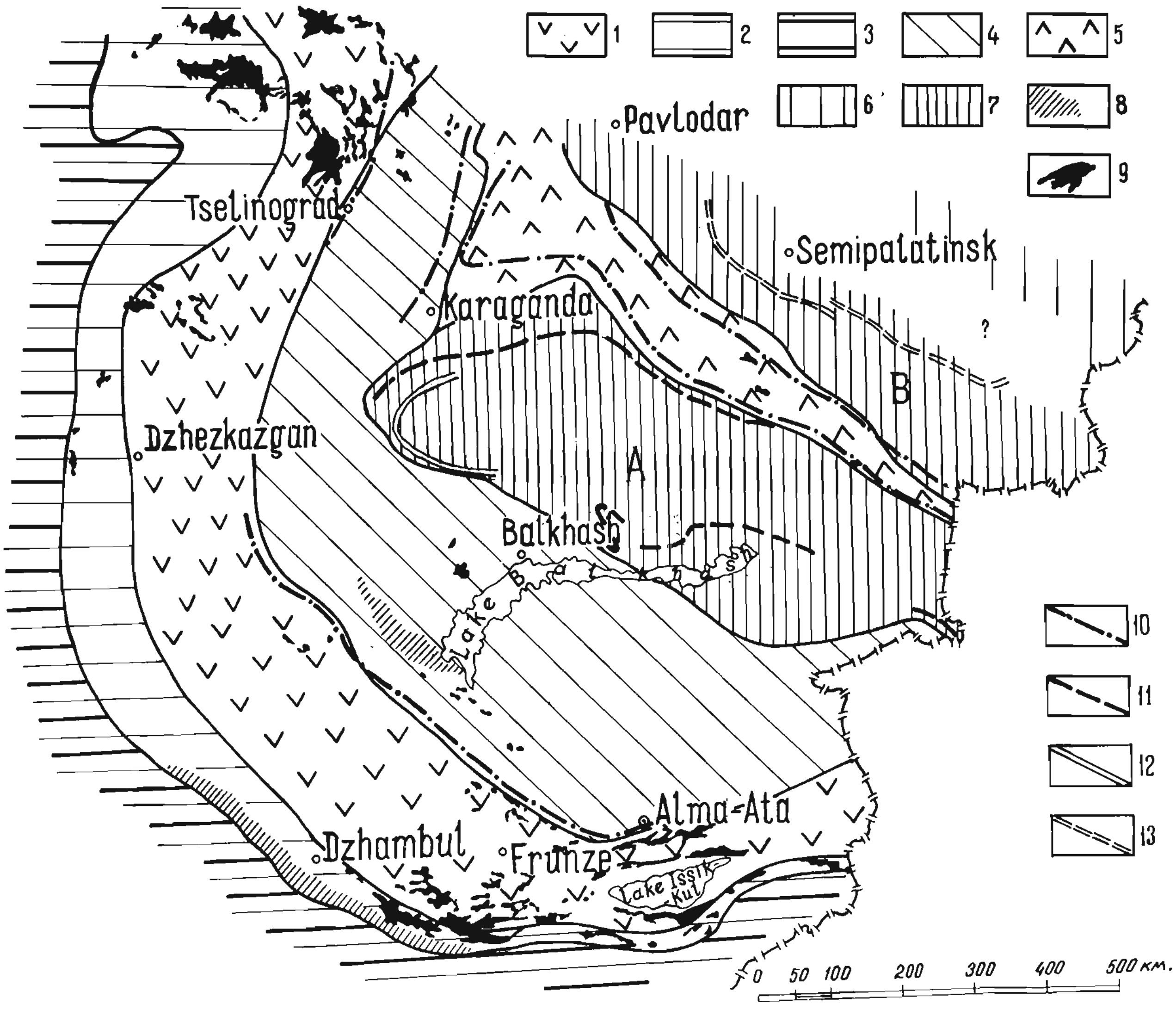

Figure 6. Paleotectonic reconstruction of Kazakhstan and North Tien Shan in the Ordovician. 1-4. Kokchetav-North Tien Shan island-arc system: 1. Stepnyak-Betpakdala-North Tien Shan volcanic arc; 2. outer relict island arc uplifts and basins; 3. continental slope during the Early Ordovician and Llanvirn; 4. inner relict island arc uplifts and basins; 5. Chingiz-Tarbagatai island-arc system; 6. Altai-Sayan shelf; 7. back-arc basins (A. Dzhungaria-Balkhash Basin; B. Zaisan Basin); 8. accretionary zones; 9. Ordovician granitoid intrusions; 10-13. spreading zones: 10. Early Ordovician and Llanvirn; 11. Early Ordovician-Llandeilo; 12. Early Ordovician-Caradoc; 13. Middle-Upper Ordovician.

forming zonally arranged flows with brown chilled crust; abundance of pyroclastic material with a wide granulometric spectrum from ash to coarse boulder, attaining 70-80 per cent of the total thickness of the volcanogenics; thick trains of volcanoclastics along the periphery of the volcanoes.

Island-arc volcanicity was durable and was characterized by the formation of both undifferentiated (basaltic) and well differentiated volcanics. In some cases basic volcanics dominated, in others, more rarely, acid volcanics prevailed. The specific peculiarity of Ordovician effusions is their increased alkalinity of both sodium and potassium types. Among basic rocks trachybasalts predominate, less common are basalts with normal alkalinity. Among the latter, calc-alkali and tholeiitic basalts are identified, which are likely to be petrochemically comparable to their equivalents from modern island arcs (Table 1). Ordovician island-arc basalts are characterized by low titanium and magnesium, moderate iron content, increased alumina content, and high alkalinity.

Ordovician island-arc volcanism developed in a complicated geodynamic environment under the conditions of alternating tensional and compressional faults. There were two main periods of volcanic activity separated by a nonvolcanic interval, the Arenig-Llandeilo and the late Caradoc-Ashgill.

The peculiar feature of the Arenig to Llandeilo period of volcanism is the predominance of basalt and andesitic basalt, with the local subordinate development of andesite and rhyolite. The volcanics form both undifferentiated and well differentiated complexes. In the Chingiz-Tarbagatai island-arc system, andesitic basalt and trachyandesite-trachybasalt complexes formed at this time. In the Stepnyak-Betpakdala-North Tien Shan island-arc system, basaltic, trachyrhyolitic, and trachybasaltic complexes predominate and andesitic complexes rarely prevail.

Late Caradoc to Ashgill volcanism is characterized by large amounts of acid eruptions, which were especially abundant in the Ashgill. Differentiated volcanic complexes are distributed here. In the Chingiz-Tarbagatai island arc at this time there were trachybasalt-andesite and basalt-rhyolite complexes. In the Kokchetav-North Tien Shan, basalt-andesite-rhyolite and basalt rhyolite complexes formed.

The amounts of Ordovician island-arc volcanogenic deposits are significant: the thickness of some volcanic complexes is 1 to 5 km, with a total thickness range of 5 to 8 km. Within the island-arc systems a displacement is observed, with volcanogenic rocks completely wedged out, both along the arcs and across them. Ordovician volcanic arcs of Kazakhstan are mature and have inherited to a considerable degree the tectonic setting of the previous Vendian-Cambrian volcanism.

The development of the Kokchetav-North Tien Shan island-arc system undoubtedly took place on continental crust, which is evident from outcrops of metamorphosed Precambrian rocks. The significant proportion of dacites and rhyolites in the Chingiz-Tarbagatai island arc likewise indicates a basement of continental crust.

The second type of Ordovician volcanism in Kazakhstan was represented by fissure eruptions along a complex system of narrow, variably oriented spreading zones, mainly at back-arc marine basins; these formed during the Arenig-Caradoc in the Dzhungariya-Balkhash and Zaisan zones, and at the early stages of the development of Kokchetav-North Tien Shan and Chingiz-Tarbagatai island-arc systems (Fig. 5). A peculiarity of volcanic formations of the spreading zones is the predominance of undifferentiated basalts (spilites), with insignificant amounts of derived rhyolite. Most of these volcanics are aphyric, have pillow structures and small amounts of pyroclastics, and may rarely have amygdaloidal structures. The thickness of the complexes is more than 2.5 km. The petrochemical composition of basalts of the spreading zones is quite uniform. They are rich in iron and titanium and low in alumina. Besides tholeiites, high sodium oxide and potassium oxide subalkaline basalts (spilite and trachybasalt) are developed widely, both of them with low silica content. The spreading zone basalts are similar to the low alumina and highly ferruginous alkali-olivine basalt and tholeiite of the

Table 1. Comparable analytical data on Kazakhstan Ordovician basalts with basalts from other modern geological structures

NN	SiO_2	TiO_2	Al_2O_3	FeO	MgO	CaO	Na_2O	K_2O
1.	49.54	0.73	17.90	9.95	5.78	9.98	2.43	0.42
2.	49.65	0.95	15.37	10.61	5.78	9.91	2.85	0.48
3.	50.80	0.61	18.10	8.63	3.42	6.85	3.77	2.26
4.	53.11	0.54	17.53	9.66	5.91	8.87	3.55	1.35
5.	49.43	0.77	18.21	8.84	5.21	8.00	3.28	0.72
6.	49.07	0.83	16.87	10.04	4.93	9.83	2.89	0.61
7.	50.60	0.68	18.60	9.26	5.42	9.61	2.56	0.76
8.	48.98	1.22	13.53	11.73	6.11	9.04	3.18	0.18
9.	48.05	1.52	14.60	12.05	6.74	10.54	5.65	0.25
10.	48.07	1.34	14.81	10.58	6.56	9.47	5.31	0.34
11.	48.80	1.80	14.87	11.43	6.24	8.75	3.04	0.37
12.	48.62	3.29	14.14	12.58	5.05	8.35	3.60	0.62
13.	49.07	1.77	14.73	10.61	5.49	8.78	4.32	0.37
14.	47.62	2.49	14.83	12.98	6.19	6.57	3.39	0.40
15.	47.84	4.51	13.39	14.33	5.75	6.89	3.07	0.26

1-7: island-arc system basalts; 1-2. tholeiitic basalts; 1. Llandeilo, Stepnyak-Betpakdala-North Tien Shan island arc; 2. island-arc sodium series (Frolova et al., 1985); 3-4. trachybasalts; 3. Llanvirn-Llandeilo, Chingiz-Tarbagatai island arc; 4. Shikotan complex of the Kuril island arc (Frolova et al., 1985); 5-7. calc-alkali basalts; 5. Arenig, Llanvirn of Chingiz-Tarbagatai island arc; 6. Caradoc of the same arc; 7. calc-alkali series of the Kuril arc (Frolova et al., 1985); 8-15: basalts of the spreading zones; 8-12: tholeiitic basalts; 13-15: subalkali basalts; 8. Llanvirn of Chingiz-Tarbagatai island arc; 9. ferruginous tholeiites of the Arabia-Indian Range; 10. Arenig-Llanvirn of the Ermentau-Chu Ili arc; 11. Llandeilo of the Dzhungaria-Balkhash back-arc basin; 12. ferruginous basalt of the Philippine Sea (Sharkov, 1985); 13. Llanvirn-Llandeilo of the Dzhungaria-Balkhash back-arc basin; 14. Caradoc of the same basin; 15. alkali-olivine basalt of the Arabia-Indian Range (Govorov, 1983).

Arabian-Indian Ridge, as well as to the ferrobasalt of the Philippine Sea. Neither calc-alkali nor high alumina rocks of island-arc type occur among the Kazakhstan back-arc basin basalt.

The wide combination of volcanic rocks with jasper, commonly with highly ferruginous sediments and ultrabasic rocks, is a particular feature of spreading zone volcanism. The peculiarities of volcanic rocks of the spreading zones and their combination with pelagic siliceous sediments are evidence of relatively deep water conditions.

There are some reasons to suggest that the migration of the spreading zones into the inner part of the basin during Arenig to Caradoc time resulted in the opening of volcanic fissures. Related to the Vendian-Cambrian volcanism, the Ordovician spreading zones of Kazakhstan are mainly neogenetic structures that are not in accordance with the disposition of the previous volcanic zones.

CONCLUSIONS

1. Kazakhstan and North Tien Shan were part of a vast Pacific-type continental margin during the Ordovician. There was a system of volcanic and nonvolcanic basins with cherty-clastic and clastic sediments. There were also back-arc basins with back-arc spreading zones (Kokelaar and Howells, 1984; Boillot, 1985) (Fig. 6).
2. The real oceanic structures, or paleo-oceans, must have occurred in the Ordovician, not only between the Urals and the Paleozoic belts of Kazakhstan in the Turgai and Priaral regions, but also in South Tien Shan.
3. Due to the composition of the Kazakhstan-North Tien Shan island-arc system, this realm must have belonged to the Siberian continental margin, and the Siberian Platform was at the core of it. This conclusion is not consistent with the Earth palinspastic reconstructions based on paleomagnetic data (Chramov, 1981; Ronov et al., 1984), which indicates a need for further studies to eliminate the existing discordances.
4. Two types of volcanic activity are singled out: volcanicity of the island-arc type and volcanicity from spreading zones in back-arc basins. There were two main periods of volcanic activity: Arenig-Llandeilo and late Caradoc-Ashgill. The predominant volcanics of Ordovician island arcs were basalt and andesitic basalt, with subordinate dacite and rhyolite and possibly some andesite. Basalt predominated in the spreading zones. The peculiarity of Ordovician volcanics is the predominance of high-alkalinity rocks.
5. The island-arc volcanism was more intensive and for a long time acted in the Chingiz-Tarbagatai volcanic arc. The volcanic activity in the Kokchetav-North Tien Shan island-arc systems was concentrated mainly in the inner part of the Stepnyak-Betpakdala-North Tien Shan arc (the active volcanic arc). The volcanic activity in the Ishim-Naryn and Yermentau-Chu Ili zones (the relict arcs) at the beginning of the Ordovician was practically absent and acted only locally. Granitoid intrusions are associated only with the Kokchetav-North Tien Shan island-arc system and mainly with its inner volcanic arc, with ages of 470-430 million years (Fig. 6).
6. The history of Ordovician sedimentation in the Kazakhstan-North Tien Shan paleobasin may be divided into three stages: 1) The Early and the earliest Middle Ordovician was characterized by the predominance of deep sea environments with pelagic and hemipelagic sedimentation in compensated basins, and the existence of narrow depositional shelf areas around islands; 2) the Middle Ordovician was a period of subsequent infilling of the deep basins by turbidites and clastics derived mainly from regions of volcanic activity; 3) the Late Ordovician was a period when the inter-arc and back-arc basins were filled by clastic sediments, and a depositional shelf was distributed over the entire area.

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Ordovician plate tectonic reconstructions

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Abstract

Plate tectonic reconstructions are presented for the Tremadoc, Arenig, Llandeilo-Caradoc, and Ashgill. During the Ordovician, the major continents were Laurentia (North America), Baltica (Northern Europe), Siberia/Kazakhstan, and Gondwana, which included China and Southeast Asia. These continents were separated by three large oceans: Iapetus, Paleotethys, and Panthalassa. Lower Paleozoic plutonism, island arc activity, metamorphism, and in some cases ophiolites, in eastern Australia, East Antarctica, western South America and North America, Arctic Canada, Mongolia, Kazakhstan, and north-central China, suggest that during the Ordovician, continents were encircled by a Panthalassic subduction zone. During the Late Cambrian or Early Ordovician, Avalonia rifted away from the northern margin of Gondwana, widening the Paleotethys and narrowing the Iapetus. The collision of the Bronson Hill-Tetagouche-Lush's Bight island arc along the eastern margin of Laurentia during the Middle Ordovician resulted in the Taconic Orogeny and the initiation of northwesterly directed subduction beneath Laurentia, which ultimately lead to the closure of the Iapetus Ocean in the middle Paleozoic. In most respects, the number, size, and rate at which the plates moved during the Ordovician were comparable to present-day plate geometry and motion.

Résumé

Les auteurs présentent des reconstitutions de la tectonique de plaques au cours du Trémadocien, de l'Arénigien, du Llandeilien-Caradocien et de l'Ashgillien. Au cours de l'Ordovicien, les principaux continents étaient la Laurentia (Amérique du Nord), la Baltica (Europe du Nord), la Siberia/Kazakhstan et le Gondwana, qui comprenait la Chine et le sud-est de l'Asie. Trois grands océans - Japet, Paléotéthys et Panthalassa - séparaient ces continents. Le plutonisme, le volcanisme d'arc insulaire, le métamorphisme et, dans certains cas, les ophiolites du Paléozoïque inférieur, qui se reconnaissent dans l'est de l'Australie, l'est de l'Antarctique, l'ouest de l'Amérique du Sud et de l'Amérique du Nord, l'Arctique canadien, la Mongolie, le Kazakhstan et le centre nord de la Chine, portent à croire qu'une zone de subduction panthalassique encerclait les continents à l'Ordovicien. Au cours du Cambrien supérieur ou de l'Ordovicien inférieur, l'Avalonia s'est détachée de la marge du Gondwana, ce qui a élargi l'océan Paléotéthys et refermé l'océan Japet. La collision de l'arc insulaire de Bronson Hill-Tetagouche-Lush's Bight le long de la marge est de la Laurentia, à l'Ordovicien moyen, a produit l'orogenèse du Taconique et engendré une subduction vers le nord-ouest, sous la Laurentia, qui a entraîné la fermeture de l'océan Japet au Paléozoïque moyen. Le nombre de plaques, leur grandeur et leur vitesse de déplacement au cours de l'Ordovicien se comparent, à presque tous les égards, à la géométrie et au mouvement actuels des plaques.

INTRODUCTION

The Ordovician is the oldest time period for which accurate plate tectonic reconstructions can be made. Plate tectonic reconstructions for the Cambrian and late Precambrian (Bond et al., 1984; Piper, 1983, 1987; Parrish et al., 1986; Kirschvink, 1990) remain speculative due to the absence of a global biostratigraphic standard, the paucity of reliable paleoagnetic results (Van der Voo et al., 1984) and the uncertainty concerning the absolute age of the Cambrian/Precambrian boundary (Cowie and Brasier, 1989; Moczydlowska and Vidal, 1988). In this paper we present plate tectonic reconstructions for the Tremadoc, Arenig, Llandeilo-Caradoc, and Ashgill epochs.

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Ordovician paleomagnetic results from North America (Van der Voo, 1988; Kent and Van der Voo, 1990), Europe (Torsvik et al., 1990), Siberia (Khramov and Rodionov, 1980; Khramov et al., 1981), South China (Lin et al., 1985a, b) and Gondwana (Bachtadse and Briden, 1990) provide the framework for the reconstructions. These paleomagnetic constraints have been supplemented with information from paleoclimatology (Scotese and Barrett, 1990), biogeography (McKerrow and Scotese, 1990), and plate kinematics. We estimate that there is a +15° uncertainty in the latitudinal positions of the major continents, and a +30° uncertainty in their relative longitudinal positions.

The poles of rotation used to reassemble the supercontinents of Gondwana and Laurentia are given in Table 1. The total finite rotation poles used to orient the major continental blocks relative to the geographic pole are listed in Table 2.

In addition to illustrating the positions of the ocean basins and continents (Fig. 1), we have attempted to plot the location of Ordovician subduction zones and regions of inferred plate divergence (arrows, Figs. 2-5). During the Ordovician there were four major convergent boundaries. Two subduction zones lay on either side of the Iapetus Ocean. One of these subduction zones ran along the northwest coast of Avalonia and a slightly younger, westward-dipping subduction zone was located along the eastern margin of Laurentia. A third major subduction zone stretched across the northern margin of Gondwana, and a fourth, the Panthalassic subduction zone, encircled Gondwana, Siberia, and the northern and western coasts of Laurentia (Fig. 1).

The areas of the plate divergence shown in Figures 2-5 are speculative. Except for ophiolitic remnants, no direct evidence exists concerning the location or orientation of ancient spreading centres, or the rates of sea floor spreading. The spreading directions shown in these figures have been drawn to account for the relative motion of the continents and to provide oceanic crust for the Ordovician trenches.

It should be noted that there have been several previous attempts to reconstruct Ordovician plate boundaries (Keppie, 1977; Zonenshain and Gorodnitsky, 1977; Kanasewich et al., 1978; Seslavinsky, 1984; Zonenshain et al., 1985, 1987). The maps presented here, although similar to these earlier attempts, are more detailed and are based on a new synthesis of paleomagnetic, paleoclimatic, and biogeographic data.

ORDOVICIAN OCEANS AND CONTINENTS

The Ordovician world, like the present-day, was made up of groups of continents separated by large ocean basins. The Iapetus Ocean separated Laurentia/Siberia from Baltica (Fig. 1). Baltica, in turn, was separated from Gondwana by the Paleotethys, which extended eastward along the northern margin of Gondwana (Fig. 1). The Panthalassic Ocean, like the modern Pacific, was bounded by subduction zones and encircled the entire ensemble of continents (Fig. 1).

Table 1. Rotation poles used to assemble Gondwana and Laurentia

A. GONDWANA (ROTATIONS WITH RESPECT TO FIXED AFRICA)			
Part 1. Central Gondwana			
Continental Block	Latitude	Longitude	Angle
South America	45.5	-32.30	58.20
Madagascar	-1.7	-87.80	22.20
Arabia	-26.50	-158.50	7.60
India	-28.10	-136.70	66.50
Sri Lanka	-18.00	-128.60	90.50
Australia	-24.63	-62.64	55.92
East Antarctica	-9.68	-31.81	58.54
Part 2. Northwest Margin			
Continental Block	Latitude	Longitude	Angle
Florida/Piedmont	62.2	-15.90	78.80
Yucatan	48.36	97.02	66.00
Iberia	-39.44	165.60	25.40
South-Central Europe	41.85	36.60	13.67
Apulia/Adria	43.78	47.30	14.62
W. Avalonia	61.68	-4.03	71.23
E. Avalonia	25.60	16.60	39.42
Turkey	-46.50	174.0	7.70
Part 3. Northeast Margin			
Continental Block	Latitude	Longitude	Angle
South China	6.39	89.93	93.34
Qiangtang/Lhasa	-32.44	-145.90	49.66
Indochina	-12.18	93.92	62.85
Burma-Malaya	1.95	102.57	86.79
North China	15.06	131.53	112.1
Part 4. Southeast and Southwest Margin			
N. New Zealand	4.83	-54.80	84.68
S. New Zealand	29.93	-58.89	103.9
Marie Byrdland	3.19	-33.40	61.47
W. Antarctic Pen.	-36.90	-19.92	81.40
B. LAURENTIA (ROTATIONS WITH RESPECT TO NORTH AMERICA FIXED)			
Continental Block	Latitude	Longitude	Angle
Greenland	-50.07	26.29	7.74
Alaska N. Slope	-70.11	51.84	78.00
Mexico	-48.60	94.10	13.00
Baja California	-51.12	80.89	5.63
Arctic Islands	-50.07	26.29	7.74
Chortis/Honduras	-42.66	88.53	43.96
N. Scotland	-82.30	-25.90	33.50
Barentsia (Svalbard)	-81.45	110.7	50.10

Table 2. Rotation poles used to reconstruct major plates relative to geographic pole

Continent stage	Latitude	Longitude	Angle[1]
A. Laurentia			
Tremadoc	-15.70	43.10	86.70
Arenig	3.02	41.64	75.51
Lland.-Car.	26.40	39.70	71.40
Ashgill	21.00	41.20	71.60
B. Baltica			
Tremadoc	-19.00	77.00	128.0
Arenig	-15.31	77.68	113.4
Lland.-Car.	-10.70	78.50	98.40
Ashgill	-9.20	72.30	88.50
C. Siberia			
Tremadoc	8.10	56.20	130.1
Arenig	13.59	57.63	124.60
Lland.-Car.	19.66	59.30	119.30
Ashgill	24.40	60.70	115.70
D. Gondwana			
Tremadoc	-6.10	86.70	152.50
Arenig	-2.66	84.32	138.90
Lland.-Car.	1.30	81.60	124.90
Ashgill	-3.40	85.50	110.60

[1] All rotations follow right-hand rule, i.e., positive counter-clockwise

Iapetus Ocean

The Iapetus Ocean probably originated when the southern margin of Baltica rifted away from the eastern margin of Laurentia during the latest Precambrian (600-550 Ma, Williams and Hiscott, 1987; Bond et al., 1984). It widened throughout the Cambrian, and was widest during Late Cambrian/Early Ordovician time. In the Early Ordovician, a transform boundary in the Iapetus Ocean converted to a southeast-dipping subduction zone bounding a northwest-facing arc (McKerrow et al., 1990), and the Iapetus Ocean began to narrow. Apart from pandemic pelagic faunas and deep water benthos, the marine faunas of the continents bordering Iapetus were distinct during the Early and Middle Ordovician, suggesting that Laurentia, Baltica and the northern margin of Gondwana were separated by oceans at least 1000 km wide (Cocks and Fortey, 1982) (Fig. 2).

In Figures 2 and 3, an island arc is shown adjacent to the southern margin of Laurentia. The Lower Ordovician benthic faunas of these island arcs (Bronson Hill, Tetagouche, and Lush's Bight) have been considered sufficiently distinct from those of Laurentia, Baltica, and Gondwana to justify a separate "Celtic Province" (Neuman, 1984). However, recent descriptions of similar fossils from both Laurentia and Avalonia (McKerrow and Cocks, 1986) would suggest that environmental factors, rather than geographic isolation, were responsible for these peculiar faunas. We believe that parts of this arc had already collided with the Grampian Highlands of Scotland prior to the Arenig, and that parts of the island arc could not have been far removed from northern Newfoundland, as collision took place there in Llanvirn time. Geographic isolation of these arc faunas from Laurentia would thus have been unlikely. The collision of these island arcs with the eastern margin of Laurentia resulted in the Taconic Orogeny.

During the Early Ordovician, the ocean floor was subducted to the southeast, beneath the Avalonian margin of Iapetus (northern Gondwana). In Figures 3 and 4, the Avalonian island arc, comprising England, Wales, southeastern Ireland, the Avalon Peninsula of eastern Newfoundland, parts of coastal New Brunswick, Nova Scotia, and coastal New England, is shown rifting from northwestern Gondwana. Calc-alkaline arc rocks appear in England, Wales, and Ireland in the Tremadoc (Kokelaar et al., 1984). This subduction zone appears to have become inactive after the Llandeilo-Caradoc (Figs. 4, 5), and a new subduction zone was initiated along the northwestern margin of Iapetus following the collision of the Bronson Hill-Tetagouche-Lush's Bight (BTL) arc with the eastern margin of Laurentia (McKerrow et al., 1990).

By the Middle Ordovician, the benthic faunas of eastern Avalonia started to lose their affinities with Gondwana and began to show more similarities to faunas from Baltica. The faunas of Avalonia and Baltica became identical in the late Caradoc (Fortey and Cocks, 1986) indicating that the Törnquist Sea between Avalonia and Baltica (Fig. 1) was narrow enough for cratonic benthos to cross easily. This connection occurred significantly earlier than the late Ashgill connection of benthic faunas across Iapetus between Baltica and Laurentia (McKerrow and Cocks, 1986; Cocks and Fortey, 1982).

Although we consider Avalonia to be composed of terranes from both sides of the Atlantic, the location of the western portions of Avalonia (Nova Scotia, and the eastern regions of Newfoundland, New Brunswick, and New England) is more uncertain. In southern Nova Scotia, the very thick clastic sequence of the Meguma Group extends upward to include Lower Ordovician sediments. No comparable sequence is known elsewhere in Avalonia, and it may be that the Meguma Terrane was originally separated from the rest of western Avalonia. Tillites have been described from the Roxbury Conglomerate of the Boston Bay Group, but their age is unknown. Tillites have also been described from Nova Scotia (Schenk, 1972). If Ashgill tillites were present in western Avalonia, western Avalonia would have been at a higher latitude than the eastern part of Avalonia. The location of the Avalon-Acadia terranes shown in Figures 2-4 is consistent with paleomagnetic results that place these terranes at temperate paleolatitudes during the early Paleozoic (Johnson et al., 1988).

Subduction beneath eastern Laurentia throughout the remainder of the Ordovician resulted in the narrowing of Iapetus, and lead to the eventual collision of Baltica with Laurentia during the mid-Silurian (McKerrow et al., 1990). The last remnants of Iapetus oceanic crust were consumed when Avalonia, which rifted away from Gondwana during the Early Ordovician, collided with Laurentia during the Late Silurian and Early Devonian (Caledonian/Acadian orogenies).

Paleotethys Ocean

Sengor (1984, 1987) recognized that the Tethys Ocean of Seuss (1893) was not a single ocean basin, but rather represented several generations of rifting, subduction, and continental collision. We use the term ''Paleotethys'' (Proto-Tethys of Ziegler, 1988) to describe the Paleozoic ocean basin separating the northern margin of Gondwana from the southern margin of Baltica and the southwestern margin of Siberia/Kazakhstan (Fig. 1).

Southern margin of Paleotethys

Like the modern western Pacific, the southern margin of Paleotethys was a major convergent plate boundary and the site of widespread back-arc basin formation (see Fig. 1, Cimmerian-Cathaysian subduction zone, cc). During the Paleozoic, numerous continental fragments rifted away from the northern margin of Gondwana and were eventually welded together to form much of south-central Europe (Ziegler, 1988), China, and Southeast Asia (Sengor, 1989; Nie et al., 1990).

Although the original position of these continental fragments is not well known, a variety of biogeographic, paleoclimatic, and paleomagnetic constraints can be used to infer their relative positions. Proceeding from southwest to northeast along the northern margin of Gondwana (Fig. 1), the major continental fragments are: south-central Europe, Turkey, Iran, Afghanistan, South China (Yangtze block) and Tarim, Indochina, Qiangtang, Lhasa, Greater India, Burma-Malaya, and North China (Sino-Korean platform).

South-Central Europe, Turkey, Iran and Afghanistan

South-Central Europe includes Iberia, France (Armorica), parts of West Germany, and the Bohemian massif. Paleomagnetic data from western France (Perroud and Van der Voo, 1985) and Spain (Perroud, 1983; Perroud et al., 1984), indicate that these areas were at high southern latitudes during the Ordovician (Torsvik et al., 1990), which is consistent with a position adjacent to the North African margin of Gondwana. Faunal evidence (Cocks and Fortey, 1982) supports the conclusion that these parts of Gondwana, together with Avalonia, bordered the southern margin of the Paleotethys Ocean. Deeper water facies, marginal to Gondwana, are present in the Ardennes of northern France, and in Bohemia, indicating

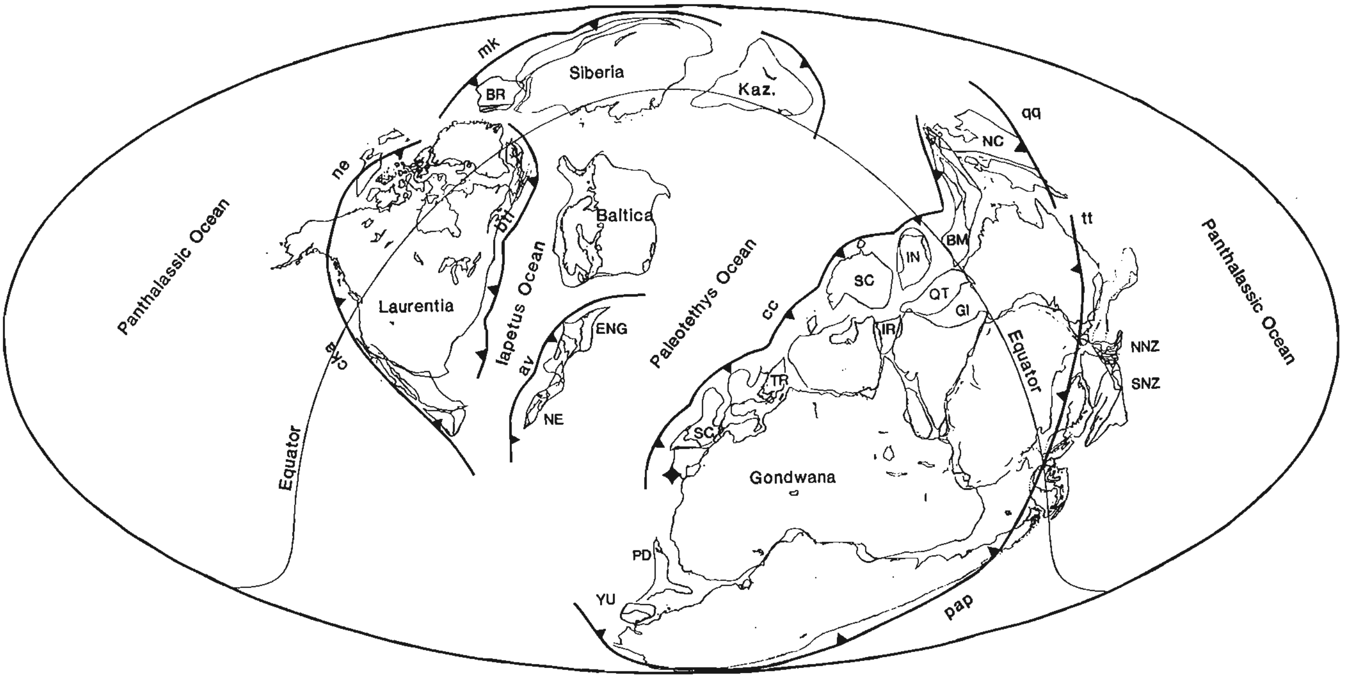


Figure 1. Key to geographic locations mentioned in text. Continental blocks: BR = Barentsia (Svalbard), ENG = England and Wales, NE = New England and Maritime Canada, YU = Yucatan, PD = Florida and Piedmont of southeastern U.S.A., SC = South-Central Europe, TR = Turkey, SC = South China, IR = Central Iran and Afghanistan (Helmand), IN = Indochina, QT = Qiangtang and Lhasa, GI = Greater India, BM = Burma-Malaya, NC = North China, NNZ = North New Zealand, SNZ = South New Zealand; Island arcs and subduction zones: cka = Chiapas-Klamath-Alexander arc, ne = North Slope Alaska arc, mk = Mongolian-Kazakhstan subduction complex, av = Avalonian arc, btl = Bronson Hill-Tetagouche-Lush's Bight arc (shown sutured to Laurentia), cc = Cimmerian-Cathaysian subduction complex, qq = Qinling-Qilian Shan Subduction zone, tt = Tasman-Trans-Antarctic subduction zone, pap = Puna-Arequipa-Perija subduction zone. The South Pole is indicated by the diamond symbol near Spain. (Middle-Late Ordovician reconstruction, Oblique Mollweide projection.)

that these parts of Europe also lay along the margin of Gondwana (Cocks and Fortey, 1982). The lower Paleozoic sequences of Morocco, Spain, France, and Bohemia are dominated by clastic sediments; there are no deposits of warm-water limestones from the Middle Cambrian through to the mid-Devonian, suggesting that these regions of Gondwana were located at high latitudes near the South Pole.

Both Turkey and Iran are composites of several continental fragments. Turkey, north of a suture that runs parallel to the Anatolian fault (intra-Pontide suture), is considered to be a rifted fragment of south-central Europe (Sengor and Yilmaz, 1981), whereas the regions south of the fault have faunas with mixed Gondwanan and Asian affinites. The location of southern Turkey during the Ordovician is not well constrained; we follow Dewey et al. (1973) and Geley (1988), who suggested that Turkey rifted from the northern coast of Egypt in the Late Triassic/Jurassic.

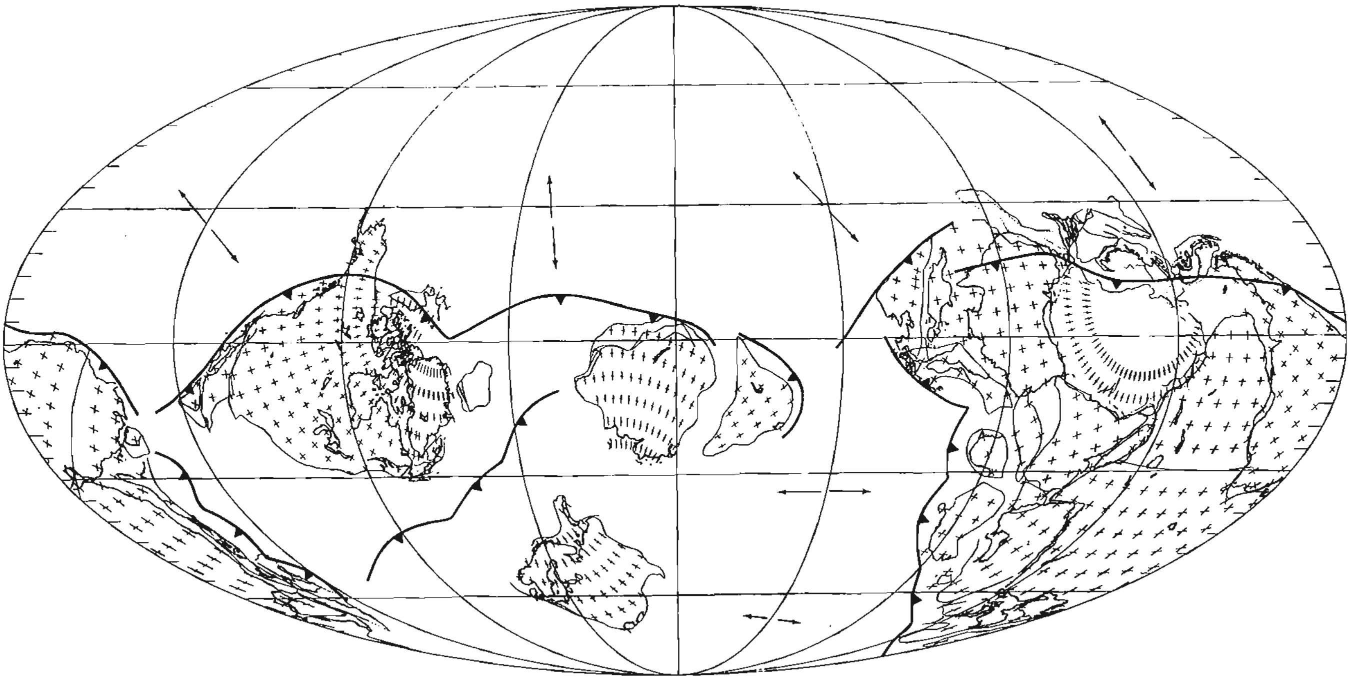
The major terranes that constitute Iran are the Alborz block, Lut block, and the Sanandaj-Sirjan zone. the Helmand (Sistan) block of Afghanistan appears to be closely associated with the Lut block of Iran. The Early Cambrian (or latest Precambrian) salt deposits of central Iran were probably once adjacent to similar salt deposits now found in the Persian Gulf, Oman, and Pakistan. In the reconstruction presented here, we place the Lut block of Iran and the Helmand block of Afghanistan in the gap between India and Arabia (Fig. 1). This position is consistent with Lower Devonian paleomagnetic data that suggests a location adjacent to India (Wensink, 1983). Mid-Devonian through Permian volcanic rocks in the Sanandaj-Sirjan zone (Berberian and King, 1981) indicate that this linear zone was either an island arc or Andean-style margin along the northern edge of Gondwana.

South China and Tarim

The position of South China and Tarim differs from the recently published reconstructions of Scotese and McKerrow (1990), and is similar to the reconstruction proposed by Burrett et al. (1990). South China is rotated 180°, so that its present-day eastern margin faces northwest. Biogeographic affinities suggest that South China was located near northwestern India and Pakistan during the Early Cambrian (Chang, 1981), although connections with North China and Australia are also maintained throughout the early Paleozoic (Burrett et al., 1990).

The occurrence of the Lower Ordovician trilobite, *Neseuretus* (Fortey and Morris, 1982), in South China indicates that there were biogeographic connections between South China, Arabia and south-central Europe. *Neseuretus* has been interpreted as a cold, shallow water fauna and its appearance in South China may reflect its southward movement into temperate latitudes during the Early Ordovician (Fig. 2). The orientation of South China shown in Figures 2-5 is consistent with Cambrian (Lin et al., 1985a, b), Silurian (Opdyke et al., 1987), and Devonian (Fang et al., 1990) paleomagnetic data.

The paleoposition of Tarim is still unknown. Late Precambrian and lower Paleozoic lithofacies, including lower Sinian tillites (Wang et al., 1985), archaeocyathids (Zhuravlev, 1986), and Lower Cambrian phosphorites (Notholt and Sheldon, 1986) indicate a closer association with South China than with North China. For these reasons we believe that Tarim was adjacent to South China and Indochina, but outboard of the Qiangtang and Burma-Malaya continental blocks (Fig. 1).



Earliest Ordovician (Tremadoc)

Figure 2. Earliest Ordovician (Tremadoc) plate tectonic reconstruction. Bold lines = subduction zones, arrows indicate areas of probable plate divergence (seafloor spreading).

Indochina, Qiangtang, Lhasa and Greater India

Very little is known about Indochina during the Ordovician, however, metamorphic rocks of early Paleozoic age (Hurley and Fairbairn, 1972) indicate that it may have formed part of the northern Gondwana subduction zone. Inboard of Indochina are Qiangtang/Lhasa, and Greater India. Qiangtang has been subdivided into eastern and western terranes (Nie et al., 1990). Western Qiangtang has strong Gondwanan affinites, including upper Paleozoic tillites and faunas (*Eurydesma*). Between Qiangtang and Greater India lies the Lhasa block, which, like Qiangtang, has Lower Permian tillites and a stratigraphy similar to northern India (Xizang Scientific Expedition, 1987). The Qiangtang and Lhasa blocks, together with Turkey, Iran, Afghanistan, and Burma-Malaya, formed the Cimmerian continent that rifted away from the northern margin of Gondwana during the late Paleozoic (Sengor, 1989; Nie et al., 1990). Cimmeria collided with Asia during the early Mesozoic, closing the Paleotethys Ocean (Sengor, 1984).

Between Qiangtang/Lhasa, the Indian Shield, and western Australia lies "Greater India". This vast region of continental crust is the northern extension of the Indian Shield. During the mid-Tertiary, Greater India was apparently subducted beneath Asia, forming part of the Tibetan plateau (Powell and Conaghan, 1973; Veevers et al., 1975).

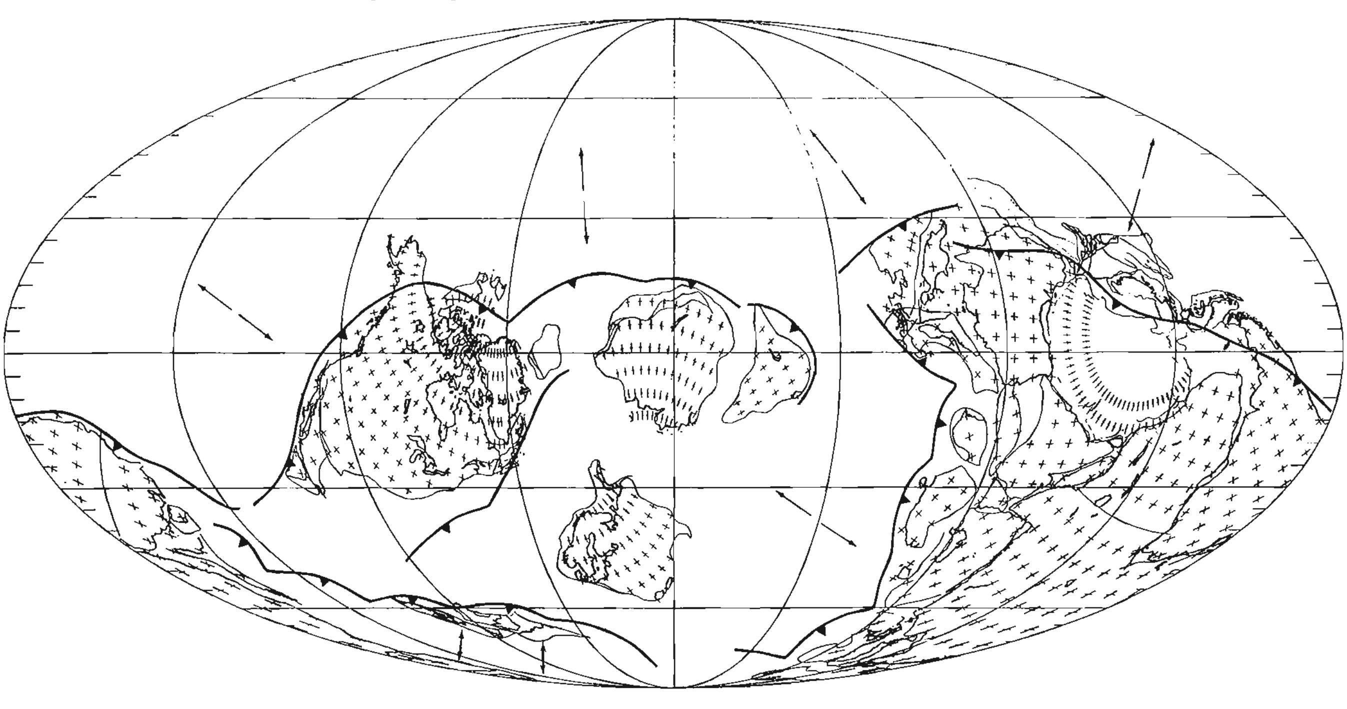
Burma-Malaya and North China

The Burma-Malaya block, or Sibumasu, or Shan-Thai, as it is sometimes called, is shown adjacent to the northwestern coast of Australia. Faunal ties between Burma-Malaya and northwestern Australia were strong during Late Cambrian and Ordovician time, as shown by the close faunal affinities of trilobites, molluscs, stromatoporoids, brachiopods, and conodonts (Burrett et al., 1990). The position of Burma-Malaya, shown in Figure 1, is consistent with the orientation predicted by Permian paleomagnetic data (McElhinny, 1981; Fang and Van der Voo, 1990). The occurrence of probable upper Paleozoic glacial diamictites (Phuket Group; Metcalfe, 1983) indicates that the Burma-Malaya block was still within the range of Permo-Carboniferous ice-rafted debris. An earlier phase of rifting between Burma-Malaya and Australia resulted in the eruption of the Antrim flood basalts, which once covered much of northwestern Australia (Veevers, 1984).

The North China block is shown adjacent to northernmost Australia. It has been rotated 180° so that the Qilian Shan subduction zone is aligned with the Tasman-Trans-Antarctic subduction zone (Fig. 1). During the Early and Middle Ordovician, North China had strong biogeographic ties with Australia and Burma-Malaya (Burrett et al., 1990). Faunal similarities with Laurentia during the late Cambrian may be because both continents occupied equatorial positions and faunas were able to cross Panthalassa via the western equatorial gyre.

Northern margin of Paleotethys

Avalonia, Baltica, and Kazakhstan bordered the northern margin of Paleotethys during the Ordovician (Fig. 1). The southern extent of Baltica is poorly known and it is likely that Baltica was a larger continent and extended farther south during the Ordovician. The Donetz Graben, located in the southeastern corner of Baltica, is the failed arm of a Devonian



Early Ordovician (Arenig)

Figure 3. Early Ordovician (Arenig) plate tectonic reconstruction. Bold lines = subduction zones, arrows indicate areas of probable plate divergence (seafloor spreading).

rift (Khain, 1985). At present we do not know which continent rifted away from Baltica, or how large this continent may have been.

Previous reconstructions (e.g., Scotese et al., 1979; Scotese, 1984), have treated Kazakhstan as a separate and independently moving paleocontinent, but we now consider Kazakhstan to be an amalgamation of volcanic arcs and far-travelled terranes that extended eastward from the Siberian plate. We propose that the relationship of Kazakhstan to Siberia was similar to the present-day relationship of southeast Asia to mainland China. The complex and often mixed aspect of the faunas of Kazakhstan may be explained, in part, by the fact that Kazakhstan is a collage of exotic and far-travelled terranes.

Panthalassic Ocean

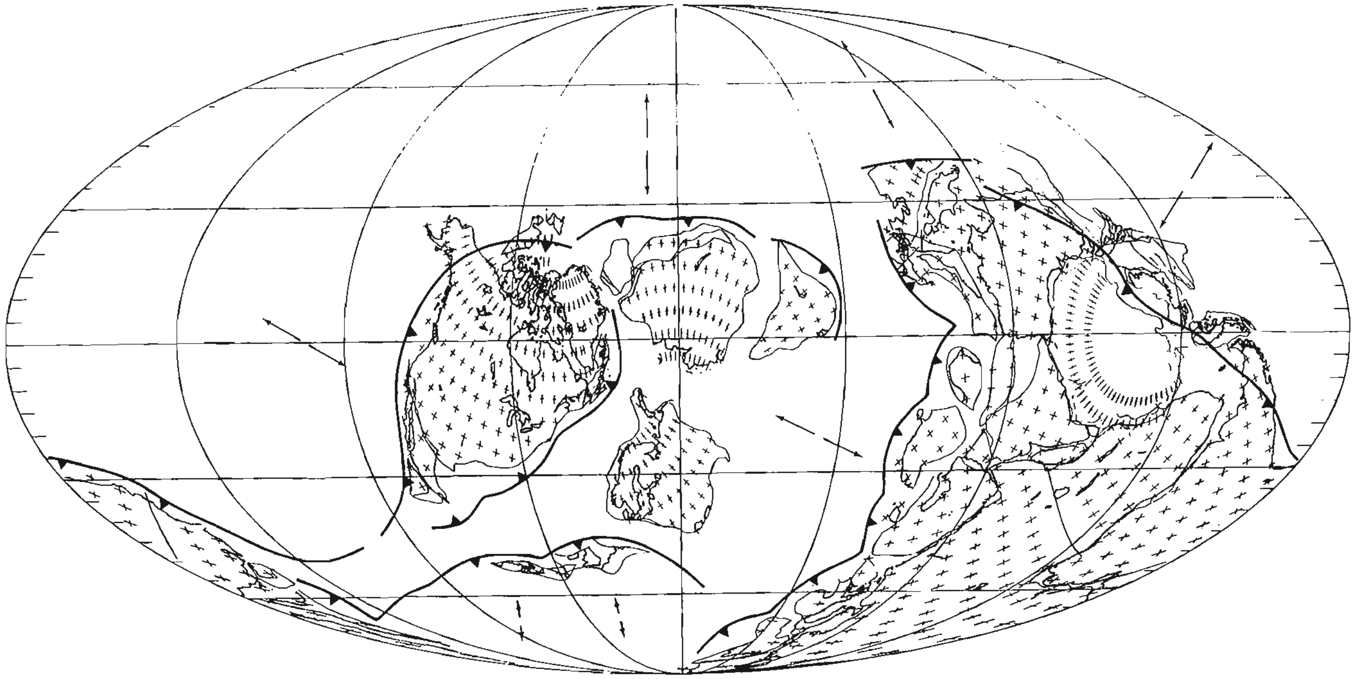
Panthalassa, the "universal sea", was the name given by the ancient Greeks to describe the vast oceanic expanse surrounding the known world. In Wegener's (1929) scheme, Panthalassa became the primordial ocean, just as Pangea was the primordial continent. Although not as vast as its Upper Paleozoic counterpart, the Ordovician Panthalassa was as wide at the equator as the present-day Pacific Ocean and included a hemispherical cap that covered the northern half of the globe (Figs. 2-5).

We propose that, like the modern Pacific Ocean basin, the Panthalassic Ocean was ringed by subduction zones (Scotese, 1987). Proceeding clockwise around Panthalassa from a startng point in eastern Australia, the components of the circum-Panthalassic subduction zone were: the Tasman-Trans-Antarctic subduction zone (tt), the Puna-Arequipa-Perija subduction zone bordering western South America (pap), the Chiapas-Klamath-Alexander subduction zone along western Laurentia (cka), the North Slope-Ellesmere subduction zone across the northern part of Laurentia (ne), the Mongolian-Kazakhstan subduction zone (mk), and the Qinling-Qilian Shan subduction zone (qq), which linked up with the northern part of the Tasman-Trans-Antarctic subduction zone to complete the circuit (Fig. 1). The best documented portions of these Ordovician convergent boundaries are the Tasman-Trans-Antarctic subduction zone (Schreibner, 1987; Veevers, 1984) and the Mongolian subduction zone (Zonenshain et al., 1971).

Much of the circum-Panthalassic subduction zone appears to have been active in the late Precambrian and although the record is incomplete, subduction along most of these margins appears to have continued through the Paleozoic and into the Mesozoic and Cenozoic. We suggest that the subduction zones ringing the present-day Pacific Ocean are the direct descendants of these ancient Panthalassic subduction zones.

Tasman-Trans-Antarctic subduction zone

The Tasman orogenic belt of eastern Australia can be divided into three regional provinces: the Kanmantoo fold belt, the Lachlan fold belt, and the New England fold belt. The westernmost belt, the Kanmantoo Orogen was tectonically active from the late Precambrian until the Early Ordovician (Veevers, 1984). The central belt, the Lachlan fold belt is the most extensive and was the site of plate convergence from



Middle–Late Ordovician (Llandeilo–Caradoc)

Figure 4. Middle-Late Ordovician (Llandeilo-Caradoc) plate tectonic reconstruction. Bold lines = subduction zones, arrows indicate areas of probable plate divergence (seafloor spreading).

the Cambrian to the Middle Carboniferous. The early Paleozoic history of the New England fold belt is uncertain, but the belt was active during the Silurian and continued to be active until the mid-Cretaceous (Veevers, 1984).

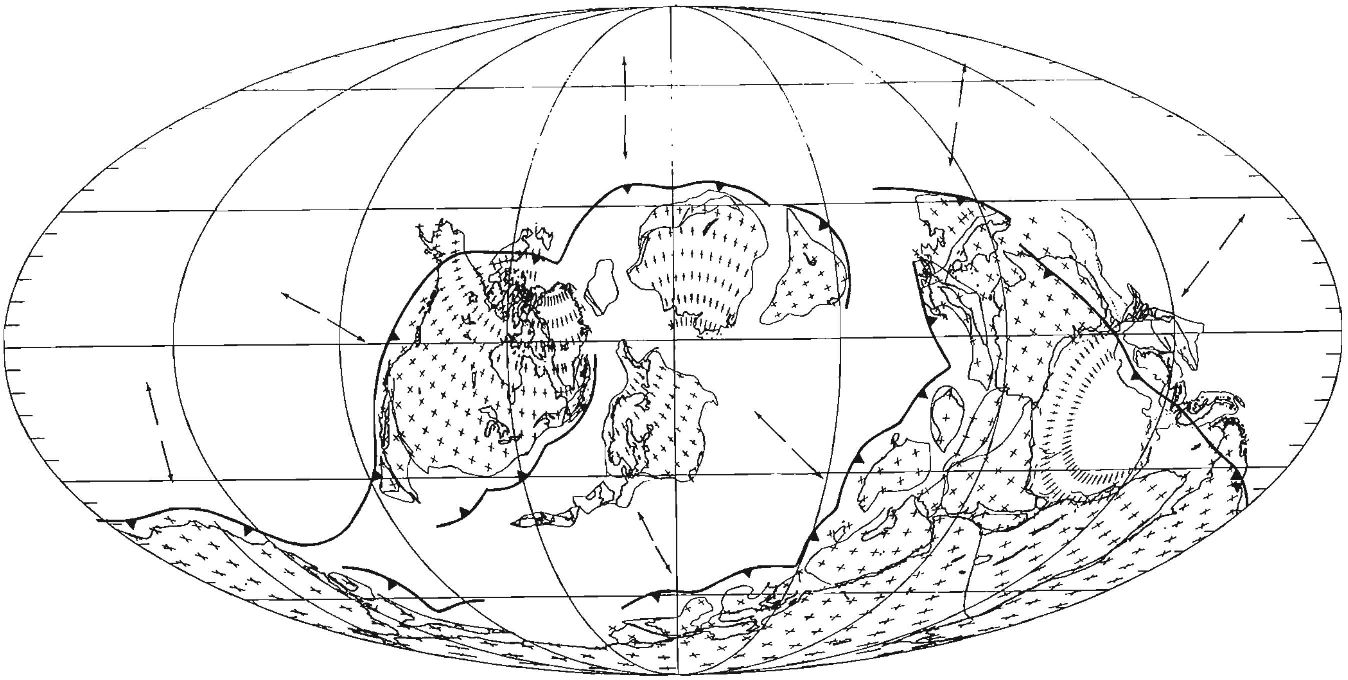
Upper Precambrian igneous and metamorphic rocks and lower Paleozoic granitic plutons of the Trans-Antarctic mountains represent the continuation of the Tasman fold belt into Antarctica (Kleinschmidt and Tessensohn, 1987). Three orogenic episodes have been recognized (Elliot, 1975): the late Precambrian Beardmore Orogeny, the Cambro-Ordovician Ross Orogeny, and a middle Paleozoic orogenic episode (Elliot, 1975). Devonian quartz diorites from Marie Byrdland (Halpern, 1968) indicate that the Trans-Antarctic subduction zone moved outward from the East Antarctic craton during the middle Paleozoic, possibly as a result of progressive continental accretion or the collision of exotic terranes. Recently, radiometrically dated metamorphic rocks of Paleozoic age from the West Antarctic peninsula indicate that this area may also have been the site of Paleozoic subduction.

Puna-Arequipa-Perija subduction zone

During the Ordovician, the Panthalassic subduction zone crossed the Falkland Plateau, traversed northwestern Argentina and exited near Concepcion on the Chilean coast. The Puna magmatic arc of northwestern Argentina is considered to be the equivalent of the Ross orogenic belt (Aceñolaza and Miller, 1982). No Precambrian rocks have been identifiedoutboard of this margin, and the suture with allochthonous Patagonian terranes is marked by an Ordovician and Silurian flysch basin that crops out in the coastal ranges of central Chile and the pre-Cordillera of Argentina (Frutos and Tobar, 1975).

Little evidence now exists for the continuation of the Panthalassic subduction zone along the western coast of South America. As noted above, the subduction zone appears to strike out to sea in the vicinity of Concepcion, Chile. The only other remnants of Panthalassic subduction are the Devonian metamorphic and plutonic rocks of the Arequipa Massif of southern Peru (Megard, 1987), and the lower Paleozoic metamorphic basement rocks of the Merida and Perija Andes in Colombia and Venezuela. In Colombia, slightly metamorphosed fossiliferous strata of Ordovician age are associated with gneiss, quartzites, phyllites, and amphiobolites, and are overlain by marine clastics of Early Devonian age (Floresta Formation; Burgl, 1973). In Venezuela a similar lower Paleozoic igneous and metamorphic basement is overlain by Devonian fossiliferous limestones (Zambrano et al., 1972).

The missing segment of the Panthalassic subduction zone may have been removed by rifting or by "tectonic erosion". Ziegler et al. (1981) suggested that much of west-central South America may have been tectonically eroded by subduction during the Mesozoic and Cenozoic. Thick accumulations of Devonian sandstone and siltstone in southern Peru and Bolivia imply that the Arequipa Massif, the source of these sediments, may originally have been more extensive (Isaacson and Sablock, 1990).



Late Ordovician (Ashgill)

Figure 5. Late Ordovician (Ashgill) plate tectonic reconstruction. Bold lines = subduction zones, arrows indicate areas of probable plate divergence (seafloor spreading).

Chiapas-Klamath-Alexander subduction zone

The continuation of the Panthalassic subduction zone along the western margin of Laurentia is recorded in the lower Paleozoic metamorphic terranes of northwestern Honduras, southwestern Mexico, as well as in the lower Paleozoic metamorphic, volcanic, and plutonic complexes of the Klamath mountains and Alexander Terrane of southwestern Alaska. Upper Precambrian and lower Paleozoic metamorphics and metavolcanics, including a high pressure, low temperature phase that has been dated as Early Ordovician (de Cserna, 1989), have been reported from Chiapas and the Acatlan Terrane (Sierra Madre del Sur) of southwestern Mexico (Ortega-Gutierrez, 1982). Subduction in this area continued through the middle and late Paleozoic (de Cserna, 1989).

The lower Paleozoic volcanic arc of the Klamath mountains in northern California represents the northward continuation of the Panthalassic subduction zone. Ordovician tonalitic and pegmatitic trondjemites from the Trinity ophiolite (Wallin et al., 1988), and associated Upper Ordovician calc-alkaline basalts with strong island arc affinites (Lovers Leap Butte volcanics, Brouxel et al., 1988) indicate that subduction along the western margin of Laurentia extended back into the early Paleozoic. This interpretation assumes that the Klamath volcanic arc is not a far-travelled terrane (Schweikert and Snyder, 1981), but rather originated in a back-arc setting (Burchfiel and Davis, 1972, 1975; Gray, 1986). Although arguments have been made for both interpretations, paleomagnetic data (Fagin and Gose, 1983) and mixed Tethyan and North American biogeographic affinities (Miller and Wright, 1987) suggest proximity to Laurentia. Although adjacent to North America, the Klamath volcanic arc was far offshore and did not influence foreland-platform sedimentation.

The volcanic arc that fringed Laurentia during the Ordovician continued northward in the Alexander terrane of British Columbia and southeastern Alaska. In these areas, Cambrian and Middle Ordovician metamorphic rocks, together with Ordovician and Silurian turbidites, calc-alkaline intrusives, and minor volcanics (Descon Formation) form the basis of the arc (Gehrels and Saleeby, 1987a). As in the case of the Klamath volcanic arc, it has been proposed that the Alexander Terrane is allochthonous and possibly far-travelled (Gehrels and Saleeby, 1987b). We find no compelling evidence for an exotic origin of the arc and agree with earlier authors (Churkin, 1974) who place the Alexander arc near the Laurentian margin.

North Slope-Ellesmere subduction zone

In the reconstruction of Laurentia presented here, the North Slope block of Alaska has been refitted against the Canadian Arctic Islands by a rotation about a pivot-point near the Mackenzie Delta (for a review see Lawver and Scotese, 1990). As illustrated in Figure 1, the Panthalassic subduction zone swung eastward across the North Slope, the eastern portion of the Canadian Arctic Islands (Ellesmere Island), and northern Greenland. The Precambrian(?) and lower Paleozoic metasedimentary and metavolcanic rocks of the Skajit allochthon of the Central Brooks Range (North Slope block) represent an island arc that fringed this subduction zone. Further evidence of island arc activity is seen in northwestern Ellesmere Island and Axel Heiberg Island where upper Precambrian through Silurian volcanics and flysch comprise the complexly deformed Pearya island arc (Trettin, 1989). Included in this sequence are a Lower Ordovician ophiolite (Thores suite) and amphibolite-grade metamorphic, which give Middle-Late Ordovician ages (452 Ma; Trettin, 1989). The deep water, lower Paleozoic sedimentary basin in northern Greenland represents a foredeep behind this volcanic arc.

Mongolian-Kazakhstan subduction zone

Numerous lower and middle Paleozoic calc-alkaline volcanics, metamorphics, and obducted ophiolites make the Mongolian-Kazakhstan subduction zone one of the best documented Paleozoic convergent plate boundaries (Zonenshain et al., 1971; Zonenshain, 1973). Built on pre-Riphean metamorphic basement (>1100 Ma), the Mongolian-Kazakhstan island arc was active from the late Precambrian (Mossakovsky and Dergunov, 1985) to the Permian, when subduction was terminated by the collision of the Tarim and North China blocks (Nie et al., 1990).

The original geometry of the arcs and subduction zones was certainly more complex than shown in Figures 2-5. Multiple ophiolite sequences (Zonenshain, 1973) indicate a succession of back-arc basin formation and collapse. Also, as proposed by several authors (Zonenshain et al., 1985; Rowley et al., 1985; Nie et al., 1990), an eastern extension of the Mongolian arc (Amuria, Zonenshain et al., 1985) was separated from mainland Siberia by a small ocean basin that did not close until the Late Jurassic.

As discussed earlier, Kazakhstan is an agglomeration of island arcs and exotic terranes. Although part of a separate subduction system, we believe that the island arcs of Kazakhstan were closely linked to the Mongolian subduction zone. As in Siberia, subduction continued throughout the Paleozoic, terminating in the Permian with the closure of the Irtysch Seaway between Kazakhstan and Siberia, and the suturing of Tarim (Tian Shan suture) and Baltica (Uralian suture).

Qinling-Qilian Shan subduction zone

Our tour of the Panthalassic subduction zone is complete as we return to Gondwana and connect the Mongolian subduction zone with the Qinling-Qilian Shan subduction zone of North China. The Qinling suture is the boundary between North China and South China (Nie et al., 1990). It is presumed to have been the site of subduction during the Paleozoic and the locus of continental collision during the Late Triassic Indosinian Orogeny (Sengor, 1984). The Qilian Mountains, located between the Ordos Basin and the Tarim block, are the westward continuation of the Qinling mobile belt. This interior mountain range consists of a series of upper Precambrian-lower Paleozoic island arcs, which collided against an Andean-style margin (Yang et al., 1986). The lower Paleozoic section includes over 10,000 metres of deep water clastics, siliceous and carbonaceous rocks, and volcanics. The peak of magmatic activity occurred during the Ordovician, with abundant intermediate and intermediate-basic volcanics (Yang et al., 1986), which were metamorphosed to greenschist facies during the middle and late Paleozoic.

In its present-day location the Qinling-Qilian Shan subduction zone appears anomalous; however, in the reconstruction presented here (Fig. 1), it forms a plausible continuation of the Tasman subduction zone. In this reconstruction (after Burrett et al., 1990), North China is shown rotated 180° and placed against the northern margin of Australia. This aligns the Qilian Shan subduction zone with the New England fold belt of eastern Australia.

CONCLUSIONS

From the plate boundaries described in Figures 2-4, it can be inferred that there were six to eight major plates during the Ordovician that were roughly comparable in size to modern plates. The rates of plate motion, based on the trajectory of the continents, were also similar to modern rates of plate motion. Based on the scenario presented here, Laurentia was the fastest moving continent, travelling eastward at rates of 15 cm/y during the Early and Middle Ordovician. Rates in excess of 10 cm/y are indicative of plates that are connected to a large subducting slab, like the modern Pacific Plate (Forsyth and Uyeda, 1975) or the Indian Plate during the Late Cretaceous and Early Tertiary (Scotese et al., 1988). In the case of Laurentia, the southern and eastern portions of the plate were subducted beneath the Avalonian arc and the Bronson Hill-Tetagouche-Lush's Bight arc (Figs. 2, 3). After collision with the Bronson Hill-Tetagouche-Lush's Bight arc and the ensuing reversal of subduction polarity (Figs. 4, 5), Laurentia's rate of motion abruptly slowed. Throughout the Ordovician, Baltica and Siberia moved northward at the moderate rate of 6 to 8 cm/y, while Gondwana moved southward over the South Pole at a similar pace.

Although preliminary, the Ordovician plate tectonic reconstructions presented here can be used as a framework for a better understanding of the paleobiogeography, paleoclimatology, and paleoceanography of the Ordovician. Only through an interactive process of revision and refinement in the light of new geological, tectonic, paleontological, and paleomagnetic data will we be able to unveil the secrets of the early Paleozoic world. Although the clues are sparse and often subject to alternative interpretations, it should be remembered that the same plate tectonic driving forces that produced the Mesozoic and Cenozoic ocean basins were also in operation during the Paleozoic. In this regard, the pattern of Paleozoic plate evolution, once it is understood, should be characterized by the same simplicity and elegance that characterizes Mesozoic and Cenozoic plate evolution.

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Oceanography in the Ordovician

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Abstract

The geographic position of major continental blocks in the Ordovician, determined primarily from lithofacies analysis and paleomagnetic studies, was the cause of a vastly different oceanic circulation pattern than that of today. The Ordovician was characterized by 1) a northern hemispheric ocean covering about one half of the Earth's surface; 2) a band of land and shallow shelf seas that occupied the tropics; and 3) a major landmass and associated shallow shelf seas in the southern hemisphere.

The trends during the Ordovician were a) the reduction of land and shallow shelves in the equatorial region and b) the movement of the southern landmass toward the south pole, culminating in a south polar ice cap during the Late Ordovician. The lack of meridional barriers in the northern hemisphere produced only zonal (east-west) surface currents: an easterly flowing current in the north polar zone; a westerly flowing current in the north temperate zone; and an easterly flowing current in the northern tropical regions.

The separation of the northern, or Boreal Ocean from the southern oceans (Iapetus and Proto-Tethys) by an equatorial barrier implies that the deep circulation of each ocean was independent. Only in the southern oceans with meridional barriers would the surface flow be similar in pattern to that of modern oceans. Trends toward an austral polar landmass, break-up of equatorial lands, and shallow shelf seas and generation of a polar ice cap during the Late Ordovician permitted establishment of connections between the oceans in the two hemispheres as well as significant deep ventilation from the edges of the polar ice cap during the Late Ordovician glaciation.

Monsoonal reversals of the equatorial currents would have occurred in the southern hemisphere due to the presence of the Gondwanan landmass. Such reversals would shift water mass boundaries on a seasonal basis and might explain apparent mixtures of marine faunal provinces (''Pacific'' = tropical; ''Atlantic'' = temperate-subpolar), especially on the Gondwanan shore or the northeastern side of the Proto-Tethys Ocean.

Résumé

La position géographique des principaux blocs continentaux de l'Ordovicien, fondée principalement sur l'analyse des lithofaciès et des études paléomagnétiques, s'est traduite par une configuration de la circulation océanique considérablement différente de celle d'aujourd'hui. L'Ordovicien était caractérisé par : 1) un océan d'hémisphère nord couvrant à peu près la moitié de la surface de la Terre; 2) une bande de terre et des mers de plate-forme peu profondes situées sous les tropiques; 3) une importante masse continentale et des mers de plate-forme peu profondes dans l'hémisphère sud. L'Ordovicien était caractérisé par les tendances suivantes : a) la réduction des terres et des plates-formes peu profondes dans la région équatoriale; b) le déplacement de la masse continentale méridionale vers le pôle sud culminant en une calotte glaciaire polaire méridionale durant l'Ordovicien supérieur. L'absence de barrières méridionales dans l'hémisphère nord n'a produit que des courants de surface zonaux est-ouest : un courant à direction est dans la zone polaire nord; un courant de direction ouest dans la zone tempérée nord; un courant de direction est dans les régions tropicales septentrionales. La séparation de l'océan nordique ou boréal des océans méridionaux (Iapetus et Proto-Tethys) par une barrière équatoriale révèle une

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circulation profonde indépendante dans chaque océan. Ce n'est que dans les océans méridionaux comportant des barrières méridionales que l'écoulement de surface était semblable à celui des océans modernes. Les tendances à la formation d'une masse continentale polaire australe, à la fragmentation des terres équatoriales et des mers de plate-forme peu profondes et à la formation d'une calotte glaciaire polaire au cours de l'Ordovicien supérieur ont permis d'établir des liens entre les océans des deux hémisphères ainsi qu'une ventilation profonde importante à partir desa bordures de la calotte glaciaire polaire au cours de la glaciation de l'Ordovicien supérieur. Des inversions dues aux moussons dans les courants équatoriaux de l'hémisphère sud auraient été causées par la présence du continent Gondwana.

De telles inversions auraient eu pour effet de déplacer les limites des masses d'eau selon les saisons et pourraient expliquer les mélanges apparents de provinces fauniques marines ("Pacifique" = tropical; "Atlantique" = tempéré-subpolaire), en particulier sur la rive du Gondwana ou le côté nord-est de la Proto-Téthys.

INTRODUCTION

This paper uses fundamental oceanographic circulation principles to speculate on the major current patterns and water masses and their implications for the biogeography of the Ordovician Period as a function of paleogeographic reconstruction of the major continental blocks. This approach elaborates on the procedures used by Stehli (1965), Ross (1975), Berggren and Hollister, 1977, and Parrish (1982) and is an extension of our work on the development of paleo-oceanographic interpretative methods (Wilde and Berry, 1986).

The primary astrophysical assumptions are 1) the Earth was the same size and shape as it is today, 2) the Earth was rotating in the same direction (west to east), 3) the Earth was rotating at about the same angular velocity (one rotation = 24 hours), 4) the path of rotation was in the plane of the ecliptic (the sun on the equator), and 5) there was some seasonality (the pole of rotation at some angle to the plane of the ecliptic). In the atmosphere of a rotating, spherical Earth, incoming solar heating would be transferred polarward in three major atmospheric cells symmetrical about the incoming solar maximum near the geographic equator. This would produce climatic zones with rising air at the meteorological equator and at 60°, and sinking air at the poles and at 30°. The major climate zones would have been bounded by these vertical air flows and would have been similar to those of today with a tropical zone extending from about 30° N to 30° S centred on the geographic equator; two temperate zones from 30° to 60°; and two polar zones from 60° to the pole. The degree designations for the boundaries are long-term mean values and would vary seasonally. The meteorological equator or thermal equator is the line that divides the polarward heat transfer in half. It will differ from the geographic equator (90° from each pole of rotation) 1) seasonally, moving with the sun north of the geographic equator in the boreal summer and south of the geographic equator in the austral summer and 2) as a function of the land-sea ratio in each hemisphere shifting into the hemisphere with the most land.

The modern open ocean is density stratified and has two layers: 1) an upper wind-mixed isothermal layer (0 to 100 m), the temperature and salinity characteristics of which are determined by local surface conditions with a general latitudinal temperature decrease from the equator to the pole and 2) a deep isothermal layer (1000 m to the bottom) of relatively uniformly cold water with gradually increasing temperature a function of distance from the area of formation at high latitudes with little or no relationship to surface latitudinal temperature gradients. These two layers are separated by a boundary layer: the main pycnocline (100 to 1000 m), where the temperature and density are transitional with depth between the upper and deep layers. For this discussion of the Ordovician ocean, only circulation in the upper mixed layer is treated in detail. Although this would cover sites of deposition of most shelf deposits, deep sea fan or continental margin deposits would intersect the transitional pycnocline, as would outer shelf sediments during high stands of sea level. In any case, most of the pelagic and benthic faunas that were likely to be fossilized would live in the upper mixed layer, although they may have become fossilized in sediments in the pycnocline.

CURRENTS

Planetary conditions

In the oceans, the movement of water is primarily the result of the interaction of frictional influence of the atmospheric winds on the ocean, and thermo-haline (density) stability. In most of the ocean, in the pycnocline and the deep layer, the effects of friction by surface winds or the bottom are negligible, so the circulation is dominated by density differences. Only in the surface mixed layer is wind friction the predominant driving force. However, the wind-driven circulation can be modified or even dominated by thermo-haline circulation in areas where density differences are sufficient to cause sinking or rising water before wind mixing and ''homogenation'' can occur. At the sea surface, air flow connecting the rising or sinking air at the cell boundaries would produce easterly surface winds in the polar and tropical zones and westerly surface winds in the temperate zones. The frictional couple between the air flow and the sea surface generates water flow in the surface ocean to the depth of the influence of friction essentially defining the surface mixed layer. Ekman (1905) mathematically described this flow, which is called the Ekman transport, and noted the net flow was 90° *cum sol* (to the right in the northern hemisphere and to the left in the southern hemisphere) to the wind. Low oceanic pressure is formed when two planetary Ekman transport vectors diverge (0° and 60°) and high oceanic pressure is formed when planetary Ekman transport vectors converge (30°). For oceanic pressure systems, like atmospheric pressure systems, the surface circulation is assumed to be geostrophic with cyclonic flow about low pressure systems and anticyclonic flow about high pressure systems. If there were no barriers in the oceans (i.e., no continents), the water circulation about such systems would be purely zonal or east-west. With meridional barriers, the north-south flow is permitted and the circulation can transfer heat to different latitudes. The Sverdrup-Ekman circulation model (Stommel, 1957) describes how Ekman transport due to planetary winds with appropriate meridional (north-south) boundaries creates quasi-circular pressure systems in the ocean.

The major currents generated by this system can also be identified with climatic zones, although the flows are connected at zonal boundaries.

In the tropics, there are two westward flowing equatorial currents astride the meteorological equator, which are the northerly and southerly limbs of the equatorial divergence (oceanic low pressure). The equatorial currents flow in the same direction due to the shift of the Coriolis deflection and the sense of geostrophic flow across the geographic equator. The equatorial currents transfer water to the west creating an accumulation of water at the western barrier. As the centre of the atmospheric low pressure at the meteorological equator is characterized by rising air, surface wind flow is diminished, as signified by the calms of the doldrums. Thus, part of the water accumulation is relieved by oceanic easterly flow embedded in the centre of the equatorial low pressure. One such current is the easterly flowing equatorial counter-current, seen best developed on the eastern side of the ocean. As this type of counter-current develops as a response to the accumulation of water on the west, the amount of transport and the extent of the counter-current would vary directly with the width of the particular equatorial ocean and the fetch of the tropical trade winds. The transport of surface counter-current is small and restricted to the narrow zone of the doldrums, so that the westward zonal flow in the tropics is dominant.

In the temperate zone, the open ocean circulation is dominated by the oceanic high pressure centred in middle latitudes at about 30°. Such systems, with meridional barriers, will produce polarward warm currents derived from the warm water piled up on the western side of the system in the tropics and equatorward cool currents on the eastern side. In general, the polarward flow is strongest as a result of western intensification (Stommel, 1948; Munk and Carrier, 1950).

In the polar regions, the circulation is cyclonic about oceanic low pressure centred at about 60°. With meridional barriers, the circulation will be cyclonic with equatorward cold currents flowing along the western barrier and polarward warm to cool currents along the eastern barrier. The temperaure of polarward flow is a function of the strength of the western intensification polarward warm current about the neighboring oceanic high pressure at midlatitudes. As the area of the polar regions is much smaller than either the temperate zone or the tropical zone, the volume of surface water formed in the polar regions is relatively small. The coherence and transport of the north-south components of circulation depends on the configuration and the meridional continuity of the barriers. Thus, a definite continental barrier such as the Americas will produce a strong western intensification as seen in the Gulf Stream.

Seasonal conditions

The planetary oceanic circulation can be modified on a seasonal basis, if the winds produced by atmospheric pressure systems oppose and override the planetary winds or are able to generate water currents counter to those produced by first-order geostrophic processes.

Equatorial-subtropical monsoons

Monsoons are produced when there is sufficient land at middle latitudes to modify the circulation derived from global oceanic conditions. The landmass or continental effect (Von Arx, 1962, p. 164; Barron, 1981) is due to the difference in heat capacity and albedo between land and water, because water has a greater heat capacity and lower albedo than land. Thus, the air temperature over land is highly variable as a function of insolation compared to water, with seasonal atmospheric high pressure created over land in the winter (sinking cold air) and atmospheric low pressure created over land in the summer (rising hot air). These continental atmospheric pressure systems reverse with season, air flows from the ocean to the land in the hemispheric summer and from the land to the ocean in the hemispheric winter. Where the configuration of the landmass results in winds flowing counter to the direction of the usual planetary winds, current reversals will occur in the ocean. In the opposite season, the monsoonal winds and the planetary winds will be in the same direction, with an increase in current speed and transport in the ocean. In the tropics, reversals of the trade winds in the hemispheric summer would occur if the continental pressure system were centred to the east of the oceanic area. Such a situation is present today in the tropical northern Indian Ocean.

Midlatitude west coast gradient currents

During the winter, off western coasts in the temperate zone, the presence of atmospheric low pressure systems offshore creates localized polarward surface winds along the coasts, and Ekman transport against them. This pile-up of water creates polarward current flow along the coast, counter to the equatorward flow on the eastern side of the oceanic high pressure system. The Davidson Current (Sverdrup et al., 1942), off the west coast of the United States, is an example of such winter seasonal currents.

WATER MASS FORMATION

Water masses are packets of water with relatively uniform or systematically varying temperature and salinity. Their distribution and coherence, therefore, would be density related. However, the distribution of surface water masses, due to the primary influence of the wind-driven circulation in the upper mixed layer, is the result of a combination of both thermohaline (density) and wind-driven factors. Surface water masses are formed and identified by the temperature-salinity conditions generally as the result of latitudinal changes in temperature (polarward cooling) and salinity (evaporation versus precipitation). Surface circulation will transport water from areas of primary formation, with gradual changes due to mixing or dilution with smaller water masses, or latitudinal temperature change, until the individual water mass encounters another water mass with differing properties. If the density difference is sufficient, the denser water mass will sink under the lighter mass at a density convergence, forming a boundary zone between the two water masses. This density convergence is different from the wind-driven convergence produced by the opposition of two Ekman transport vectors. Thus the boundaries of water masses are only subparallel with the climatic boundaries where the flow is zonal, and diverge significantly where the flow is meridional at continental barriers. Water mass nomenclature used here follows that in Sverdrup et al. (1942) and Tchernia (1980).

Equatorial water masses are of relatively low density, because they are warmer and generally less saline than other oceanic water, due to high precipitation in the tropics. More dense but basically tropical masses, or central water, show a wider range of warm temperatures, as a result of meridional flow and mixing and generally higher salinity at its polarward edge, as the evaporation maximum occurs at the latitude of the centre of the oceanic high pressure system. Surface waters in temperate to subpolar latitudes are cool with relatively low salinity due to higher precipitation at the boundary between the temperate and polar zones. The paired equatorial convergences define the region of the doldrums and highest rainfall. This least dense water floats on top of more dense tropical waters and is carried in the equatorial countercurrent. As the salinity difference caused by precipitation is slight, due to the enormous volume of the oceans, the paired equatorial convergences are weak. The equatorial convergences lie embedded within the major wind-driven planetary divergence associated with the oceanic low pressure near the equator (Sverdrup, 1947). Accordingly, at the global scale of the maps figured here, the equatorial water masses are merged with the central water masses. At the subtropical convergence, the warmer but more saline central water sinks with the colder but less saline subpolar waters. This boundary generally is polarward of the centre-line between the oceanic middle latitude high with warm central water flowing westward on the equatorward part of the anticyclonic oceanic high, and cooler subpolar water flowing east on the other side of the convergence. At the subtropical convergence, central water sinks, forming the upper part of the pycnocline and subpolar water sinks, forming intermediate water at the base of the pycnocline. During present glacial climates, there is a major convergence in the polar regions, where dense, cold polar water sinks, forming deep and bottom water below the pycnocline. The high latitude wind-driven planetary divergence occurs between the subtropical convergence and the polar convergence at the centre of the oceanic low-pressure system.

Water masses, because of both the uniformity and the continuity of variation in temperature and salinity within each water mass, commonly define major biogeographic regions (e.g., La Fond and La Fond, 1971; McGowan, 1972). In this paper, the figures depict the equatorial and central water masses as warm water, subpolar water masses as cool water, and polar water masses as cold water.

PALEOGEOGRAPHIC RECONSTRUCTIONS

The representations of significant continental blocks given in the figures presented here are based on the reconstructions of Scotese (1986). As most of the fossil record is in marine rocks, the outline of the continental blocks presented by Scotese (1986) must include the continental shelf and even parts of the slope and rise, with sea level somewhere inside the given outline. For the Sverdrup-Ekman model, the meridional barriers are not coincident with sea level, but with the edge of the continental block on the slope and rise. Thus, lack of detailed knowledge of the location of sea level on a given block at any given time is not a critical problem in reconstruction of the major global currents. Such imprecision is a problem for coastal currents, particularly for seasonal monsoonal and gradient currents. We have chosen to illustrate currents flowing over continental blocks seaward of the position of the major shield and platform areas, assuming these areas would have been shallow enough to be considered a barrier in the Sverdrup-Ekman model. However, the depicted currents would not flow necessarily over shallow platforms and regions of continental shelf shallower than 100 m. Areas within a continental block boundary known to be deep water continental margin, such as the Upper Ordovician section at Dobs' Linn, Scotland, are indicated as oceanic.

Kirshvink (1988, pers. comm.) has developed paleogeographic reconstructions that show differences in block orientation, particularly for Kazakhstan and North China, compared with Scotese (1986). Such differences would not affect the major currents, so are not included here.

RELEVANCE TO THE ORDOVICIAN

Imparting climatic zones and seasonality to the Ordovician is a risky business, particularly using analogues of the modern interglacial climate with a relatively high temperature contrast between the polar and the tropical latitudes. Although the Upper Ordovician is considered to have had a glacial climate, most of the Ordovician is considered to have had a milder climate than today (Frakes, 1979; Brenchley, 1984; Barnes, 1986). Also, the position of the continents and other landmasses, which might affect physical oceanography, is different from today (Scotese, 1986; Cocks and Fortey, 1982). Thus, the position of the climatic belts is conjectural. The approach used here is to assume that the oceans, at least since the late Precambrian, covered a greater surface area than the land. Accordingly, the primary driving force for surface circulation is that of wind blowing over water. A first-order circulation model would be that of a limitless ocean, similar to the Hough-Goldsborough models described by Stommel (1957). Thus, the influence of the varying positions of the continental blocks would be to impose modifications to the primary pattern, chiefly in permitting meridional flow, as described by the Sverdrup-Ekman equations.

For times when the climate was nonglacial, it is difficult to postulate the extent of the variability of the seasons. However, with a rotating spherical Earth, there always will be latitudinal differences in insolation and heat flux away from the zone of maximum insolation at the meteorological equator. Seasons occur as a result of the angle of the pole of rotation to the plane of the ecliptic, causing a variation in the length of the diurnal light-dark cycles both annually and latitudinally. Only if the pole of rotation was perpendicular to the plane of the ecliptic, so that light-dark cycles would be the same latitudinally, would there not be any seasons. However, there still would be climatic belts and a latitudinal difference in insolation with the heat flux from the equator to the poles. Vanyo and Awramik (1985), using data derived from stromatolites, indicated that the obliquity of the Earth's rotation to the ecliptic at 850 Ma was similar to the modern. Thus, seasonality in the Ordovician similar to modern conditions is not unreasonable.

The Ordovician was prior to the development of vascular land plants that reduce the land albedo. The terrestrial albedo in the Ordovician was probably high, similar to that of deserts. Therefore, even with a relatively mild global climate, if the landmasses were in the appropriate locations, regional modifications of the wind patterns, such as monsoons, could occur. The extent of modification would be a function of the seasonality and the areal extent of the land and its latitudinal position. Because large contiguous landmasses occurred on the Gondwanan continent in the Ordovician, it may be assumed that monsoons also would have occurred then. In this paper, due to lack of land in the northern hemisphere, figures designated northern hemispheric summer typify mean annual nonseasonal conditions. Due to the presence of land in the southern hemisphere, figures designated northern hemispheric winter or austral summer show the maximum reversal effects of the austral monsoons.

Both the Ekman transport and the velocity of geostrophic currents vary with the angular rotation of the Earth or planetary vorticity (Defant, 1961, p. 383). If the number of days in a year was significantly more in the Paleozoic (410 days in a year in the Silurian [Mazzullo, 1971]), then the geostrophic current velocities would be about 15 per cent more for the same pressure gradient seen today and net transport of water and circulation would be more vigorous. No current velocities or transport volumes are expressly quantified here, so any change in the angular rotation of the Earth would have a relative effect.

The International Union of Geological Sciences (IUGS) has published correlations charts for Australia, New Zealand, and Antarctica (Webby et al., 1981); Canada (Barnes et al., 1981); China (Shen-Fu, 1980); Kazakhstan and Middle Asia (Nikitin et al., 1986); the Near and Middle East (Dean, 1980); South America (Aceñolaza and Baldis, 1987); southwestern Europe (Hamman et al., 1982); and the United States (Ross et al., 1982). The major lithofacies patterns for intervals within the Ordovician may be constructed on the paleogeographic maps from data from these charts. Thick carbonate sections are found in blocks within the tropical regions as plotted by Scotese (1986). Throughout the Ordovician, North America, Australia, North China, Siberia (Taimyr) and parts of Kazakhstan were sites of carbonate sequences presumably deposited in shallow shelf seas. Starting in the Caradoc, the Balto-Scandia-Russian Platform also showed shallow shelf carbonate deposition. Thick shallow shelf sea accumulations of calcium carbonate only form in tropical and subtropical latitudes as a function of maximum light availability, because year-around light refraction drops markedly at about 35 degrees from the equator, which is the present poleward limit of Bahamian environments.

As Berry and Boucot (1973) discussed, based on Latest Ordovician-Early Silurian worldwide correlations, glacial and glaciomarine strata developed across North Africa, eastern South America, the modern Middle East, and some sites in Spain during the Late Ordovician. Simultaneously, glacioeustatic sea level lowering worldwide led to exposure of wide areas of former continental shelf. The effects of major continental glaciation about a pole in central Africa are seen in the stratigraphic record both in the form of glacial and glaciation-related deposits, and in the form of a major depositional hiatus. The geographic location of the glacial beds is consistent with the plate reconstructions from remanent magnetism indicated by Scotese (1986).

Ordovician graptolite faunal provincialism has been related to oceanic surface water temperatures by Skevington (1974, 1976). He pointed out that the most species-rich faunas are those that occur about the craton-platform areas that were sites of thick accumulations of shallow shelf carbonates. The species-rich faunas, those of the "Pacific" Faunal Region, were most likely tropical. The other graptolite faunas of the Ordovician (e.g., Erdtmann, 1984) have been grouped in the "Atlantic" Faunal Region. Such appellations using modern geographic names can be misleading, particularly if the the faunal region was based on climatic conditions. According to the present reconstructions, the "Pacific" region actually encompassed the tropical and subtropical areas, whereas the "Atlantic" region was in middle to high southern latitudes.

Based on considerations of prominent lithofacies patterns and on a consistent biogeography for graptolites, the rock and fossil data regarding the latitudinal (climatic zonal) positions for major cratons and shelf seas are consistent with paleomagnetic data. Accordingly, the maps showing major marine surface currents and water masses for the various depicted time intervals from the Late Cambrian to Early Silurian reflect consideration of three lines of evidence bearing on paleogeography: paleomagnetism, lithofacies patterns, and biogeography.

Figures 1 to 6 depict the principal features of the surface oceanography from the Late Cambrian through the Early Silurian. The following text concerning each of the maps is, in effect, an expanded figure caption showing how the basic oceanographic principles are used to reconstruct the oceanoaphic features. Specific references to sections or columns from the IUGS compilation charts or other compilations are given in the text to demonstrate the interaction among paleogeography and lithofacies to reconstruct the paleo-oceanography.

LATE CAMBRIAN

Figures 1A, B represent the surface oceanic current patterns in the Late Cambrian, and set the scene for the Ordovician. The major feature in the north polar region is open ocean, with continental blocks all south of 35°N.

Primary circulation

A zonal Boreal Ocean covered the north polar and north temperate zones. This ocean had two embayments extending to the equator between northern Gondwana (New Guinea and Australia) and Laurentia, each with a longitudinal extent of almost 180°. Depending on the level of these blocks in the northern hemisphere, the actual Boreal Ocean may have extended to the equator and covered the whole northern hemisphere. Temperate to cool water dominated in the Boreal Ocean due to the lack of significant meridional barriers to advect warm water from the tropical areas. Clockwise circulation in the two weak high pressure systems in the two boreal embayments advected cool water equatorward along the west coast of North America and New Guinea, and warm water polarward in the Arctic Islands and south Australia.

Emergent land on the equator included the Laurentian, Siberian, and Australian-Antarctic shields. Thus, the global oceanic low pressure systems probably were fragmented, with the major systems south of the high pressure centres of the Boreal Ocean embayments. The water current flow was east to west, parallel to the equator.

The African, South American, and Antarctic shield areas were in the southern hemisphere as part of Gondwana. A relatively small Austral Ocean was divided by separate blocks consisting of Siberia, Kazakhstan, and Balto-Scandia and the Russian Platform. These barriers produced two counterclockwise oceanic high pressure systems, which moved warm water south along eastern North America and Siberia and Kazakhstan, and cool water north along western Siberia and through Turkey, Iran, and South China. The Austral Ocean was connected to the Boreal Ocean on the west through a seaway between Laurentia and Gondwana (South America). This would produce a third counterclockwise oceanic high pressure circulation gyre in the southern hemisphere. The Gondwanan shields form the only continuous meridional barrier on the eastern side of the Austral Ocean. Thus, the easternmost oceanic high pressure gyre would be the most coherent. The two western high pressure gyres have limited eastern barriers, so would have relatively weak northern meridional flow on the eastern side. The two western-most gyres are commonly combined into the Iapetus Ocean, whereas the easternmost gyre is included in the Proto-Tethyian Ocean.

The Austral Ocean was open to the south pole with the Balto-Scandian barrier at about 60°S. Circulation was clockwise, governed by the oceanic low pressure.

Seasonal patterns

Figure 1A shows the effects of monsoons produced by the concentration of continental blocks in the southern hemisphere. During the southern hemisphere summer, anti-trade monsoon winds would be generated along the coastal and platform areas of Gondwana. In contrast to the equatorial currents shown in Figure 1B, these monsoonal currents would drive warm water eastward and polarward through North China, Borneo, Sumatra, New Guinea, and South China, countering the cool flow from the midlatitude oceanic high pressure systems. In the northern hemisphere, cool water would be advected southward along New Zealand and the east coast of Australia.

Divergences (upwelling)

Tropical divergences would occur along the equator with the most persistent in the Arctic Islands and southern Siberia. Due to monsoonal current reversals, upwelling occurred along the equator during the northern hemisphere summer in North China, New Zealand-Australia, and southern Laurentia. Near 60°S, planetary divergences occurred in North Africa, Iberia-Amorica, Balto-Scandia, and Turkey and the Near East. Seasonal divergences near 30°S in the austral summer produced upwelling off western coasts in South China. Upwelling would persist in the austral winter as a result of offshore advection due to the meeting of monsoonal warm and equatorward cool currents.

EARLIEST ORDOVICIAN: TREMADOC

The major paleogeographic changes from the Upper Cambrian were 1) the shift of both Laurentia and Gondwana to the south, increasing the latitudinal extent of the zonal Boreal Ocean; 2) the northward drift of a) Balto-Scandia, closing the Iapetus gap and increasing the width of the Tornquist Sea between Balto-Scandia and the parts of Western Europe attached to North Africa and the main Gondwanan continent and of b) Siberia, maintaining the open ocean connection between Iapetus and Proto-Tethys in the Austral Ocean (Figs. 2A, B).

Primary circulation

The zonal portion of the Boreal Ocean expanded to about 30°N, limiting meridional flow in the northern hemisphere to the tropic zone with a weak development of oceanic high pressure in subtropical to southern north temperate regions. With the shift of the major landmasses to the south, the westward flowing north equatorial current expanded meridionally in the tropical regions. The preponderance of zonal flow in the northern hemisphere would produce a temperature distribution in the surface ocean, with isotherms parallel to lines of latitude with minor deflections of cold water to the south on the eastern side (Western Canada and New Guinea), and of warm water to the north on the western side (Arctic Islands and New Zealand) of the boreal embayments.

In the southern hemisphere, the westward flowing south equatorial current was segmented by the the Laurentian, Siberian, and Gondwanan shields. The northern movement of Balto-Scandia and the southerly shift of South America combined to merge the oceanic high pressure gyres of the Upper Cambrian into one gyre in the Iapetus Ocean with weak northerly flow on the eastern side. Circulation in the midlatitude Proto-Tethys oceanic high would be strong on the western side and weak on the southerly limb at the juncture with Iapetus.

At high southern latitudes, the Tornquist Sea and its Proto-Tethys counterpart were astride the planetary oceanic low pressure system producing strong clockwise circulation centred at about 60°S.

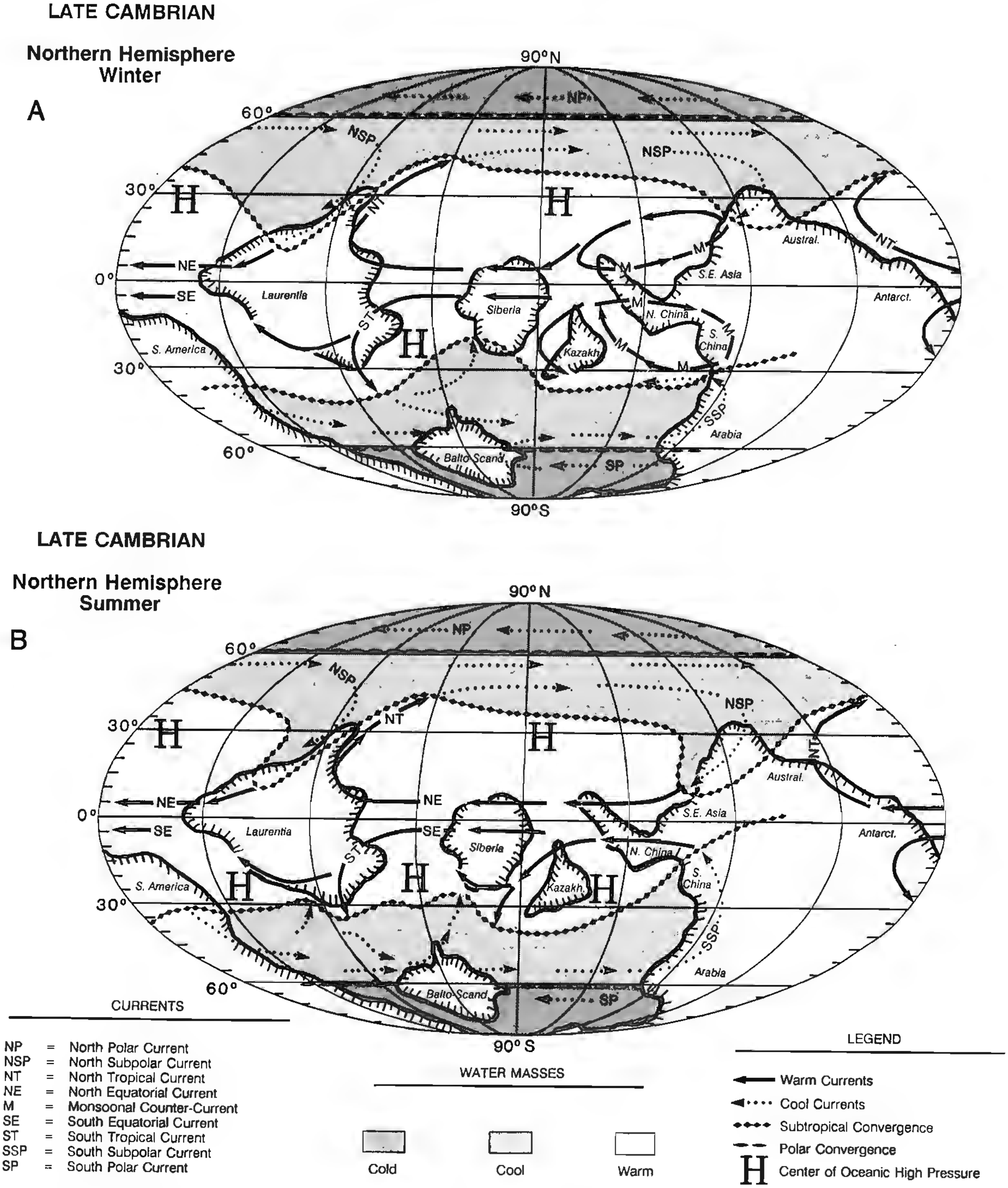

Figure 1. Surface current circulation and surface water masses in the Late Cambrian. Paleogeography from Scotese (1986). **A,** Northern hemispheric winter and southern hemispheric summer. **B,** Northern hemispheric summer and southern hemispheric winter.

Seasonal patterns

During the austral summer in the Tremadoc (Fig. 2A) current reversals would develop in the subtropical regions along the shores of the Gondwanan continent similar to those in the Upper Cambrian. Although the direction of the equatorial current during the austral monsoon would be reversed (actually, by modern analogy: the equatorial countercurrent) to flow east; there would be little thermal change in the northern hemisphere due to the already weak meridional flow. The major changes in temperature would be in the southern hemisphere in South China where cool water advecting from the south would be replaced by warm tropical water driven by the monsoon.

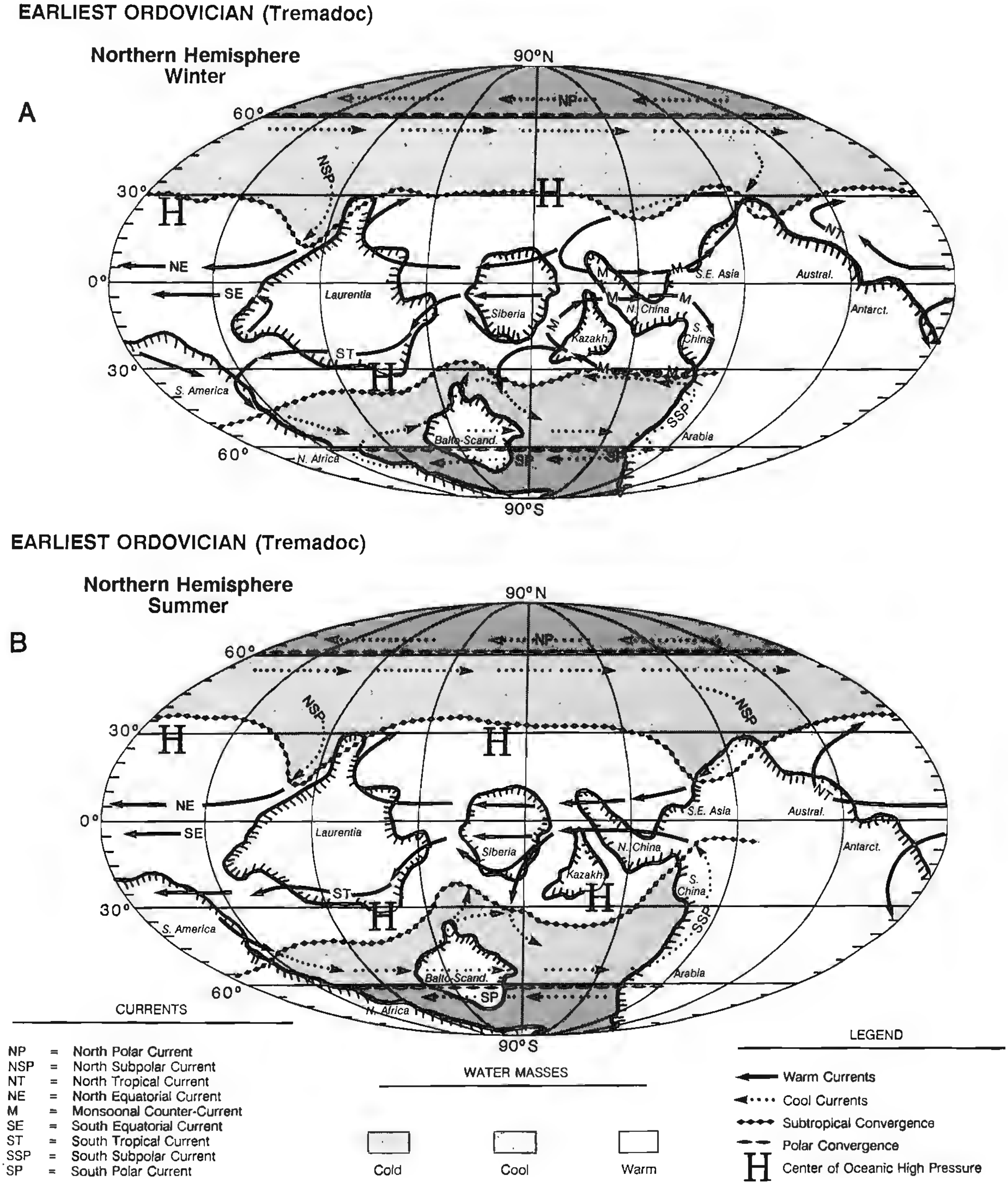

Figure 2. Surface current circulation and surface water masses in the Earliest Ordovician (Tremadoc). Paleogeography from Scotese (1988). **A,** northern hemispheric winter and southern hemispheric summer. **B,** northern hemispheric summer and southern hemispheric winter.

Divergences (upwelling)

Major tropical divergences would occur along the equator in western North America (Mayflower Schist, Ross et al. [1982], no. 14) and Arctic Islands in Laurentia, Siberia, North China (Hsienerhtai Fm, Erdtmann [1982], no. 19), Weddel Sea (Antarctica), and southernmost South America. In the southern hemisphere, the planetary divergence at 60°S would affect North Africa, Maguma (Barnes et al. [1981], no. 68), Avalonia (Barnes et al. [1981], no. 87), South Ireland and Britain, the Baltic area (Barnes et al. [1981], no. 66), the southern Ukraine, Asia Minor (Karadere Formation, Dean [1980], no. 4), the Levant, and northwestern Arabia.

Less intensive seasonal upwellings would occur in the summer in southwestern China (Pingtou Formation, Shen-Fu [1980], no. I) to Iran south of 30°S, augmented by offshore flow caused by the Gondwanan monsoonal currents merging with the cool eastern gyre of the oceanic Proto-Tethyan high. Depending on the latitude of Laurentia, potential upwelling would occur on the east side of the Iapetus high pressure gyre from eastern Canada (Levis Formation, Barnes et al. [1981], no. 66), Scotland, and northern Newfoundland (Cow Head, Barnes et al. [1981], no. 74). In the northern hemisphere, upwelling might occur in New Zealand (Webb Formation, Webby et al. [1981], no. 59) as a result of offshore flow where the northern branch of the Gondwanan monsoonal current flowing south meets the easterly deflection of the north equatorial current.

EARLY ORDOVICIAN (ARENIG)

The major change in the Arenig (Figs. 3A, B) from the Tremadoc was the continued northerly drift of Balto-Scandia and Siberia, and the southerly drift of Gondwana. Southern Ireland and Great Britain, and Iberia-Amorica began to drift northward, away from North Africa and Gondwana.

Primary circulation

Zonal flow continued to dominate the circulation in the north polar and temperate belts. The movement of the Siberian Shield astride the equator deflected the west-flowing north equatorial current slightly. In the southern hemisphere, the twin barriers of Balto-Scandia and Siberia emphasized the independent oceanic midlatitude high pressure systems in the Iapetus and Proto-Tethyian oceans. However, some transport probably occurred, of cool water from Iapetus eastward across the north side of Balto-Scandia, and of warm water from tropical Proto-Tethys westward across Siberia and Kazakhstan. At high latitudes, clockwise circulation about the oceanic low pressure system centred at 60°S and the Tornquist Sea persisted, connecting Iapetus and Proto-Tethys on the south. The increase in land in the south polar zone probably reduced the transport volume in the southern part of the gyre, which would cause increased northerly advection along the west coast of Balto-Scandia, or more likely, against the Gondwanan barrier to the east of Balto-Scandia.

Seasonal patterns

The effect of the Gondwanan monsoon probably vanished in the northern hemisphere due to the continued southerly drift of the main Gondwanan continent and concentration of landmasses in the southern hemisphere. However, as a consequence of this, monsoonal conditions intensified in the subtropical regions of the southern hemisphere. During the austral summer, monsoonal warm water would flow south through South China and into Central Asia, whereas during the austral winter the monsoon would intensify the planetary oceanic high pressure flow of cool water northward into the subtropics even as far as North China.

Divergences (upwellings)

With landmasses on the equator, the planetary equatorial divergence was segmented into three parts. The longest extends from the west coast of Laurentia (Phi-Kappa Formation, Ross et al. [1982], no. 9 and Kanosh Formation, Ross et al. [1982] no. 25) to the east coast of Gondwana (Victoria, Australia, Webby et al. [1981] no. 23; Lancefield, Erdtmann [1982], no. 17). Smaller segments of the equatorial oceanic low pressure zone were between Laurentia and Siberia and between Siberia and Gondwana, over what is now Kazakhstan, North China, Sumatra, and Borneo. Upwelling in the Gondwanan sector would be reduced by the monsoon, especially south of the equator.

High latitude divergence at about 60°S in the southern hemisphere would be seen in Venezuela (Cerrojon Formation, Aceñolaza and Baldis [1987], no. 59); southern Appalachia (Suwanee Basin, Ross et al. [1982], no. 100), northwest to west Africa, Nova Scotia (Halifax, Barnes et al.[1981], no. 67); Avalonia (Wabana Formation, Barnes et al.[1981], no. 83); Wales, Denmark to the Ukraine on Balto-Scandia (Lindström and Vortische [1953]); northwestern Arabia-Asia Minor (Bosporus, Dean [1980], no. 4; Sedisehir Formation, Dean [1980], nos. 5,6,7,8-12).

Summer seasonal upwelling in the southern hemisphere would occur at midlatitudes off the western coasts of Central America-southern Mexico, and northern Norway-Spitzbergen (Valhalfonna, Archer and Fortey [1974]; Fortey [1984]), with the best developed system in western South China (Ningkou Shale, Shen-Fu [1980], nos. XV,XVI; Chungyi Formation, Shen-Fu [1980], no. XVII, and Indo-China).

MIDDLE ORDOVICIAN (LLANDEILO-CARADOC)

In the Middle Ordovician (Figs. 4A, B), Gondwana continued its southern drift. The major change from the Arenig was the formation of an eastern meridional barrier in the Iapetus Ocean, because of the merging of Laurentia and Siberia in the tropics and the near attachment of Balto-Scandia to Siberia. In high southern latitudes, the Rheic Ocean opened to separate the Meguma, Avalonia, South Ireland-Great Britain, Belgium sliver from Iberia-Amorica and North Africa. Accordingly, the Iapetus and Proto-Tethyan oceans were more sharply delineated, although a seaway connection probably existed between them.

Primary circulation

In the northern hemisphere, zonal circulation continued in the Boreal Ocean north of 30°N, potentially extending to the equator with only weak meridional flow west of Laurentia and east of New Zealand-Australia. Due to the shift of Gondwana toward the south, the north equatorial current flowed westward, relatively

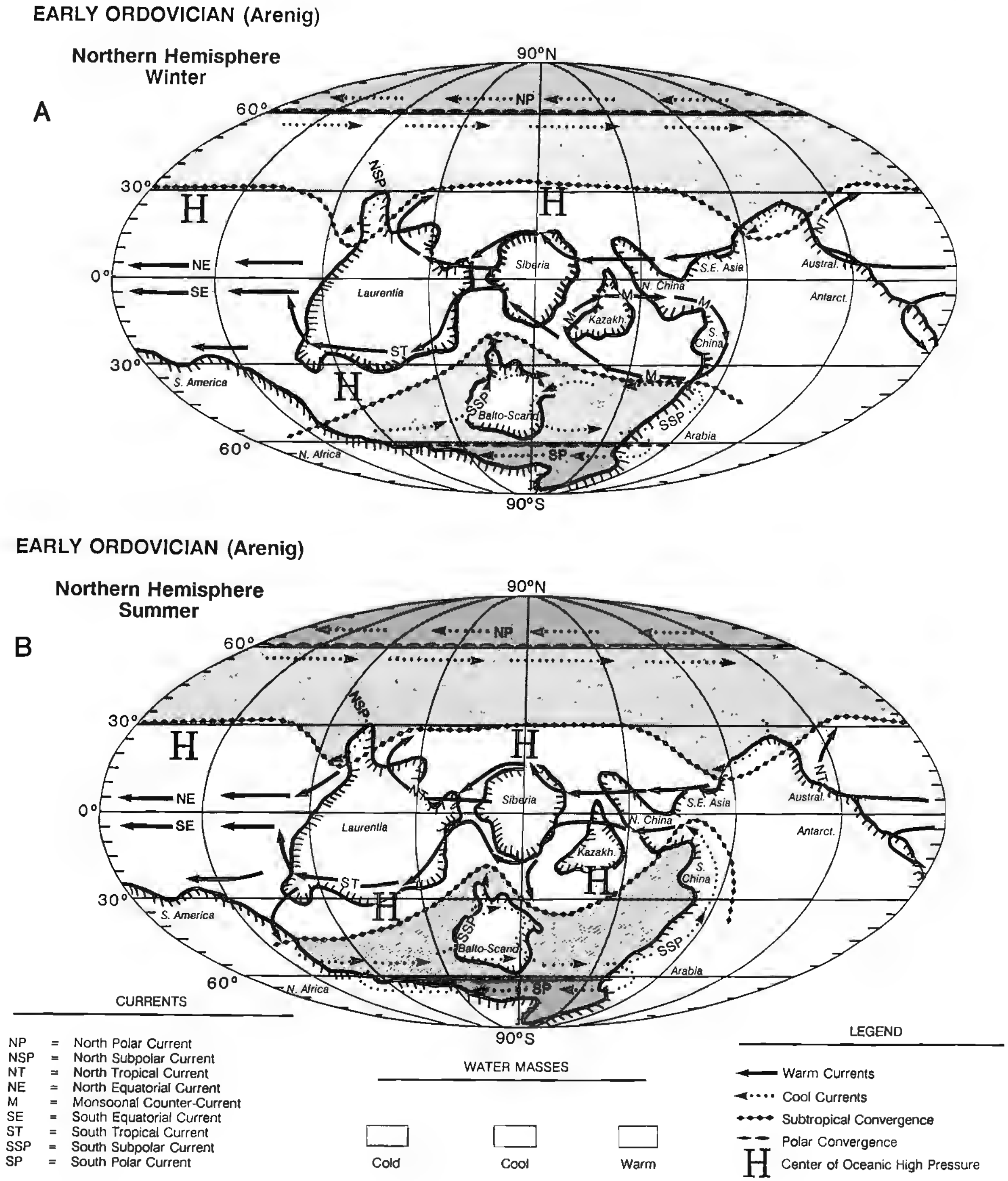

Figure 3. Surface current circulation and surface water masses in the Early Ordovician (Arenig). Paleogeography from Scotese (1988). **A,** northern hemispheric winter and southern hemispheric summer. **B,** northern hemispheric summer and southern hemispheric winter.

unimpeded from the west coast of Laurentia to Siberia with only the Laurentian and Siberian shields under shallow water or emergent. The south equatorial current was blocked on the west by the Australian-Antarctic shields, but re-formed in Southeast Asia through North China and Kazakhstan until the current was blocked by the Siberian Shield.

At midsouthern latitudes, three oceanic high pressure gyres were maintained. With South America drifting south of 30°S, a distinct high pressure system could develop in the Proto-Pacific. This system was linked with the Iapetus anticyclonic gyre in the gap between Mexico and South America south of 30°S. At this time, both Iapetus and Proto-Tethys had

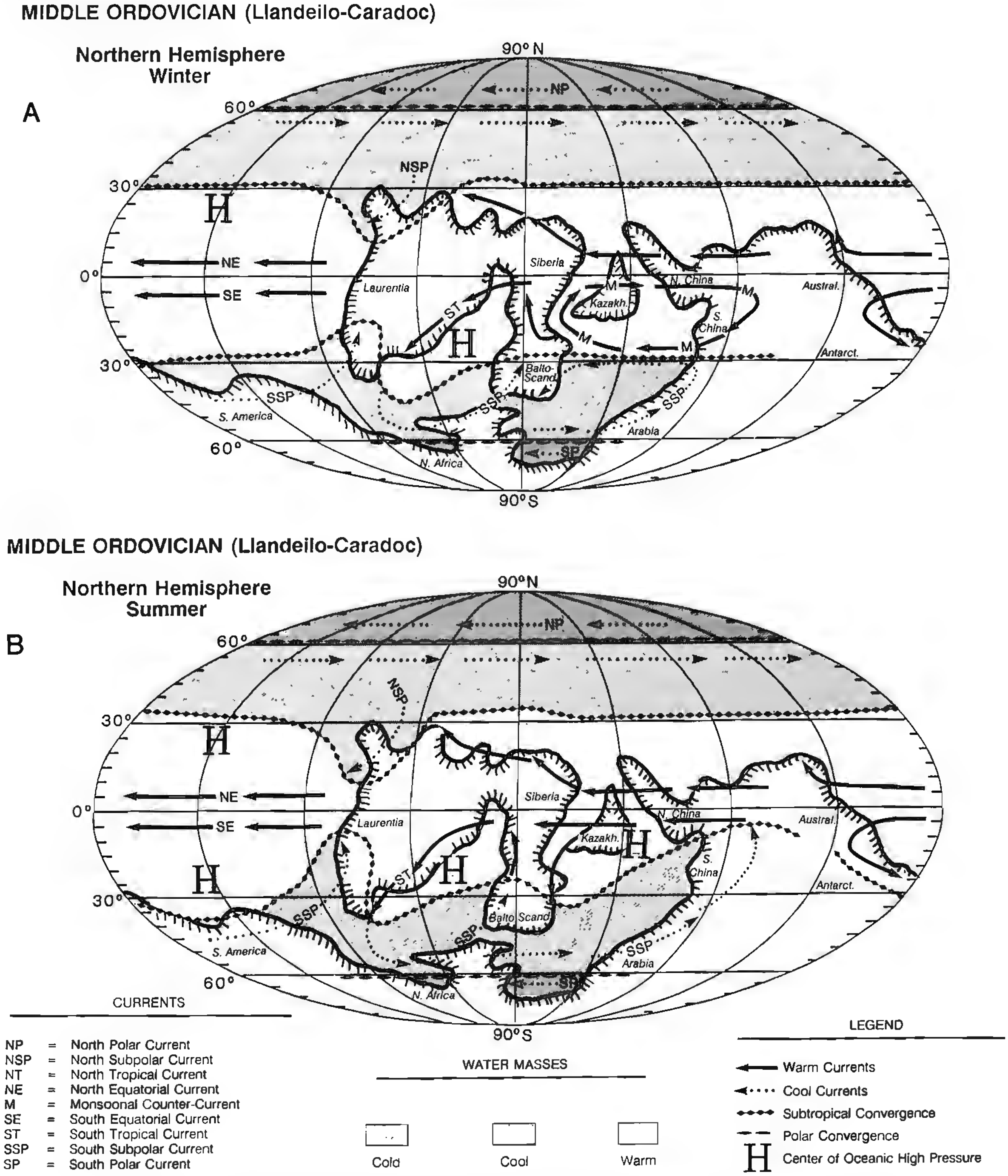

Figure 4. Surface current circulation and surface water masses in the Middle Ordovician (Llandeilo-Caradoc). Paleogeography from Scotese (1988). **A,** northern hemispheric winter and southern hemispheric summer. **B,** northern hemispheric summer and southern hemispheric winter.

well developed meridional barriers, although eastward flow from Iapetus probably leaked into Proto-Tethys via the Tornquist Sea. The circulation in the Rhaeic Ocean and the southernmost part of Proto-Tethys were governed by cyclonic flow about the 60°S oceanic low pressure system.

Seasonal patterns

Monsoonal current reversals during the austral summer would occur along the northwest shores of Gondwana (Fig. 4A) due the continuing presence of the large landmass to the south.

Divergences (upwelling)

The planetary divergence at the equator was divided into two major oceanic low pressure systems with the most persistent open ocean system located between the west coast of Laurentia (Phi Kappa, Ross et al. [1982], no. 9; Vimini/Valmay, Ross et al. [1982], nos. 12,13,14,17,18; in Canada: Skoki Formation, Barnes et al. [1981], nos. 4,5,7,8,9 and Road River Formation, Barneset al. [1981], nos. 10,12,13) and the north coast of Gondwana (Victoria, Webby et al. [1981], nos. 23,26,27). The system between Gondwana and Siberia would be modified in the southern hemisphere by the austral winter monsoons with modified upwelling in eastern Kazakhstan and central North China.

In the vicinity of 60°S, upwelling along the planetary divergence would occur in southern Appalachia (Suwanee, US, no. 100); west Africa, Iberia (Hamman et al. [1982], nos. 27,28); west Armoricia (Urville, Le Pissot, Hamman et al. [1982], nos. 3,4,5; Schistes d'Angers, Hamman et al. [1982], nos. 7,8); Belgium, and northwest Arabia (Jordan, Dean [1980], no. 20; Arabia Hanadir Shale, Dean [1980], no. 21).

Seasonal upwellings in the summer along western coasts of middle latitude oceanic high pressure systems would occur in the southern hemisphere of Central America, western Norway, and western South China and Indo-China (Shen-Fu [1980], nos. XIII,I). The seasonal upwellings in China at middle latitudes would be augmented by the divergence of the currents caused by the Gondwanan summer monsoon.

LATE ORDOVICIAN (ASHGILL)

Gondwana continued to move south with the South American, African and parts of the Antarctic shields in high southern latitudes astride the south pole (Figs. 5A, B). The Tornquist Sea closed, linking the Meguma through Belgium blocks with Balto-Scandia. Iapetus narrowed with a seaway at the equator separating Laurentia-Siberia from Balto-Scandia. The Rhaeic Ocean widened, which acted as the major water communication between the widening South Pacific and Proto-Tethys.

Primary circulation

Zonal flow in the Boreal Ocean continued above 30°N. The alignment of the Laurentia, Siberia, and Balto-Scandia blocks, which reduced Iapetus essentially to an embayment, created a meridional barrier in the tropical regions with significant meridional flow on both sides of that block. The most persistent equatorial currents flowed to the west between Laurentia and northern Gondwana. The major oceanic high was in the South Pacific, with smaller highs in Proto-Tethys and Iapetus. With ice in the south polar regions and limited tropical circulation in the northern Iapetus and Proto-Tethys, the surface waters in the southern hemisphere, except in the western South Pacific, were noticeably cooler than in the Middle Ordovician. Circulation about the 60°S oceanic low was limited by the extensive landmasses and the possibility of ice shelves covering significant parts of the gyre. With glaciation in the southern hemisphere and the lowering of sea level, the areal extent of the shelf seas was limited so that the meridional barriers would be more uniform. These conditions would lead to more intense circulation in the open ocean.

Deep circulation

Glaciation in the Ashgill produced conditions favoring the formation of cold deep water during the austral winter, which could affect the depositional environment on the continental rise as seen in deposits on deep sea fans. Prior to the glaciation, deep sea fan deposits such as at Dob's Linn, Scotland were anoxic black shale. For pre-Ashgill times, in tropical areas, the waters in the lower part of the mixed layer and the pycnocline were derived from the sinking of relatively warm central water at the subtropical convergence. However, the upper Ordovician deposits are more oxic mudstone, indicating ventilation of middle depths. The narrowing of Iapetus and the position of the Gondwanan shore near the pole suggests that Iapetus was ventilated by cold deep water sinking during the formation of sea ice in the austral winter. Also, during glaciation, subpolar water must have been colder and more oxygen-rich. This water, also sinking at the subtropical convergence, may have chilled the overlying central water and ventilated it by vertical advection. Thus, although Dob's Linn was associated with tropical Laurentia, at this time, with warm surface waters, the water in the pycnocline overlying the offshore fans was cold and relatively oxygen-rich. On the other hand, deep sea fan deposits on the continental slope and rise from the late Ordovician of China continue to be anoxic black shale. Accordingly, cold water sinking from the Gondwanan ice shelf at the south polar convergence might have been blocked from entering the Proto-Tethyan Ocean due to the bathymetric ridge-rise associated with the opening of the Rhaeic Ocean. The ridge apparently was high enough to, at least partially, block subpolar water sinking at the subtropical convergence, permitting dilution by the larger volume of central water formed in the larger Proto-Tethys. Alternately, this deep water may not have contained sufficient oxygen to vertically ventilate the pycnocline or to compensate for the organic productivity in Proto-Tethys.

Seasonal patterns

Monsoonal circulation in the Indo-Australian-Antarctic regions of Gondwana likely intensified during glaciation, particularly along the northeastern subtropical coast.

Divergences (upwellings)

Equatorial upwelling along the planetary divergence would persist from the west coast of Laurentia (Antelope Valley, Ross et al. [1982], no. 17; Road River Formation, Barnes et

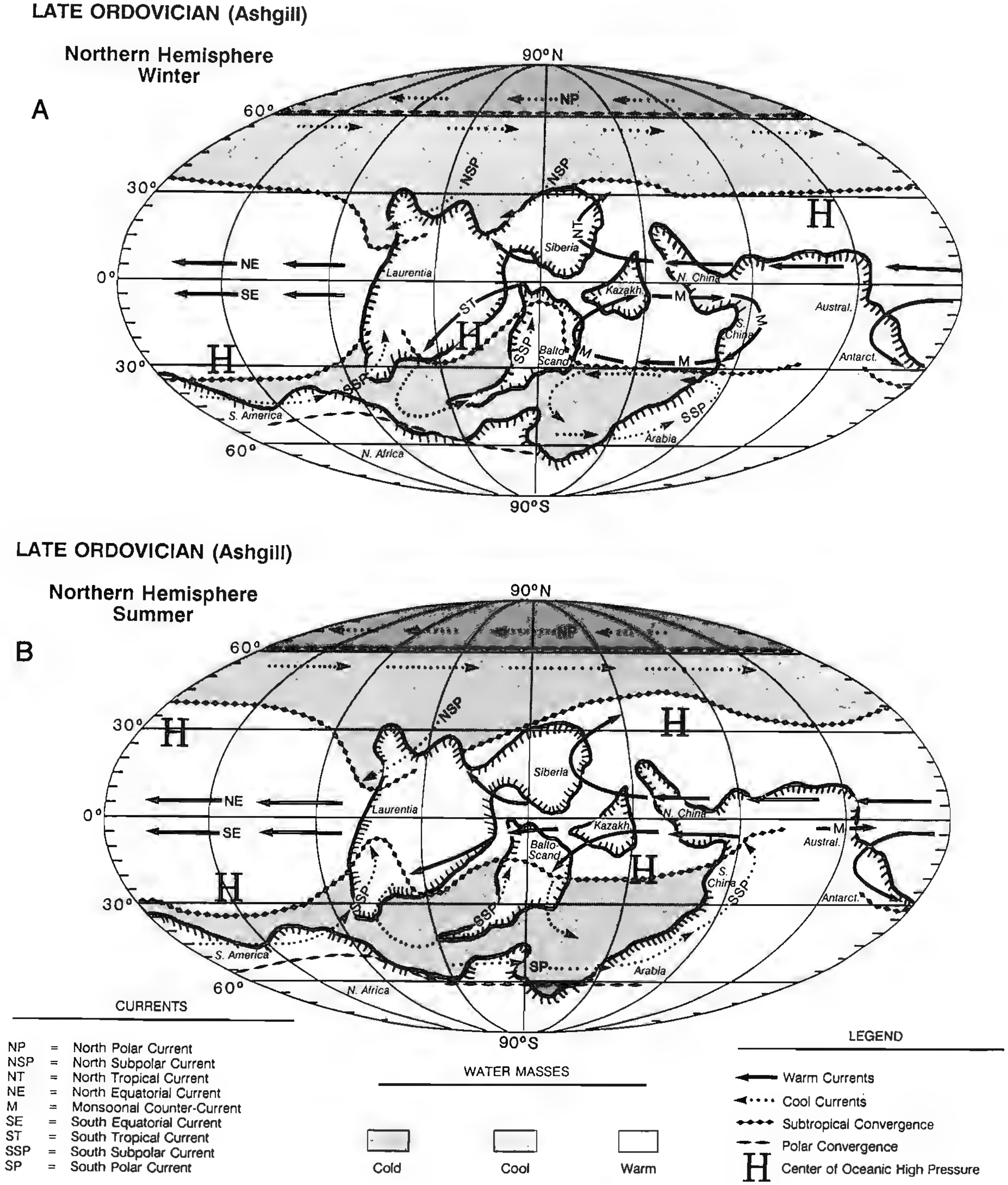

Figure 5. Surface current circulation and surface water masses in the Late Ordovician (Ashgill). Paleogeography from Scotese (1988). **A,** northern hemispheric winter and southern hemispheric summer. **B,** northern hemispheric summer and southern hemispheric winter.

al.[1981], no. 12) to the east coast of Gondwana (Australia, Melbourne, Webby et al. [1981], nos. 25,26,27; Victoria [Webby et al. [1981], nos. 29,30,31). The austral polar latitude divergence was limited by ice and land with upwellings seen in Iberia (Orea Shale, Hamman et al. [1982], nos. 23,24,26); western France (glaciomarine pelites and fragments, Hamman et al. [1982], nos. 1,2,3); and northwest Arabia (Syria, Dean [1980], no. 19).

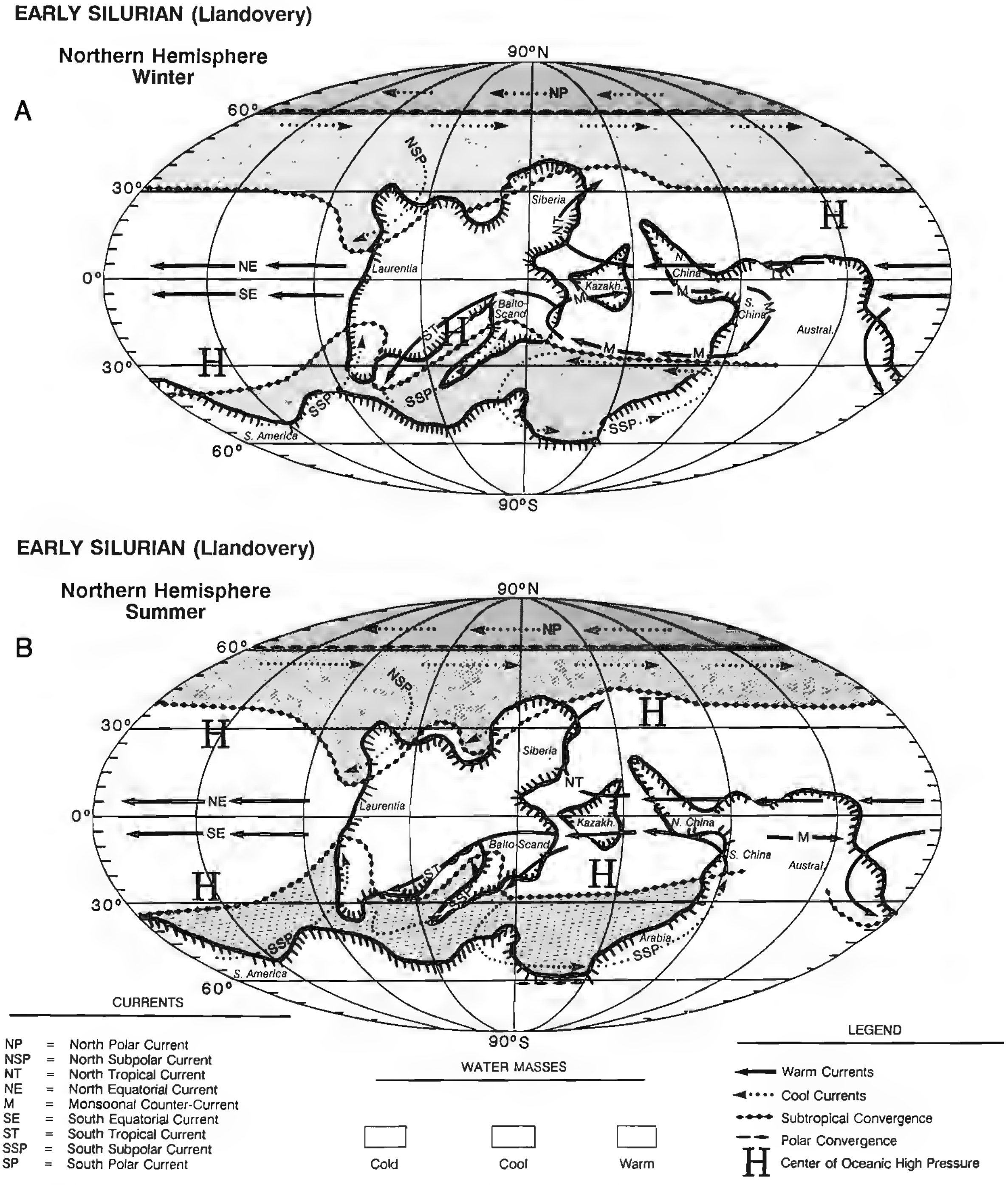

Figure 6. Surface current circulation and surface water masses in the Early Silurian (Llandovery). Paleogeography from Scotese (1988). **A,** northern hemispheric winter and southern hemispheric summer. **B,** northern hemispheric summer and southern hemispheric winter.

LOWER SILURIAN (LLANDOVERY)

Deglaciation occurred with the shift of Gondwana away from the south pole (Figs. 6A, B) Siberia moved above 30°N, reducing the zonal Boreal Ocean. Laurentia moved south as South America moved north, constricting the connection between the Proto-South Pacific with the austral Iapetus and Rhaeic oceans. This constriction would force more cool water along the coast of Central America and Mexico in a manner similar to the modern Peru-Chile (Humboldt) Current, which is enhanced by the constriction of the Drake Passage. The limited supply of cool water moving into the southern Iapetus and Rhaeic oceans permitted more warm water from the tropics to flow into these areas, particularly on the western sides of the oceans.

SUMMARY

Given the basic assumptions discussed above, oceanographic conditions in the Ordovician have been reconstructed. The characteristic features of oceanic conditions in the Ordovician include:

1. Zonal circulation in the polar and most of the temperate regions of the northern hemisphere due to a globe girdling Boreal Ocean.
2. Meridional barriers in the tropical northern hemisphere, which permitted weak north-south flow in the sub-tropics.
3. Westward-flowing equatorial currents were discontinuous until the Late Ordovician as the result of numerous equatorial landmasses and shallow seas. Although landmasses would impede equatorial zonal flow, a mixture of tropical and subtropical waters would produce vast areas of central water with relatively uniform temperature conducive to carbonate deposition in shallow areas. Planetary divergence and upwelling near the equator would be limited by the reduced wind fetch caused by the discontinuity of true oceanic depths there.
4. Three small to modest-sized oceans existed at middle latitudes in the southern hemisphere, with at least partial meridional barriers. The easternmost, the Proto-Tethys, had the most persistent barrier on its eastern shore. As a consequence of the north-south barriers, warm water flowed polarward on the western side of the oceans and cool water flowed equatorward along the eastern side. Seasonal upwellings would enhance productivity at middle latitudes along the eastern coast of these oceans.
5. Monsoonal conditions produced by the concentration of landmasses in Gondwana produced seasonal current reversals on the subtropical Gondwanan shore of Proto-Tethys. This permitted mingling of tropical (old Pacific) and subpolar/temperate (old Atlantic) faunas.
6. Austral polar and subpolar areas were land during most of the Ordovician. Increased productivity at the planetary divergence at the polartemperate boundary would have occurred in embayments of the three southern oceans.
7. Pycnoclinal and deep circulation was governed chiefly by sinking of warm central water until the Late Ordovician. During glaciation, colder and denser intermediate and deep waters ventilated the pycnocline, particularly in the narrow Iapetus Ocean. However, tectonic ridge-rise systems, such as those that produced the Rhaeic Ocean could block pycnoclinal ventilation in tropical regions of Proto-Tethys.

Consequences of the evolving Ordovician oceanic circulation patterns would be: 1) enhanced upwelling (and increased organism diversity) on the southeastern side of a bounded southern ocean (Iapetus in the early Ordovician); 2) inhibited upwelling at the equatorial planetary divergence during the presence of the equatorial barrier; 3) significant deep ventilation in the global ocean during glaciation; and 4) low new nutrient availability in an isolated Boreal Ocean.

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Bathymetric biosignals and Ordovician chronology of eustatic variations

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Chen, Junyuan, Bathymetric biosignals and Ordovician chronology of eustatic variations; in *Advances in Ordovician Geology, C.R. Barnes and S.H. Williams (ed.), Geological Survey of Canada, Paper 90-9, p. 299-311, 1991.*

Abstract

The Ordovician Period has long been regarded as a time when the number of plates and the length of the ridge systems were maximal, and thus sea level was high, with a prolonged 350 m rise from middle Arenigian to Caradocian. Although the high latitude disposition of Gondwanaland was favourable for ice and snow, triggering the formation of large ice sheets at the end of the Ordovician, large lowland ice sheets were generally absent due to the existence of a greenhouse state. The ice sheets may have been actively enlarged due to increased precipitation during episodes of anomalously enhanced volcanic activity. Rapid formation and subsequent melting of the large ice sheets may have led to rapid changes in sea level on the order of between 50 m and 300 m.

Emphasis is placed on the interpretation of the bathymetric biosignal of the chambered cephalopod animals, and its use in establishing the history of quantitative bathymetric variations in broadly separated areas. Thus, a chronology of eustatic variations has been established quantitatively from the evidence of the bathymetric fluctuations in certain areas, assuming these fluctuations to be a mirror of eustatic variation.

Résumé

L'Ordovicien est depuis longtemps considéré comme une période pendant laquelle le nombre de plaques et la longueur des systèmes de dorsales ont atteint des niveaux jamais dépassés; c'est pourquoi cette période a été marquée par des niveaux de mer élevés avec une augmentation prolongée de 350 m de l'Arénigien moyen au Caradocien. Même si la latitude élevée du Gondwana était favorable à la formation de glace et de neige, donc de vastes calottes glaciaires à la fin de l'Ordovicien, l'effet de serre a généralement empêché la formation de grandes calottes glaciaires dans les basses terres. Les calottes glaciaires ont pu s'accroître par suite d'une augmentation des précipitations durant des épisodes d'activité volcanique anormalement intense. La rapide formation et la fusion subséquente des grandes calottes glaciaires ont eu des répercussions sur les niveaux de la mer qui ont subi des changements rapides de l'ordre de 50 m à 300 m.

L'accent porte sur l'interprétation du biosignal bathymétrique des céphalopodes cloisonnés et de son utilisation pour établir l'histoire des variations bathymétriques quantitatives dans des zones très éloignées l'une de l'autre. Par conséquent, une chronologie des variations eustatiques a été établie à partir de données quantitatives des fluctuations bathymétriques dans certaines zones, en supposant que ces fluctuations sont le reflet de variations eustatiques.

INTRODUCTION

Sea level is the most important reference level on the earth's surface. Evidence for its synchronous changes are remarkable in the Ordovician Period. The synchroneity of significant variations of sea level no longer seems enigmatic, and the variations are recognized as the combined changes in the volume of water and the volume of the oceanic basins (Hays and Pittman, 1973). The Ordovician Period has long been regarded as thalassocratic (Jaanusson, 1984), when the

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number of plates and the length of the ridge system were maximal, and thus sea level was high with a prolonged rise from middle Arenigian to Caradocian time of approximately 350 m. The greenhouse state characterized the Ordovician Period, with low gradients in latitude temperature and high mean temperature. Although high latitude disposition of Gondwanaland (Erdtmann, 1982, 1986; Ziegler, 1981) was favourable for ice and snow and a potential trigger for the formation of large ice sheets over the Ordovician Period, large lowland ice sheets were generally absent for most of the period. The ice sheets may have been actively enlarged due to increased precipitation during episodes of increased solar flux, or due to intense veils of dust in the high atmosphere during episodes of abnormally enhanced volcanic activity. The rapid formation and subsequent melting of the large ice sheets led to rapid changes in sea level.

The purpose of this paper is to provide both a new method of quantitative bathymetric analysis and a working hypothesis for approaching the quantitative consideration of sea level changes during the Ordovician Period. Emphasis is placed on the interpretation of the bathymetric biosignal of the cephalopod shell, and its use in establishing the history of bathymetric variations in broadly separated regions. Quantitative portrayal of global changes in sea level no longer seems difficult, because eustatic change can be illustrated quantitatively from bathymetric fluctuations in certain regions, assuming these fluctuations were a reflection of the eustatic fluctuations.

QUANTITATIVE BIOSIGNAL OF BATHYMETRY

Implosion depth and first failure

Chambered cephalopod animals possess subatmospheric pressure within their chambers (Fig. 1), and thus their shells yield a signal of some finite depth known as the implosion depth. The animals would be killed by implosion at depths beyond the implosion depth. The shell of chambered cephalopods may show three areas of failure under high pressure: implosion of the outer shell wall, inward implosion of the last septum, and explosion of the siphuncle into the chambers. The question of which part of the chambers would fail first under high pressure in living *Nautilus* specimens has been debated by different authors. Denton and Gilpin-Brown (1966) believed that the connecting rings would fail first under high pressure. But Chamberlain and his collaborators (Chamberlain and Chamberlain, 1985; Chamberlain and Pillsbury, 1985) suggested that the septal sutures were the weakest points. Studies on implosion depths and the nature of implosion in the fossil record (Westermann, 1973; Chen, 1988a,b) have previously demonstrated that 1) the shell wall is usually much thicker, and hence stronger, than the septa because it also has to withstand the bending movement of the entire shell, and sudden attack; 2) connecting rings probably performed multiple functions other than withstanding the ambient pressure; 3) septa are favourable indicators of implosion depth, because they are frequently imploded under high water pressure during postmortem sinking.

SSI (Septal Strength Index) and its implication

The strength of the septa against implosion is estimated from the ratio of thickness (δ) to radius (R), and it is calibrated on experimental data for the implosion of *Nautilus* and *Spirula* (Fig. 2) (Westermann, 1973). Assuming that fossil cephalopods behaved similarly, a septal strength index (δ/R x 1000) of 10, corresponding to a depth of 300 m, suggests an actual depth limit for the animal of about 200 m (Fig. 3) (Westermann, 1973). Chen (1988a,b) has made systematic observations of the SSI and its distribution using several thousands of fossil cephalopod samples. The results of comparative analysis of SSI distribution between different fossil assemblages have led to the conclusions that 1) cephalopods originated in

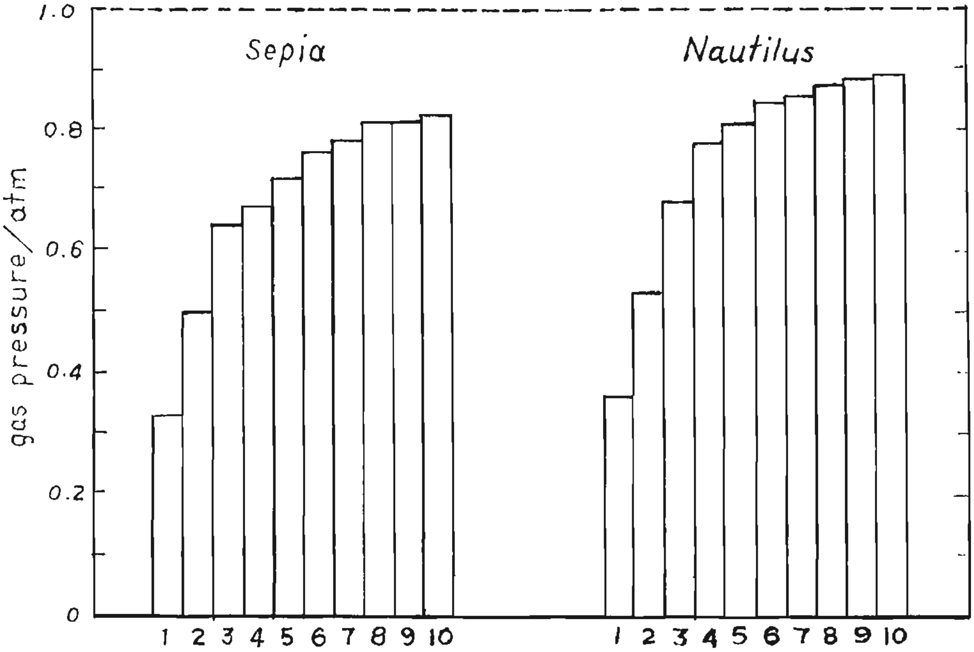

Figure 1. Air pressure distribution in successive air chambers of living cephalopods (from Denton, 1974).

shallow seas, carrying an SSI of approximately two in the early cephalopods, indicating they were inhabitants of shallow water (depths up to 40 m); 2) all the early Paleozoic cephalopods were continental shelf inhabitants; 3) all the populations of cephalopods lived in the same shelf sea, but adapted themselves to a wide areal dispersion within, but not beyond, the shelf break. Hence, all the taxa within a single fossil assemblage, and all the fossil assemblages within a narrow time slice of an epeiric sea, show about the same implosion depths related to the lower limit of the epeiric sea. The lower Paleozoic cephalopods under our investigation were dominantly nektobenthic. Their tendency of spatial dispersion for feeding, mating, and daily and seasonal locomotion activities resulted in maximal dispersal in the epeiric sea. SSI observations show no tendency for areal extension beyond the shelf break to over the slope region; its topographic gradient would be unproductive for cephalopods (Fig. 4).

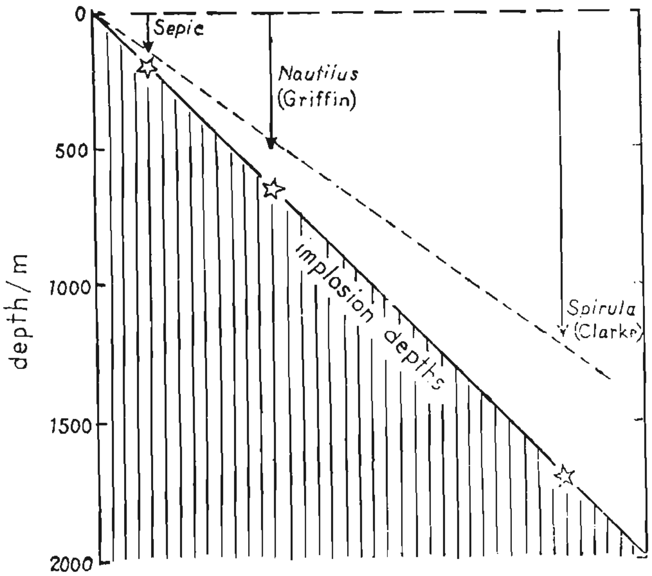

Figure 2. Relation between shell strength (implosion depth) and depth range of living cephalopods (from Denton, 1974).

Postmortem distribution

Views of the possibility of postmortem transportation of cephalopod shells for unknown distances from the actual habitat of the animals are broadly divergent among researchers. The results of field investigations of the fossil record show a variable postmortem history of cephalopods, but show little evidence of marked postmortem drifting after the animals died near or on the seafloor. The animals would have become waterlogged within hours of their death and their remains would have taken longer to decay if the water had been cool and deficient in oxygen. The possibility that the animals were buried in situ, without postmortem drifting, has been firmly established by the fossil evidence of *Cyclolituites*, which was a compressed coiled nautiloid from the Middle Ordovician Datienba Formation (Fig. 5). The fossil was buried in a vertical position with the living chamber upward, providing evidence of a postmortem history of the animal that 1) it died on the seafloor, becoming waterlogged soon after its death; 2) the soft body in the living chamber decayed slowly and the chamber gradually filled with gas due to decay of the body; 3) the shell rotated from a downward position to an upward position of the living chamber because of the effect of negative buoyancy on the soft body; 4) the shell was buried before the entire loss of soft parts, and the soft parts may have taken longer to decay.

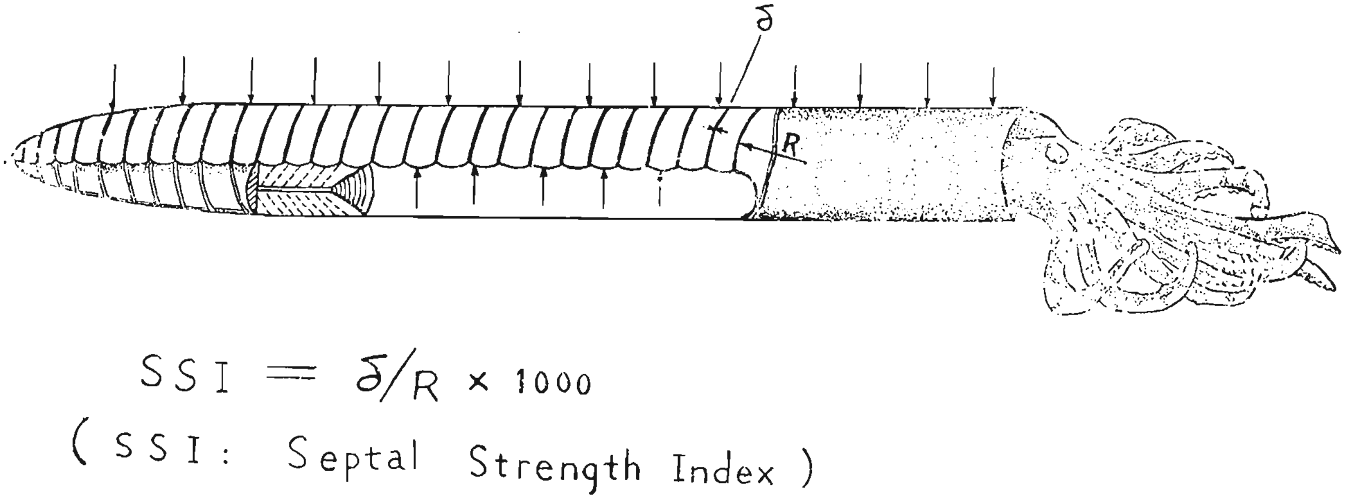

Figure 3. Schematic explanation of cephalopod functional problem in withstanding the ambient hydrostatic pressure of water, and the strength of the septa against implosion; and septal strength index (SSI). , thickness of septa; R, radius of septa.

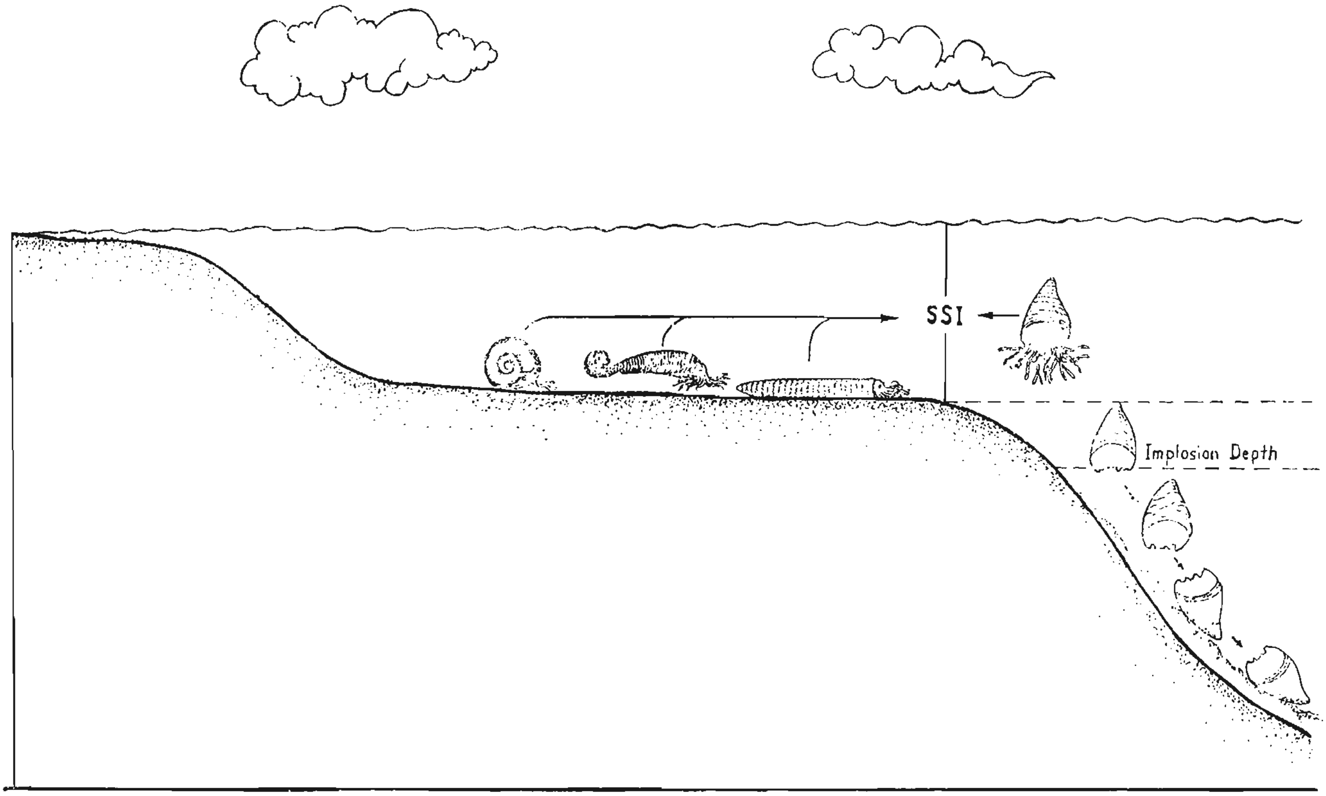

Figure 4. Interpretative relationship between nektobenthic and epipelagic cephalopod Septal Strength Index (SSI) and submarine topography; and postmortem history of epipelagic animal and its septae broken by inrushing water at or below its implosion depths during postmortem sinking.

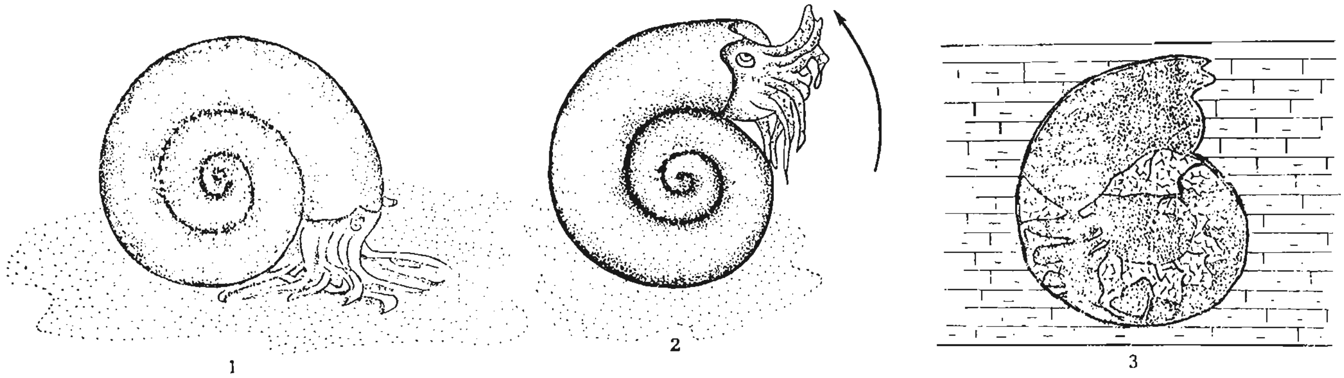

Figure 5. Reconstruction of postmortem history of a vertically embedded, Middle Ordovician coiled nautiloid, *Cyclolituites* from the Datienba Formation, Songtao, Guizhou province (Fig. 7, loc. 13). 1, The animal died on the seafloor and became waterlogged soon after its death; 2, the shell rotated into an upward position, from a previously downward position, of the living chamber due to the change to negative buoyancy of the soft body during decay; 3, burial in situ with a vertical position without marked transportation.

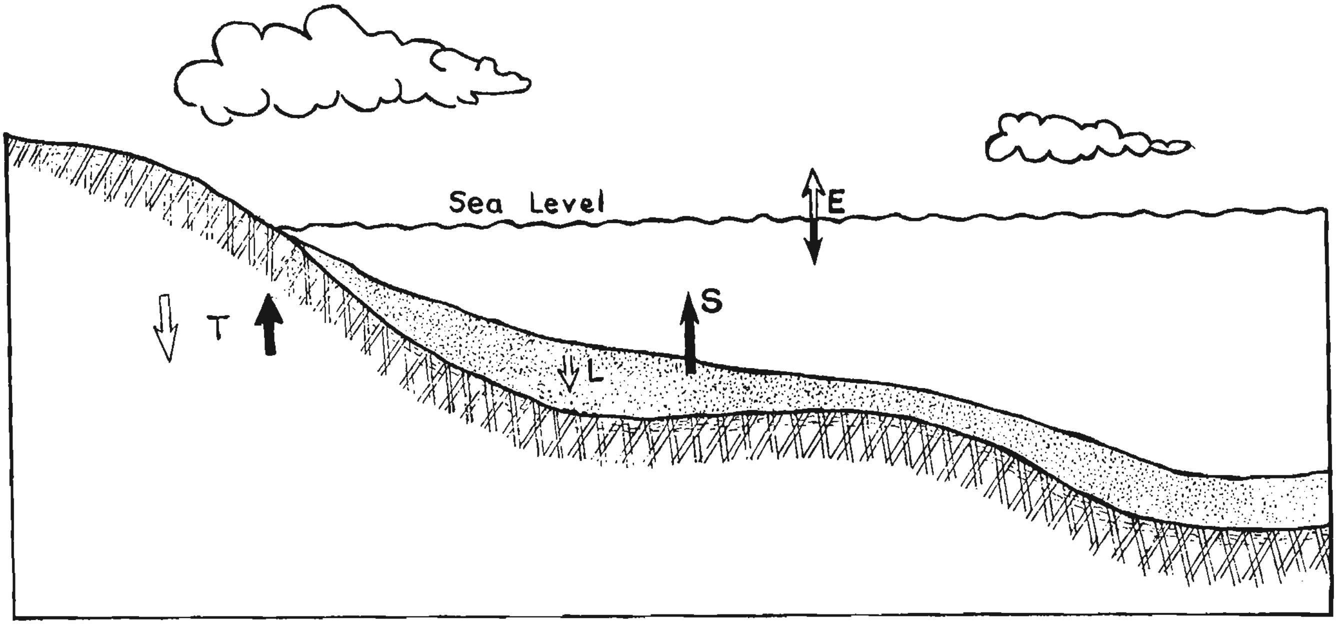

Figure 6. Schematic explanation of bathymetric dynamics and multiple interactions. T, tectonic rising or subsidence; L, subsidence due to sedimentary loading; E, eustatic rising or lowering; S, sediment infilling. Open arrows represent positive factors and filled arrows represent negative factors that affect paleodepths.

Evidence of marked postmortem transportation appears to be uncommon in our collections. Downslope transportation of animal remains from shallow water into adjacent, markedly deeper environments has been broadly recognized in turbidity and debris flow deposits in the southern margin of the Yangtze Platform. Downslope transportation is readily correlated with episodic occurrences of low sea level stands.

Epipelagic animals were frequently buried in markedly deeper environments, and their septae were broken by inrushing water during postmortem sinking at or below their implosion depths (Westermann, 1985). The specimens that had broken septae, but intact shell walls, are a rare but informative phenomenon in our collections. They all belong to a single fossil assemblage collected from a single debris flow bed in the upper Caradocian Taoqupo Formation in Yaoxien, Shaanxi Province, which was deposited on the southwestern margin of the Sino-Korean Platform, a part of the fore reef slope environment during the late Caradocian series (Chen and Zou, 1984). The animals consisted overwhelmingly of oncocerids, and were probably epipelagic. The cephalopod SSI investigation suggests that they were hovering within the upper 90 m water layer in the fore reef slope where the water depth was in excess of their implosion depth of 135 m. However, the septal debris in the phragmacone shows no marked loss or dislocation, suggesting the shell descended only a short distance after the septae imploded (Fig. 4).

DYNAMIC CHANGES OF EPEIRIC BATHYMETRY AND THE DETECTION OF EUSTATIC CHANGES

Multiple interactions

The bathymetry of epeiric seas changes constantly in response to the activities related to sediment infilling, tectonism, and eustacy (Fig. 6). Sediment infilling, tectonic uplift, and sea level lowering are negative factors, whereas tectonic and sediment loading subsidence, and eustatic rise are positive factors, which periodically affect the paleodepths of epeiric seas. These positive and negative factors vary through time, and there may occur a phased pattern of bathymetric variations from a deepening phase, to a depth-stable phase, to a shallowing phase. The three phases may also occur independently (Chen, 1988a,b).

The bathymetric changes and their multiple interactions can be expressed mathematically by: $D_t=D_0-S_t+L_t+T_t+E_t$, where D_t is depth at a proposed time, t; S_t, the total thickness of sediment infilling during t-interval; L_t, the subsidence due to sedimentary loading during t-interval; T_t, tectonic rising (-) or subsidence (+) during t-interval; E_t, eustatic rising (+) or lowering (-) during t-interval (Fig. 6).

Influence of tectonism

The epeiric seas under our investigation were under constant slow subsidence apparently due to the combined effects of thermal cooling and sediment loading. The combined value of $T_t + L_t$ is determinable, and estimates of the order of 1.5-2 cm/1000 y. can be determined during much of the Ordovician for the Sino-Korean Platform, and for the Tremadocian and early Arenigian of the Yangtze Platform. The value seems larger along the western margin of the Sino-Korean Platform where it was on the order of 7.5 cm/1000 y. from the middle Arenigian to the late Cardocian (Figs. 7, 9).

Bathymetric dynamics of the neritic realm

The rate of sedimentation varies greatly through time and in different regions in the neritic (shallow water) realm. It might be extremely high, as much as 30 m/1000 y. (Wilson, 1975) and thus sediment infilling (S_t) would be the most active factor to affect negatively the epeiric bathymetry. Nondeposition and erosion would be commonly expected as sediment infilling built the seafloor up to, and above, the sea level. The occurrence of fluctuations on the order of a few metres near sea level seems rather common in neritic regions, with shallowing-upward sequences, tidal sequences, or supratidal sequences developing under the following conditions: 1) slow eustatic rise; 2) minimal fluctuations of eustacy; 3) slow and constant tectonic subsidence. The marked deepening trend, which seems a rare phenomenon in shallow water sequences, is unnecessary in relation to eustatic highs or eustatic rise, but it may be attributed to an active phase of eustatic rise with a high rate of sedimentation that cannot keep pace. As the deepening trend slowed with time, the depth stable phase would occur, accompanying the appearance of an open circulation environment and a sharp increase in normal marine faunas. As the rate of eustatic rise slowed, the shallowing phase would resume, and sediment infilling would build the seafloor up to, and above, sea level within a few thousand years; the effect of tectonic subsidence would be incomparably small — only a few to ten centimetres during the period. Thus, the thickness of the shallowing sequences provides useful evidence for estimating the paleodepth in the local area.

Eustatic changes

In areas where the activities of tectonic subsidence and sediment infilling were minimal, the bathymetric variations would be dependent on eustatic variations. Thus, the detailed investigation of the quantitative variations of bathymetry provide a quantitative interpretation of the eustatic variations through time.

BATHYMETRIC FLUCTUATIONS OF EPEIRIC SEAS

Yangtze area

Metamorphic and intrusive events caused this area (Fig. 7) to become cratonic about 840 Ma, and thermal cooling mechanics have governed a constant subsidence of the Yangtze area since 740 Ma. The thermal cooling mechanics may have declined during the Ordovician Period and become negligible

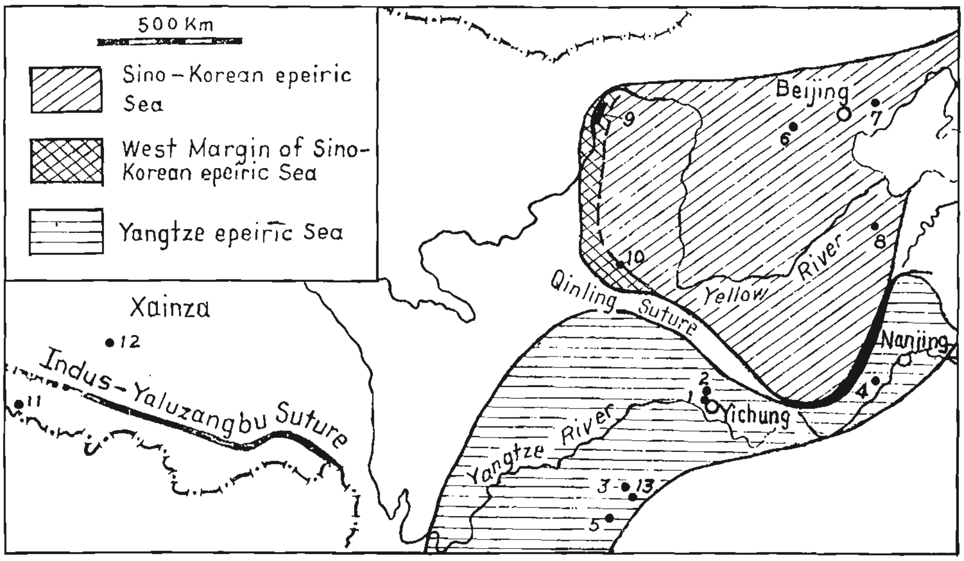

Figure 7. Index map showing the following localities mentioned in the text: 1, Huanghuachang, Yichang, Hubei province; 2, Fengxiang town, Yichang, Hubei province; 3, Yanhe, Guizhou province; 4, Wuwei, Anhui province; 5, Shiqian, Guizhou province; 6, Magezhuang, Laishui, Hebei province; 7, Tangshan, Hebei province; 8, Badou and Xinwen areas of Shandong province; 9, Haibowan area, Inner Mongolia Autonomous Region; 10, Yaoxien, Shaanxi province; 11, Nylam, Tibet Autonomous Region; 12, Xainza, Tibet Autonomous Region; 13, Songtao Guizhou province.

with time since the middle Arenigian (Fig. 8). This area (Fig. 7) was covered broadly with epeiric seas and the bathymetric fluctuations were basically of a eustatic nature since the middle Arenigian. Because it was a suitable environment for cephalopods, the quantitative reconstruction of the Ordovician history of bathymetry and eustacy is possible.

Tremadocian-early Arenigian

This interval is represented by a thick sequence of shallow water carbonates. The rate of subsidence was slow, on the order of 15 cm/1000 y. and there was always sufficient carbonate accumulation to equal or exceed sea level rise, with numerous nonsedimentation or erosion periods. Cephalopods appeared repeatedly in response to transgression periods. SSI study on the cephalopod assemblages provides evidence of epeiric sea levels that may have fluctuated within a narrow spectrum of 60 m throughout the period of the Tremadocian and early Arenigian (Fig. 9).

Mid-Arenigian to mid-Caradocian

In much of the Yangtze area, this period was represented by a prolonged trend of deepening, interrupted by two brief regression events in the upper *Didymograptus hirundo* Zone, equivalent to the late Arenigian, and in an equivalent of the lower part of the *Glyptograptus teretiusculus* Zone of the terminal Llanvirnian (Fig. 9). The net sedimentation rate was extremely slow, with an estimate of 0.25 to 0.1 cm per 1000 years. The extensive occurrence of hardground surfaces indicates the repeated removal of material, yielding cemented resistant surfaces, accompanied by subsequent solution, which removed skeletal and consolidated material. Upper portions of cephalopod shells often show evidence of heavy loss of skeletal material. The bathymetric changes in the epeiric sea were documented to include three major deepening events with magnitudes of approximately 100 m, 200 m, and 300 m, with two intervening shallowing events with magnitudes of 40 m and 100 m (Figs. 9, 10). The bathymetric changes of this period are considered to be a reflection of eustatic changes due to minimal tectonic subsidence and extremely slow rates of sedimentation.

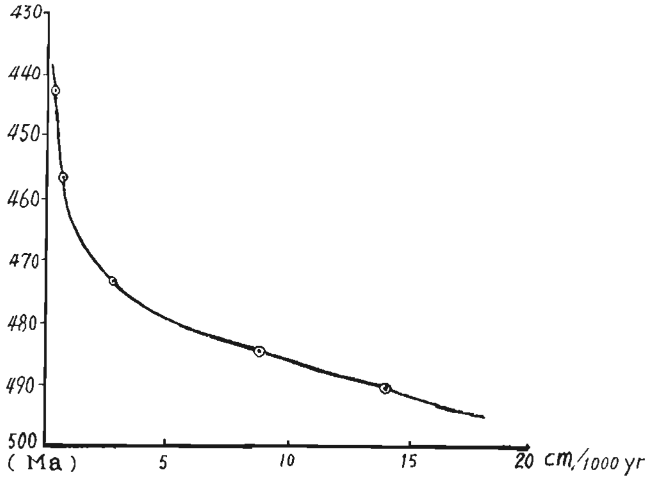

Figure 8. Declining trend in the rate of tectonic subsidence of the Yangtze area during the Ordovician Period.

Late Caradocian

This period was characterized by a great shallowing trend with a magnitude of 350 m, estimated from evidence of facies and cephalopod SSI changes in the upper Pagoda and Linhsiang formations.

A brief transgressive event in the lower Wufeng Formation (W_1-W_3) was proposed by Chen (1988a,b) with the interpretation of an abrupt facies change of shelly calcareous mudstone into black graptolitic siliceous shale in response to a sudden eustatic rise in sea level during the deglaciation episode.

In the period of the upper Wufeng Formation (W_4-W_5), a shallowing trend has been recognized by evidence such as 1) occurrence of the benthonic shelly *Hirnantia* Fauna; 2) reduced diversity of taxonomic composition of graptolites (Mu, 1974); 3) size of the habitat area of the *Hirnantia* Fauna increasing upward from the *Diceratograptus mirus* Zone and reaching a climax near the top of the Wufeng Formation; 4) evidence of subaerial emergence; 5) introduction of mass flow deposits on the continental slope and the adjacent oceanic basin.

Lower reach of the Yellow River: bathymetric dynamics

Ordovician rocks in this area (Fig. 7) are represented by shallow water carbonate sequences that accumulated intermittently at rates higher than those of subsidence to form vertical sequences of three different types: 1) cyclic sequences; 2) shallowing-upward sequences; and 3) supratidal sequences. The cyclic sequence consists of (from the base): 1) deepening unit; 2) depth-stable unit; 3) shallowing unit. The deepening unit seems to have formed during episodes of rapid rise in sea level when the rate of eustatic rise plus the rate of subsidence was greater than the rate of sedimentation. The depth-stable unit is marked by the occurrence of normal marine faunas such as cephalopods, trilobites, stromatoporoids, and brachiopods; it seems to be a period of open circulation and maximal water depth. The shallowing unit corresponds to a period when the sedimentation rate was greater than the subsidence rate plus the rate of sea level rise; its thickness is a good indicator of paleodepth at the transgressive climax of the cyclic sequence. These cyclic sequences have been broadly recognized in the Beianzhuang Formation of the midle to upper Arenigian, and in the lower Machiakou Formation (Llandeilian and lower Caradocian). They are interpreted as being related to the peak of the eustatic rise event.

Carbonate accumulation under minor bathymetric fluctuation conditions are represented by shallowing-upward sequences, as observed in the upper Yehli Formation of late Tremadocian age.

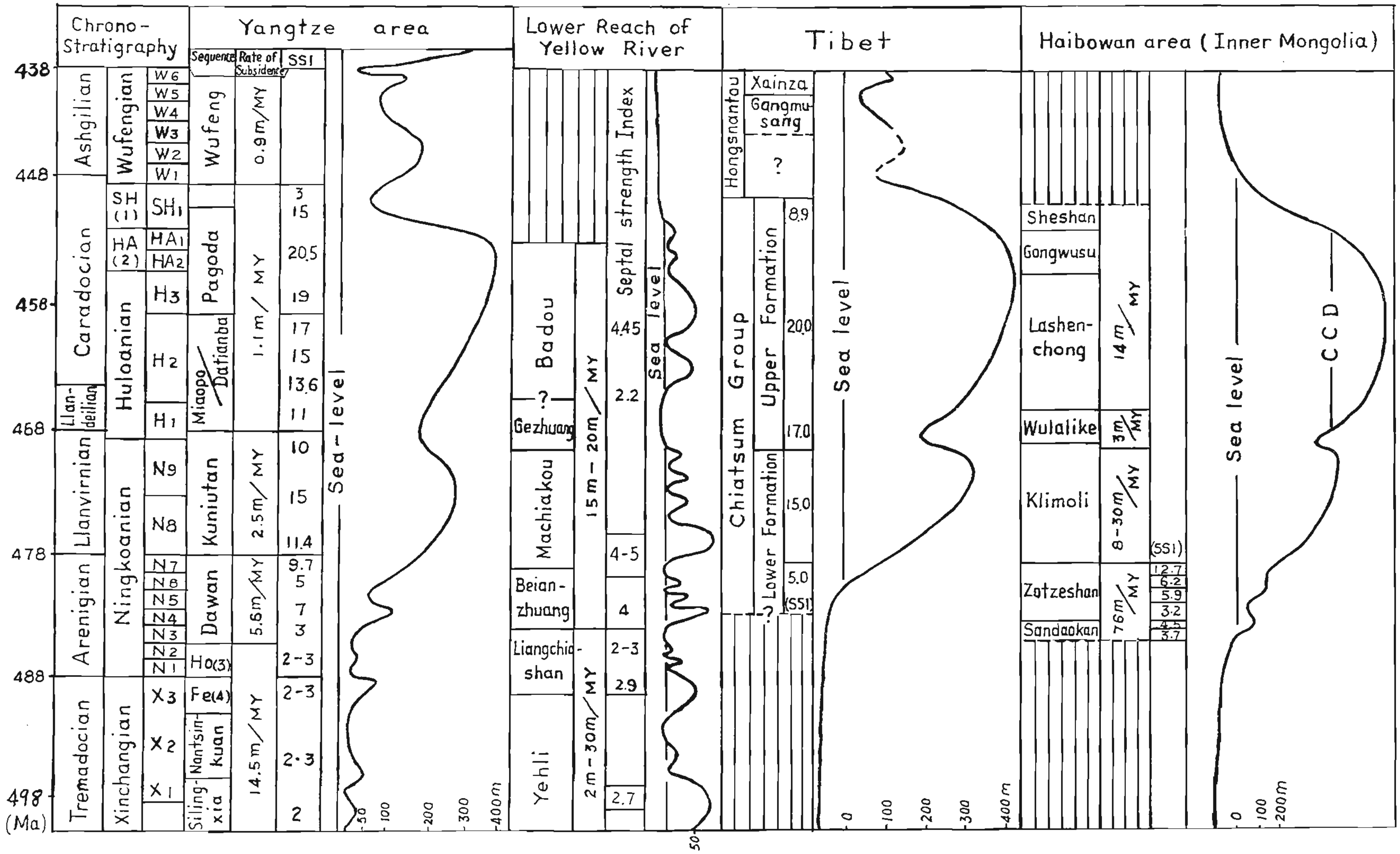

Figure 9. Schematic explanations of 1) bathymetric variations; 2) net rate of subsidence; 3) cephalopod Septal Strength Index (SSI) during Ordovician Period among several tectonically separated regions. Abbreviation explanation: 1, Shikounian; 2, Hanjiangian; 3, Honghuayuan; 4, Fenghsiang. The scheme of graptolite-based chronostratigraphic subdivisions in the chart is after Mu (1974).

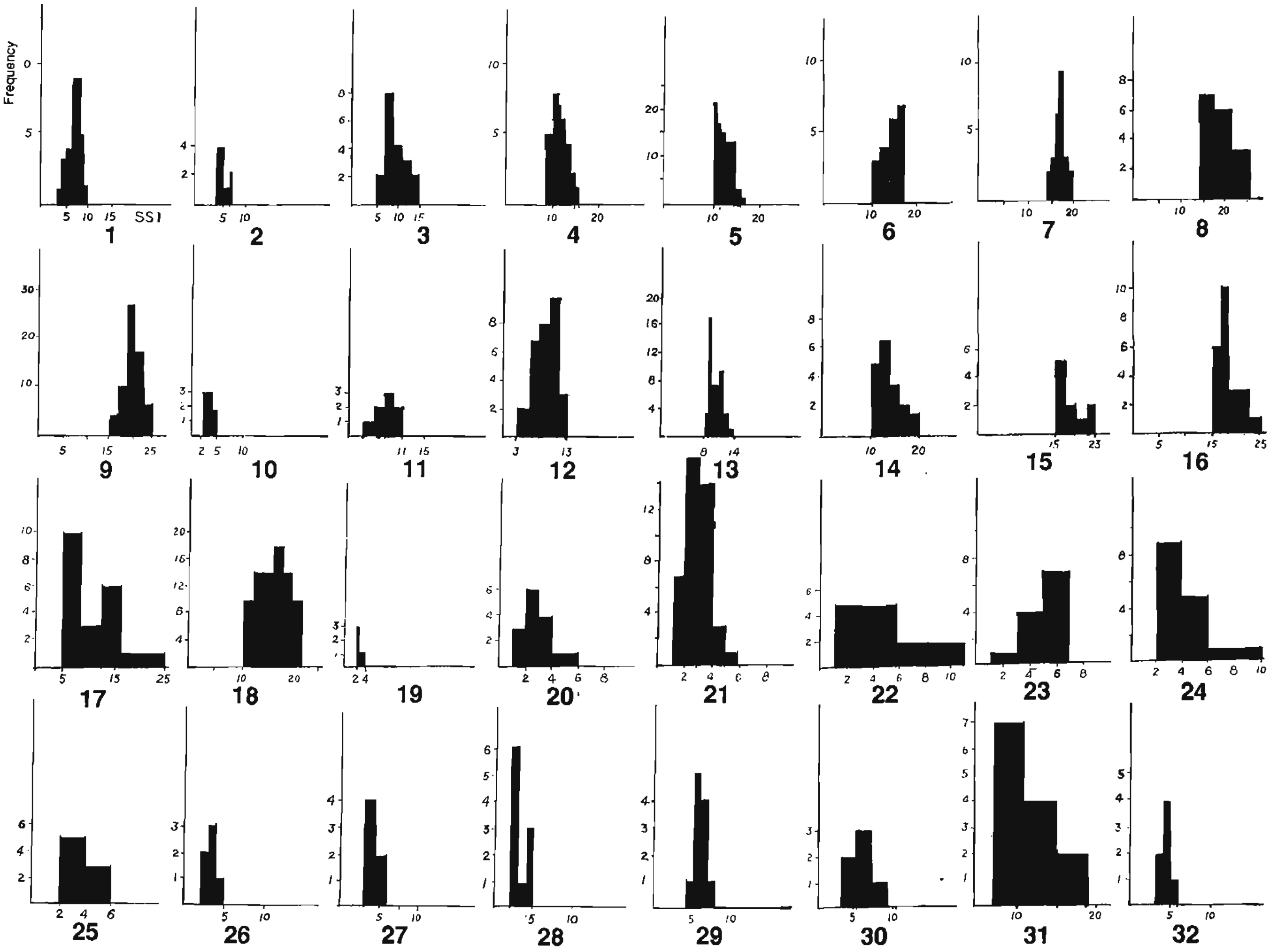

Figure 10. Histograms 1-32 showing frequency (number of specimens) versus Septal Strength Index (SSI) of the cephalopod assemblages. 1-9, all from Locality 1 (Fig. 7), arranged stratigraphically in ascending order : 1. Collection Number (Coll. No.) ACC 332, the *Protocycloceras deprati* Zone of the Dawan Formation, an equivalent of the *Oncograptus* Zone of the Australia graptolite sequence; 2. Coll. No. ACC 334, the upper *Protocycloceras deprati* Zone, an equivalent of the *Cardiograptus* Zone of the Australian graptolite sequence; 3. Coll. No. ACC 335, covering an interval including the *Glyptograptus sinodentatus* Zone below and **G.** *austrodentatus* Zone above in the uppermost Dawan Formation; 4. Coll. No. ACC 337, the lower Kuniutan Formation; 5. ACC 338, the upper Kuniutan Formation; 6. ACC 339, the Miopo Formation. 10-15, arranged stratigraphically in ascending order: 10. the fossil assemblage is from the southwestern margin of the Yangtze Platform in the *Bathmoceras* Zone, which lies between the *Azygograptus suecicus* Zone below and the *Protocycloceras deprati* Zone above; 11. ACC 311 from Locality 2 (Fig. 7) of the *Protocycloceras deprati* Zone of the Dawan Formation; 12. ACC 315, from Locality 2 (Fig. 7) in the upper *Protocycloceras deprati* Zone; 13. ACC 314, Locality 2 (Fig. 7), in the basal Kuniutan Formation; 14. ACC 311, Locality 2 (Fig. 7), the uppermost Kuniutan Formation; 15. ACC 320, Locality 2 (Fig. 7), the Pagoda Formation. 16-19, arranged stratigraphically in ascending order: 16. ACC 88, Locality 3 (Fig. 7), the upper Kuniutan Formation; 17. Locality 4 (Fig. 7), the uppermost Kuniutan Formation; 18. Locality 4 (Fig. 7), the Datianba Formation; 19. Locality 5 (Fig. 7), the Linhsiang Formation equivalent. 20-25, all from the Lower Reach of the Yellow River, and arranged stratigraphically in ascending order: 20. Locality 6 (Fig. 7), in the basal Yehli Formation; 21. Locality 7 (Fig. 7) in the basal Liangchiashan Formation, 22. Locality 8 (Fig. 7) in the Beianzhuang Formation; 23. Locality 8 (Fig. 7) in the Machiakou Formation; 24. Locality 7 (Fig. 7) in the Machiakou Formation; 25. Locality 8 (Fig. 7) in the Badou Formation. 26-32, all from Locality 9 (Fig. 7), and arranged stratigraphically in ascending order: 26. the *Pseudowutinoceras* Zone in the lower Sandaokan Formation; 27. the *Parakogenoceras* Zone of the upper Sandaokan Formation; 28. the *Polydesmia zuezshanensis* Zone of the lower Zotzeshan Formation; 29. the *Ordosoceras* Zone of the lower Zotzeshan Formation; 30. the lower part of the *Pomphoceras* Zone in the upper Zotzeshan Formation; 31. the upper *Pomphoceras* Zone in the uppermost Zotzeshan Formation; 32. the Sheshan Formation.

Sedimentary accumulation of the supratidal sequences took place persistently, within a narrow zone from sea level up to storm tidal level, with durations of up to several million years. These have been observed in the top of the Liangchiashan Formation, in the top of the Beianzhuang Formation, and in the Gezhuang Formation; they seem to indicate a time of extreme stability in sea level (Figs. 8, 9).

Haibown area (Inner Mongolia)

Along the western margin of the Sino-Korean Platform (Fig. 7), Ordovician rocks of middle Arenigian to late Caradocian age rest unconformably on upper Cambrian strata. They comprise a depositional cycle deepening-upward from shallow shelf carbonate to hemipelagic, thinly bedded carbonate, to an alternation with graptolitic shale, to basinal graptolitic shale deposits, and then shallowing-upward through clastic turbidites to shoal deposits. The subdivisions of the Ordovician strata include (from the base): 1) Sandaokan Formation: *Pseudowutinoceras* Zone; *Parakogenoceras* Zone; 2) Zotzeshan (Zuozishan) Formation: *Polydesmia zuezshanensis* Zone; *Ordosoceras* Zone; *Pomphoceras* Zone; 3) Klimoli Formation: *Amplexograptus confertus* Zone; *Pterograptus elegans* Zone; 4) Wulalike Formation: *Glyptograptus teretiusculus* Zone; 5) Lashenchong Formation: *Nemagraptus gracilis* Zone with *Syndyograptus* Subzone above and *Climacograptus bicornis* Subzone below; 6) Gongwusu Formation: *Amplexograptus gansuensis* Zone; 7) Sheshan Formation: *Eurasiaticoceras-Sheshanoceras* Zone (Chen et al., 1984).

Investigation of cephalopod SSI (Fig. 10) shows evidence of a deepening trend in the mid-Arenigian equivalent (Sandaokan Formation) (Figs. 9, 10). The Zotzezhan Formation, upper Arenigian, shows a slow deepening trend of the epeiric sea from 60 m to 120 m in much of the formation. A sharp change in bathymetry was documented in the top of the Zotzeshan Formation (uppermost Arenigian), with depths in the epeiric sea of approximately 250 m as indicated by cephalopod SSI (Fig. 10, histogram 31). The evidence of a continuous deepening seems remarkable in the Klimoli Formation, which is interpreted as representing a hemipelagic slope environment in its major part, with a constant trend of deepening upward to depths near or below the upper limit of the CCD (Carbonate Compensation Depth) at its top.

A brief shallowing event in the latest Llanvirnian has been registered by an introduction of gravity flow deposits at the base of the Wulalike Formation. The main part of this formation contains black graptolitic shale. Its sedimentary accumulation is thought to have occurred in a marginal basinal environment at a depth below the CCD.

The Laschenchong Formation is represented by black pelagic graptolitic shale in its lower part, a *Syndyograptus* Subzone equivalent, which was succeeded by an alternating sequence of siltstone, shale, and pebbly limestone represented in the upper part of the *Climacograptus bicornis* Subzone and in the overlying Gongwusu Formation. The Sheshan Formation is the highest unit, consisting of a sequence of skeletal and oolitic limestone, and arenaceous shale representing a very shallow, turbulent depositional environment. The SSI of the cephalopod assemblage suggests the lower limit of the epeiric sea was about 100 m during the late Caradocian (Fig. 10, histogram 32).

Mt. Jolmo Lungma region (Tibet)

The shallow water carbonates have been broadly recorded in this region (Fig. 7), suggesting an extensive epeiric sea during the Ordovician Period. The Chiatsun Group in Nylam area (Figs. 9-11) contains a thick sequence of carbonates up to 833 m thick, of late Arenigian to late Caradocian age, and lies unconformably on the metamorphic rocks of the Juchiatsun Group (Mu et al., 1973). The lower 726 m of the Chiatsun Group consist of massive limestone with interbeds of siltstone and sandstone; they are thought to have been laid down at a considerably higher rate under shallow water conditions as a response to a profound transgressive event in the late Arenigian. The SSI of the cephalopod assemblages provides evidence of maximal depths of the epeiric sea of approximately 100 m during the late Arenigian, deepening progressively to 300 m in the Llanvirnian. The upper 97 m of the Chiatsun Group is composed of reddish limestone with rich cephalopod faunas (Chen, 1975). The SSI of these cephalopod assemblages indicates a major transgressive and subsequent regressive event during the Caradocian (Fig. 8). The changes of biofacies and bathymetry are comparable with those of the Yangtze area (Chen, 1988a,b). The bathymetry indicated by SSI changes from 340 m to 400 m, and then a shallowing phase followed by the occurrence of the *Richardsonoceras* Fauna in the top beds of the upper Caradocian when the water depth was about 180 m.

The Hongshantou Formation consists of a 70 m thickness of brown shale, with extremely rare fossils, which is considered to represent the regressive event in the Late Ordovician.

Xainza area (Tibet)

This area (Fig. 7, loc. 12) is tectonically separated from the Mt. Jolmo Langma region by the Indus-Yaluzangbu Suture. The Ordovician rocks include three units, in ascending order: the Kordo, Gangmusang, and Xainza formations. The Kordo Formation is about 400 m thick, and consists mainly of carbonates yielding a rich cephalopod fauna of Yangtze area affinity with an age probably ranging from Llandeilian to Caradocian. The Gangmusang Formation consists of shallow water carbonates probably of late Ashgillian age, regarded as representing the regressive phase during the period equivalent to the *Diceratograptus mirus* Zone and the *Paraorthograptus uniformis* Zone of the upper Wufeng Formation. The Xainza Formation contains the top 14.09 m of the Ordovician sequence (Ni et al., 1981). Its lower 5.27 m consists of greyish graptolitic shale containing a rich *Diplograptus bohemicus* Fauna (Mu and Ni, 1983), whereas the upper 8.82 m is represented by a shallow water, grey argillaceous limestone with the associated shelly *Hirnantia* Fauna.

EUSTACY AND ITS CAUSES

Eustatic changes in sea level are dependent on a number of interrelated controls and they can be brought about by changes in volume either of ocean water, or of ocean basins (e.g., Pitmann, 1978; Schopf, 1980). A large, high latitude Gondwana landmass existing near the south pole during Ordovician time (Ziegler, 1981; Erdtmann, 1982, 1986) could have been a place for permanent ice and snow. The growth and decay of ice sheets may have appreciably altered the water volume in the oceans within narrow time intervals.

Stable period with episodic fluctuations

Sea level remained stable from the latest Cambrian to the early Arenigian, apart from episodic fluctuations (Fig. 11). The fluctuations are considered to have been glaciation-driven, were rapid, and had characteristic sequences of a regressive and a succeeding transgressive event (Fig. 11). These events were known as 1) Lange Ranch Eustatic (regressive) Event (Miller, 1984), succeeded by the basal Tremadoc (transgressive) event (Chen, 1988a,b); 2) Black Mountain Eustatic (regressive) Event (Miller, 1984), followed by a transgressive event at end of the early Tremadocian (Chen, 1988a,b); 3) late Tremadoc eustatic (transgressive) episode (Barnes, 1984; Fortey, 1984), which was marked in China by a flood of *Adelograptus-Clonograptus-Kiaerograptus* Fauna into the cratonic succession, and was preceded by a regressive event in the middle Tremadocian (Chen, 1988a,b).

The evidence of the characteristic sequence of regressive and succeeding transgressive events together with their short duration suggests that the cause of the eustatic fluctuations in this period may have been related to changes in solar radiation. Assuming an increase in solar radiation, the ice sheet may have advanced; an increase in solar radiation would have resulted in increased wind strength circulation and evaporation, and a sudden increase in the amount of snow on the landmasses in high latitudes. With a further increase in solar radiation, wet snow would melt and a post-glacial eustatic rise would follow (Lamb, 1977).

Eustatic rise with sudden lowering episodes

For a duration of roughly 40 m.y. (middle Arenigian to Caradocian), there were three transgressive cycles with related rises in sea level amounting to approximately 80 m in the middle Arenigian, to 200 m in the latest Arenigian and Llanvirnian, and to 200 m in the Llandeilian and Caradocian. The rate of sea level rise is estimated to be on the order of 20 m per m.y. (Chen, 1988a,b). The two sharp reversal episodes in sea level occurred in the late Arenigian and in the latest Llanvirnian-early Llandeilian, respectively. An estimated few tens, to 100 m, in the lowering of sea level may have taken place in roughly one million years during each of the reversal episodes (Figs. 10, 11).

The cause of the prolonged process of eustatic rise was previously interpreted to have been driven by rather rapid seafloor spreading (Chen, 1988a,b) and in this paper the author emphasizes the idea of an increase of the length of spreading ridges that could have changed sufficiently to raise sea level by some hundreds of metres during the Ordovician Period. It is known that the number of lithospheric plates varies over geological time, and the total length of the ridges changes from maximal during continental dispersal to minimal during aggregation periods (Fischer, 1984). Thus, sea level may be expected to be high during continental

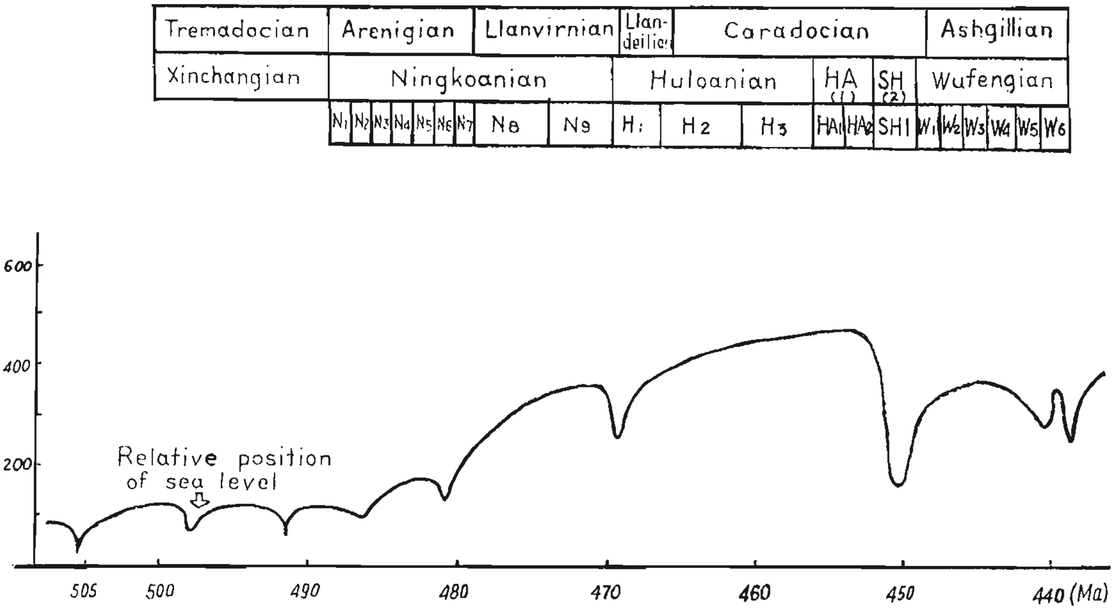

Figure 11. Schematic explanation of the changes in sea level during the Ordovician Period. SH, Shikounian; HA, Hanjiangian.

dispersion periods and low during pangeal episodes. The great rise in sea level from the middle Arenigian to Caradocian did indeed occur in a time of continental dispersion. Jaanusson (1984) interpreted the Ordovician Period as pronouncedly thalassocratic, which was in great contrast to widespread "epeirocratic" emergence known in the late Precambrian pangeal period.

Continental dispersion and maximal activity may have been accompanied by maximal volcanism during the Ordovician Period (Stillman, 1984). The appearance of intense veils of dust in the high atmosphere would probable have accompanied the abnormally enhanced volcanic activity; it is possible that it caused the accumulation of ice sheets and sharp reversal trends of sea level in the late Arenigian and in the latest Llanvirian and early Llandeilian.

Late Ordovician eustacy and glaciation

The author (1988a,b) has argued that the sea level during the Late Ordovician might have fluctuated successively in a sequence of 1) a latest Caradocian lowering event; 2) an early Wufengian rising event; 3) a Hirnantian lowering event; 4) a *bohemicus* rising event; 5) a terminal Wufengian lowering event (Fig. 11). The latest Caradocian seems to have had a major glaciation episode, during which the ice cap grew at a rapid rate, with a related 350 m lowering of sea level as documented in the Yangtze Epeiric Sea. The evidence of the early Wufengian deglaciation and eustatic rising event are strong, bringing a flood of graptolitic faunas almost everywhere within the Yangtze Epeiric Sea. The Hirnantian lowering event is indicated by a change of biofacies in the shelf succession, with the shelly facies and associated brachiopods of the *Hirnantia* Fauna increasing in areal extension. The *bohemicus* rising event was proposed by Chen (1988a,b) to represent a sea level change at the top of the Wufengian Stage. The brief transgressive event, which brought graptolite faunas onto the shelf sequence in the Yangtze area and in the Xainza area (Tibet), can be correlated precisely with the lower part of the *Diplograptus bohemicus* Zone of the uppermost Wufengian. Evidence of the terminal Wufengian lowering event was documented broadly in many sections of separated regions in the Tibet and Yangtze River areas, China (Chen, 1988a,b) as well as in Europe (Brenchley and Newall, 1984). This terminal Wufengian lowering event can be correlated closely with the upper *Diplograptus bohemicus* Zone.

CAUSES OF GLACIATION UNDER THE GREENHOUSE EFFECT

The general aspects of the paleoenvironments seem to indicate a greenhouse effect during the Ordovician Period, in which an essentially nonglacial world was briefly plunged into glacial conditions. Because of the existence of a greenhouse effect during that period, the glaciations seem to have been rapidly developed due to an abundance of atmospheric moisture, but were difficult to sustain with the onset of a warm climate. The deglaciations appear to proceed rapidly and an anoxic marine environment was generally widespread.

Cause of glaciation

The glaciations appear to have resulted from increased solar influx from the latest Cambrian to the early Arenigian, as mentioned above. Plate activity and accompanying volcanism apparently climaxed in the middle Arenigian to Caradocian. An intense veil of dust in the high atmosphere would be expected to raise the earth's albedo to a level at which much solar energy was reflected, and the glaciation may thus have resumed in the following separate periods: 1) within the late Arenigian; 2) latest Llanvirnian-early Llandeilian; 3) late Caradocian.

ACKNOWLEDGMENTS

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Evidence for isotopic change associated with Late Ordovician glaciation, from brachiopods and marine cements of central Sweden

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Abstract

The carbon and oxygen stable isotopic composition of brachiopods and marine cements from the Boda Limestone carbonate mud mounds (Ashgillian) of central Sweden has provided evidence for the effect of the Late Ordovician-Early Silurian glaciation on the isotopic composition of contemporaneous sea water.

Brachiopods and marine cements were analyzed from different stratigraphic levels within the mounds. Isotopic and trace element results show that primary marine compositions are preserved. The $\delta^{18}O$ *and* $\delta^{13}C$ *values are approximately constant throughout the pre-Hirnantian mound limestone (mean* $\delta^{18}O = -4.5$ ‰, $\delta^{13}C = +1.8$ ‰*), but brachiopods and cements from coquinas of Hirnantian age at the top of the mounds show significant enrichment in* $\delta^{18}O$ *and* $\delta^{13}C$ *(mean* $\delta^{18}O = -2.5$ ‰, $\delta^{13}C = +6.3$ ‰*) when compared to the mound components.*

A drop in water temperature of approximately 10°C could explain the 2‰ positive $\delta^{18}O$ *shift. Paleoecological considerations, however, render this unlikely. It is therefore proposed that the shift marks an increase in sea water* $\delta^{18}O$*, perhaps caused by a glacial "ice volume" effect.*

The careful sampling of brachiopods and cements rules out biological vital effects and diagenetic alteration as potential mechanisms of ^{13}C *enrichment. The shift is interpreted as representing a significant change in the* $^{13}C/^{12}C$ *ratio of the dissolved bicarbonate in Hirnantian sea water.*

Résumé

La composition en isotopes stables de carbone et d'oxygène des brachiopodes et des ciments marins des monticules de boue carbonatée du calcaire Boda (Ashgillien) dans le centre de la Suède a fourni des indices sur les effets de la glaciation de l'Ordovicien supérieur au Silurien inférieur sur la composition isotopique de l'eau de mer contemporaine.

On a analysé des brachiopodes et des ciments marins provenant de différents niveaux stratigraphiques au sein des monticules. L'analyse des isotopes et des éléments traces montre que les principales compositions marines ont été conservées. Les valeurs ^{18}O *et* ^{13}C *sont approximativement constantes dans tout le calcaire des monticules antérieur à l'Hirnantien (*^{18}O *moyen = -4,5* ‰, $^{13}C = +1{,}8$ ‰*) mais les brachiopodes et les ciments de lumachelles d'âge hirnantien au sommet des monticules indiquent un enrichissement important en* ^{18}O *et* ^{13}C *(*^{18}O *moyen = -2,5* ‰, $^{13}C = +6{,}3$ ‰*) lorsqu'on les compare aux composantes des monticules.*

Une chute de la température de l'eau d'environ 10 C pourrait expliquer le changement positif de 2‰ de ^{18}O*. Des considérations paléo-écologiques, toutefois, rendent cette explication improbable. Il est donc proposé que ce changement indique une augmentation de* ^{18}O *dans l'eau de mer, causée peut-être par un effet glaciaire dit «volume de glace».*

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L'échantillonnage soigné des brachiopodes et des ciments font ressortir comme mécanisme possible d'enrichissement en ^{13}C des effets vitaux biologiques et une altération diagénétique. La différence est interprétée comme un changement important du rapport $^{13}C/^{12}C$ du bicarbonate dissous dans l'eau de mer hirnantienne.

INTRODUCTION

This paper presents one aspect of the research in progress on the sedimentology and diagenesis of the Kullsberg and Boda limestones of the Siljan district, central Sweden (Fig. 1). The aim of this study is to document changes in the stable isotopic composition of Ashgillian sea water through the transition from nonglacial to glacial times.

The isotopic analysis of skeletal and inorganic minerals precipitated in isotopic equilibrium with the ambient sea water represents the most accurate method of estimating the $^{18}O/^{16}O$ and $^{13}C/^{12}C$ ratios of ancient oceans. Lowenstam (1961) established that brachiopods do not exert vital effects (the biological fractionation of isotope species during biomineralization) on oxygen during the secretion of their low-Mg calcite shell. However, there is evidence to suggest that biological controls may affect the carbon isotopes (Veizer et al., 1986; Rowse, 1988). To complement the brachiopod data, inorganically precipitated marine cements were sampled from corresponding levels within the limestone.

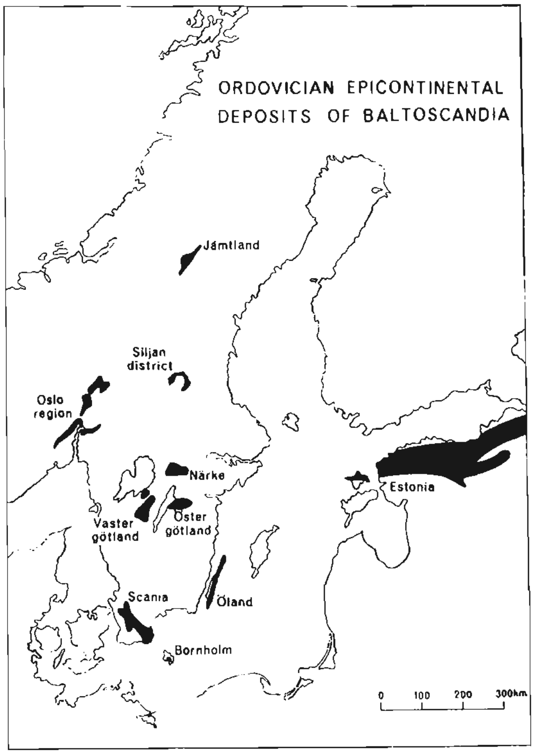

Figure 1. Ordovician epicontinental deposits of the Baltoscandian Basin, from Jaanusson (1982).

Modern marine cements are high-Mg calcite or aragonite and are more likely to be subject to chemical and mineralogical changes during diagenesis. Although controversy therefore exists as to the validity of using ancient marine cements as indicators of marine geochemistry, Given and Lohmann (1985) and Kendall (1985) present evidence that such phases may be primary precipitates and may therefore record "original" signatures.

This paper reports values from brachiopods and marine cements that are significantly different from those for Ordovician limestones published by Veizer et al. (1986) and Popp et al. (1986).

GEOLOGICAL SETTING

The Kullsberg and Boda limestones are two stratigraphic levels of carbonate mud mounds of Caradocian and Ashgillian age, respectively (Fig. 2). The mud mounds developed as local carbonate accumulations within the nodular mudstone succession of the central Baltoscandian epicontinental shelf (Männil, 1966). This paper is restricted to data from the Boda mud mounds only.

The Boda mounds reach a maximum size of 1000 m in diameter and 140 m in thickness (Jaanusson, 1982). Mounds are composed of a massive core facies containing characteristic calcite cement-filled stromatactoid cavities. The core is surrounded by fossiliferous bedded flank deposits that interdigitate with the normal off-mound shelf facies. The top of the Boda mounds is marked by a gradational facies change (Fig. 2); bioclastic packstones and grainstones replace cavity-bearing mound limestone. A brachiopod coquina horizon, containing the distinctive *Holorhynchus* Fauna occurs above the facies change, indicating a Hirnantian age for the bioclastic sequence (Jaanusson, 1982). Both the bioclastic units and underlying mound limestone are dissected by a karstic unconformity (Fig. 2).

The eustatic regression associated with the Upper Ordovician glaciation (Berry and Boucot, 1973; McKerrow, 1979) resulted in drastic alterations in facies patterns, faunal zonality, and distribution of land areas within the Baltoscandian Basin during the Hirnantian stage (Männil, 1966; Brenchley, 1988). This sea level fall is interpreted as responsible for the facies change and eventual karstification of the Boda mounds. Graptolitic black shale of the upper Llandoverian *sedgwicki* Zone (Thorsland and Jaanusson, 1960) lies unconformably on top of the Hirnantian units.

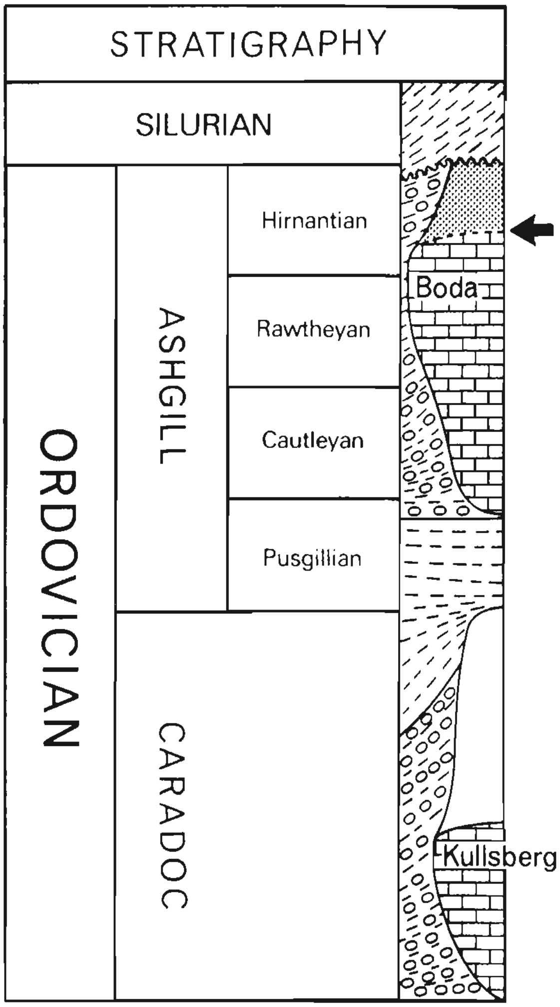

Figure 2. Stratigraphic section of part of the Ordovician succession in the Siljan district, depicting the onlapping relationships between the mounds and the basinal sediments. The gradational facies change from Boda mound limestone (bricked ornament) to bioclastic units (stippled) is arrowed. Note also the karstic unconformity between the top of the Hirnantian bioclastic units and the Silurian shales.

SAMPLE SELECTION AND METHODS

Brachiopods

Brachiopods and marine cements were sampled from three Boda mound localities: Osmundsberg, Kalholn, and Östbjörka.

The impunctate orthids *Glyptorthis* sp., *Platystrophia* sp., and *Nicolella* sp. were selected for sampling from the numerous brachiopod genera that are found within the Boda Limestone (Jaanusson, 1982). These three genera occur commonly within the argillaceous units that interdigitate with the bedded bank facies. Orthids are present within the Hirnantian coquinas but proved impossible to separate from the enclosing marine cement. The impunctate spiriferid *Hindella* sp. was therefore selected for sampling from the coquina unit.

All brachiopod samples were separated from their host rock and sectioned perpendicular to the hinge line in order to expose the thickest portion of shell material. The shell was then photographed in transmitted light and under cathodoluminesence so any later diagenetic components present in the skeletal calcite could be avoided during sampling (cf. Popp et al., 1986; Rowse, 1988). Unaltered, nonluminescent skeletal calcite was extracted using a modelling drill fitted with a fine tungsten-carbide bit, care being taken not to drill deeper than 1 mm. For the orthids, it was found necessary to drill the outer shell to obtain sufficient powder for analysis. The shell surface was picked clean of matrix and then immersed in 20% HCl for 20 seconds to expose clean skeletal calcite.

Marine cements

One of the most striking features of Paleozoic mud mounds is the ubiquitous presence of calcite cement-filled stromatactoid structures, which may comprise 50 per cent of the mound limestone (Bathurst, 1959; Ross et al., 1975; Wallace, 1987). In the Boda Limestone, the stromatactoid cement consists of one or more isopachous layers of radial, coarse fibrous, variably inclusion-rich crystals, which exhibit undulose extinction (Pl. 1, fig. 1). Some individual crystals may demonstrate the characteristics associated with radiaxial fibrous calcite (Bathurst, 1959, 1975; Kendall, 1985) or fascicular-optic calcite (Kendall, op. cit.). However, the bulk of these cements are best described as radial-fibrous calcite, transitional between the above two end members. Specimens of marine cement were collected at levels in the mounds to correspond with the brachiopod samples.

As the stromatactoid fabric is restricted to the mound limestone and is not present in the overlying bioclastic beds, marine cements precipitated on the inside and outside of the *Hindella* brachiopods were analyzed. These cements are present as 2 to 3 mm thick fringes of inclusion-rich, radial fibrous calcite crystals showing curved cleavage planes and associated undulose extinction (Pl. 1, fig. 2).

By analogy with modern reef cements and their ancient equivalents, both sets of cements studied are interpreted as being marine precipitates.

Cement samples were sectioned, polished, and photographed in transmitted light and cathodoluminescence to aid sampling. Areas considered suitable for analysis, i.e., nonluminescent and unveined, were drilled out as described for the brachiopods above.

Analysis

All powders were roasted in a Bio-Rad E2000 Plasma Asher for 4 hours, to remove all traces of organic matter.

Analyses of oxygen and carbon stable isotopic ratios were carried out after reaction of 3 mg samples with 100% orthophosphoric acid at 25°C. The resultant CO_2 was analyzed on a Sira 12 triple collecting mass spectrometer against an internal reference gas. Results were corrected by standard methods (Craig, 1957), and a fractionation factor for the reaction with orthophosphoric acid of 1.01025 was applied

(Friedman and O'Neill, 1977). All results are reported relative to PDB. Reproducibilities for both $\delta^{18}O$ and $\delta^{13}C$ are within 0.1°/₀₀.

Trace element analyses were carried out on an Inductively Coupled Plasma Spectrometer. Ten mg of calcite powder was dissolved in 5 ml of 5% ARISTAR HCl and run against mixed trace element standards of varying concentrations. Resulting concentrations were corrected for acid impurities and machine drift. The reproducibility for both Mn and Sr was calculated as 1%.

This study reports the results of over 120 isotopic analyses and 60 trace element analyses of brachiopods and marine cements.

TRACE ELEMENT RESULTS

Alteration

The interpretation of geochemical data from the analysis of limestones and fossils requires an accurate knowledge of the degree of preservation of the specimens involved. Investigation into the diagenetic history of the Boda mounds indicates that the influx of meteoric water, associated with subaerial exposure, was limited to regions in close proximity to the unconformity surface (Fig. 2). Such areas were consequently avoided during sampling.

In general, heavy isotopic values and low Mn concentrations characterize samples that show no petrographic sign of alteration. Where possible, the sparry cement phases were avoided during sampling of brachiopods and marine cements. However, the total exclusion of these secondary phases from analyzed samples was not feasible as they commonly form intricate alteration fabrics with the primary phases (Pl. 1, figs. 3, 4). All the sparry cements have light $\delta^{18}O$ values and are variably depleted in Sr and enriched in Mn; low $\delta^{18}O$ and Sr values and/or high concentrations of Mn in marine cements and brachiopods can therefore be attributed to contamination.

In the plot of Sr against Mn for the brachiopods (Fig. 3), most of the *Hindella* brachiopods are in the field of Sr and Mn concentrations found in modern brachiopods (Popp et al., 1986). The orthids, however, plot on the alteration path between primary shell material and burial cements. The *Hindella* brachiopods therefore appear to have escaped extensive diagenetic alteration, maintaining almost original Sr and Mn concentrations. This assumption is consistent with cathodoluminescence results that show the *Hindella* brachiopods to contain smaller amounts of brightly luminescing burial cements than the sampled orthids (compare Pl. 1, fig. 5 with Pl. 1, fig. 6).

The extent to which the marine cements have been altered is more difficult to assess. Under cathodoluminescence, the cement is a mosaic of dull to nonluminescing coarse fibrous crystals containing numerous inclusions of brightly luminescing calcite, each crystal enclosed by thin veinlets of the same Mn-rich calcite (Pl. 1, figs. 3, 4). The nature of the fibrous cements suggests that the fabric observed today has been at least partially diagenetically altered by recrystallization and intercrystalline cementation. Although such radial-fibrous cements may be considered as primary marine precipitates (Kendall, 1985), it is obvious from the cathodoluminescence fabrics (Pl. 1, figs. 3, 4) and the trace element data (Fig. 4) that the cements have been altered to a greater degree than the brachiopods.

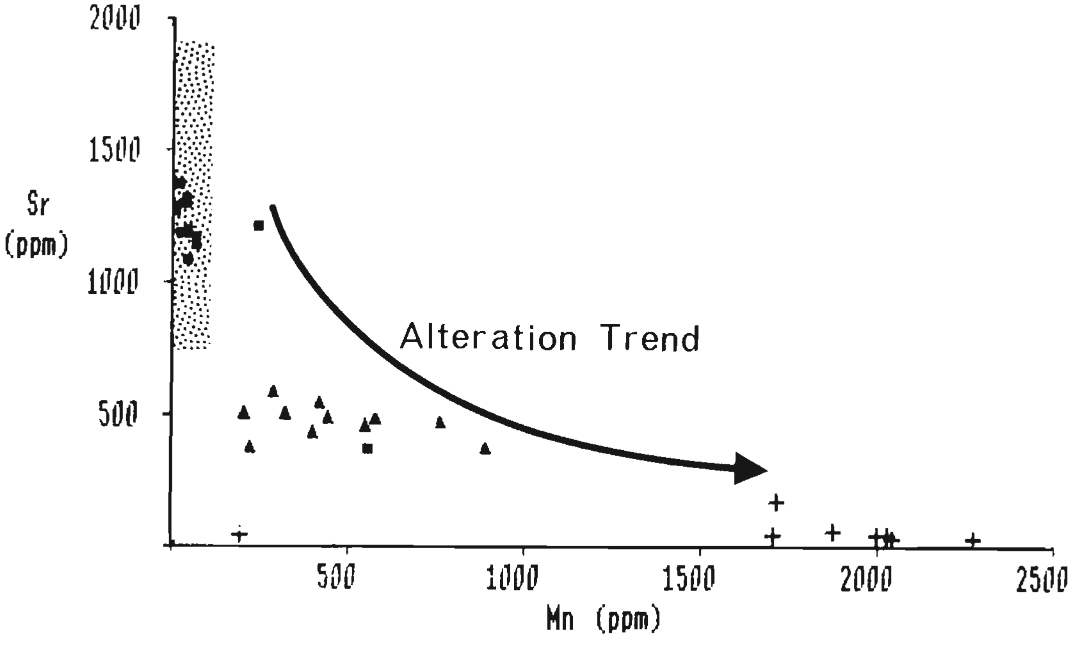

Figure 3. Brachiopod Sr **versus** Mn plot and geochemical alteration trend. Closed squares, *Hindella* brachiopods; closed triangles, mound orthids; crosses, diagenetic calcite. Stippled field represents the range of Sr and Mn concentrations in modern day brachiopods (Popp et al., 1986).

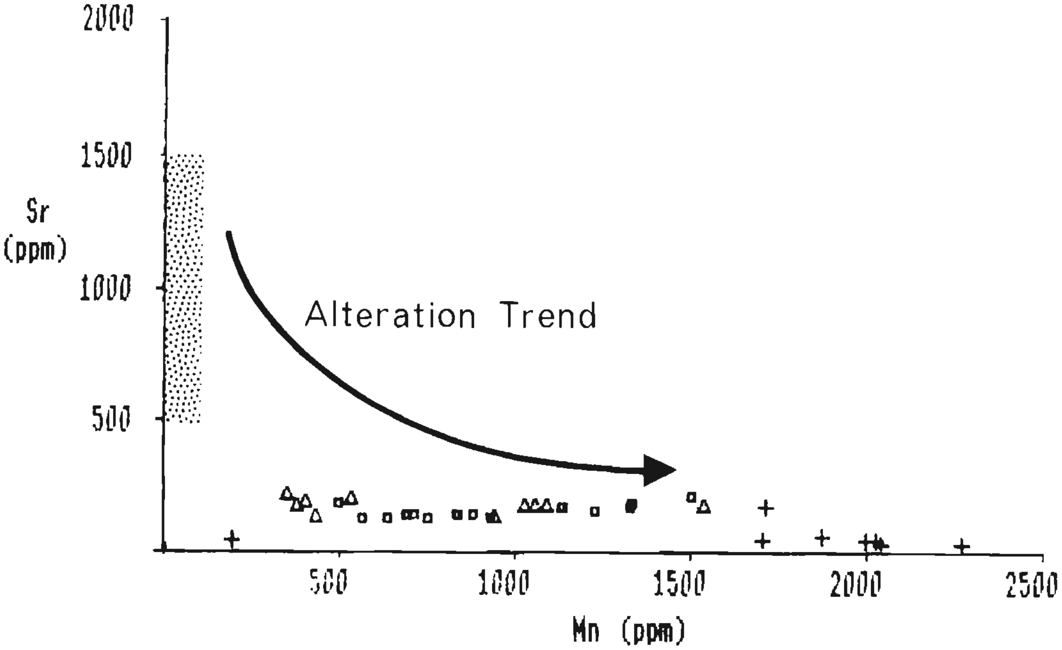

Figure 4. Marine cement Sr **versus** Mn plot and alteration trend. Open squares, Hirnantian coquina cements; open triangles, mound stromatactoid cements; crosses, diagenetic calcite. Stippled field represents the range of Sr and Mn concentrations for inorganically precipitated calcite from seawater (Arthur et al., 1983; Rowse, 1988).

ISOTOPE RESULTS

Brachiopods

Brachiopod stable isotope data plot in two well defined fields (Fig. 5). Orthids, from all levels within the mound limestone, plot in a relatively tight field with the least negative $\delta^{18}O$ values and most positive $\delta^{13}C$ values being -3.5°/₀₀ and +2.2°/₀₀, respectively (average values of $\delta^{18}O$= -4.5°/₀₀ and $\delta^{13}C$= +1.5°/₀₀). The Hirnantian age *Hindella* brachiopods exhibit very different isotopic signatures to those of the mound orthids (Fig. 5). The heaviest $\delta^{18}O$ and $\delta^{13}C$ values are -1.4°/₀₀ and +6.8°/₀₀, respectively (average values $\delta^{18}O$= -2.5°/₀₀ and $\delta^{13}C$ =+6.3°/₀₀).

The inclusion of varying amounts of diagenetic secondary calcite within the brachiopod shell material is indicated by the spread of isotope data along alteration trends from an unaltered marine composition toward the $\delta^{18}O$ and $\delta^{13}C$ values associated with the burial cements (Fig. 5).

Marine cements

Although the data fields are less well defined, $\delta^{18}O$ and $\delta^{13}C$ values from the marine cements fall into the same general area as their stratigraphically equivalent brachiopods.

Relative to the orthids (Fig. 5), the mound stromatactoid cement isotope data (Fig. 6) is enriched in $\delta^{13}C$ and has a wider range of $\delta^{18}O$ values (average values $\delta^{18}O$= -5.2°/₀₀ and $\delta^{13}C$= +2.4°/₀₀).

The coquina cements (Fig. 6) have positive $\delta^{13}C$ values as high as the coquina brachiopods, but are slightly depleted in ^{18}O. This is compatible with the cathodoluminescence results (Pl. 1, fig. 5) and the trace element data (Fig. 4), which illustrate that the cements have incorporated greater amounts of secondary calcite than the brachiopods.

As with the brachiopod data (Fig. 5), there is some scatter in the cement data due to the presence of diagenetic calcite (Fig. 6).

DISCUSSION

Figure 7 compares the average isotopic composition of the brachiopods and cements obtained in this study to the isotopic values reported by Veizer et al. (1986) and Popp et al. (1986) for material of similar age. The mound components correspond well with the oxygen isotope data of Veizer et al. (op. cit.) and Popp et al. (op. cit.), although $\delta^{13}C$ values for the cements are enriched by about 1°/₀₀. The similarity between the stable isotopic composition of mound orthids and values reported in the literature suggests that carbonate in isotopic equilibrium with pre-Hirnantian sea water had a composition of approximately $\delta^{18}O$ = -4.4°/₀₀ and $\delta^{13}C$ = +1.5°/₀₀ (Fig. 7). The coquina *Hindella* brachiopods have highly elevated isotopic signatures and compared to the data of Veizer et al. (op. cit.) and Popp et al. (op. cit.), they are the most enriched in ^{16}O and ^{13}C of well preserved Ordovician material obtained to date.

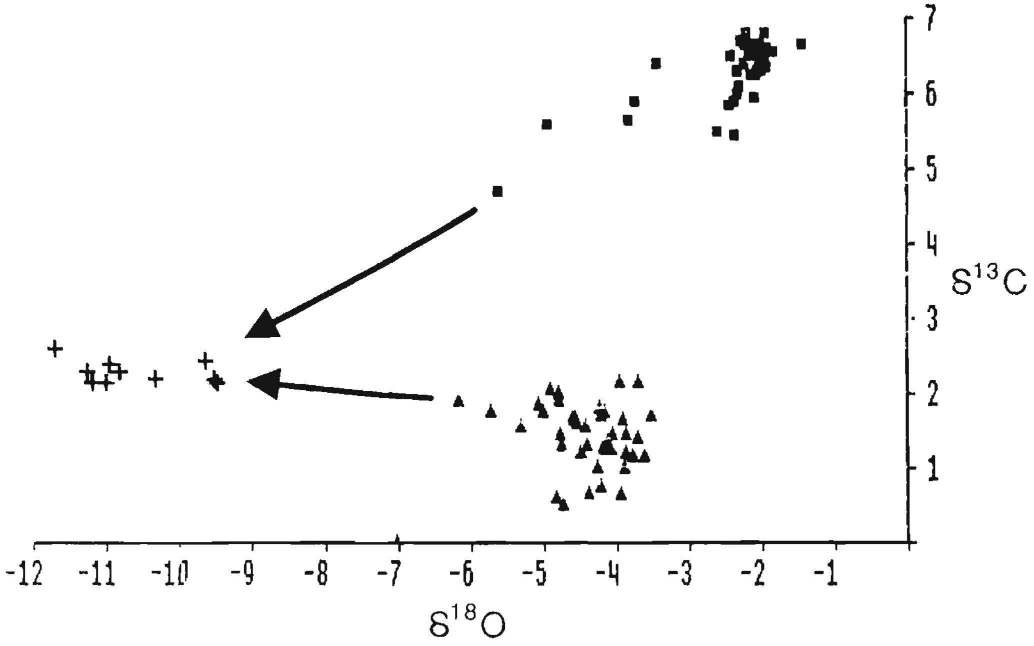

Figure 5. Brachiopod $\delta^{18}O$ **versus** $\delta^{13}C$ plot. Closed squares, *Hindella* brachiopods; closed triangles, mound orthids; crosses, diagenetic calcite. Primary marine signatures are shifted toward the diagenetic calcite field (arrowed) due the incorporation of variable amounts of burial cement into the brachiopod shell material.

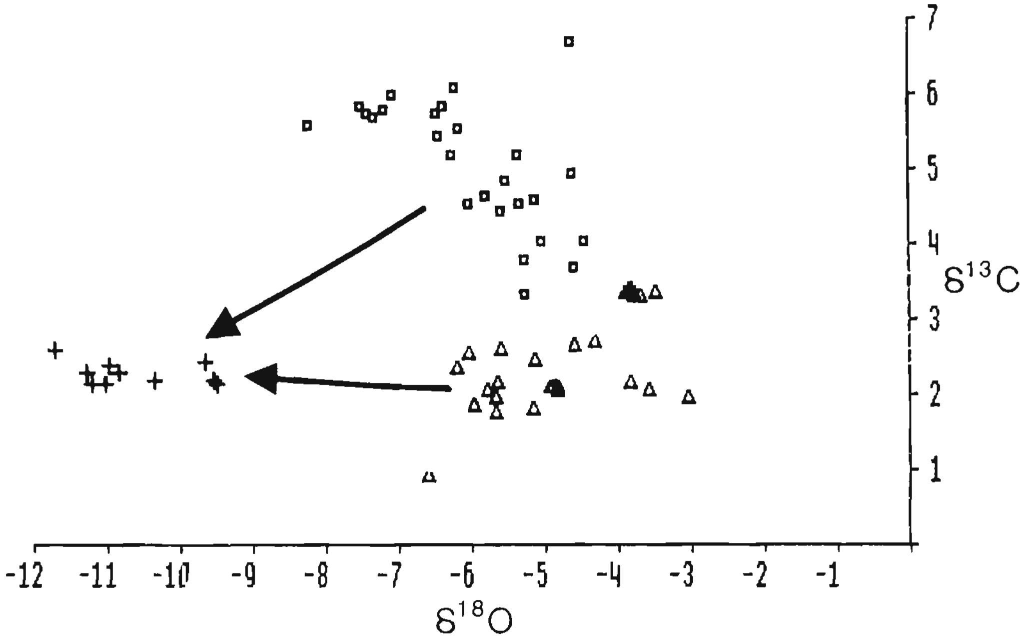

Figure 6. Marine cement $\delta^{18}O$ **versus** $\delta^{13}C$ plot. Open squares, Hirnantian coquina cements; open triangles, mound stromatactoid cements; crosses, diagenetic calcite. The marine cements have undergone a relatively greater degree of diagenetic alteration (compared to the brachiopods), resulting in the large range in $\delta^{18}O$ and $\delta^{13}C$ values, and a trend toward the diagenetic calcite field (arrowed).

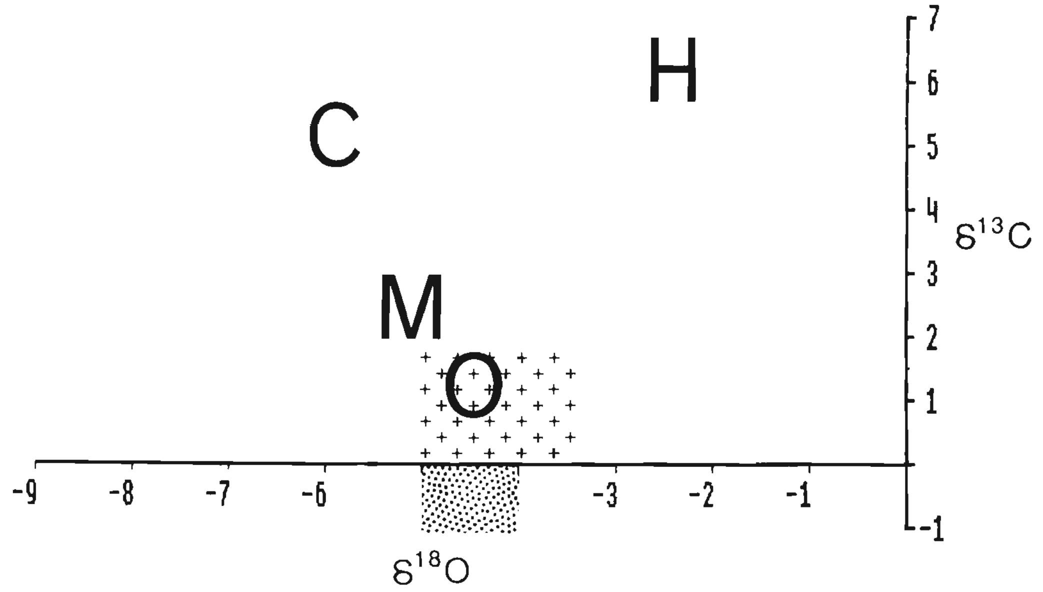

Figure 7. $\delta^{18}O$ **versus** $\delta^{13}C$ plot schematically representing the isotope data in Figures 5 and 6, and comparing it with other published isotope data from the upper Ordovician. H, *Hindella* brachiopods; C, Hirnantian coquina cements; O, mound orthid brachiopods; M, mound stromatactoid cements; crossed ornament, Ordovician brachiopod data (Popp et al., 1986; Veizer et al., 1986); stippled ornament, Ordovician marine cement data (Veizer et al., 1986).

The correspondence between stratigraphically equivalent brachiopod and marine cement data suggests that the isotopic values furthest from the burial cements (in Figs. 5, 6) are essentially unaltered primary marine values.

Comparing the two sets of primary signatures, an obvious shift occurs between the $\delta^{13}C$ values of the mound cements and brachiopods, and those of the Hirnantian age coquinas. In addition, the $\delta^{18}O$ values of the *Hindella* brachiopods is approximately 2°/₀₀ more positive than the mound orthid signatures. These shifts coincide with the glacially induced facies change from mound limestone to Hirnantian bedded bioclastic units and it is suggested that the changes in isotope composition are also linked to the glaciation.

Oxygen data

The 2°/₀₀ shift of average $\delta^{18}O$ values recorded by the brachiopods may partially be explained in terms of a change in ocean temperature. Using Craig's (1965) paleotemperature equation, a drop in ocean water temperature of approximately 10°C could account for the $\delta^{18}O$ enrichment. Such a large temperature change is thought unlikely because of the diverse nature of the biota present in both the mounds and the bioclastic beds, the presence of Hirnantian age coral-stromatoporoid buildups and oolites elsewhere on the platform, and the subequatorial location of Baltica (Brenchley, 1988). An alternative hypothesis involves a glacial "ice volume" effect (Hudson, 1977). Growth of the Gondwanaland icecap in Late Ordovician time would have preferentially entrapped ^{16}O within the ice and caused the remaining ocean water to be relatively enriched in ^{18}O. However, the size of the shift may be too large to be accounted for solely by this process. It is therefore proposed that a combination of a change in seawater isotopic composition and a small decrease in ocean temperature appears the likely cause, although the relative importance of each process remains unclear.

Carbon data

On the basis of the brachiopod data alone, the $\delta^{13}C$ shift could be interpreted as having been caused by a "vital effect", particularly as it proved impossible to sample a single genus of brachiopod in each facies. However, the fact that the shift is recorded in inorganically precipitated marine cements as well as skeletal calcite rules out any form of biological fractionation as the cause of the marked increase of $\delta^{13}C$ values. Slight difference in $\delta^{13}C$ values between mound orthids and their adjacent cements may, however, be due to minor vital effects (Figs. 5, 6).

Because the majority of samples analyzed exhibit some signs of diagenetic alteration, the effects of diagenesis must, therefore, be considered when attempting to explain the $\delta^{13}C$ shift between the mound limestones and the overlying Hirnantian beds. The petrography and geochemistry of the mound and bioclastic facies show they have an identical diagenetic history, thus ruling out any form of facies related diagenesis controlling the $\delta^{13}C$ shift. Furthermore, the occurrence of a $\delta^{13}C$ shift at exactly the same stratigraphic level in all three localities studied does not support a diagenetic origin.

Positive shifts in $\delta^{13}C$, similar to the one described in this paper, have been reported from elsewhere in the stratigraphic column (e.g., Weissert et al., 1979; Scholle and Arthur, 1980; Tucker, 1986; Carson, 1988). Such excursions are generally associated with periods of enhanced sedimentation of organic matter: the removal of isotopically light organic carbon leaves the remaining dissolved marine carbon relatively enriched in ^{13}C. There is, however, no evidence from the Baltoscandian Basin for the synchronous deposition of black shale or organic-rich sediment (Jaanusson, pers. comm.). An additional problem comes from faunal evidence that suggests generally low levels of planktonic productivity (Brenchley, 1984).

Despite these problems, the removal of ^{12}C-rich organic matter is favoured as the cause of the carbon shift. One potential solution is the generation of a ^{13}C enriched water mass outside the Baltoscandian Basin. Wilde and Berry (1984) conclude that the onset of glacial climatic conditions could induce oceanic destabilization, the subsequent overturn of dysaerobic or anaerobic water masses onto continental margins aiding in the preservation of organic-rich sediments. Such a scenario may be associated with the Late Ordovician glaciation, the Hirnantian brachiopods and marine cements recording the passage of a ^{13}C enriched water mass into the Baltoscandian Basin.

This paper represents the first documentation of a Hirnantian age positive carbon shift. The presence of a shift of this magnitude has important implications in the interpretation of the paleoceanography and paleoecology of Hirnantian times. The isotopic analysis of equivalent age sediments is required to ascertain whether the shift is merely a local effect, restricted to the seas around Baltica, or a widespread (possibly global) event.

A further discussion of oxygen and carbon isotope data for the Kullsberg and Boda limestones can be found in Marshall and Middleton (1990).

CONCLUSIONS

1. Brachiopods from coquinas capping the Boda mounds have the least negative $\delta^{18}O$ values and the most positive $\delta^{13}C$ values reported to date for unaltered Ordovician marine cements and brachiopods.

2. The similarity between the stable isotopic composition of the brachiopods and cements from within the mounds and published data for Ordovician carbonates suggests the stable isotopic signature of carbonate in equilibrium with nonglacial Ashgillian sea water was approximately $\delta^{18}O = -4.4‰$ and $\delta^{13}C = +1.5‰$.

3. Hirnantian brachiopods and associated marine cements record a positive carbon shift of approximately 4‰ and a positive oxygen shift of approximately 2‰. These shifts coincide with the facies change from mound limestone to bedded bioclastic units.

4. The shift in $\delta^{18}O$ is attributed to a combination of temperature decrease and a glacially induced change in the isotopic composition of the sea water.

5. Sedimentological and geochemical reasons preclude biological fractionation and diagenetic alteration as possible causes of ^{13}C enrichment. The most suitable explanation for the observed shift is a major change in the distribution of carbon between organic and inorganic carbon reservoirs.

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PLATE 1

Figure 1. Photomicrograph, in plane polarized light, of the isopachously layered coarse, radial fibrous marine cement that infills stromatactis cavities. Scale bar = 1 mm.

Figure 2. Photomicrograph, in plane polarized light, of radial fibrous coquina marine cements (C) forming an isopachous fringe on a *Hindella* brachiopod valve (H). Scale bar = 1 mm.

Figure 3. Cathodoluminescence photomicrograph showing the same field of view as figure 1. Scale bar = 1 mm.

Figure 4. Cathodoluminescence photomicrograph showing the same field of view as figure 2. Key as in figure 2. Scale bar = 1 mm.

Figure 5. Cathodoluminescence photomicrograph of a cross-section through a *Hindella* shell (H). Scale bar = 1 mm.

Figure 6. Cathodoluminescence photomicrograph showing the sectioned umbo (O) of a mound orthid (*Glyptorthis*). Scale bar = 1 mm.


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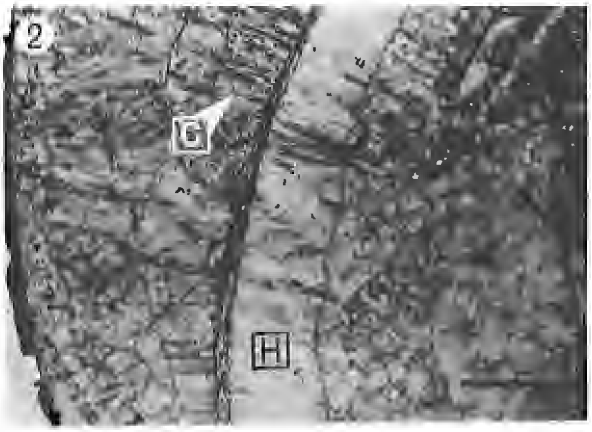
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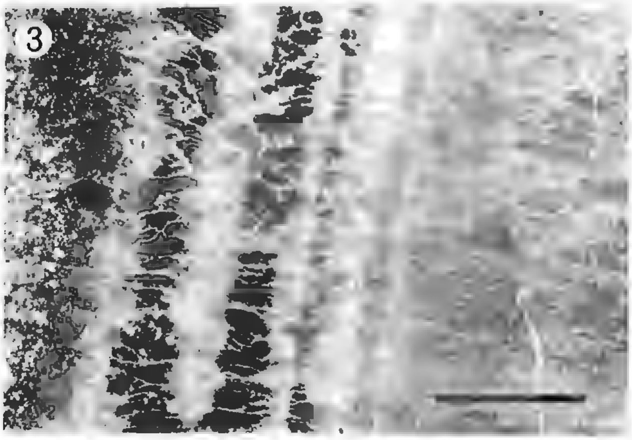
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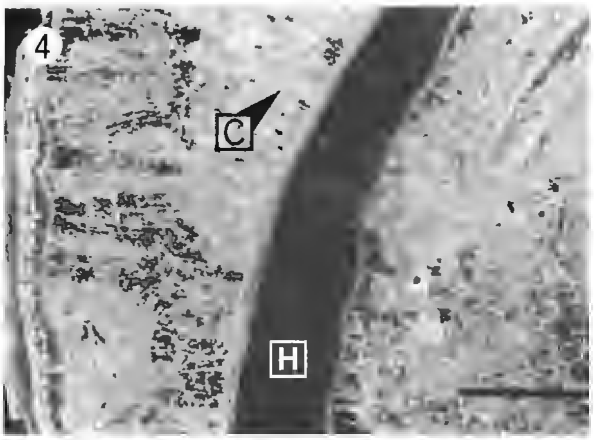
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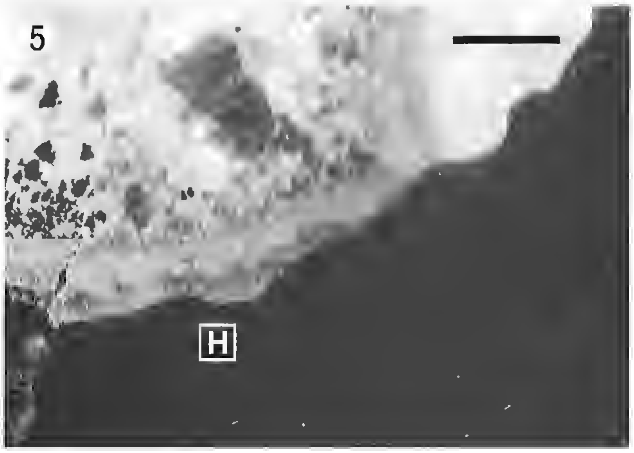
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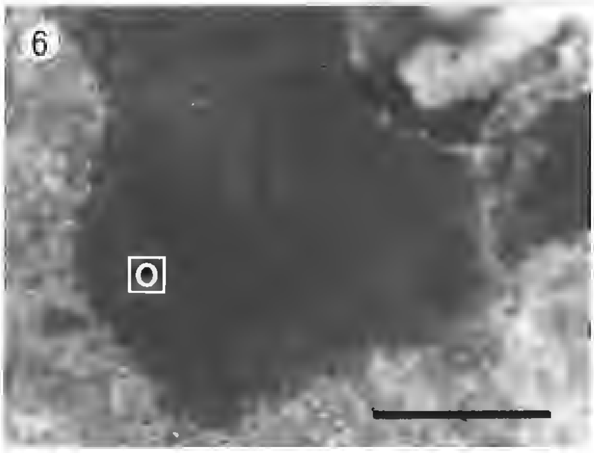
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Hirnantian glaciomarine diamictites — evidence for the spread of glaciation and its effect on Upper Ordovician faunas

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Abstract

Upper Ordovician sequences in Portugal and in the Prague Basin, Czechoslovakia, have glaciomarine diamictites that are Hirnantian in age. Upper Ordovician diamictites in Morocco are likewise of Hirnantian age, as probably are those elsewhere in Europe. Thus, although the continental glaciation of Saharan Africa cannot be accurately dated, the extension of the ice into peripheral areas appears to have been confined to the Hirnantian. This is consistent with Hirnantian eustatic sea level changes.

The sequence in the Prague Basin suggests that cold climates with floating marine ice developed early in the Hirnantian, before the main glacio-eustatic regression, whereas in Portugal, deposition from marine ice was somewhat later and postdated the regression.

The diamictites in the Prague Basin were probably deposited from an ice shelf or possibly from sea ice, whereas the diamictites in Iberia may have been deposited mainly from floating glacier ice.

The expansion of cold climates at the start of the Hirnantian coincides with the first and major phase of Late Ordovician extinction, and precedes the maximum low stand of sea level during the Hirnantian regression. It is concluded that lowered temperatures at the start of the Hirnantian may have played a major role in the extinction, particularly among high latitude faunas and floras.

Résumé

Au Portugal et dans le bassin de Prague, en Tchécoslovaquie, des séquences de l'Ordovicien supérieur contiennent des diamictites glaciomarines qui remontent à l'Hirnantien. Des diamictites de l'Ordovicien supérieur qui se trouvent au Maroc et vraisemblablement ailleurs en Europe datent eux aussi de l'Hirnantien. Par conséquent, bien qu'il soit impossible de dater avec précision la glaciation continentale de l'Afrique saharienne, la glace n'a vraisemblablement envahi les zones périphériques que durant l'Hirnantien, ce qui est compatible avec les fluctuations eustatiques du niveau marin à l'Hirnantien.

La séquence qui se rencontre dans le bassin de Prague porte à croire que des climats froids, caractérisés par la présence de glace marine flottante, se sont développés au début de l'Hirnantien, avant la principale régression glacio-eustatique. Par opposition, au Portugal, l'accumulation de sédiments en provenance de la glace marine a eu lieu un peu plus tard, après la régression.

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Les diamictites du bassin de Prague proviennent vraisemblablement d'une plate-forme de glace ou encore de la glace de mer, tandis que celles de la péninsule Ibérique pourraient s'être accumulées principalement à partir de glaciers flottants.

L'expansion des climats froids au début de l'Hirnantien coïncide avec la première et principale phase d'extinction de l'Ordovicien supérieur; elle précède la phase principale de bas niveau de la régression hirnantienne. Les basses températures au début de l'Hirnantien auraient vraisemblablement été une des causes principales de l'extinction, notamment des faunes et des flores des hautes latitudes.

INTRODUCTION

The existence of a Late Ordovician continental glaciation centred on the African part of Gondwanaland is now well established through the work of Beuf et al. (1971) and others. The glaciation was probably of "antarctic type, based on a large peneplaned continent, under the pole" (Spjeldnaes, 1981). Land-based ice occurred in parts of Saharan Africa and South Africa, and floating ice in peripheral regions (Fig. 1). The age and nature of the glaciation has been the subject of reviews by Spjeldnaes (1981), Hambrey (1985), and Robardet and Doré (1988). Spjeldnaes noted that it was difficult to determine the age of the continental glaciation because fossils are generally lacking, and even where they are present they are possibly derived from the underlying sequence. The peripheral glaciomarine deposits have yielded an Upper Ordovician (Hirnantian) fauna in Morocco (Destombes, 1968b) and elsewhere have been shown to be of broadly Late Ordovician age, suggesting that they may be roughly contemporaneous (Havlícek, 1974, 1976; Spjeldnaes, 1981; Robardet and Doré, 1988).

In this paper we use evidence from sequences with diamictites in western Iberia and the Prague Basin (Czechoslovakia) (Fig. 3) to constrain more closely the age of the peripheral glacial deposits. We then briefly review the likely nature of the depositional processes and finally comment on the likely influence of a rapid climatic cooling on Upper Ordovician faunas.

STRATIGRAPHY

The age of continental glacial deposits in Africa is not well constrained. In Saharan Africa, glacial deposits of the Felar-Felar Formation lie on an erosional surface above various units of different ages. Fossils of Caradoc or possibly Ashgill age are found in nodules in the green silts above the erosional surface in the Hoggar region, but as Spjeldnaes (1981) and Legrand (1985) pointed out, they are likely to be derived from the underlying sequence. Only in Morocco are there glacial deposits with intercalated marine horizons yielding a fauna of Hirnantian age (Destombes, 1968a, b, 1981; Destombes et al., 1985).

The upper limit of the continental glacial deposits is broadly indicated by the overlying black graptolitic shales. In Algeria, Legrand (1986) has recognized the presence of graptolites, probably equivalent to the *persculptus* Zone, and possibly the top of the underlying *extraordinarius* Zone, above the glacial deposits. In Morocco, the *persculptus* Zone is known only from one locality; more usually *acuminatus* is the lowest zone recorded above the diamictites (Destombes and Willefert, 1988). Farther south in the Hodh area of Mauritania, Willefert (1988) documented the *persculptus* Zone occurring in a sequence of microconglomerate, clay, and sandstone in the upper part of the glacial sequence.

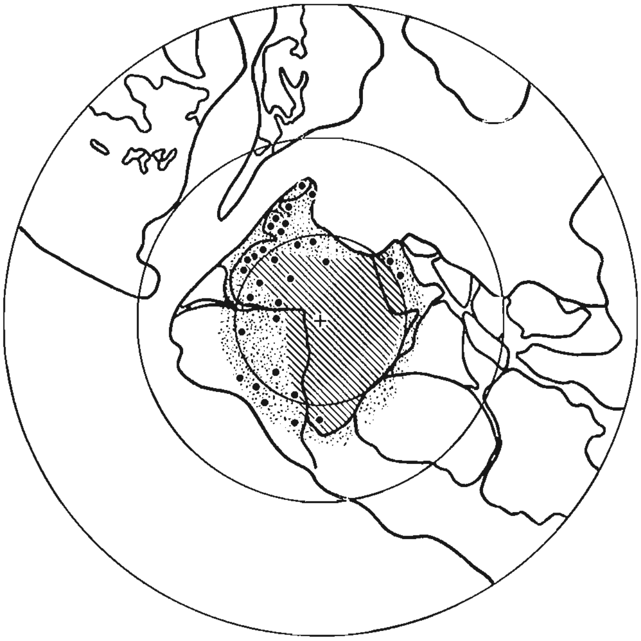


Figure 1. Distribution of Upper Ordovician glacigenic deposits. Dots indicate recorded localities. Stippled area has glaciomarine deposits dominant, hachured area has evidence for land-based ice. Some localities interpreted here as being of Late Ordovician age have been previously referred to the Early Silurian. (Base map from Ashgill reconstruction of Scotese and McKerrow, 1990)

Although there may have been a small polar ice cap throughout much of the Ordovician, if part of Gondwana lay in polar latitudes, the stratigraphic evidence from continental deposits of Saharan Africa only demonstrates that the glacial deposits were laid down sometime during the Caradoc or Ashgill.

In the peripheral areas, dating of the glaciomarine diamictites depends upon two different strands of evidence, one biostratigraphic, the other event-stratigraphic. Biostratigraphic evidence for correlation at this stratigraphic level is relatively good because the uppermost Ashgill stage (the Hirnantian) (Fig. 2) is characterized by the distinctive *Hirnantia* fauna. This has now been recognized widely throughout Europe and China, in North America, Asia and the U.S.S.R. (Rong, 1984) and more recently in South America (Benedetto, 1986). The fauna provides good evidence for identifying the presence of rocks of Hirnantian age. However, in sections where there is not a complete sequence of faunas or the underlying faunas are endemic, it may not be possible to determine the precise position of stage boundaries. Where faunal evidence is inadequate, event stratigraphy, using the evidence of sea level changes, can be more effective.

It has been shown that at many places around the world (Berry and Boucot, 1973; Brenchley, 1988), where uppermost Ordovician rocks are present, there is a regressive phase near the base of the Hirnantian followed by a transgressive phase at approximately the base of the *Glyptograptus persculptus* Zone (Fig. 2). Sea level changes that can be shown to be of approximately the same age on several separate plates are a good indication that the sea level changes are eustatic rather than tectonic. The effects of eustatic movements will be synchronous, and can be used for correlation except where they have been modified by particularly strong local tectonic effects. The first signs of the late Ashgill eustatic regression occur at, or close to, the base of the Hirnantian (Brenchley and Newall, 1980; Brenchley and Cullen, 1984; Brenchley, 1988), whereas evidence for the subsequent transgression is found below or within the *Glyptograptus persculptus* Zone, i.e., in the upper part of the Hirnantian (Fig. 2). Consequently, we have used the first evidence of regression to identify the base of the Hirnantian, and the transgression to identify the upper part of the stage, where suitable fossils are not present.

AGE AND ORIGIN OF THE DIAMICTITES

Outside Morocco, the age of the glaciomarine diamictites has not been precisely determined, though it has been suggested that they are late Ashgill in age (Havlícek, 1974). In Celtiberia (Spain, Fig. 3), for example, diamictites (the Orea Shales) occur above a karst surface on the Urbana Limestone (= Cystoid Limestone), which Hafenrichter (1980) suggested is of Rawtheyan age. In Normandy, Upper Ordovician diamictites overlie beds of Caradoc age and at Caen contain reworked fragments of Ashgill limestone (Robardet and Doré, 1988).

New evidence from Portugal and the Prague Basin suggests a Hirnantian age for the diamictites in these two regions.

Diamictites in Portugal

Diamictites are known from several areas of northern and central Portugal. Although diamictites have been recognized in Iberia for some considerable time, a recent review of the Upper Ordovician stratigraphy of central Portugal (Young, 1985) has allowed their reappraisal within this sector of the peninsula.

The upper member of the Sobredo Formation (Fig. 4) of the Valongo/Arouca area of northern Portugal (Romano and Diggens, 1976) consists of 1.5 m of laminated mudstone, which lies abruptly, possibly unconformably, over the sandstone of the lower member, and is overlain by more than 100 m of mainly massive "pebbly greywacke" grading upward into less pebbly lithotypes (Fig. 5, column V). At a few horizons there is a faint lamination with laminae up to a few millimetres thick. The diamictites are muddy, fine grained sandstone, bearing clasts of sandstone, quartz, and granite of 1-50 mm diameter, usually 2-10 mm (Romano and Diggens, 1976). The clasts are generally thinly dispersed through the sandstone and are seen on exposed surfaces as one in a square metre or less.

The biostratigraphic control on the diamictite is poor. The first fossils below the diamictite, from the Valongo Formation, are of Llandeilo, or possibly early Caradoc age (Romano, 1982,

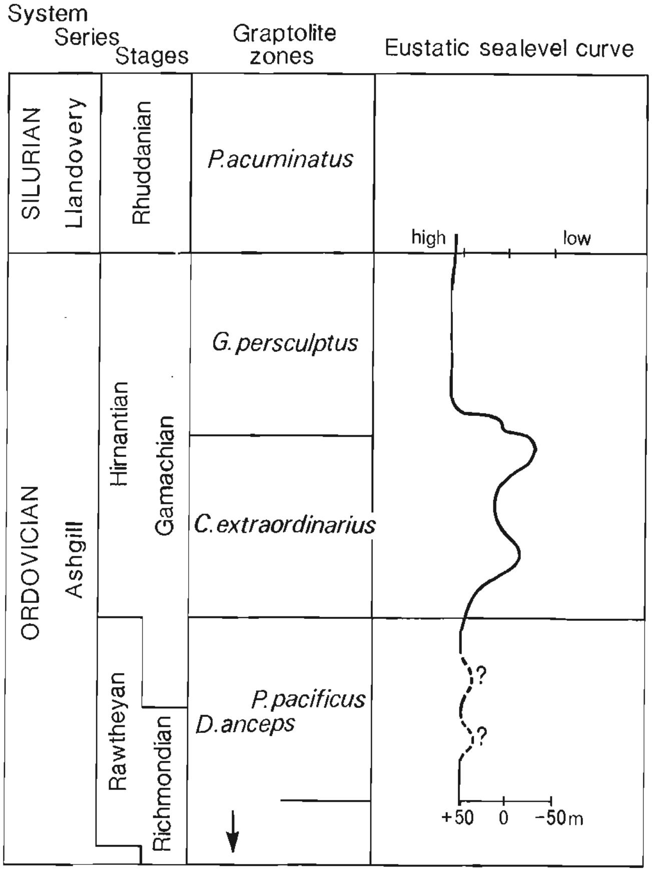


Figure 2. Upper Ordovician stratigraphy and diagrammatic illustration of the timing of glacio-eustatic sea level change in the Late Ordovician.

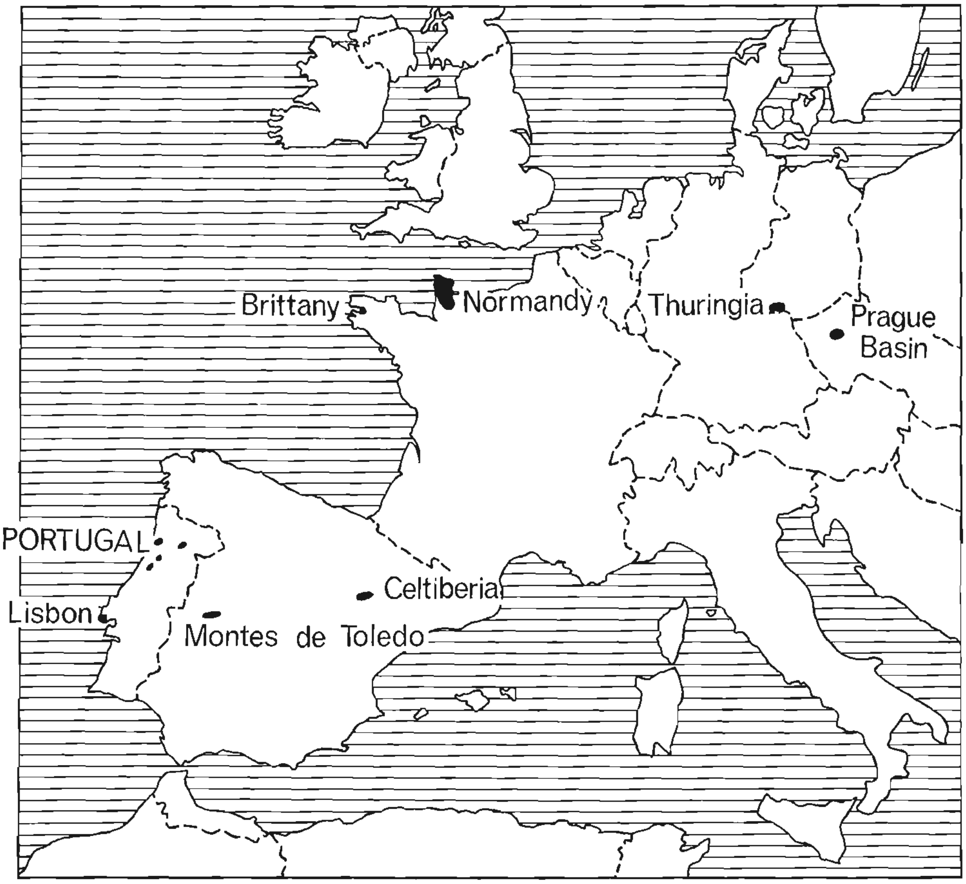


Figure 3. Areas in Europe with Upper Ordovician diamictites (solid black areas).

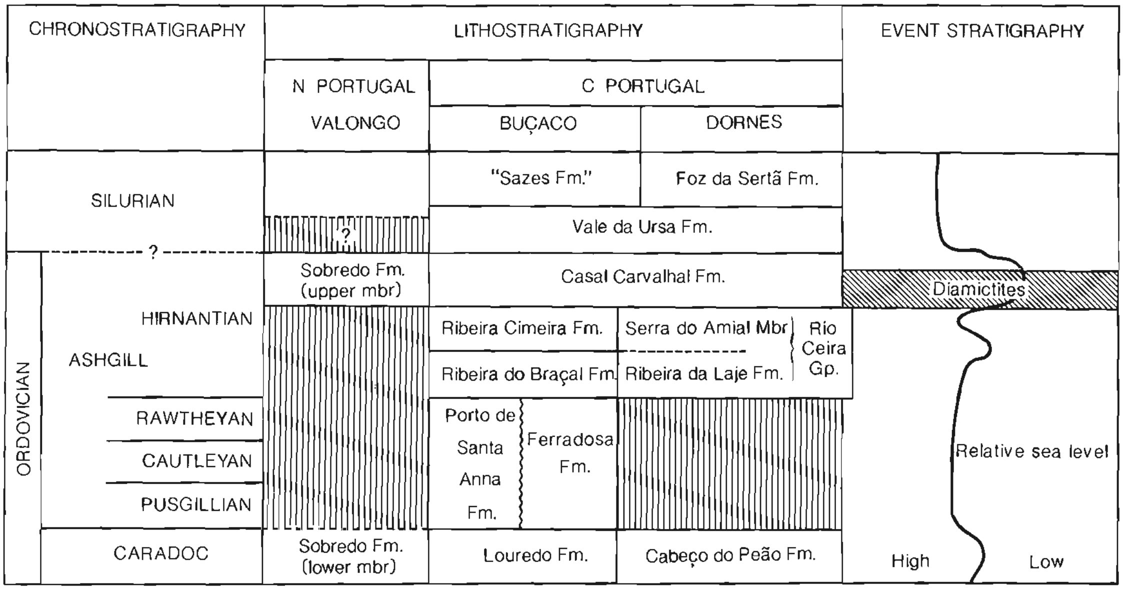


Figure 4. Lithostratigraphy of Upper Ordovician deposits in north and central Portugal, showing correlation of units with chronostratigraphic scale and eustatic and glacial events.

and work in progress) and the earliest Silurian graptolites above the diamictites are of middle to late Llandovery age (Romariz, 1963).

In central Portugal, a more complete sequence of Late Ordovician age is exposed (Figs. 4-6). Diamictites generally overlie a regressive sequence, the Rio Ceira Group (Fig. 4) (Young, 1988). This sequence (Fig. 5, between lines 1 and 3) includes mudstone and siltstone toward the base (the lower part of the Ribeira da Laje Formation of south-central Portugal and the Ribeira do Braçal Formation of the Buçaco Syncline), but the frequency and thickness of intercalated sandstone beds increase upward through these formations, culminating in the Serra do Amial Member of south-central Portugal (e.g., Dornes). In the Buçaco area, there was a lower to middle Ashgill volcanic centre (Porto de Santa Anna Formation), with associated bioclastic carbonates (Ferradosa Formation). Where these deposits are thin, in the south of the Buçaco area, they are overlain by a regressive sequence (Ribeira do Braçal Formation), but around the margins of the thicker developments the volcanics are unconformably overlain by the conglomeratic sandstone of the Ribeira Cimeira Formation (Figs. 4, 5). This conglomeratic sandstone also lies unconformably on the Ribeira do Braçal Formation in the southern part of the syncline. Some comminuted conglomeratic debris has also been recorded in the Serra do Amial Member of south-central Portugal. The clasts in the conglomerate are almost exclusively derived from the Porto de Santa Anna Formation.

The diamictites overlie the Ribeira Cimeira Formation and the Serra do Amial Member and are interbedded near the contact in some areas (Fig. 5). They lie within the lower part of the Casal Carvalhal Formation. In the Dornes area, the base of the diamictites consists of a massive, hard, pebbly sandstone, with clasts up to 20 cm in diameter. The diamictites at most localities are less than 30 m thick and consist of massive, poorly sorted, muddy sandstone, bearing isolated large sand grains (Pl. 1, fig. 4) and clasts of up to 5 cm in diameter, although rare clasts of up to 15 cm in diameter have been recorded. The lithotypes represented in the clasts include sandstone, limestone, and ooidal ironstone. In the Buçaco Syncline, the dominant clast lithotype is a decalcified impure limestone bearing chamosite ooids. In many areas, there are well sorted, laminated sandstones, probably of storm origin, intercalated within the lower part of the diamictites. These show evidence of significant soft sediment deformation, particularly load casts.

The diamictites grade upward into dark mudstone with sandy laminae, which form the upper part of the Casal Carvalhal Formation. In the uppermost part of the formation, these laminae become thicker, and distinct load-casted sandstone beds are intercalated in the few metres immediately below the base of the Vale da Ursa Formation. This formation consists of black, pyritic sandstone, often showing considerable load casting. The upper part of the Vale da Ursa Formation contains finely laminated sandstone, the Serra dos Aguilhões Member, and is overlain by micaceous muddy sandstone with mudstone rip-up clasts, the Junqueira Member, at the base of the Foz da Serta Formation. The Junqueira Member is generally less than 4 m thick and is sharply overlain by black graptolitic mudstone, which forms the bulk of the Foz da Sertã Formation.

The Upper Ordovician sequence in the Rio Ceira Group of central Portugal is interpreted as recording glacio-eustatic regression (Fig. 5, between lines 1 and 3), followed by the deposition of glaciomarine diamictites. The diamictites are interpreted as being of glaciomarine origin because of the

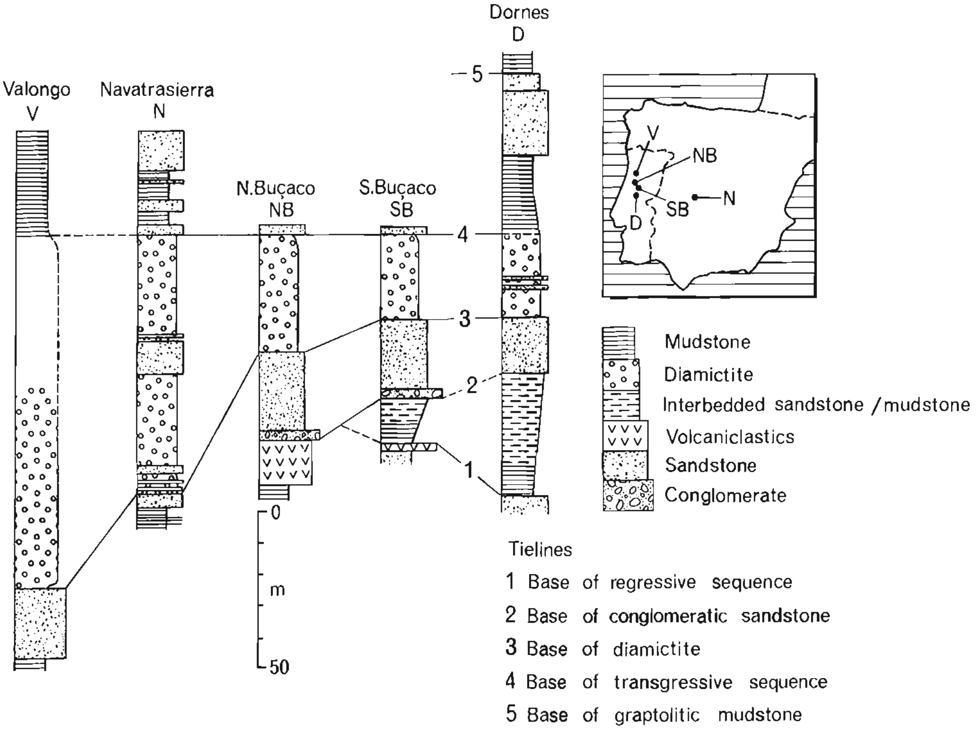


Figure 5. Selected diamictite-bearing sequences of Late Ordovician age in the Iberian Peninsula. The localities in Portugal are 20 to 50 km apart. Navatrasierra is about 250 km from Buçaco.

combination of their texture, as well as their very wide extent, forming a blanket deposit over a very large area, and the absence of any demonstrable lateral facies change (see discussion below). Most of the pebbles recovered from the diamictites of central Portugal are subangular to subrounded. Striated pebbles have not been found, although a faceted morphology is not uncommon among the larger clasts. The clast lithotypes are compatible with relatively local derivation. The similarity between the impure limestone with chamositic ooids seen as clasts in the Buçaco area, and the lithotype of the lower part of the Porto de Santa Anna Formation in the southern part of the syncline is striking. Young (1985, 1988, 1990) has described a belt of reduced subsidence during the Ordovician across central Portugal, the Dornes/Amendoa "rise", lying between the southern part of the Buçaco Syncline and the Dornes area. This belt could have acted as a source for all the pebble lithotypes seen, and might have been a slightly positive topographic feature at the time. The presence of striated pebbles (Pl. 1, fig. 1) in central Iberia (Navatrasierra in the Montes de Toledo), demonstrates some input of glacially eroded debris, but the subrounded nature of the clasts, together with their inferred local derivation, suggest that sea ice might have played a role, as well as glacier ice, during the deposition of the diamictites.

The upper part of the Casal Carvalhal Formation is interpreted as being deposited during the glacio-eustatic transgression. The gradual passage of the diamictites upward into better-sorted sediments suggests a slowing of the sedimentation rate and increased reworking. The Vale da Ursa Formation and the base of the Foz da Sertã Formation are the products of reworking on the sediment-starved shelf after sea level rise. It is only in the upper Llandovery that sedimentation resumes, albeit with organic-rich black graptolitic mudstone.

The central Portugal area provides some useful biostratigraphic data on the diamictites (Young, 1990). In the southern part of the Buçaco Syncline, the Porto de Santa Anna Formation is reduced to just a few metres in thickness. Its base contains a fauna (Fig. 6, assemblage 1) that is of Pusgillian age, but the top yields a fauna of small brachiopods (Fig. 6, assemblage 2) showing some affinity to the *Proboscisambon* Community (Havlícek, 1982; Havlícek and Mergl, 1982; Cocks and Rong, 1988) from the uppermost Králuv Dvur Formation (Rawtheyan) of the Prague Basin. These Rawtheyan sediments are abruptly overlain by siltstone at the base of the Ribeira do Braçal Formation, bearing a low diversity shelly fauna (Fig. 6, assemblage 3) with elements related to those of *Hirnantia* faunas in other regions. These sediments, which also bear large isolated quartz sand grains, mark the earliest sediments deposited in the area during the glacio-eustatic regression and are interpreted as being of Hirnantian age.

In the south-central Portugal region (e.g., Dornes) the Serra dos Aguilhões Member contains, near its base, a poorly preserved fauna of graptolites (Fig. 6, assemblage 4), indicating a Rhuddanian age. The overlying Junqueira Member is a condensed deposit, yielding middle to late Llandovery age graptolites (Fig. 6, assemblage 5). The base of the graptolitic mudstone above the Junqueira Member yields a graptolite fauna of late Llandovery (*griestoniensis* Biozone) age.

Diamictites in the Prague Basin

The Ashgill of the Prague Basin consists of a monotonous sequence of grey mudstone, the Králuv Dvur Formation, overlain by sandstone and shale of the Kosov Formation. The Králuv Dvur has a trilobite and sparse brachiopod fauna belonging to benthic assemblage 5 or 6, suggesting a deep shelf or slope environment (Havlícek, 1982), whereas the sandstone of the Kosov Formation has wave ripples and hummocky cross-stratification, implying a shallow shelf environment (Brenchley and Storch, 1989).

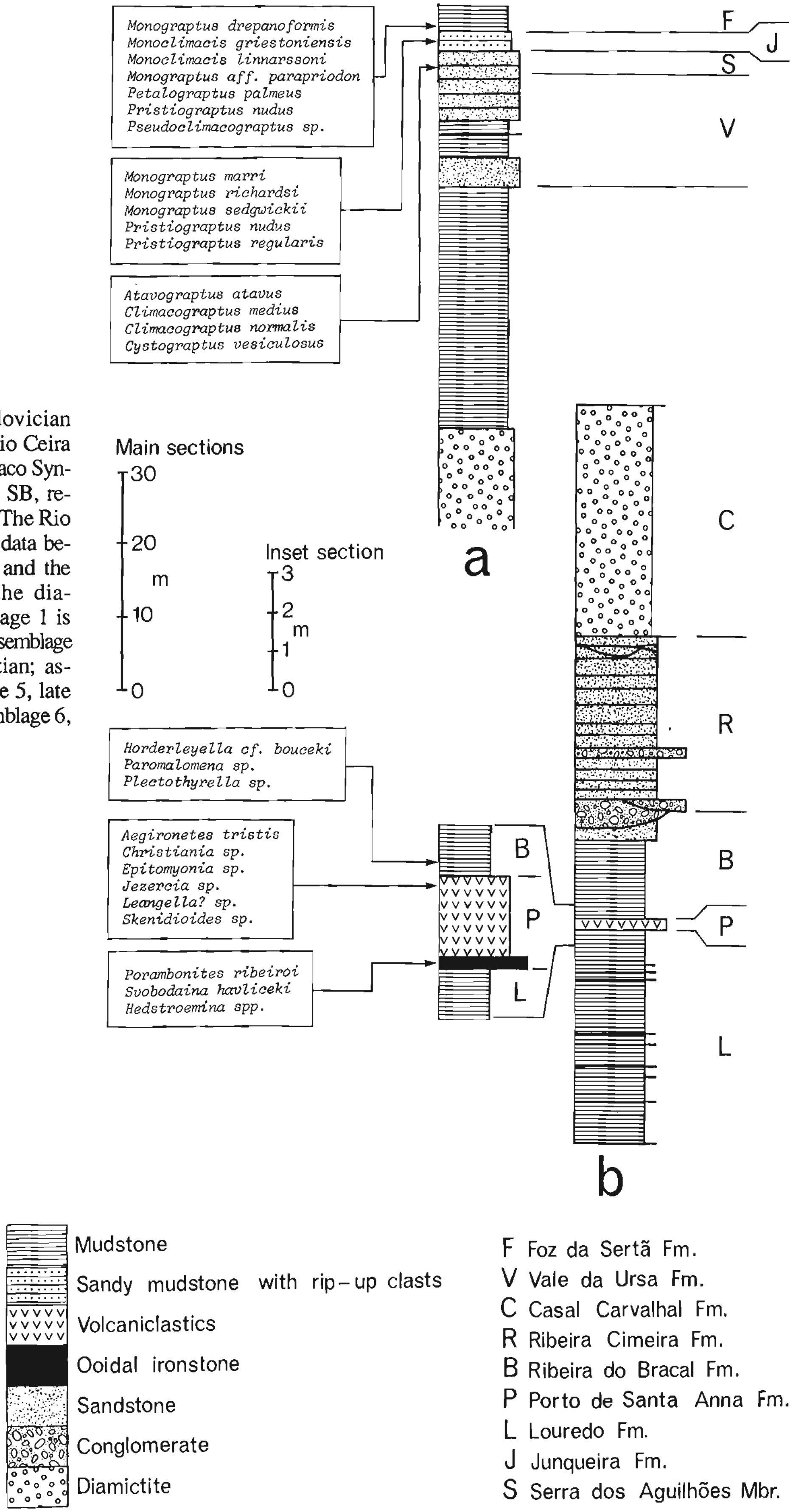
The Kosov Formation has a fauna of Hirnantian age in its upper part, but is mainly unfossiliferous in its lower and middle portions, which prevents the identification of the lower Hirnantian boundary on purely biostratigraphic evidence. There are, however, sedimentological and faunal changes in the top few metres of the Králuv Dvur that suggest the initiation of the late Ashgill regression. Within 3 m of the top of the Králuv Dvur is the Perník Bed (Fig. 7, Unit A), a thin (<10cm) calcareous mudstone/siltstone, with a diverse fauna of more than 70 genera including many trilobites and brachiopods. In the succeeding few centimetres (Fig. 7, Unit B) the rich fauna disappears, to be replaced by a low diversity *Mucronaspis* association, containing mainly *Mucronaspis*, ostracodes, and bivalves (Storch and Mergl, 1989). At the same level there is a change in the colour and texture of the mudstone, from the calcareous Perník Bed (Fig. 7, Unit A), through brown-grey silty mudstone, to silty, grey shale with sparse nodules, to silty grey shale without nodules (Fig. 7, Unit B).

The *Mucronaspis* association, though probably not diagnostic of the Hirnantian, is a common deep shelf association in the Hirnantian (cf. Lespérance and Sheehan, 1981) and its presence, taken together with the lithological change, is believed to identify the base of the Hirnantian at a few centimetres above the Perník Bed.

The lower diamictites (Fig. 7, Unit C), which have traditionally been taken to mark the base of the Kosov Formation, occur 3 to 4 m above the Perník Bed and about 3 m below the sandstone of the Kosov Formation, which reflects the peak of the first phase of the Hirnantian regression. The lower diamictite horizon is 10 to 20 cm thick and consists of four or more thin, structureless, ungraded beds. The sediments have well rounded to subrounded quartz grains (<2 mm in diameter) in a fine sand/silt/mud matrix. Sparsely dispersed through the rock are larger clasts, usually less than 2 cm in diameter.

The upper diamictite (2 to 3 m thick, Fig. 7, Unit E) is separated from the lower one by about 2 m of mudstone. The diamictite consists of several beds, 10 to 30 cm thick, passing up into very thinly bedded diamictite (1.0 to 10 mm thick), separated by thin partings of mudstone and siltstone (Pl. 1, fig. 2). Clasts, usually 1 to 2 cm in diameter but more rarely up to 5 cm in diameter, are thinly dispersed through the diamictite, especially in the thicker beds (Pl. 1, fig. 3). The clasts are mainly siltstone, sandstone, or oolitic ironstone derived from the underlying formations, though granitoids and clasts of chert and mildly metamorphosed shale were probably derived from the surrounding Proterozoic basement, suggesting the possibility of glacial erosion (Havlíek, 1974).

Figure 6. Sections of Upper Ordovician deposits from a. Dornes and b. the Rio Ceira inlier in the southern part of the Buçaco Syncline (see also Fig. 5, columns D and SB, respectively), both in central Portugal. The Rio Ceira area provides biostratigraphic data beneath the diamictite (assemblages 13) and the Dornes area provides data above the diamictite (assemblages 4-6). Assemblage 1 is interpreted as being of Pusgillian age; assemblage 2, Rawtheyan; assemblage 3, Himantian; assemblage 4, Rhuddanian; assemblage 5, late Aeronian-early Telychian; and assemblage 6, Telychian, ***griestoniensis*** *Biozone*.



The diamictites are thought to be glaciomarine in origin because 1) the very poorly sorted texture and structureless bedding is common among glaciomarine sediments and closely resembles that found in approximately contemporaneous glaciomarine diamictites in Normandy (Dangeard and Doré, 1971); 2) Storch, 1990 has recently discovered facetted clasts and dropstones within the diamictite; 3) the beds are unlikely to have been deposited as debris flows because some are only a few millimetres thick. Furthermore, the diamictites have such a closely similar stratigraphy on both margins of the basin, 15 km apart, that they are unlikely to have been gravity-flow deposits. Although the texture and bedding of the sediments and the presence of striated clasts suggests deposition from floating ice, the sand grains in the diamictites do not show dense pitting on the surface, which is commonly found in glacially transported sediment. The presence of mainly locally derived clasts implies erosion of nearby areas. The well rounded sand fraction in the diamictites appears to have been derived unmodified, from adjacent shorelines. The implication is either that ice passed from nearby platform

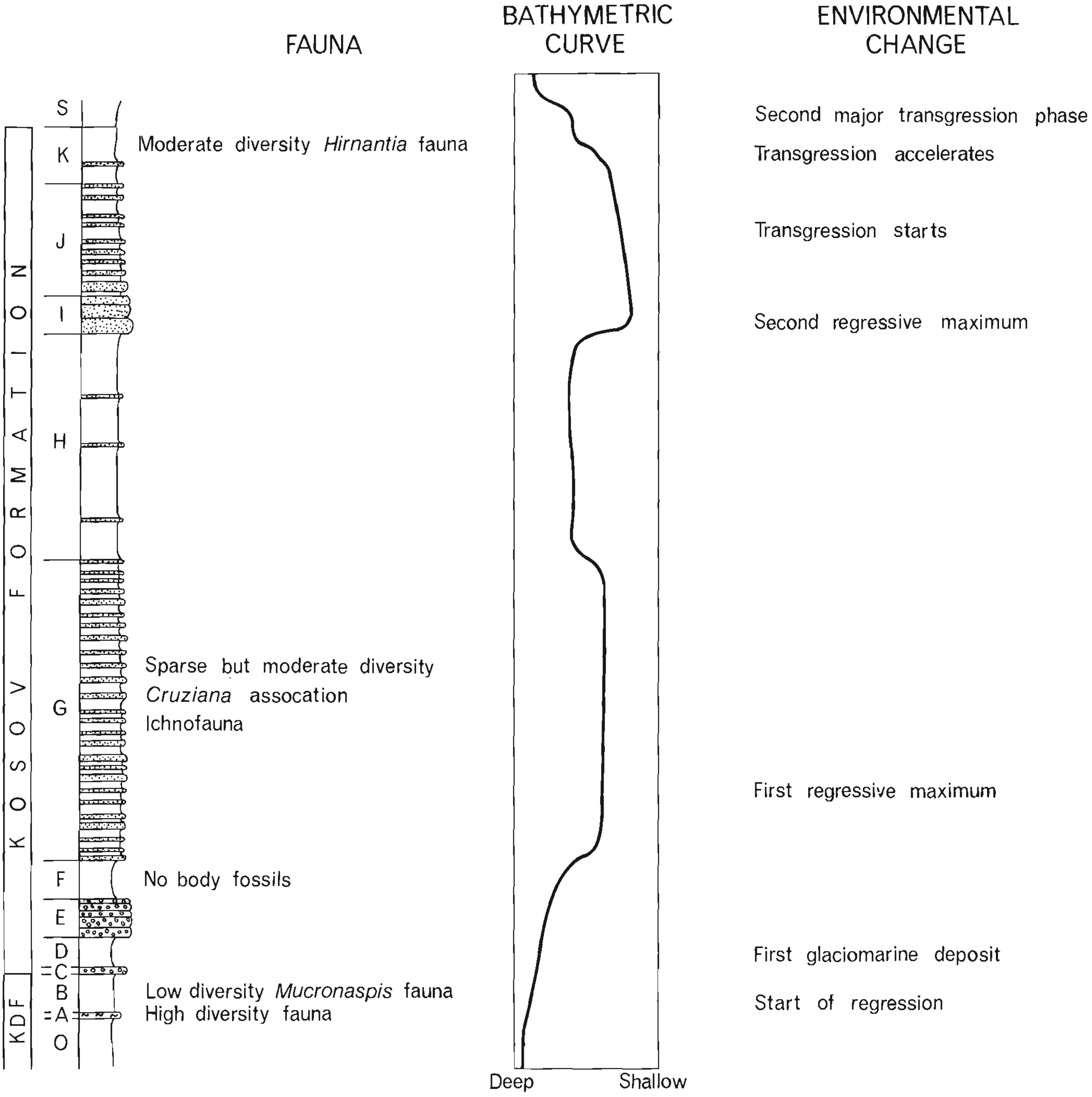


Figure 7. Upper Ordovician sequence in the Prague Basin, with environmental and bathymetric interpretation. The generalized sequence is based on Storch and Mergl (1989).

areas across contemporary shorelines and left deposits subglacially, or possibly sometimes sediment was deposited from seasonal ice (Spjeldnaes, 1973, 1981; Clark and Hanson, 1983), in the adjacent marine basinal areas.

Age of the diamictites: discussion

It has been shown that the diamictites in the Prague Basin were deposited early in the Hirnantian, prior to the first major regressive phase. In Portugal, the diamictites occur after the regressive phase, suggesting they were deposited somewhat later. Elsewhere, there is some evidence that the diamictites are of slightly different ages within the Hirnantian. In Normandy, diamictites occur before any markedly regressive event, whereas in Celtiberia the diamictites occur above a karst surface (Hafenrichter, 1980), implying that deposition postdated the regression, as in Portugal. Whether the diamictites of different age represent local advances of ice belonging to the same glacial advance, or whether deposits reflect different glacial phases, is not easy to determine.

Origin of diamictites: discussion

The diamictites occur on an extensive platform fringing Gondwanaland. The Ashgillian sediments underlying the diamictites are generally middle to deep shelf mudstone of uniform facies or are sandstone capping a regressive sequence. There is no evidence of marginal or intrashelf mountain areas, nor of steep slopes marginal to, or within the platform. The Hirnantian glaciomarine sea level changes place the diamictites in an overall regressive context, although there were probably also minor sea level fluctuations within the Hirnantian.

The diamictites are characteristically massive or indistinctly laminated, and of uniform facies laterally. We reject the possibility that they are debris flows for the following reasons: 1) they occur on an extensive, flat platform; 2) they are closely associated with shallow marine deposits; 3) they occur, with similar facies, on both flanks of the Prague Basin; 4) there are no associated turbidites.

The diamictites do not appear to be the products of continental glaciation because there is commonly no erosional surface at their base; they lie between marine deposits and are interbedded with marine deposits in some places; they lack any lenticular sands indicative of subglacial melt waters; and there is no apparent clast fabric.

The diamictites are unlikely to have been deposited proximally where glacier ice passes out into the sea. The deposits in such environments show evidence of rapid sedimentation from streams or subglacial melt waters or fallout from ice and deposition from debris flows. The products typically show rapid vertical and lateral facies changes (Eyles et al., 1985). More probably, the diamictites are the products of floating ice on an ice shelf or of seasonal winter ice, but their exact mode of origin is difficult to determine because there appear to be no described Recent analogies. Most modern ice shelves draw some of their ice from a glaciated hinterland and commonly contain erratics drawn from an uplifted source region. In contrast, most of the clasts contained in the Hirnantian diamictites have come from Ordovician rocks of nearby areas, whereas the well rounded coarse sand fraction appears to have been derived from contemporary shorelines. The ice that deposited the diamictites either passed from a region of recently emerged marine sediments, crossed a shoreline and floated free as shelf ice, or the debris was accreted at the shoreline by seasonal winter ice (Spjeldnaes, 1973, 1981; Clark and Hanson, 1983).

The implications seem to be that a wide, flat, emergent platform developed its own ice cover, which may have been driven by Gondwanan ice. That the platform ice flowed at least locally as glacier ice is indicated by the presence of striated clasts in some places (Normandy, Prague Basin, and central Spain).

Marine diamictites normally would have been formed when ice sheets advanced and global sea level fell. The regressive sequence in Portugal is interpreted as the result of such a glacio-eustatic fall. The presence of marine diamictites above the regressive sequence therefore appears anomalous. It can, however, best be explained by the effect of loading by an advancing or growing ice sheet, which being grounded on the topographic highs may have isostatically depressed the whole platform area and re-established marine conditions in the basinal depressions. We believe it was this ice cover over a platform of low topography, with areas of grounded ice and intervening areas of floating ice, which gave the distinctive character to the Upper Ordovician diamictites.

ENVIRONMENTAL AND FAUNAL CHANGES

The presence of glaciomarine diamictites a small distance above the base of the Hirnantian implies that cold climates were spreading into lower latitudes early in Hirnantian times. If the peak of the first regressive phase reflects the first maximum growth of ice caps, this was somewhat later in the Hirnantian. The diamictites of the Prague Basin could reflect the initial cooling, whereas those of Iberia could reflect the maximum spread of ice.

The most pronounced phase of Late Ordovician extinction in high and middle latitudes occurred at the base of the Hirnantian (Brenchley, 1984a, b), before the Hirnantian regression had a major effect on many shelf environments. On the other hand, there is now good evidence that the climate was becoming colder in early Hirnantian times. A significant fall in sea surface temperature could account for the sudden disappearance of a major part of the planktonic fauna and flora (graptolites, acritarchs, conodonts) outside the tropics, and might also have affected the benthic fauna. Whether there was a contemporaneous intrusion of cold bottom waters into oceanic regions, causing an oceanic overturn bringing toxic, unconditioned water to the surface is debatable (cf. Wilde and Berry, 1984) but changes in seawater chemistry could also have had a major and almost instantaneous effect on both plankton and benthos.

There is no evidence that the lowered temperatures had an immediate effect on tropical environments, where carbonate deposition apparently continued without interruption into the Hirnantian. It is possible that the very extensive shallow marine platforms of tropical regions were affected by falling seawater temperatures at a slightly later stage, or the Hirnantian regression had a major effect in these areas (Sheehan, 1973, 1975). A relatively small fall in sea level could have diminished the marine area considerably, and might have seriously affected the "perched faunas" by a reduction in habitable area (Johnson, 1974).

In conclusion, we identify the importance of lowered temperatures in the early Hirnantian extinction, and thus support the contention of Stanley (1988) that global cooling was a major factor in Paleozoic mass extinctions. However, our hypothesis differs from that of Stanley in that it seems premature to identify one single cause for the terminal Ordovician extinction. It seems more likely that there were a number of related factors such as change in temperature, seawater geochemistry and habitable area, each linked to the growth and decay of the Upper Ordovician ice caps, but each having its own effect on the faunas.

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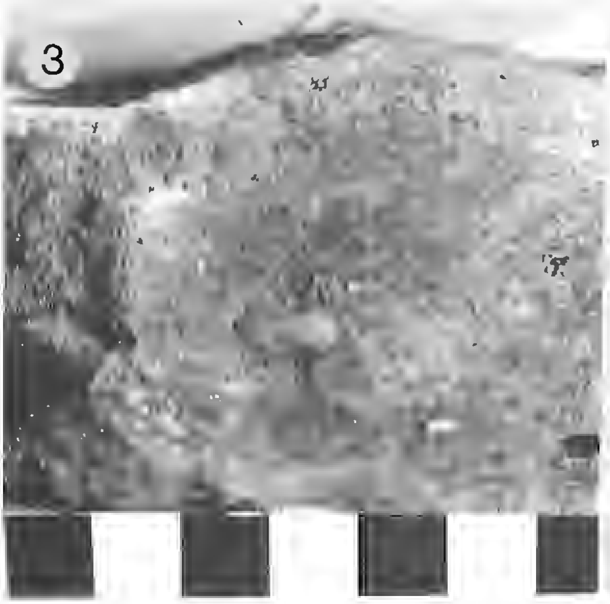

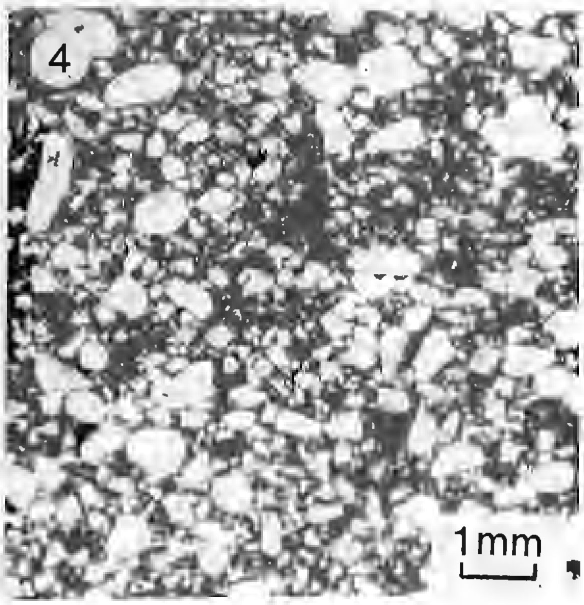

PLATE 1

Figure 1. Striated quartzite clast in the diamictite sequence in the Guadarranque Syncline near Navatrasierra, west-central Spain.
Figure 2. Thinly interbedded diamictite and mudstone, Bechovice section, Prague Basin.
Figure 3. Diamictite showing clasts in sandy mudstone, levin section, Prague Basin.
Figure 4. Photomicrograph of diamictite from Rio Ceira section, central Portugal.