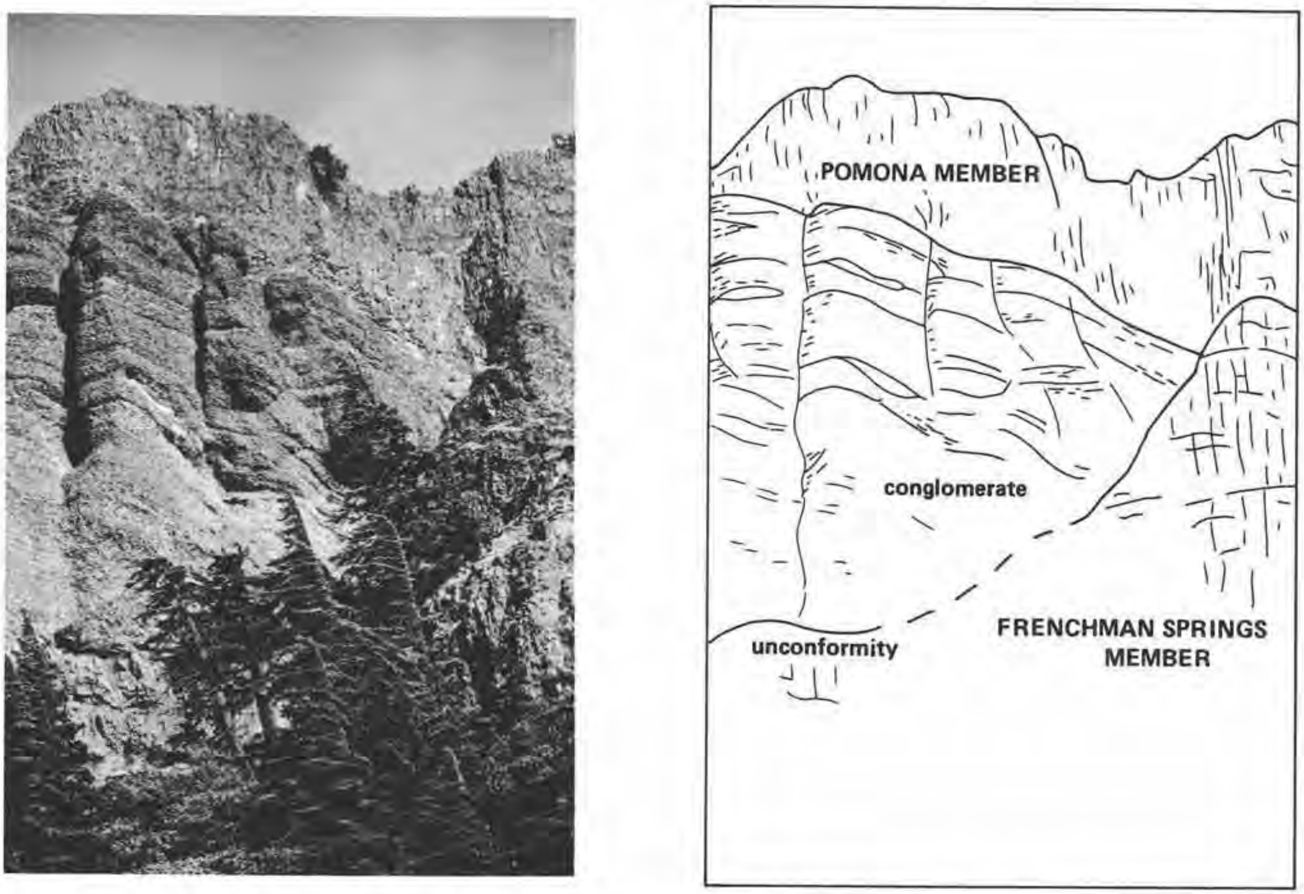

Figure 11.–Contact relations of the Pomona Member intracanyon flow in the Viento Park area (shown in Fig. 10). View to the southwest from Interstate Highway 84. White lenses in the conglomerate are a quartzofeldspathic sandstone.

ric, it commonly has a greater (by a factor of 2 to 3) concentration of phenocrysts than that in less phyric areas. Where rich in phenocrysts, this unit takes on a truly porphyritic character. Abundant phenocrysts and the presence of megascopic olivine led some early mappers to misidentify the Pomona Member as an olivine basalt of Cascadian origin, an understandable error.

The Pomona Member, like the Priest Rapids Member, has reversed paleomagnetic polarity. However, the chemistry of these two units is quite different (Table 1). The Pomona Member has relatively lower FeO, TiO_2, P_2O_5, and K_2O contents than other flows in the section and has higher concentrations of MgO and CaO than the others. These differences are summarized in detail in Swanson and others (1979a).

The jointing of the Pomona Member intracanyon flow (Fig. 11) is very similar to that observed at the type locality. It has well-developed basal columns and an entablature that appears to be hackly jointed but is better described as being made up of narrow, irregular columns 5 to 15 cm wide. This flow outwardly resembles older Columbia River basalt flows in each of the areas discussed in this paper. It particularly resembles the Grande Ronde Basalt, which is by far the dominant exposed unit within the Columbia River Gorge. This resemblance has contributed to past confusion regarding the identification and recognition of intracanyon relations (Allen, 1932; Wells and Peck, 1961; Sceva, 1966; Kienle, 1971; Waters, 1973; Beaulieu, 1977).

KLICKITAT RIVER AREA

In the Klickitat River area (Fig. 10), a canyon approximately 1 km wide and more than 100 m deep contains not only the Pomona Member (12 m.y. old) but also the Elephant Mountain Member (10.5 m.y. old). These units are exposed where a bend in the paleodrainage is transected by the Klickitat River canyon. Both units are in contact with flows of the Frenchman Springs Member across steep buttress unconformities. The exposures of the paleocanyon suggest a width of approximately 1.25 km and a minimum depth of 100 m below the highest known paleotopography at the top of the uppermost Frenchman Springs flow. It is probable that the top of this Frenchman Springs flow closely approximates the paleocanyon rim because units normally within the intervening stratigraphic interval are known to lap out 5 to 12 km to the south, on the south side of the Horse Heaven Hills uplift (Anderson, 1987; Swanson and others, 1979a). The depth of the paleocanyon is unknown due to insufficient depth of exposure. However, the basal colonnade of the Pomona Member is exposed at river level, suggesting that the flow base is not very far beneath the river level, assuming that relations are similar to those in areas where the base is exposed, such as at Mitchell Point. Whether gravel deposits are present beneath the Pomona at the Klickitat River area remains unconfirmed. However, such deposits would be consistent with all other known localities.

Sedimentary deposits are exposed at the top of the Pomona Member and underlying the Elephant Mountain Member. These deposits consist of a well-cemented quartzofeldspathic micaceous sandstone unit overlain by pebble to cobble conglomerate with collective thickness of approximately 20 m. Significantly, the top of the underlying Pomona Member is preserved, and the basal sandstone is cemented with silica. This cementation is strongest near the underlying lava flow, suggesting that it resulted from ground-water flow along the relatively impermeable flow top. The composition of both the conglomerate clasts and the sandstone suggests a source exotic to either the nearby ancient Cascade Range or to the Columbia Plateau itself. Quartzite cobbles and pebbles in the conglomerate are less abundant than are basalt clasts. However, quartzite and mica are major constituents of the underlying sandstone. These sediments are present at the same stratigraphic horizon as the Rattlesnake Ridge Member of the Ellensburg Formation.

The Elephant Mountain Member, in intracanyon exposure, has a prismatic to blocky basal colonnade 1 to 2 m thick and an entablature approximately 10 to 15 m thick. It is an aphyric flow that appears to be relatively fine grained in hand sample. Geochemically, it is distinguishable from the underlying Pomona Member on the basis of its higher total Fe, TiO_2, K_2O, and P_2O_5 and lower CaO and Al_2O_3 contents (Table 1). It can be distinguished from the underlying Frenchman Springs Member on the basis of its lower P_2O_5 concentration. Samples from this flow yielded reversed natural remanent polarity, as in the Columbia Plateau. Two flows of this type have been reported in the Pasco basin area. The upper flow was described as having platy columns and normal natural remanence, whereas the lower flow is described as having altered entablature with vesicle sheets and reversed natural remanent magnetic polarity (Myers and others, 1979). The Elephant Mountain Member is transitional when subjected to alternating field demagnetization in the laboratory (Swanson and others, 1979b). However, this has not been confirmed for the intracanyon samples at the Klickitat River. The single Elephant Mountain flow in the Klickitat River area is tentatively correlated here with the lower of the two flows described in the Pasco basin on the basis of paleomagnetism.

The presence of fluvial deposits between the Pomona and Elephant Mountain Members clearly indicates that the river that occupied the Pomona paleodrainage re-occupied the same channel after the eruption. This conclusion is supported by the observation that the Pomona Member flow top is located well below the minimum elevation of the paleocanyon rim, indicating that the flow underfilled the canyon. Preservation of the top of the Pomona Member flow clearly reflects the establishment of a fluvial depositional environment. Whether this depositional environment persisted for the entire duration of the 1.5-m.y. hiatus between eruptions is not clear, based upon present evidence. The Elephant Mountain Member rests conformably on the interstratified sediment at the north side of the paleocanyon and is in contact with flows of the Frenchman Springs Member across a steep buttress unconformity. However, the picture is less clear on the south side of the canyon where the contact is poorly exposed.

A possibility that cannot be ruled out is that, after an initial period of deposition, the river incised its channel along the south side of the canyon to some unknown depth. The conformable exposures could, under this interpretation, be viewed as overflow adjacent to the main channel. The relative thinness of the Elephant Mountain member suggests that this interpretation may be correct. Absence of this member to the west could mean that (1) the observed thinness reflects lap-out resulting from volume attenuation due to cessation of eruption, or (2) that the channels filled by the two Saddle Mountains members separate at a point downstream. The former interpretation would assume that no adjacent channel was present and that deposition indeed characterized the 1.5-m.y. interim between eruptive episodes. The latter interpretation implies that erosional downcutting occurred after a period of deposition and that remnants of the Elephant Mountain intracanyon flow may yet be discovered within the Cascade Range.

In the Klickitat River area, the Saddle Mountains paleocanyon appears to be filled above the Elephant

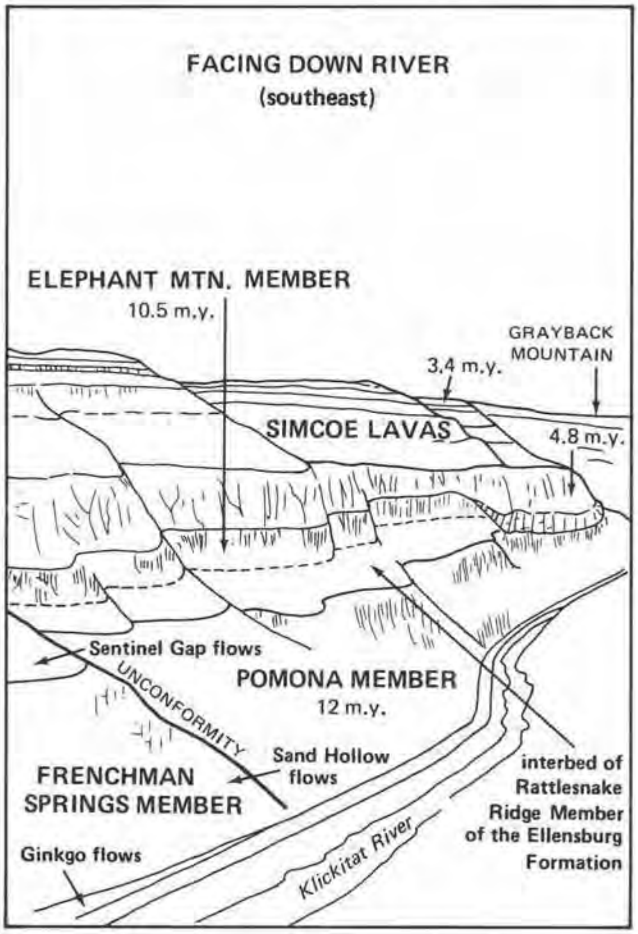

Figure 12.–Contact relations of Pomona Member intracanyon flow in the Klickitat River area (location shown in Fig. 10). Sand Hollow and Gingko are informally named flows (Beeson and others, 1985). The labeled "interbed" is part of the Rattlesnake Ridge Member of the Ellensburg Formation. View to the southeast.

Mountain Member by flows of the Simcoe volcanic field. Figure 12 shows the apparent contact relations between the Columbia River Group flows and the younger overlying flows. The lowest and oldest of these flows is an olivine basalt flow with jointing very similar to that of the underlying Saddle Mountains Member intracanyon flows. The paleomagnetic polarity of this flow, like that of the Pomona Member, is reversed. The chemistry and lithology, however, are distinctive; the flow is a high-alumina basalt (Kuno, 1966) containing abundant olivine. The chemistry of this oldest Simcoe Lava flow closely resembles that of the basaltic suite overlying it. A K-Ar age obtained from the basal colonnade of this flow yielded 4.8 ± 0.22 m.y., indicating an approximate 5.7-m.y. period between the eruption of the Elephant Mountain Member and this earliest Simcoe lava flow. Contact relations elsewhere suggest that the flow did not fill the revitalized post-Elephant Mountain paleodrainage but, instead, occupied a canyon that crosses the Saddle Mountains canyon. This younger canyon, of an ancestral Klickitat River, crosses the Horse Heaven Hills uplift at Grayback Mountain and appears to be antecedent to that ridge.

The paleocanyon cut in Saddle Mountains time does not cross the Horse Heaven Hills uplift at the same point as the younger Simcoe paleocanyon. Field evidence suggests that the cross-over point is at least 10 km to the west. The Saddle Mountains paleocanyon probably predates the major part of the ridge deformation; no angular discordance is evident between flows within the canyon and flows forming the canyon walls. The same is also true of the other localities discussed in this paper. However, none of these localities lies close enough to the most intensely deformed part of the uplift to be certain that relatively localized uplift was not under way in pre-Saddle Mountains time. Limited uplift is suggested by the approximate parallelism between the position of the canyon and folds of the Horse Heaven Hills trend and by localized channel overflow in some areas and general confinement elsewhere.

The occurrence of the early Pliocene and late Miocene paleocanyons together in the Klickitat River area is thus considered to be a fortuitous intersection of paleodrainages of different ages and different trends.

MOSIER SYNCLINE AND BINGEN ANTICLINE

Conformable exposures of Pomona Member have long been known to be present within the Mosier syncline, a northeast-trending low between the Horse Heaven Hills uplift on the north and the Columbia Hills uplift on the south (Schmincke, 1964, 1967; Rietman, 1966). This Saddle Mountains flow reached the Mosier syncline via a paleodrainage that extended into the area from the north across the Bingen anticline uplift (Fig. 10). The area at the crest of the Bingen anticline was part of a topographic low that had greater northward extent than the Mosier syncline.

To clarify the picture, we will consider exposures of the Pomona Member that occur at the crest of the Bingen anticline (Fig. 10), immediately north of the Mosier syncline. This anticline forms part of an anticlinorium that includes the Horse Heaven Hills uplift along its north edge. Conformable exposures of the Pomona Member extend up the south flank of this fold from the adjacent syncline. Exposures at the crest, however, are clearly unconformable. A paleocanyon more than 50 m deep and of unknown width overlies boulder and cobble conglomerate of dominantly basaltic composition. The latter is a clast-supported deposit with a sandstone matrix and has a thickness of at least 5 m. Rounded boulders as much as 1 m in diameter are present, and most are of Wanapum Basalt (Fig. 13). Mixed lithologies including quartzite are present in this deposit in some areas.

The position of this Pomona Member paleodrainage is clearly outside of what is commonly referred to as the Mosier syncline but was apparently located within the low area filled by the lava flow. This strongly suggests that the pre-Pomona lowland was more extensive than the present syncline and that the positions of the two low areas may coincide only in part. Exposures of the Pomona Member on the south side of the Mosier syncline also suggest greater extent for the pre-Pomona Member lowland. Anderson (Swanson and others, 1981) has demonstrated that the Pomona Member is present at least 8 km to the south

Figure 13.–Cobble conglomerate that underlies the Pomona Member in the Bingen anticline area (Fig. 10). Includes abundant clasts of Wanapum Basalt and rare quartzite.

of the syncline axis in the upper Mosier Creek area. Figure 10 shows the extent of the mapped Pomona Member in comparison to the position of presently observed folds.

Two additional points regarding the role of the Mosier syncline in localizing the Pomona Member are worthy of note. First, nowhere are outcrops known that demonstrate onlap of the member against older members. Flatirons are present along the limbs of adjacent folds, and some thinning is evident. However, clear evidence of a lap-out of Pomona Member against present anticlines or bounding faults is lacking. Second, pillow basalt and interstratified sediment beneath the Pomona Member is notably lacking in the Mosier syncline. Thin interbeds are present in some parts of the syncline but are local in extent.

All the above observations are consistent with the interpretation that a relatively well-drained low area formerly existed in the approximate position of the present Mosier syncline, but that it was significantly more extensive than the later fold. The older low was a very shallow feature that collected very little sediment; discordancy between flows flooring this shallow warp and the Saddle Mountains flow that later filled it is so slight that it is unmeasurable in all presently known exposures. The low was crossed near its presumed northern margin by a channel that deepened significantly to the north and to the west where it entered and exited the low, respectively. The floor of the pre-Saddle Mountains low was apparently an upland surface adjacent to the paleodrainage that was covered by the Pomona Member when the channel overflowed. The conformable exposures are therefore interpreted to be spill-over deposits fed by the intracanyon flow. It is possible that similar spill-over deposits once extended well to the north of the paleochannel but have been eroded away in this uplifted area.

VIENTO PARK AREA OF THE COLUMBIA RIVER GORGE

The Pomona Member occurs in spectacular exposures at a promontory called Mitchell Point in the Columbia River Gorge west of the town of Hood River, Oregon (Fig. 10), and extends westward along the steep walls of the gorge above Viento State Park to near Shellrock Mountain. This area is directly north of Mount Hood and is therefore approximately at the central axis of the present Cascade Range.

The base of the Saddle Mountains paleocanyon is best exposed in the Viento Park area, where the flow fills a channel at least 100 m deep with a minimum width of at least 1.5 km (Anderson, 1980). Figure 11 shows the exposed channel filled with bedded gravel and the overlying Pomona Member. The intracanyon flow is approximately 65 m thick in this area and appears to lap out against older Columbia River Basalt Group flows to the south. The sedimentary fill is a well-cemented basaltic cobble conglomerate with a sandstone matrix (Sceva, 1966; Anderson, 1980). The conglomerate contains rare quartzite pebbles, scattered wood fragments, and thin layers of interstratified bedded micaceous sandstone. Angular blocky boulders of basalt derived from the Frenchman Springs Member are present along the channel wall and are considered to be the product of mass wasting during alluviation. The blocks reflect the steepness of the canyon walls during deposition of the gravel. The conglomerate is at least 70 m thick, indicating a significant alluvial event following a period of downcutting.

Palagonitic sandstone is locally present at the base of the Pomona Member and overlying the conglomerate described above. This deposit is 1.5 to 3.0 m thick and underlies the unpillowed base of the lava flow. Sediment of this type probably indicates the presence of minor quantities of water in scattered pools along the floor of the paleocanyon. Greater quantities of water would have caused the generation of pillow basalt that would now be exposed at the base of the flow. The mechanism involved in the formation of this palagonitic material could be analogous to the Priest Rapids scenario, only in miniature. Exposures of the base of the lava flow, west of Mitchell Point and east of Perham Creek near the abandoned Columbia Gorge highway, lack palagonitic sandstone deposits. Instead, sediment there consists of shaly siltstone overlying pebble and cobble conglomerate. Pillow basalt is absent at the base of the intracanyon flow.

The Mitchell Point exposures and those to the west suggest relatively dry conditions prior to the arrival of the lava flow. It could be argued that the described exposures are simply spill-over deposits adjacent to a channel cut into the underlying conglomerate. The gravel deposits in this scenario would then be terraces along the sides of the paleodrainage, a possibility that we do not rule out here. However, the absence of soil at the top of the alluvial fill, particularly where fine-grained deposits are present, and the presence of palagonitic sandstone together suggest that the surface covered by the lava flow was being actively alluviated. Deposits at the Bridal Veil, Oregon, area and across the Columbia River near Cape Horn in Washington display a similar lack of both pillow basalt and palagonitic sandstone (Tolan, 1982; Tolan and Beeson, 1984).

The apparent dryness of the Saddle Mountains paleodrainage is strikingly different from the Wanapum drainage. A possible explanation for this is based upon paleogeographic considerations. The data presented in this paper, together with other published data on the Pomona Member intracanyon flow (Swanson and Wright, 1976; Swanson and others, 1977, 1980; Camp, 1981), indicate that the paleodrainage filled by this flow extended entirely across the Columbia Plateau and had its headwaters in Idaho. The Pomona flow apparently temporarily beheaded the occupant stream where the flow entered the paleodrainage in the eastern Columbia Plateau, perhaps causing water to pond behind the resultant lava dam until the eruption ceased. Cut-off of the occupant stream would explain why there is little evidence of dynamic interaction with water within the paleocanyon downstream (Swanson and others, 1979b; Camp, 1981). One would anticipate under this interpretation that side streams would continue to flow and that local accumulations of pillow basalt would result where these streams were dammed and backfilled by the advancing lava flow. Pillow basalt occurs in association with the Pomona Member in non-intracanyon exposures in the Columbia Plateau (Schmincke, 1964, 1967). These deposits, together with underlying lacustrine and paludal sediments, indicate that the Pomona Member advanced into shallow lakes and ponds in some areas, a picture similar to that for the Priest Rapids Member. The difference between the two may lie in the throughgoing nature of the paleodrainage filled by the Pomona Member. This flow was almost invariably backfilling lowlands adjacent to a paleodrainage. The Priest Rapids flow, in contrast, crossed most of the Columbia Plateau, a poorly drained surface, as a sheet flood. It entered a well-defined canyon only on the western perimeter of the plateau. The Pomona Member was obliged to fill a shallow canyon or channel before spreading laterally out of it as spill-over flow. It covered a generally better drained surface than did the Priest Rapids and was constrained, to some extent, to follow the paleodrainage across that surface.

IMPLICATIONS FOR UPLIFT AND SUBSIDENCE

The two intracanyon basalt units discussed in this paper present an opportunity to examine the general uplift of the Cascades in the area north of Mount Hood, Oregon, after the eruption of these flows approximately 14 and 12 m.y. ago. The top of each of these flows theoretically provides a datum approximately parallel to the original grade of an antecedent stream channel that extended at two different times across the entire width of the evolving late Miocene Cascade Range. The discussion here is limited to upper flow surfaces due to the greater number of available data points for estimating post-eruptive elevations.

The Pomona Member encountered local highs and lows in the drainage it followed. This is demonstrated by overflow in low areas and underfilling at high areas. The Priest Rapids Member, in contrast, overflowed its canyon in each of the areas discussed. The Pomona Member overflowed its canyon in the Bingen anticline area but was confined at various depths within the canyon in all other areas. The highs and lows in the southwest Columbia Plateau did not coincide precisely with present structural highs and lows. The Pomona Member, as mentioned previously, occupied a low in the Bingen anticline area at what is now the crest of the fold. The low areas occupied by the Priest Rapids Member, on the other hand, appear to have coincided to some extent with structural lows coincident with mapped synclines along trends parallel to the Yakima Ridges.

In the Bull Run area, the Priest Rapids Member appears to have followed the Bull Run syncline, a structure that has been undergoing active minor subsidence since at least late Grande Ronde Basalt time (Vogt, 1981; Beeson and others, 1982). In The Dalles area, the unit appears to have been broadly confined to The Dalles-Umatilla syncline, although at least one flow, as mapped by Anderson (Swanson and others, 1981), extended across the Columbia Hills trend to near the Horse Heaven Hills trend on the north. No thinning of the Priest Rapids Member across the ridge has yet been demonstrated. However, the uppermost of two flows, the Lolo flow (Wright and others, 1973), appears to be localized within the Mosier syncline and the eastern part of the Swale Creek syncline, north of the Columbia Hills anticline; south of the anticline, this flow appears to have been primarily localized within The Dalles-Umatilla syncline. This suggests that the Columbia Hills anticline was rising, at least locally, in the post-Rosalia pre-Lolo period and that the amplitude of the uplift was at least as great as the thickness of the latter lava flow. The Lolo flow in the The Dalles and Mosier syncline areas is usually less than 15 m thick.

The area of the present crest of the Bingen anticline, where channels of both the Pomona and Priest Rapids Members are located, was originally the northern edge of a general low that became progressively narrower through time as the anticline rose and gained amplitude. Folding appears to have occurred at a rate that permitted these drainages to persist. The observed channel incision at the crest of the Bingen anticline may represent the initial stages of erosion due to antecedence.

Assessing the magnitude of relative uplift and subsidence within the southwestern Columbia Plateau since the eruption of the Pomona Member requires that the post-eruptive elevation be estimated and then compared to the present elevations determined from mapping. It is difficult to estimate the original post-eruptive elevations because of extensive deformation of the western Columbia Plateau and Cascade Range that has produced net relief ranging from well below sea level in the deepest basins of the Columbia Plateau to substantially above sea level within the anticlinal Yakima Fold Belt. None of the surface in this area can be considered a reliable datum for comparison purposes. The eastern plateau, however, is much less deformed. The gentle tilt of parts of southeastern Washington has been interpreted as roughly equivalent to a paleoslope at the top of the Columbia River Basalt Group rocks (Swanson and others, 1979b). This surface suggests a paleogradient of approximately 1 m/km (S. P. Reidel, personal commun.).

The eastern Washington paleosurface, when projected toward the west at 1 m/km, intersects the Oregon coast near present sea level. Considering the assumptions and distances involved, this value is probably a reasonable one. For example, increasing or decreasing the gradient by 0.5 m/km would result in a range at the coast of plus or minus several hundred meters with respect to modern sea level. Because sea level could have been lower but was probably not much higher during the Miocene than at present, the 1 m/km value is considered to be a reasonable minimum gradient to which uplift and subsidence can be compared.

Figure 14 is a profile along the top of the intracanyon Pomona Member from the western edge of the area covered by sheet floods in the central Columbia

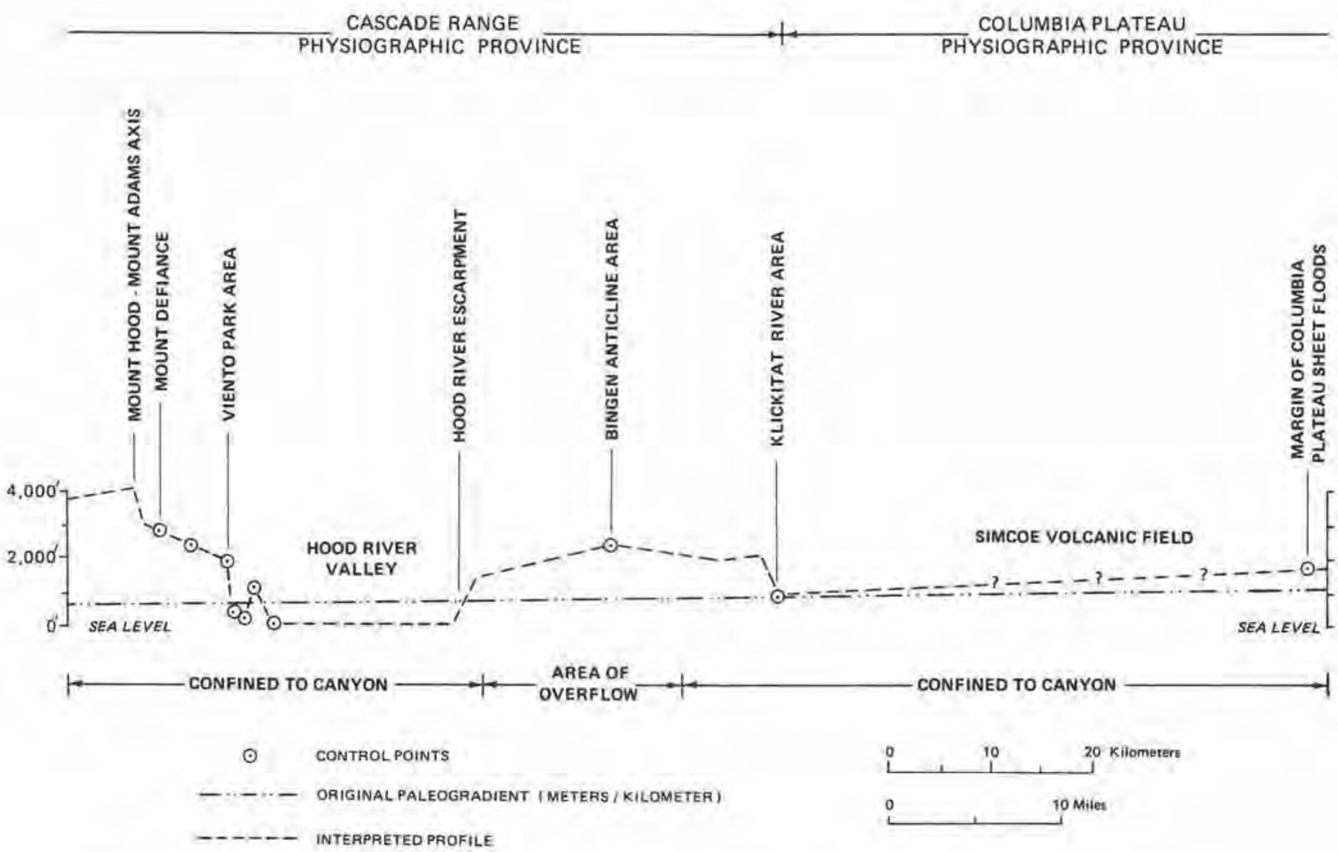

Figure 14.–Interpreted profile (dashed line) at the top of the Pomona Member intracanyon flow compared to an estimated original paleogradient (dotted dashed line) at the top of the Columbia River Basalt Group.

Plateau on the east to the Shellrock Mountain area on the west. This profile, therefore, traces the paleodrainage from within the Columbia Plateau to the center of the Cascade Range north of Mount Hood, Oregon. The hypothetical paleogradient of 1 m/km derived from the eastern plateau is superimposed on this plot. Comparing the two reveals that there are points of maximum uplift (1) near Mount Defiance, (2) in the Bingen anticline area, and (3) west of the Klickitat River, from west to east. A fourth area of uplift may be present under the Simcoe volcanic field but is poorly constrained because of uncertainty regarding the route followed by the paleodrainage beneath the field. These points of uplift occur at 10- to 40-km intervals along the paleodrainage and in all places are in part or wholly the result of oblique crossing of structures along the Horse Heaven Hills trend. It should be stressed that the profile is not transverse to structure. Minimums at or below grade occur at the Hood River Valley and Klickitat River. In the Hood River valley, subsidence is considered to be the result of faulting along the Hood River escarpment, a "Cascadian" structure. Significantly, these maximums and minimums are of approximately equal magnitude and are not in excess of uplift or subsidence observed elsewhere in the Yakima fold belt. Uplift associated with later arching of the Cascade Range, is not discernable on the basis of the paleodrainage data.

The top of the Pomona Member intracanyon flow immediately east of the Mount Hood-Mount Adams axis lies near the top of the Frenchman Springs Member, the youngest of the underlying Columbia River Basalt units. Canyon relief relative to the Pomona Member flow top was therefore minimal. Canyon relief measured in the same manner to the west of the Mount Hood-Mount Adams axis was considerably greater. Available data indicate that the intracanyon flow entered a much deeper canyon in the western part of the Cascade Range, one with a depth of more than 244 m and a width of 2.4 km, at Bridal Veil, Oregon (Tolan, 1982; Tolan and Beeson, 1984). The top of intracanyon flow in this area was reported by Tolan (1982) to be more than 200 m below the canyon rims. The increased topographic relief along the paleodrainage appears to occur abruptly near Shellrock Mountain as the result of faulting.

Stratigraphic separation of Columbia River Basalt Group units older than the Pomona Member is estimated to be 400 m (up on the west) near Shellrock Mountain across a fault located along the northwest-trending Wind River alignment. If all the paleocanyon deepening observed to the west is assignable to this fault, then the increased depth is quantitatively equal to the fraction of fault separation that occurred prior to the eruption of the Pomona Member 12 m.y. ago. The 200-m depth to the top of the Pomona Member observed by Tolan (1982), for example, would amount to half of the 400 m of fault separation. The precise depth of the canyon at Shellrock Mountain has yet to be determined.

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EARLY AND MIDDLE CENOZOIC STRATIGRAPHY OF THE MOUNT RAINIER–TIETON RIVER AREA, SOUTHERN WASHINGTON CASCADES

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ABSTRACT

Field observations and isotopic dating of the Eocene to early Miocene stratigraphic sequence of the Mount Rainier–Tieton River area provide new insights into the paleogeographic, magmatic, and structural evolution of the southern Washington Cascades. Early and middle Eocene continental sandstones, basalts, and rhyolites unconformably overlie deformed Jurassic to Cretaceous sedimentary, volcanic, and crystalline rocks that comprise the Tieton inlier. The sequence of Eocene rocks west of the inlier is much thicker than that exposed to the east. Isotopic ages (55-42 million years) and lithology of the Eocene units indicate broad correlation with the early Eocene Swauk Formation and the middle Eocene Naches Formation to the north and the Puget Group to the west.

A middle Cenozoic volcanic sequence consisting of two major units overlies the Eocene strata. The regionally extensive basal unit, the Ohanapecosh Formation of Oligocene age, consists of well-bedded andesitic to dacitic pyroclastic and volcaniclastic sedimentary rocks and much smaller volumes of andesite and rhyolite flows and silicic ignimbrites. The Ohanapecosh sequence is interpreted here as deposited subaerially and in shallow lakes on an aggradational surface of its own construction. In the type area, a thick (3,000 meters) Ohanapecosh section dated at 36 to 28 million years lies conformably on the Eocene sequence west of the Tieton inlier, whereas a thin sequence of middle and upper Ohanapecosh disconformably overlies the Eocene strata east of the inlier. The absence of Paleocene strata and the marked thinning of both the Eocene and Ohanapecosh strata across the inlier indicate that it acted as an intermittent structural high during early Cenozoic time. In the late Oligocene the Ohanapecosh underwent weak folding followed by uplift and erosion.

The upper unit of the middle Cenozoic volcanic sequence consists of late Oligocene and early Miocene silicic ignimbrites, andesitic lavas, and associated pyroclastic rocks that overlie the Ohanapecosh Formation unconformably. The ignimbrites occur both below and interbedded with the andesites. They comprise a number of large, lithologically distinct, ash flows derived from several different source vents and range in age from about 27 to 22 million years. A caldera in the Mount Aix area is the probable source of several of these ignimbrites. The basal ignimbrites of this sequence have been traditionally correlated with the Stevens Ridge Formation. Our work shows, however, that the basal silicic units in the study area differ widely in age and lithology and that the type Stevens Ridge Formation is younger than the type Fifes Peak Formation. The andesites are less well dated, but range in age from 27 to 20 million years. The andesites form a series of large composite cones and intervening lava and volcaniclastic sequences. The Fifes Peak Formation was mildly deformed and eroded prior to accumulation of the overlying flows of the middle Miocene Columbia River Basalt Group east of the present Cascade crest.

The Eocene strata of the Tieton area are characterized by the association of a bimodal, basalt-rhyolite, suite of volcanic rocks interbedded with nonvolcanic continental sandstones. Rapid lateral changes in thickness and facies in this sequence, as in the correlative Puget Group and Naches Formation, appear to reflect the faulting, differential uplift and subsidence, and local volcanism that characterize the Eocene of central and northern Washington. The Eocene tectonic regime ended by 37 million years ago when magmatic activity in Washington State became restricted to the narrow north-south Cascade magmatic arc defined by the distribution of the Ohanapecosh-Fifes Peak sequence and coeval shallow plutons.

INTRODUCTION

A record of magmatic activity that began in early Eocene time and persists to the present is preserved in the volcanic and plutonic rocks of the Cascade Range of Washington State. Much of this record is represented in a west-east transect across the Cascades from Mount Rainier through White Pass to the upper Tieton River (Fig. 1). Detailed geologic mapping has been done in a number of key areas (for example, Fiske and others, 1963; Swanson, 1964, 1978; Ellingson, 1959, 1972; Schreiber, 1981; Clayton, 1983), and the broad stratigraphic relations are now relatively well known.

Cenozoic rocks in the area can be assigned to four major supracrustal sequences shown diagrammatically below. The oldest Cenozoic sequence consists of sparsely exposed Eocene sedimentary and volcanic rocks that are broadly correlative with units elsewhere in western Washington. The Eocene beds are overlain, conformably and disconformably in different areas, by middle Cenozoic rocks of the Cascade magmatic arc. This mid-Cenozoic sequence comprises two subdivisions: the basal unit, the Ohanapecosh Formation, which consists of intermediate volcaniclastic sediments of early and middle Oligocene age; and the upper unit, the Fifes Peak Formation of latest Oligocene and early Miocene age, which consists of andesitic lavas and fragmental rocks underlain by and interbedded with silicic ash flows. The third sequence is distal Grande Ronde flows of the Columbia River Basalt Group, which are of middle Miocene age and unconformably overlap the older units on the east flank of the Cascade Range (Swanson, 1967). The fourth sequence, lavas and pyroclastic rocks of Pliocene and Quaternary age, unconformably overlies the older rocks and is most extensive in the Mount Rainier and White Pass areas (Fiske and others, 1963; Clayton, 1983).

IV. LAVAS AND PYROCLASTIC ROCKS
(PLIOCENE AND QUATERNARY)

III. COLUMBIA RIVER BASALT GROUP
(MIDDLE MIOCENE)

II. FIFES PEAK FORMATION
(LOWER MIOCENE)
OHANAPECOSH FORMATION
(OLIGOCENE)

I. VOLCANIC AND SEDIMENTARY ROCKS
(EOCENE)

This paper addresses the problems of the stratigraphic relations, geochronology, and correlation of the two older stratigraphic sequences of the Mount Rainier–Tieton area. The stratigraphic record of this area is important for several reasons. First, it represents a relatively complete sequence of early Eocene through early Miocene age with only minor stratigraphic gaps. It is, thus, an extension into the middle Cenozoic of the stratigraphy described by Tabor and others (1984) for the central Cascade Range of Washington. Second, because several of the units extend outside the study area, relations here have regional application. Finally, interpretation of this record in the light of its geochronology is critical for understanding the Cenozoic structure, paleogeography, and patterns of magmatism in the Pacific Northwest and their relation to associated tectonic events. Although our discussion relies heavily on previous work, we have freely reinterpreted it in the light of our own field observations and new age determinations.

Apart from a review and re-evaluation of the stratigraphy of the Mount Rainier–Tieton area, our chief contribution is 32 isotopic age determinations, most of which are fission-track ages on zircon. Sample localities and (or) age data are shown on Figure 2, Table 1, and Appendix I. These dates establish a chronology for several major regional magmatic episodes and help resolve problems of definition and correlation of several widespread Cascade stratigraphic units.

The fission-track method has yielded a generally consistent chronology as confirmed by: (1) similar ages from the same unit at different localities; (2) the sequence of ages relative to stratigraphic position; and (3) cross-dating by other methods. Our age determinations show that the fission-track method is well suited for dating volcanic rocks that have been subjected to low-temperature alteration and that may yield erratic results from K-Ar dating. Fission-track dating was done by the external detector method on zircon by following the procedures described by Naeser (1976). Note that the standard error in the fission-track ages is a function of the number of tracks counted (McGee and others, 1985); it is about 10 percent of the age for most of our samples. The errors are given in Table 1; they are omitted in the text and figures for the sake of simplification, but clearly are important in evaluating the accuracy of the ages.

By processing 10 to 20 kg of material from four silicic tuff samples we were able to recover sufficient zircon (about 100 mg) to carry out U-Pb isotopic dating. For each sample, a minimum of one coarse-grained and one fine-grained fraction was analyzed. Methods and results are summarized in Appendix II. Three of the samples dated by U-Pb yielded rather precise ages (±0.1 to 0.5 m.y.). (Fission-track ages on these same samples are concordant with the U-Pb ages.)

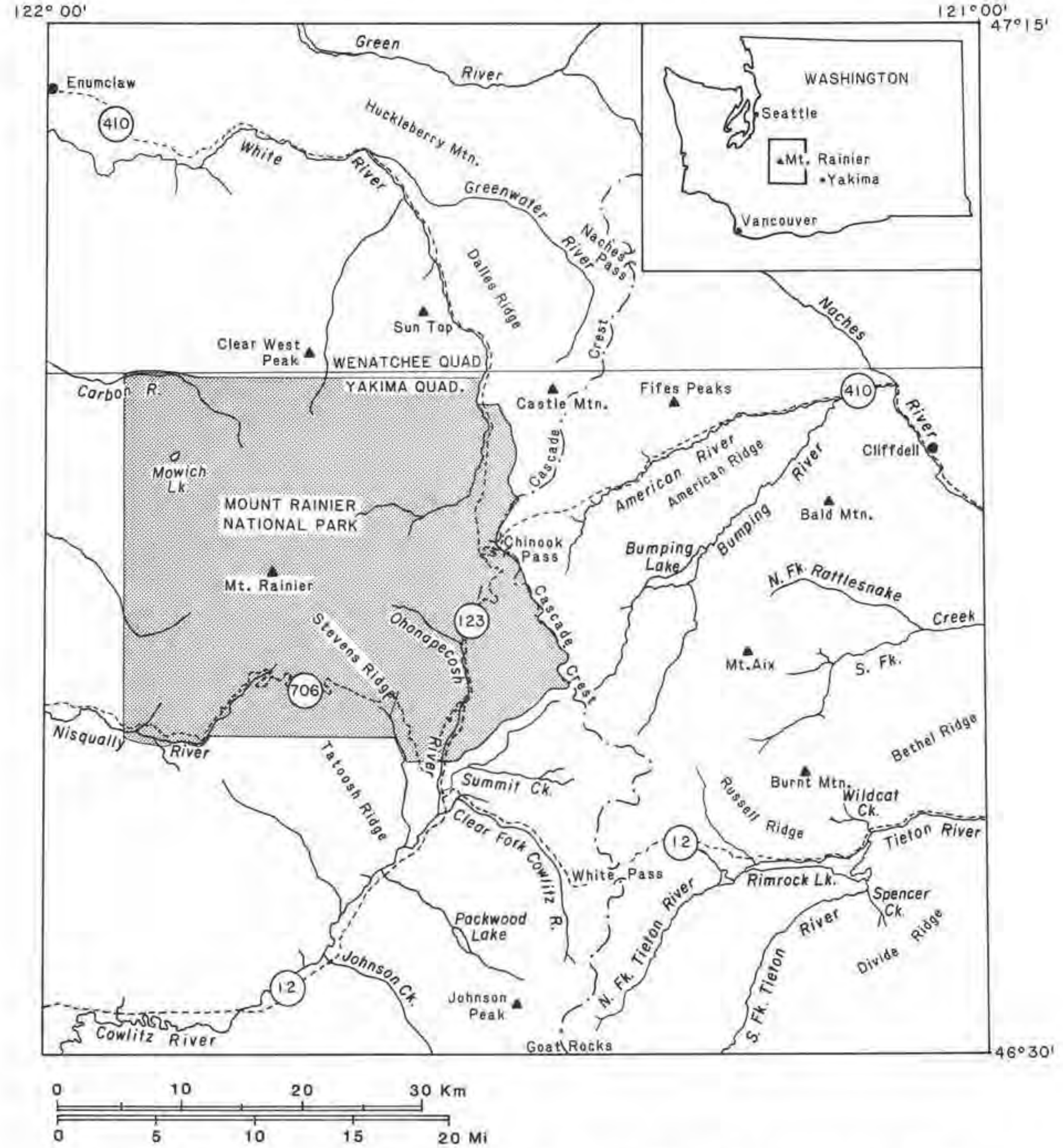

Figure 1.–Localities in the Mount Rainier–Tieton area referred to in the text. Numbered dashed lines are State Highway routes.

EOCENE OF THE TIETON AREA

Introduction

In the White Pass–upper Tieton River area, Eocene volcanic and sedimentary rocks crop out discontinuously along the margin of an inlier of Mesozoic basement, the Tieton inlier. The inlier is a domal uplift consisting of upper Jurassic to lower Cretaceous oceanic rocks (turbidites, ribbon cherts, and pillow basalts of the Russell Ranch Formation) and late Jurassic plutonic rocks (Indian Creek amphibolite [Ellingson, 1972] and Peninsula tonalite [Swanson, 1967]). Eocene strata are best exposed along the west, south, and east flanks of this basement structural high and dip radially off it (Fig. 2). This area is the southernmost exposure of pre-Tertiary basement rocks in the Washington Cascades.

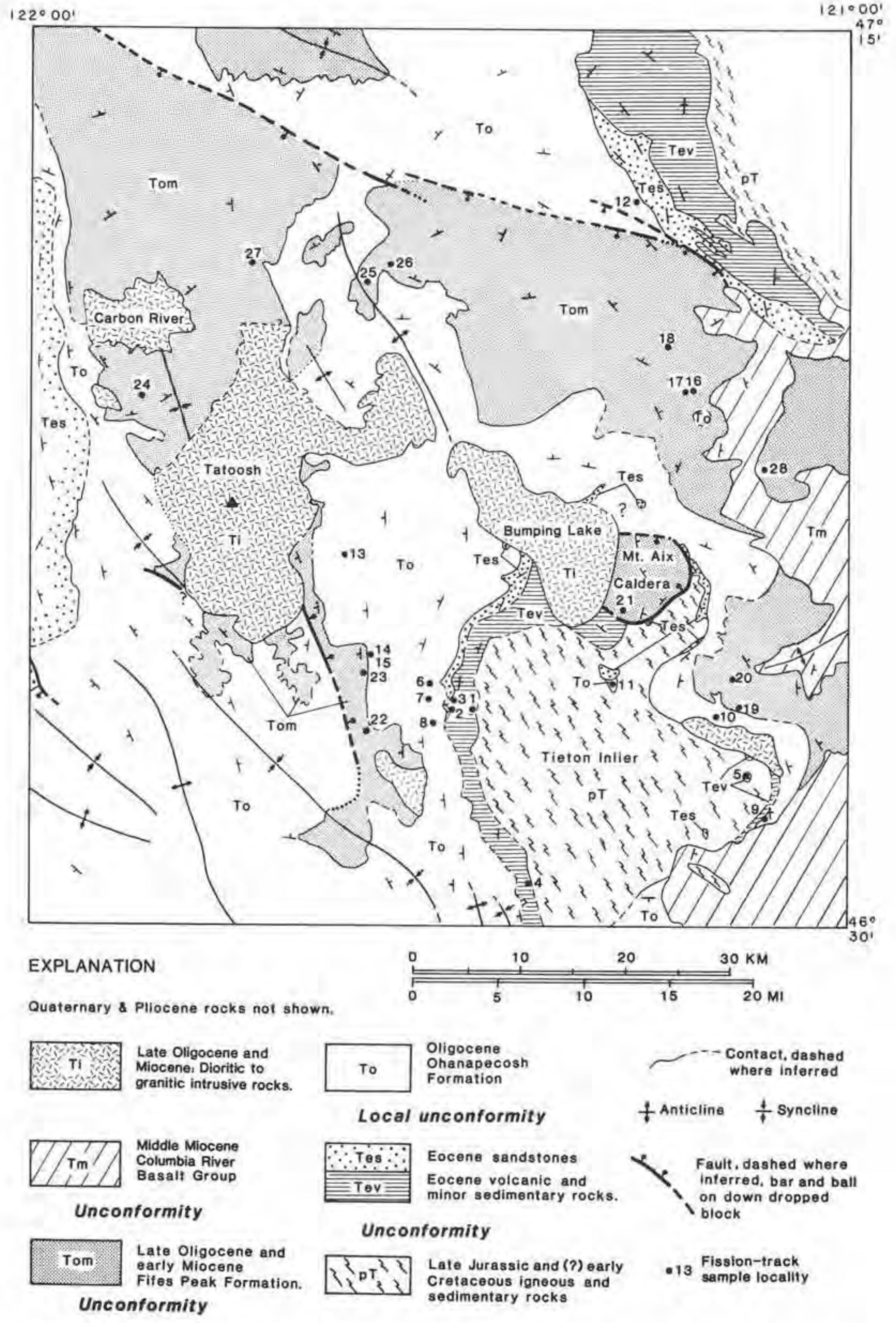

Figure 2.–Generalized geologic map and localities of the dated samples. Same area as Figure 1. Pliocene and Quaternary rocks not shown. Modified from: Abbott (1953); Ellingson (1959); Fiske and others (1963); Swanson (1967, 1978); Hammond (1980); Schreiber (1981); Clayton (1983); Frizzell and others (1984), and our mapping.

Table 1.–Zircon fission-track and U-Pb ages from Tertiary volcanic rocks in the Mount Rainier–Tieton River area, Washington. See also Appendix I and Figure 2 for further data.

Locality number	Sample	Comment	Grains counted	Spontaneous Tracks Density ($\times 10^6/cm^2$)	Spontaneous Tracks Number counted	Induced Tracks Density ($\times 10^6/cm^2$)	Induced Tracks Number counted	Thermal neutron dose ($\times 10^{15}/cm^2$)	Age (m.y. $\pm 2\sigma$)
				Eocene Units					
1	JV 310	U/Pb age							55 ± 3
2	JV 229	Summit	8	5.23	1340	3.65	935	1.03	44.0 ± 3.9
3	JV 232	Creek section	13	5.46	2910	3.64	1939	1.03	46.1 ± 2.7
4	TRG 2	North Fork Tieton River	6	8.17	1210	5.98	886	1.04	42.4 ± 3.7
5	JV 128	Spencer Cr.	6	7.38	1125	5.63	858	1.07	41.8 ± 3.8
				Ohanapecosh Formation					
6	JV 125	Lower	8	5.60	1201	4.67	1000	1.02	36.5 ± 3.1
7	JV 126	Ohanapecosh	6	7.90	888	6.27	785	1.08	36.4 ± 3.6
8	JV 94B		8	3.08	889	2.59	748	1.00	35.5 ± 3.6
9	CT 6	"Wildcat	6	2.85	944	2.85	944	1.13	33.7 ± 3.2
10	JV 139	Creek"	5	4.74	791	4.40	733	1.00	32.2 ± 3.3
11	JV 233	beds	8	4.55	1698	4.40	1642	1.03	31.8 ± 2.2
12	JV 96	Little Naches River	6	4.16	424	8.90	453	1.15	32.1 ± 4.3
13	JV 3	Upper	6	4.26	788	8.70	806	1.04	30.4 ± 3.0
14	JV 269	Ohanapecosh	6	5.08	705	5.38	747	1.01	28.5 ± 3.0
15	JV 124		6	6.46	711	7.37	811	1.08	28.3 ± 2.9
				Fifes Peak Formation					
16	JV 62	Bumping	6	2.80	233	6.02	251	1.00	27.7 ± 5.0
17	JV 241	River tuff	6	3.75	790	4.29	903	1.02	26.6 ± 3.6
18	JV 239	U/Pb							25.5 ± 0.1
	JV 239	Fifes Peak	6	2.96	661	3.67	809	1.07	26.2 ± 2.8
19	JV 226	Burnt Mtn. tuff	6	3.30	791	3.76	901	0.94	24.6 ± 2.4
20	JV 140	Tieton volcano	6	3.01	988	4.20	1379	1.09	23.3 ± 2.0
21	BP	Mount Aix	12	6.25	2893	7.30	3379	1.03	26.3 ± 1.3
22	JV 67	U/Pb Age							24.8 ± 0.3
	JV 67	Stevens	8	5.89	1176	7.11	1420	1.07	26.5 ± 2.1
23	JV 32	Ridge	6	3.63	639	8.70	735	1.05	27.3 ± 2.9
24	JV 92	Mowich Lake	6	3.96	495	11.42	714	1.17	24.2 ± 1.1
25	JV 165	U/Pb Age							$22.2^{+0.3}_{-0.5}$
	JV 165	Sun Top	6	6.56	1200	7.77	1421	0.97	24.5 ± 2.0
26	JV 36		6	2.89	548	7.40	704	1.03	24.0 ± 2.8
27	JV 163	Clear West Peak	6	2.71	695	3.88	995	0.96	20.0 ± 2.0
28	JV 354	Cliffdell	7	2.64	635	3.69	887	1.09	23.3 ± 2.4

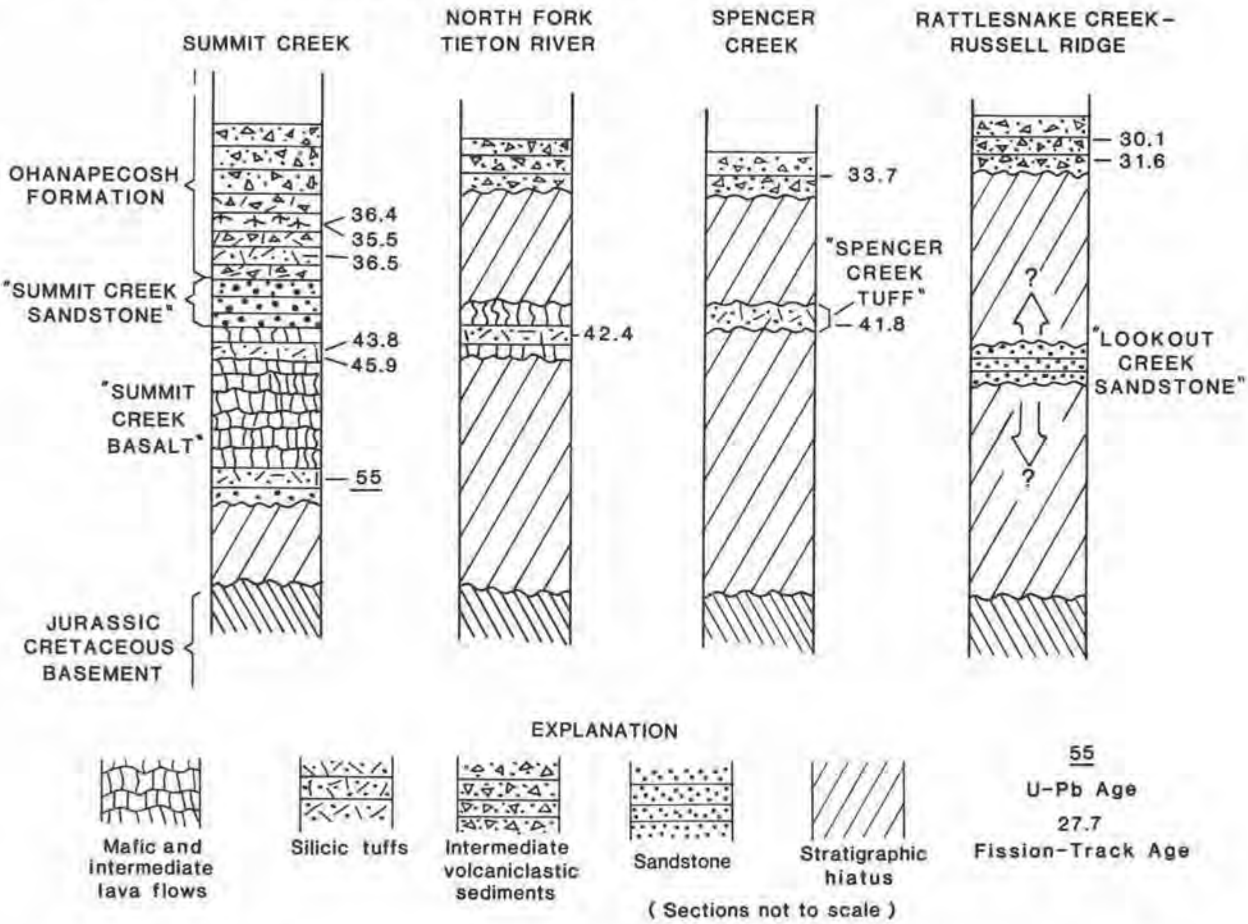

Figure 3.–Schematic stratigraphic column summarizing ages and correlation of Eocene strata in the Mount Rainier–Tieton River area. See text for discussion and Table 1 for age data. Unit names in quotation marks are informal.

Reconnaissance examination of several well exposed Eocene sections permits discussion of their stratigraphy. The dominant Eocene rocks are subaerial basalt flows, micaceous quartzose sandstones, and silicic pyroclastic rocks. Isotopic dating and lithology indicate that the Eocene rocks of the Tieton area are broadly correlative with the Puget Group to the west and with the Swauk and Naches Formations to the north. The stratigraphy of the Tieton Eocene beds is briefly summarized below for four relatively well studied sections: Summit Creek; North Fork of the Tieton River; Spencer Creek; and Rattlesnake Creek (Fig. 3). Tentative correlation is then made among these sections and the sections outside the area.

Summit Creek Section

In the canyon of Summit Creek west of the pre-Tertiary inlier, dense, sparsely phyric subaerial basalt flows dominate a steeply west-dipping 2,000-m-thick basal unit that unconformably overlies highly deformed Jurassic turbidites. Silicic tuff, argillite, and arkosic and volcaniclastic fluvial sediments occur as interbeds. Sandstones and pebble conglomerates interbedded in the lower part of this section contain clasts of graywacke, chert, vein quartz, and silicic plutonic rocks derived from the Mesozoic basement. Zircon from an ash-flow tuff (JV 310) interbedded with conglomerate near the base of the Summit Creek section gave a U-Pb model age of 55 ± 3 m.y., indicating an earliest Eocene age. South of Summit Creek, silicic ash flows on strike with the upper part of the basalt section yielded fission-track ages of about 44 and 46 m.y. Thus, the age of the lower part of the Summit Creek section lies in the range of about 55 to 45 m.y.

A unit of fluvial arenites about 750 m thick overlies the basalts of Summit Creek with apparent conformity. This unit was informally named the Summit Creek sandstone by Ellingson (1959, 1972). In Summit Creek canyon these beds have near vertical dips and are locally overturned as much as 20°. Massive and cross-bedded lithic and feldspathic arenites make up about 80 percent of this unit; the remainder is mudstones, coal seams (with leaf fossils), and silicic tuffs. The top of the unit grades conformably upward into the Ohanapecosh Formation through an interval about 100 m thick, in which conglomerates and quartzose sandstones derived from pre-Tertiary source rocks are intercalated with andesitic and dacitic fluvial volcaniclastic sediments. The Summit Creek sandstone is older than about 36 m.y., based on fission-track ages obtained from three zircon samples from rhyolites in the lower part of the Ohanapecosh Formation. Thus, its age is late middle to late Eocene. Sediments similar in appearance and stratigraphic position to the Summit Creek sandstone form a poorly exposed NNE-trending outcrop belt between the White Pass highway (State Route 12) and Bumping Lake, a distance of about 20 km. At Crag Mountain, 7 km NNE of Summit Creek, the sandstone is well exposed. Here, the unit dips gently westward and displays southwesterly paleocurrent directions. The sandstone thins rapidly south of Summit Creek and is absent in the Eocene section south of the White Pass highway. Sandstones stratigraphically below the Ohanapecosh reappear, however, in the Johnson Creek area about 20 km SSW along this trend (Winters, 1984).

Section on the North Fork of the Tieton River

Altered, highly vesicular subaerial basalt flows crop out in the area of the upper North Fork of the Tieton River at the southwestern margin of the Tieton inlier. The basalt flows form a section about 200 m thick that overlies Mesozoic turbidites and is overlain by about 250 m of interbedded mafic to intermediate lavas and silicic volcaniclastic rocks (Fig. 3). A thin quartz-phyric vitric tuff immediately above the basalt yielded a fission-track age of about 42 m.y. The Eocene section dips about 15° southwest and appears to be conformable with the overlying Ohanapecosh Formation.

Spencer Creek Section

Swanson (1964, 1978) mapped and described local exposures of volcanic and volcaniclastic sedimentary rocks that overlie the pre-Tertiary basement near Spencer Creek at the margin of the inlier south of Rimrock Lake. The most prominent member of this section is a welded ash-flow tuff, referred to informally below as the Spencer Creek tuff. We obtained a fission-track age of about 42 m.y. from this tuff. In addition, zircons from two samples of the volcaniclastic sedimentary rocks in the Spencer Creek area, the sandstone of Spencer Creek (Swanson, 1964, 1978), were dated. Zircon from a thin bed of clay-rich tuff that appears to overlie the welded tuff gave an age of about 34 m.y. and is thus age correlative with the Ohanapecosh Formation. Four zircons from a second clayey sand sample yielded an average age of about 137 m.y. This age indicates that these zircons are detrital and evidently were derived from Jurassic tonalites of the Tieton inlier, which have given a fission-track age of about 132 m.y. (Clayton, 1983).

Rattlesnake Creek Section

Schreiber (1981, 1982) mapped quartzose arenites near Lookout Creek, a tributary of Rattlesnake Creek, which unconformably overlie plutonic rocks, turbidites, and pillow lavas along the eastern margin of the Tieton inlier. This Tertiary unit, informally called the Lookout Creek sandstone by Schreiber, consists of thin-bedded, commonly cross-laminated, fluvial lithic-quartzose sandstones in a sequence that attains a thickness of as much as 200 m. Volcaniclastic sediments correlated below with the Ohanapecosh Formation disconformably overlie the sandstone, which dips gently to moderately east off the flank of the Tieton inlier.

Age and Correlation

Radiometric ages (summarized in Fig. 3) for the early Cenozoic strata that unconformably overlie the Jurassic and (?)Cretaceous basement rocks of the Tieton inlier range from about 55 to 42 m.y., or from early Eocene to late middle Eocene. Lithologic similarities, stratigraphic position, and radiometric ages support correlation of the Eocene beds of the White Pass–Tieton area with stratigraphic units exposed nearby to the west and north (Fig. 4). The principal correlatives west of the area are the continental sediments and intercalated volcanic rocks of the Puget Group (Fisher, 1961; Gard, 1968; Vine, 1969). Correlative units north of the area in the Cascades in central Washington have been described by Tabor and others (1984).

The thickest and most complete section of Eocene rocks is exposed in Summit Creek, west of the inlier. The 55-m.y. U-Pb model age determined near the base of this section suggests that these rocks are coeval with the lower parts of the Swauk and Chuckanut Formations (Tabor and others, 1984; Johnson, 1982, 1984) in the central and northwestern Washington Cascades respectively. Silicic tuffs interbedded in the upper part of the thick basalt section just south of Summit Creek have given fission-track ages of 46 and 44 m.y., thus bracketing the age of the basalt in the range of about

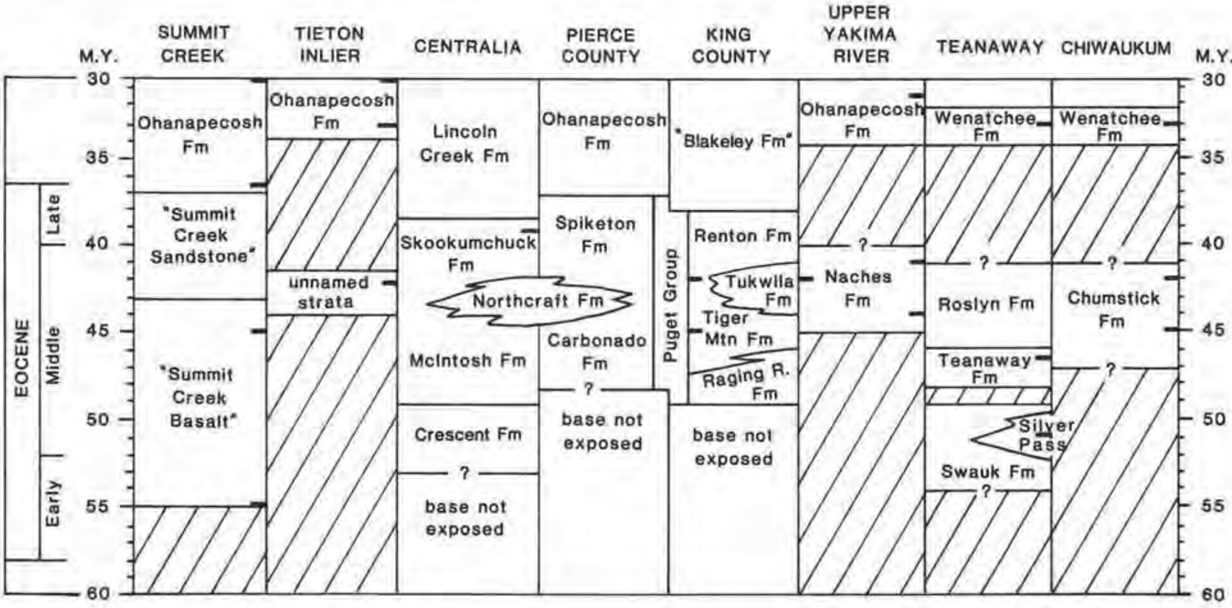

Figure 4.–Correlation of the Eocene rocks of the Mount Rainier–Tieton River area with those from nearby areas. The bars indicate radiometric age control. The chart is based on the following sources: Summit Creek and Tieton inlier, this paper; Centralia (Snavely and others, 1958); Pierce County (Gard, 1968); King County (Vine, 1962, 1969; Turner and others, 1983; and Frizzell and others, 1984); Upper Yakima River and Teanaway (Tabor and others, 1984; Frizzell and others, 1984); and Chiwaukum (Gresens and others, 1981).

55 to 45 m.y. The upper part of the basalt sequence may in part be age correlative with the middle Eocene Teanaway Formation (about 47 m.y.) exposed to the north in the central Washington Cascades (Tabor and others, 1984). This correlation is consistent with the observation that both the Summit Creek and Teanaway lavas are overlain by coal-bearing quartzose continental sandstones, those of the Summit Creek area and the Roslyn Formation, respectively. Alternatively, the upper basalts may correlate in part with the late early to early middle Eocene volcanic rocks of the Silver Pass area east of Snoqualmie Pass, with the volcanic and arkosic strata of the late middle Eocene Naches Formation to the north, or with middle Eocene volcanic units in the Puget Group to the west and northwest. The age of the Summit Creek sandstone, comprising the upper part of the Eocene sequence in the Summit Creek section, is bracketed by dates of about 45 m.y. for the underlying volcanic rocks and ages near 36 m.y. for the lower part of the conformably overlying Ohanapecosh Formation. The Summit Creek sandstone occupies the same stratigraphic position as the upper sandstone members of the Puget Group, the Spiketon Formation (Gard, 1968) west of Mount Rainier, the Renton Formation (Vine, 1969) of the King County coalfield area to the north, and apparently equivalent sandstones in the Johnson Creek area southwest of the Tieton inlier (Winters, 1984).

The lower Cenozoic strata exposed around the Tieton inlier display marked lateral changes in thickness and lithology. The Summit Creek section is a thick and relatively complete Eocene sequence. However, in most sections lower Eocene rocks are absent, and middle Eocene strata directly overlie the pre-Cenozoic basement. The 42-m.y. age from the thin basalt section of the upper North Fork of the Tieton River is essentially the same as that from the welded tuff at Spencer Creek. The analytical uncertainty in the fission-track ages permits the interpretation that these rocks are age equivalents of the lithologically similar upper part of the volcanic sequence in the Summit Creek section. These middle Eocene rocks closely resemble and are probably correlative with the coeval Naches Formation, an association of continental sediments interbedded with basalts, silicic ignimbrites and lavas, and andesites, which crops out extensively to the north (Tabor and others, 1984). Based on its lithology, the sandstone unit described by Schreiber (1981, 1982) in the Lookout Creek area of the Rattlesnake Creek drainage is probably Eocene, but no evidence is available for a more precise age assignment.

OHANAPECOSH FORMATION

Introduction

The Ohanapecosh Formation, defined by Fiske and others (1963) in their study of the geology of Mount Rainier National Park and its environs, is the lowest of three middle Tertiary units exposed in the park—the Ohanapecosh, the Stevens Ridge, and the Fifes Peak Formations. These three formations had previously been mapped as the Keechelus Formation (Smith and Calkins, 1906), a unit since abandoned (Waters, 1961). Well-bedded pumice- and lithic-rich tuffs and lapilli tuffs characterize the Ohanapecosh Formation. These include airfall pyroclastic, pyroclastic-flow, and water-laid deposits. Beds range from massive and unsorted to better sorted and crudely graded. Though pervasively altered, mostly in zeolite facies (Fiske and others, 1963; Hartman, 1973), the abundance of pumice and the preservation of relict clinopyroxene and intermediate plagioclase indicates a dacitic to andesitic composition. Local accumulations of lava, dominantly andesitic, are near-vent facies. Some of the larger lava accumulations represent volcanic centers from which the pyroclastic material may have been erupted. In contrast to the underlying Eocene strata that are characterized by quartzose sandstones, nonvolcanic detritus is not common in the Ohanapecosh stratigraphic section and younger strata of the Cascades. Fiske and others (1963) described an Ohanapecosh stratigraphic section approximately 3,000 m thick in exposures along Stevens Canyon Road (State Route 706) in Mount Rainier National Park and the White Pass Highway to the south. They interpreted the volcaniclastic sediments of the Ohanapecosh as debris flows and turbidites generated by subaqueous eruptions.

More recent work shows that the Ohanapecosh Formation is widely exposed in southern Washington between the Columbia River Gorge and Snoqualmie Pass (Ellingson, 1968; Wise, 1970; Fischer,1970; Hartman, 1973; Hammond, 1980; Frizzell and others, 1984). The Ohanapecosh represents the first volcanic expression of the Tertiary Cascade magmatic arc in southern Washington and marks the end of the Eocene tectonic regime in the Pacific Northwest (Vance, 1982). New data on the age, stratigraphy, depositional environment, and correlation of the Ohanapecosh Formation are presented below.

The Age of the Ohanapecosh Formation

The thick Ohanapecosh section studied by Fiske and others (1963) is a homoclinal sequence dipping between 50° and 30° to the west. The Ohanapecosh is overlain by silicic ash-flow tuffs of the Stevens Ridge unit. Fiske did not study the lower 450 m of the Ohanapecosh because it was covered by young basalt flows along State Route 12 west of White Pass. The basal part of the Ohanapecosh Formation is well exposed, however, in the canyon of Summit Creek and in the area between Summit Creek and the highway. We place the base of the Ohanapecosh at the first appearance of volcaniclastic sediments above the nonvolcanic Summit Creek sandstone. In the canyon of Summit Creek, the lower approximately 100 m of Ohanapecosh consists of thin-bedded, well-sorted, volcanic sandstones and siltstones with subordinate interbeds of nonvolcanic sandstone and conglomerate. Sedimentary structures indicate a continuation of the fluvial environment in which the Summit Creek sandstone was deposited. More typical Ohanapecosh volcaniclastic sediments with minor intercalations of silicic ash flows and lava flows overlie this gradational interval. A plagiophyric ash-flow tuff about 300 m above the base of the unit yielded a fission-track age of about 36 m.y. About 250 m higher in the sequence two laminated rhyolite flows (one of which is exposed near the base of Fiske's measured section on State Route 12 west of White Pass) also dated near 36 m.y.

We determined three fission-track ages from the upper Ohanapecosh in Mount Rainier National Park. The large rhyolite flow north of Stevens Ridge at Indian Bar yielded an age of about 30 m.y. Ages of about 28 m.y. were determined for two silicic ash flows exposed at Backbone Ridge on the Stevens Canyon Road (State Route 706) about 120 m stratigraphically below the basal Stevens Ridge ash-flow tuff. Thus, the Ohanapecosh Formation spans the age range of about 36 to 28 m.y., and probably somewhat more because neither the highest nor lowest beds were dated.

Bear Creek Beds

We determined a single fission-track age of about 32 m.y. on zircon from an ash flow from Bear Creek in the Little Naches River area north of the present study area. As noted by Frizzell and others (1984), this date and lithologic association suggest that the Bear Creek rocks are equivalent to the Ohanapecosh.

Wildcat Creek Beds

Swanson (1964, 1978) mapped a series of tuffs and volcaniclastic sediments north of Rimrock Lake in the Tieton area. This unit, here informally referred to as the Wildcat Creek beds, is about 700 m thick

(Bruce Lander, personal commun., 1984) and consists of well-bedded andesitic to dacitic volcaniclastic rocks. Typical of the unit are graded beds, which pass upward from a base of lithic- and pumice-rich lapilli tuff to ash tuff, which commonly bears accretionary lapilli. These rocks resemble the Ohanapecosh, but they tend to be somewhat finer grained and better bedded, and they lack interbedded lavas. At Wildcat Creek the unit contains vertebrate fossils of Oligocene age (Grant, 1941). Schreiber (1981, 1982) mapped the unit about 20 km farther north, in the Rattlesnake Creek area, where it overlies Eocene sandstones. Clayton (1983) described an erosional remnant of similar rocks which overlies Eocene quartzose sandstones at McNeil Peak on Russell Ridge (locality 11, Fig. 2) in the northern part of the Tieton inlier.

We determined a fission-track age of about 32 m.y. from a pumice flow in the middle of the Wildcat Creek section. Schreiber (1981, 1982) obtained a fission-track age of 30.1 ± 3.0 m.y. from Wildcat Creek beds in the Rattlesnake Creek area. We determined a fission-track age of about 34 m.y. on volcaniclastic sedimentary rocks mapped in the Spencer Creek area south of the Tieton River by Swanson (1964, 1978), thus supporting correlation with the Wildcat Creek beds. The lithology of the Wildcat Creek beds and the absence of lavas suggests that the unit is a distal facies of the Ohanapecosh Formation.

In view of the ages and relative thinness of the section at Wildcat Creek, rocks equivalent to the lower Ohanapecosh appear to be absent above the Tieton inlier. A fission-track age of about 32 m.y. determined on an erosional remnant of the Wildcat Creek rocks overlying thin Eocene sandstone beds at McNeil Peak on Russell Ridge on the crest of the Tieton inlier suggests the existence of a major unconformity. In the Spencer Creek area, the discrepancy between the age of the welded tuff (about 42 m.y.) and that of the overlying volcaniclastic sediments (about 34 m.y.) clearly implies a large stratigraphic hiatus.

The apparent absence of rocks of the lower Ohanapecosh Formation on the crest and east flank of the pre-Cenozoic inlier suggests either that the inlier was a topographic high in early Oligocene time and that the lower Ohanapecosh was not deposited, or that the lower Ohanapecosh was uplifted and erosionally removed prior to the middle Oligocene. The contrast between relatively thin, incomplete sections of Oligocene beds of the inlier and the thick and more complete section in the type area to the west parallels the relations in the Eocene strata. A further episode of major up-arching of the dome-like Tieton inlier postdates the Ohanapecosh, which is draped over the structure.

Depositional Environment

On the basis of the well-developed bedding, grading, the lateral continuity of individual units, the presence of sedimentary slump structures, the absence of fluvial sediments, and the similarity to volcanic sediments of clearly marine origin in Japan (later described by Fiske and Matsuda, 1964), Fiske (1963) concluded that most of the Ohanapecosh sediments were deposited as debris flows and turbidites shed from the flanks of subaqueous volcanoes. We believe, however, that a number of observations indicate that much of the unit was deposited subaerially: (1) Andesitic lava flows are locally abundant in the Ohanapecosh, but pillows have not been reported; (2) Silicic ash-flow tuffs, in part welded, present near both the base and top of the unit probably reflect subaerial eruptions; (3) Pumice is mixed with lithic lapilli in many of the tuffs, not sorted out as would be expected in subaqueous deposition; (4) Accretionary lapilli are abundant in some beds, implying subaerial formation; moreover, these fragile structures would not remain coherent during lengthy transport or settling through water; and (5) Abundant fragments of carbonized wood and locally preserved nonmarine vertebrate fossils indicate a terrestrial source, if not subaerial deposition.

These observations lead us to the conclusion that the Ohanapecosh sediments were deposited as airfall debris and mud flows, in part in a subaerial environment and in part in shallow water. The lateral continuity of some beds reflects deposition on an aggradational surface of low relief probably produced by voluminous and long-lived regional explosive volcanism.

Criteria for Recognition

The age range that we have established for the Ohanapecosh Formation is probably valid for the unit elsewhere in Washington outside the type area. However, rocks suitable for dating by either the K-Ar or the fission-track method appear to be lacking in many Ohanapecosh sections. In such places, correlation must be made on the basis of similar lithology and position in stratigraphic section. These criteria, however, must be applied with caution.

Well-bedded volcaniclastic sediments of intermediate composition are the most distinctive Ohanapecosh rocks in the Mount Rainier–White Pass–Tieton area. Similar beds are present locally in younger volcanic units of the region (for example, Hartman, 1973), but none are known to be as thick as 1,000 m. The Ohanapecosh Formation in southern Washington overlies distinctive units such as the Puget Group and the Naches Formation, which consist of interbedded quartz-rich continental sandstone and calc-alkaline volcanic rocks. The Ohanapecosh is overlain locally by distinctive silicic ignimbrites and related sediments and andesitic lavas and pyroclastic rocks correlative with the Fifes Peak Formation. Position in the stratigraphic section, where a complete section is present, combined with lithology will generally permit recognition of the unit. In some areas, however, the Ohanapecosh is dominated by nondiagnostic rocks, commonly andesitic lavas or silicic ignimbrites, and confident identification of the unit may not be possible without radiometric dating.

FIFES PEAK FORMATION

Introduction

The designation Fifes Peak Andesite was given by Warren (1940, 1941) to a series of dominantly andesitic rocks lying below the Columbia River basalts on the eastern flank of the southern Washington Cascades. Warren clearly distinguished the near-vent andesitic rocks of the Fifes Peak from the distal flood basalts of the overlying Columbia River Basalt Group. He also recognized the existence of a mild angular unconformity between these units and the presence of moderate relief on the buried Fifes Peak paleolandscape. Abbot (1953) included silicic tuffs in the Bumping River area in the Fifes Peak Andesite. Waters (1961) excluded basal silicic beds from the unit and renamed it the Fifes Peak Formation.

Volcanic sequences similar in lithology and stratigraphic position to the Fifes Peak Formation are widely distributed in the southern Washington Cascades. Fiske and others (1963) traced the Fifes Peak west discontinuously into Mount Rainier National Park. Fischer (1970) and Hartman (1973) mapped exposures just north of the park, and recent mapping carries it and its equivalents north to the Cedar Lake area (Frizzell and others, 1984). Swanson (1966, 1978) mapped two Fifes Peak eruptive centers east of the Tieton inlier. Harle (1974) studied andesitic lavas in the Council Bluff area west of Mount Adams which appear to be Fifes Peak correlatives, and reconnaissance mapping by Hammond (1980) indicates that Fifes Peak equivalents extend as far south as the Columbia River.

Fiske and others (1963) established a three-fold subdivision of the middle Cenozoic volcanic rocks in the area of Mount Rainier National Park. The three units are: a basal sequence of well-bedded volcaniclastic rocks, the Ohanapecosh Formation discussed above; a middle interval of silicic ash-flow tuffs and associated sediments, the Stevens Ridge Formation; and an upper unit of dominantly andesitic lavas and breccias, which they correlated with the Fifes Peak Formation. However, stratigraphic relations on the broader scale of the southern and central Washington Cascades and our isotopic dating indicate that regional relations are more complicated than those in the park. The main problems arise from uncertainties as to the definition, regional extent, and correlation of the Stevens Ridge Formation. Field relations, petrographic study, and isotopic dating indicate that a number of silicic pyroclastic units that have loosely been correlated with the Stevens Ridge Formation differ in stratigraphic position, lithology, or age from the type Stevens Ridge and, thus, cannot be its physical equivalents.

Silicic pyroclastic rocks similar to those of the Stevens Ridge are not unique to a single stratigraphic interval in the Washington Cascades, but were erupted from a number of separate vents between about 28 and 22 m.y. ago. These silicic rocks occur: (1) as interbeds in the upper Ohanapecosh Formation (as discussed above and by Frizzell and others, 1984); (2) between the Ohanapecosh and the overlying andesitic Fifes Peak rocks; (3) interbedded within the Fifes Peak sequence (Fischer, 1970; Frizzell and others, 1984); and (4) independently of these other associations in the north-central Cascades, where Ohanapecosh volcaniclastic rocks and andesitic Fifes Peak rocks are absent (Vance and Naeser, 1981). Some of these silicic sequences are characterized by quartz-phyric ash-flow tuffs very similar to those of the type Stevens Ridge. Thus, lithology alone is not a reliable basis for correlation of silicic pyroclastic units with the Stevens Ridge Formation. Conversely, however, significant differences in lithology or age of these silicic rocks rule out physical equivalence. Our isotopic dating indicates that silicic pyroclastic rocks of "Stevens Ridge" lithology lying stratigraphically above the Ohanapecosh Formation and below or near the base of the andesitic Fifes Peak section in the Mount Rainer–Tieton area range from about 27 to 22 m.y. in age, whereas the andesitic Fifes Peak ranges from about 27 to 20 m.y. The silicic pyroclastic rocks that have been loosely correlated with the Stevens Ridge are, thus, essentially the age equivalent of the Fifes Peak Formation. Furthermore, precise U-Pb ages of 25.5 ± 0.1 m.y. on a pumice lapilli tuff interbedded with the type Fifes Peak andesites in the American River area and 24.8 ± 0.3 m.y. on rocks physically continuous with the type Stevens Ridge in the Mount Rainier area indicate that the age relations are the reverse of those in the park. Similarly, the silicic tuffs that underlie the type Fifes Peak in the

Bumping River area are significantly older than the type Stevens Ridge.

Isotopic dating and field relations indicate the following history for Fifes Peak volcanism. The long-lived subsidence that marked accumulation of the Ohanapecosh Formation on a surface of low relief had ended by about 27 m.y. ago with folding, uplift, and erosion. Fifes Peak activity was initiated by violent, large-volume eruptions of silicic pyroclastic material from a number of separate vents, at least one of which was a major caldera. These basal silicic units, which range over at least 3 m.y. in age and consist of rhyodacitic ash-flow tuffs and related airfall and water-laid deposits, commonly occur as valley fillings unconformably overlying eroded Ohanapecosh. The development of moderate topographic relief on the Ohanapecosh would explain the absence of basal ash-flow tuffs in many sections and why the tuffs are of much more restricted areal extent than either the Ohanapecosh or the voluminous andesitic rocks that dominate the Fifes Peak. Local, generally low-angle discordance between the Ohanapecosh and overlying Fifes Peak units indicates that mild warping or folding accompanied the uplift (Fiske and others, 1963).

In each of the areas described below, deposition of the early silicic pyroclastic rocks was followed by dominantly andesitic activity from a series of scattered volcanoes. Coalescence and overlapping of these volcanoes produced a regionally extensive unit. A number of andesitic eruptive centers of Fifes Peak age have been identified. Swanson (1966) described two andesitic stratovolcanoes in the Tieton River area. At least two other Fifes Peak volcanoes are present in the American River–Naches River area; one is the type area at Fifes Peaks, the other is at Cliffdell (Carkin, 1985). Elsewhere, the unit consists of monotonous andesitic and basaltic andesite lava sequences as much as 1,000 m thick. We agree with Waters (1961) that these are probably shield volcanoes. Large composite cones in varying stages of growth and erosion rose above lower shield volcanoes and intervening plains of lava and pyroclastic rocks. Silicic ash-flow tuffs related to calderas, to the stratovolcanoes, and possibly to the venting of shallow plutons, together with lahars and other volcaniclastic debris accumulated on the lower flanks of the volcanoes and on the surrounding lava plains. Unconformities separate many of the major Fifes Peak members due to the primary dips of overlapping volcanic units (Swanson, 1966; Carkin, 1985).

We have dated rocks from four relatively well known sections of Fifes Peak strata. Each section overlies the Ohanapecosh and several are marked, at least locally, by a basal unit of silicic tuffs. The stratigraphy and age data for these sections are briefly reviewed below and summarized in Figure 5. Several large and distinctive silicic ignimbrites in the Fifes Peak Formation have been mapped in detail. Because these are useful stratigraphic markers, we have retained the names previously used for them and suggest informal names for several others.

American River Area

This is the type area of the Fifes Peak Formation (Warren, 1941). The Fifes Peak section here unconformably overlies well-bedded volcaniclastic sediments of the Ohanapecosh Formation. The lower part of the Fifes Peak section in this area includes a distinctive sequence of vitric ash-flow tuffs about 300 m thick, here informally named the Bumping River tuff. In most places, the Bumping River tuff overlies andesitic beds as much as several hundred meters thick, but locally it rests directly on Ohanapecosh strata. Two samples of the tuff yielded fission-track ages of about 27 m.y. The Bumping River tuff is overlain by andesitic lavas and pyroclastic and volcaniclastic rocks and minor intermediate to silicic pumice flows, ash flows, and air-fall tuffs, interpreted as the distal deposits of Fifes Peak volcano. Fifes Peak volcano is a large dissected cone about 18 km in diameter consisting of andesitic breccias and subordinate lavas. A 2-m-thick bed of pumice-lapilli tuff (JV 239) exposed on Fifes Ridge about 5 km east of Fifes Peak yielded a zircon U-Pb age of 25.5 m.y. and a fission-track age of about 26 m.y. Zircon from the upper part of the Fifes Peak section south of Cliffdell (Carkin, 1985) was dated at 23.3 ± 2.4 m.y. Laursen and Hammond (1974) reported a K-Ar age of 30 ± 3 m.y. on plagioclase from the Fifes Peaks area, an age inconsistent with our dates and probably too old.

The Fifes Peak Formation in this area is unconformably overlain by lower (Grande Ronde) flows of the Columbia River Basalt Group that have been dated elsewhere at about 16 m.y. (McKee and others, 1977).

Tieton River–Rattlesnake Creek Area

The stratigraphy of the Tieton River area was studied by Swanson (1964, 1978). The Fifes Peak here unconformably overlies the Ohanapecosh Formation (Wildcat Creek beds) which has given a fission-track age of about 32 m.y. The formation in this area is unconformably overlain by the Columbia River basalt.

The Fifes Peak consists of a basal unit of silicic ash-flow tuffs about 300 m thick, the Burnt Mountain tuff of Swanson (1964, 1978), and two overlying stratovolcanoes, the Tieton volcanoes. A zircon fission-track age of about 25 m.y. was determined for the Burnt Mountain tuff and an age of about 23 m.y. from

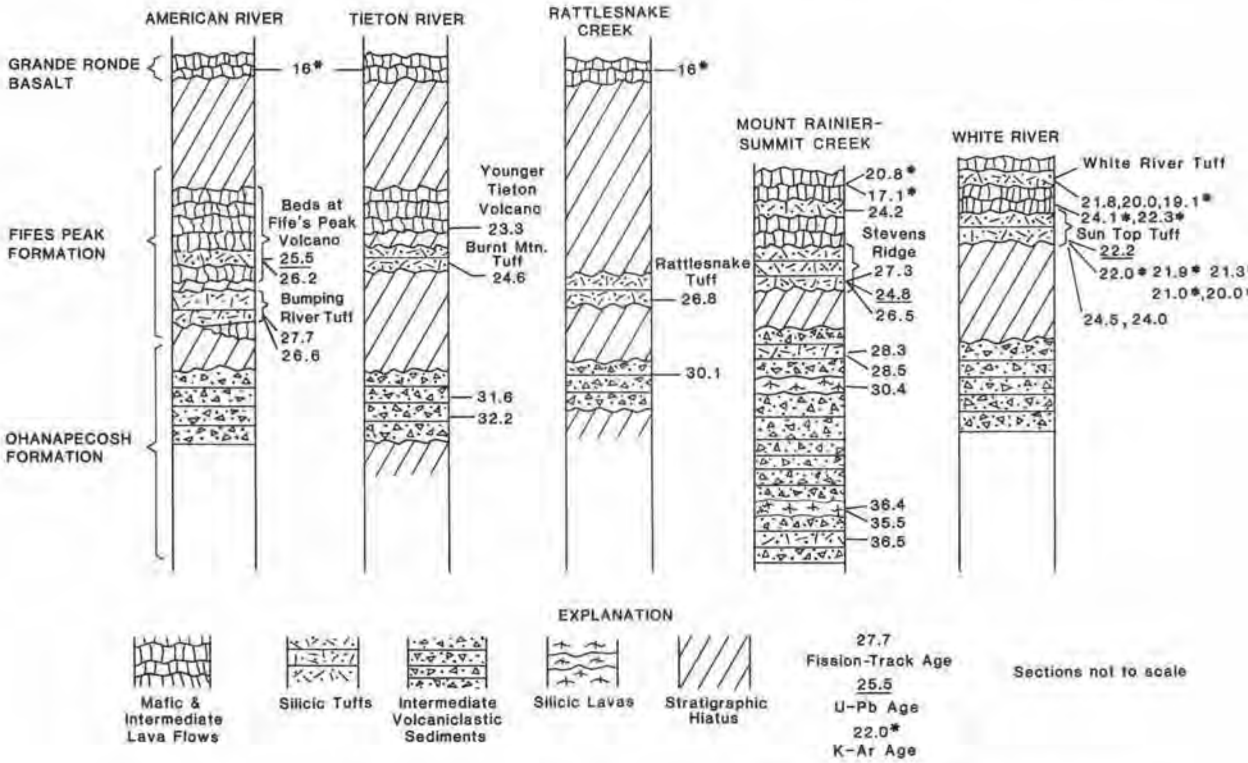

Figure 5.–Summary of ages and stratigraphic relations of the Ohanapecosh and Fifes Peak Formations. See text and Table 1.

a pumice flow interbedded with distal lavas of the younger Tieton volcano.

Schreiber (1981, 1982) studied the stratigraphy of the Rattlesnake Creek area about 10 km north of the Tieton River and determined several fission-track ages. The Ohanapecosh (Wildcat Creek beds) in this area gave an age of about 30 m.y. The unit is overlain by poorly exposed vitric ash-flow tuff, referred to informally by Schreiber as the Rattlesnake Creek tuff and dated by her at about 27 m.y. This tuff unit is petrographically similar to and may prove to be equivalent to the Burnt Mountain tuff.

In the Mount Aix area Schreiber mapped the east and south margins of a complex eruptive and collapse structure about 7 km in diameter, which she interpreted as a caldera. This structure, referred to below as the Mount Aix caldera, is not yet fully mapped, but it may be the source of several major Fifes Peak ash-flow sequences exposed around its periphery. These tuffs include those at Bumping River, Burnt Mountain, and Rattlesnake Creek. Schreiber (1981, 1982) obtained a date of about 28 m.y., and we determined an age near 26 m.y. on different samples of ash-flow tuffs from the caldera fill complex. These quartz-phyric, crystal-rich tuffs closely resemble the type Stevens Ridge ash-flow tuffs in Mount Rainier National Park and, given the 10 percent uncertainty in the fission-track ages, may possibly be equivalent to the Stevens Ridge. More detailed mapping of the caldera, as well as investigations of the petrography, distribution, and lateral variation in thickness and particle size of these several tuff units will be needed to more securely establish their correlation and source. Because some of the tuffs of the Mount Aix area differ widely in lithology, several eruptive episodes must be represented.

Mount Rainier Area

Fiske and others (1963) studied the Stevens Ridge–Fifes Peak sequence of Mount Rainier National Park. The Stevens Ridge Formation consists of thick rhyodacitic ash-flow tuffs overlain by silicic volcanic sediments. The ash flows are rich in crystals, dominantly plagioclase and commonly large quartz phenocrysts, and typically contain abundant lithic volcanic fragments and pumice in a matrix of devitrified shards. Stevens Ridge sections in the park range widely in thickness to a maximum of about 900 m, thus suggesting accumulation on a surface of considerable topographic relief and an unconformable relation with the underlying Ohanapecosh Formation (Fiske and others, 1963).

Zircon from the prominent exposures of the basal Stevens Ridge ash flow on State Route 12 just south of the park gave a U-Pb age of 24.8 m.y. and a fission-track age of about 26 m.y. A sample from the type area on the Stevens Canyon Road in the park gave a fission-track age of about 27 m.y. The source of the Stevens Ridge pyroclastic rocks is unknown.

The Stevens Ridge unit in the park is conformably overlain by remnants of a Fifes Peak andesitic lava sequence as much as 700 m thick. These andesites are interbedded with minor volcaniclastic sediments and, near Mowich Lake, with an ash-flow tuff that gave a fission-track age of about 24 m.y. Hartman (1973) reports K-Ar ages of 20.8 ± 2.6 and 17.1 ± 4.3 m.y. (corrected to the new decay constant) on plagioclase from an andesite collected near Mowich Lake.

Fiske and others (1963) described a welded ash-flow tuff of local extent at the Palisades in the northeastern corner of the park. This unit is noteworthy in that it reportedly overlies its source, a shallow intrusion which broke through its cover of Ohanapecosh Formation and vented to the surface. Mattinson (1977) determined a U-Pb age of 25.1 m.y. for this tuff. Although the U-Pb age of the Palisades rocks is similar to that of the Stevens Ridge unit in the park, these units are not physically correlative because they differ significantly in mineralogy.

White River Area

The geology of the White River area, lying immediately north and northeast of Mount Rainier National Park, has been studied by Fischer (1970), Hartman (1973), and Frizzell and others (1984). Both the basal silicic beds and the upper, dominantly andesitic members of the Fifes Peak Formation are widely exposed. Silicic ash flows and overlying sediments make up the base of the section in the area of the Dalles and Sun Top Peak at the northeast corner of the park. We here refer to these beds as the Sun Top tuff. Outcrop patterns show that the tuff fills a valley in the eroded underlying Ohanapecosh. The Sun Top tuff was correlated by Hartman with the Stevens Ridge, but its age and mineralogy—it contains abundant brown hornblende that is lacking in the Steven Ridge—rule out this correlation. A sample from near the summit of Sun Top Peak gave a U-Pb age of 22.2 m.y., a hornblende K-Ar age of 22.0 ± 0.5 m.y. (Frizzell and others, 1984), and a fission-track age of about 24 m.y. Another Sun Top sample collected west of the confluence of the West Fork with the White River gave K-Ar ages of 21.3 ± 0.7 and 22.0 ± 0.8 m.y. on hornblendes and biotite respectively (Frizzell and others, 1984). A third sample from just northeast of the park yielded a fission-track age of about 24 m.y. Hartman (1973) reports K-Ar ages of 21.0 ± 1.5 and 20.0 ± 1.8 m.y. (corrected for the new decay constant) on plagioclase from another Sun Top sample. These data indicate an age near 22 m.y., significantly younger than both the type Stevens Ridge and the type Fifes Peak. Fischer (1976) reported K-Ar ages of about 35 and 24 m.y. (corrected for the new decay constant) respectively on hornblende and biotite from a tuff which he correlated with the Stevens Ridge in the Clear Fork area. This tuff is probably equivalent to the Sun Top tuff. The hornblende age is contradicted by all other age data for the unit and is clearly too old.

The ash flows at Sun Top are overlain successively by: silicic volcaniclastic sedimentary rocks, here also included in the Sun Top unit; an andesitic flow sequence; a silicic tuff unit here referred to as the White River tuff (Frizzell and others, 1984); and more andesite flows. Whole-rock K-Ar ages of 22.3 ± 1.9 and 24.1 ± 1.4 m.y. (corrected for the new decay constants) were determined on a sample of Fifes Peak andesite from Castle Mountain northeast of the park (Hartman, 1973). The age of the lower andesites of the Fifes Peak north of the park is about 22 m.y. as it overlies the Sun Top tuff, which is well dated near 22 m.y., and is older than a U-Pb age of about 21.8 m.y. on the flow-laminated intrusive rhyodacites of Clear West Peak (Mattinson, 1977), the apparent source of the White River tuff (Fischer, 1970; Hartman, 1973; Frizzell and others, 1984). We determined a fission-track age of about 20 m.y. on zircon, and Hartman reported a K-Ar whole-rock age of 19.1 ± 0.4 m.y. (corrected for the new decay constants) from other samples of the Clear West Peak intrusion.

Summary and Conclusions

The Fifes Peak Formation overlies the Ohanapecosh Formation with regional unconformity. Fifes Peak activity was marked by the construction of major volcanic edifices producing greater topographic relief than that of the Ohanapecosh interval. At several localities Fifes Peak volcanism was initiated by silicic pyroclastic activity. This early silicic volcanism was diachronous as indicated by the following dates: Bumping River tuff, older than 25.5 m.y. (U-Pb), probably about 27 m.y. (fission-track); Stevens Ridge, about 24.8 m.y. (U-Pb); and Sun Top tuff about 22 m.y. (U-Pb and K-Ar). Andesitic volcanism had begun by 25.5 m.y. (U-Pb), probably at about 27 m.y., in the Fifes Peak area and, accompanied by sporadic silicic activity, continued to at least 21.8 m.y. (U-Pb) in the White River area. The upper age limit of the Fifes Peak Formation is not well defined; its stratigraphically highest parts in the Naches Pass area have not been dated. However, in the Bethel Ridge and Naches River areas in the east slope of the southern Washington Cascades, the Fifes Peak Formation was mildly warped and eroded prior to accumulation of the overlying flows of the Columbia River Basalt Group dated at about 16 m.y. Fifes Peak activity had ceased by that time.

As discussed in detail above, regional stratigraphic relations and radiometric age determinations indicate that andesitic and silicic volcanism were essentially coeval in the stratigraphic interval overlying the Ohanapecosh Formation. Moreover, silicic ignimbrites in the lower part of the sequence differ widely in age and lithology, and thus are not correlative with the Stevens Ridge unit of the Mount Rainier National Park area. We believe that this stratigraphic problem can best be resolved by redefining the Fifes Peak Formation more broadly to include both the silicic pyroclastic and andesitic rocks stratigraphically above the Ohanapecosh in the Fifes Peak Formation. The Stevens Ridge would be reduced to member status and restricted to the type section and to demonstrable equivalents. Other silicic units, formerly loosely correlated with the Stevens Ridge, could then be given separate site-specific names and designated as members of the Fifes Peak Formation, as can the andesites of the various Fifes Peak eruptive centers. This approach is consistent with the practice of several recent workers of assigning separate names to silicic and andesitic units within this broad stratigraphic interval (Swanson, 1964; Fischer, 1970) and will accommodate rigorous stratigraphic subdivision of the Fifes Peak Formation as more detailed studies are carried out.

CONCLUSIONS

The pre-Tertiary rocks of the Tieton inlier are the basement to a sequence of Eocene fluvial sedimentary rocks and basaltic and rhyolitic volcanic rocks locally preserved along its periphery. A thick Eocene section is exposed in the Summit Creek area west of the inlier. The lower part of this sequence consists of about 2,000 m of subaerial basalt flows with minor interbedded argillite, arkosic conglomerate, and silicic ash flows that yielded a zircon U-Pb age of 55 m.y. at the base. Zircon from ash flows high in the unit gave fission-track ages near 45 m.y. The top of the Eocene sequence consists of coal-bearing fluvial arkosic sediments about 750 m thick whose age is bracketed between about 45 and 36 m.y. Eocene strata locally preserved on the crest and along the south and east flanks of the inlier are less than 300 m thick and are of late middle Eocene age at the two localities where they have been dated. The older Eocene basaltic sequence of the Summit Creek area is the age equivalent of the Swauk–Silver Pass unit and Teanaway Formation in the central Cascades. The late middle to late Eocene rocks are correlative with the Naches Formation to the north and the Puget Group to the west.

The contrast between the thick and relatively complete Eocene sequence of the Summit Creek section and the thin, incomplete Eocene sequences in the Tieton inlier suggests that the inlier was a basement high during Eocene time. This marked and abrupt contrast in thickness suggests the existence of a hingeline between a rapidly subsiding basin on the west and the more slowly subsiding, intermittently uplifted Tieton block on the east. It is possible that this hingeline may have been fault controlled, though our reconnaissance mapping has failed to reveal any major faulting in this zone. Fault control has been proposed for similar, localized, thick Eocene nonmarine sequences in western Washington (Johnson, 1982; Tabor and others, 1984).

Eocene volcanism in the White Pass–Tieton River area, and the Puget Group, Taneum, Teanaway, and Naches volcanism constitute the westernmost expression of Challis magmatism in the Pacific Northwest. The Challis episode (Armstrong, 1978) lasted from approximately 53 to 42 m.y. and was distinguished by widely scattered calc-alkaline volcanic and shallow plutonic activity through much of Washington, Idaho, northeastern Oregon, and southern British Columbia. Challis magmatism was accompanied by development of fault-controlled basins characterized by thick volcanic and sedimentary sequences (Ewing, 1980).

The Cascade magmatic arc in Washington State was established as a relatively narrow NNE-trending belt by about 36 m.y. ago, early Oligocene (Vance, 1982; Vance and others, 1986). Cascade activity followed an interval of diminished magmatism indicated in the Mount Rainier–Tieton area, and elsewhere, by an absence or paucity of radiometric ages in the range 41-37 m.y. Cascade volcanism began at about 36 m.y. ago with the accumulation of the intermediate pyroclastic and volcanic rocks of the Ohanapecosh Formation. The tectonic regime that controlled Eocene deposition in the Mount Rainier–Tieton area remained active into early Ohanapecosh time. West of the Tieton inlier, a thick and complete Ohanapecosh section, which included early Oligocene rocks (36-34 m.y.), conformably overlies Eocene beds. By contrast, the Ohanapecosh of the Tieton inlier lacks earliest Oligocene strata, is markedly thinner, and unconformably overlies pre-Tertiary and Eocene rocks. These relations imply that the inlier was a structural high in early Ohanapecosh time and that a hingeline existed at the boundary of the inlier with the rapidly subsiding Ohanapecosh basin to the west. Ohanapecosh activity extended from about 36 to 28 m.y., early and middle Oligocene. It was dominantly andesitic to dacitic in composition, but rhyodacitic ash flows and lavas are locally present in both the lower and upper parts of the type section in the Mount Rainier–Summit Creek area. The Ohanapecosh Formation reflects an interval of regional subsidence and low relief. This tectonic regime ended with an episode of mild folding, uplift, and erosion that preceded the deposition of the overlying Fifes Peak Formation. Major up-arching of the Tieton inlier occurred at this time, producing steep to moderate dips in the thick Eocene-Ohanapecosh sequence west of the inlier and gentler dips on the east and south flanks.

Fifes Peak volcanism was diachronous. It began in the type area prior to 25.5 m.y. ago, probably at about 27 m.y., with eruption of andesites followed by silicic tuffs. In the Mount Rainier and Sun Top areas, it began with silicic tuffs erupted at about 24.8 and 22 m.y. respectively. Later Fifes Peak activity was dominantly andesitic and basaltic and related to a series of scattered stratovolcanoes and shield volcanoes. Large stratovolcanoes imply significant topographic relief on the Fifes Peak landscape. Fifes Peak volcanism, including sporadic silicic activity, continued until at least 20 m.y ago. Gentle folding occurred after deposition of the Fifes Peak Formation, continuing the up-arching of the domal Tieton inlier. This folding predates the deposition of the unconformably overlying flows of the Columbia River Basalt Group that are dated at about 16 m.y.

As discussed in Appendix II, a significant component of inherited zircon is present in three of the four silicic tuff samples dated by the U-Pb method. The age of this inherited lead is in the range 1,150-1,650 m.y. The presence of this old zircon implies that pre-existing crustal rocks, either Precambrian crust or younger crust containing Precambrian detrital zircons, played a part in the genesis of the silicic magmas by partial melting or assimilation. In either case, this isotopic signature indicates the presence at depth of silicic crustal rocks which contain old zircon and which are quite different from the Jurassic sedimentary and igneous rocks exposed in the Tieton inlier.

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APPENDIX I.–Localities and descriptions of dated samples

Eocene Units

Locality 1 Sample JV 310

lat. 46° 40.9′ N., long. 121° 27.6′ W.

Road-cut about 1 km up US Forest Service (USFS) road 1449 which links Cortright Creek with Laurel Hill. First large outcrop on north side of road as one begins the westward ascent toward Laurel Hill from the sharp "U" turn that marks the eastern end of the Cortright Creek road. Eastern part of outcrop is a Miocene intrusion. Tuff overlies steeply dipping conglomerate and underlies laminated mudstones 50 m up section to the west.

Greenish, quartz-phyric lapilli tuff. About 7% quartz, 5% feldspar altered to carbonate and zeolites, 5-10% lithic lapilli in devitrified glass matrix.

Locality 2 Sample JV 229

lat. 46° 40.9′ N., long. 121° 29.9′ W.

From south side of Cortright Creek valley in a road cut along USFS road 1439. Platy to blocky jointed, pale gray rhyolite.

Dense laminated quartz-phyric welded tuff?

Locality 3 Sample JV 232

lat. 46° 41.4′ N., long. 121° 29.5′ W.

From a small spur road off USFS road 145, about 150 m west of their intersection. A resistant knob of massive to platy jointed, white to gray, banded rhyolite.

Dense laminated crystal-poor welded tuff?

Locality 4 Sample TRG 2

lat. 46° 32.1′ N., long. 121° 23.6′ W.

From a small creek where it plunges over the northern rim of the gorge of the North Fork Tieton River. The basal, 20-cm-thick tuff bed of a thick sequence of well-bedded volcaniclastic rocks and lava flows. The tuff overlies deformed thin-bedded mudstones and vesicular olivine basalts.

Quartz-phyric vitric tuff.

Locality 5 Sample JV 128

lat. 46° 37.7′ N., long. 121° 7.9′ W.

Outcrops of the Spencer Creek tuff at bridge at mouth of South Fork Tieton River at Rimrock Lake.

Black glassy welded tuff with about 15% plagioclase and quartz crystals and large flattened pumice clasts.

Locality 6 Sample JV 125

lat. 46° 42.3′ N., long. 121° 31.0′ W.

USFS road 1400A on the south side of Summit Creek, 2 km east of Summit Creek Campground. Ash flow from lower Ohanapecosh.

Abundant pumice lapilli to 1 cm, about 10% small volcanic lithic fragments, 5% plagioclase phenocrysts in a matrix of shards. Laumontite and prehnite alteration.

Locality 7 Sample JV 126

lat. 46° 41.5′ N., long. 121° 31.1′ W.

Laurel Hill (ridge between Cortright and Summit Creeks) USFS road 1449, 0.5 km SE of hill 4640. On strike with the base of the type section of the Ohanapecosh.

Laminated gray rhyolite flow with about 2% plagioclase phenocrysts.

Locality 8 Sample JV 94B

lat. 46° 40.3′ N., long. 121° 30.8′ W.

State Route 12, 0.2 km west of milepost 143. Near base of Fiske's type section of the Ohanapecosh.

Laminated gray rhyolite with 3% plagioclase crystals.

Locality 9 Sample CT-6

lat. 46° 35.9′ N., long. 121° 06.5′ W.

From about 0.5 km south of a hairpin turn on USFS road 1324 where it turns north-south at altitude 4,300 ft above Spencer Creek. Interbedded tuffaceous shales, sandstone, and conglomerate exposed along the steep valley sides of Spencer Creek.

White crystal vitric airfall tuff with ash altered to clay.

Locality 10 Sample JV 139

lat. 46° 40.5′ N., long. 121° 06.5′ W.

End of USFS road 1481, just north of Wildcat Creek. A marker bed mapped by Swanson (1978) in the middle of the Wildcat Creek section.

Greenish pumice lapilli tuff contains about 10% basalt and andesite fragments up to 1 cm across, and 1% plagioclase crystals.

Locality 11 Sample JV 233

lat. 46° 42.3′ N., long. 121° 17.2′ W.

From near the trail along the crest of Russell Ridge about 2 km northwest of McNeil Peak.

Brownish to buff crystal-poor pumice lapilli tuff.

Locality 12 Sample JV 96

lat. 46° 6.4′ N., long. 121° 15.9′ W.

Prominent outcrop of Ohanapecosh Formation on USFS road 1911, Bear Creek, Little Naches area.

Welded ash-flow tuff. Flattened pumice fragments to 3 cm in diameter, about 15% plagioclase crystals, and 10% basalt and andesite clasts.

Locality 13 Sample JV 3

lat. 46° 48.8′ N., long. 121° 37.1′ W.

Rhyolite flow at Indian Bar; Cowlitz Divide trail just south of point 5914′. White to buff rhyolite with flow lamination and scattered spherulites.

Contains about 3% plagioclase phenocrysts.

Locality 14 Sample JV 269

lat. 46° 43.4′ N., long. 121° 35.2′ W.

Upper Ohanapecosh on the east side of Backbone Ridge on Highway 706 (Stevens Canyon Road).

Ash-flow tuff. About 13% lithic volcanic fragments with scattered pumice lapilli in a matrix of altered shards. Laumontite-prehnite alteration.

Locality 15 Sample JV 124

lat. 46° 43.4′ N., long. 121° 35.2′ W.

Just south of previous locality and about 100 m higher in the Ohanapecosh section.

Ash-flow tuff. About 20% volcanic lithic fragments, 10% plagioclase, and 7% quartz crystals and abundant small fragments in an altered ash matrix. Veining and patchy alteration to carbonate common.

Fifes Peak Formation

Locality 16 Sample JV 62

lat. 46° 56.3′ N., long. 121° 12.0′ W.

Bumping River tuff from prominent cliff along Bumping River Road, 7 km south of confluence of Bumping and American Rivers.

Ash-flow tuff. About 7% volcanic lithic clasts, 5% plagioclase crystals, abundant pumice lapilli (<1.0 cm) in a matrix of slightly compressed glass shards.

Locality 17 Sample JV 241

lat. 46° 56.5′ N., long. 121° 11.9′ W.

On USFS road 1711, 0.5 km north of the previous locality. Massive pinkish lapilli tuff.

Pumice and basaltic and andesitic lapilli tuff. Pumice as much as 8 cm across, about 7% plagioclase crystals in a pinkish to buff vitric matrix.

Locality 18 Sample JV 239

lat. 46° 58.7′ N., long. 121° 13.8′ W.

Type Fifes Peak Formation. USFS road 1800 on Fifes Ridge at head of west Quartz Creek. 2-m-thick white pumice lapilli tuff interbedded in andesitic breccia.

About 15% andesite and basalt clasts, 2% plagioclase crystals, and less than 1% crystals of green clinopyroxene in matrix of pumice lapilli (to 2 cm) and shards.

Locality 19 Sample JV 226

lat. 46° 41.0′ N., long. 121° 08.3′ W.

Burnt Mountain tuff of Swanson (1978) from about 1.5 km north of Lightning Lake on the ridge crest at the eastern end of a northern branch (USFS road 1426) of USFS road 144 in the Wildcat Creek drainage basin.

Pale- to dark-brown, quartz-poor, pumice and lithic lapilli tuff.

Locality 20 Sample JV 140

lat. 46° 42.4′ N., long. 121° 08.9′ W.

Pumice lapilli tuff from the lowest of the conspicuous light-colored pumice flows of the Tieton volcano at western Bethel Ridge (Swanson, 1978) from a ridge crest trending due south from Cash Prairie on western Bethel Ridge.

White to buff, crystal-poor, rhyolitic pumice lapilli and accessory block tuff.

Locality 21 Sample BP

lat. 46° 45.7′ N., long. 121° 16.7′ W.

From massive cliffs of rhyodacitic pyroclastic rocks on the southeast side of Bismark Peak just above the last saddle below the summit. Occurs as a vertical dike-like, cross-cutting body.

A dense crystal-rich tuff containing plagioclase and distinctive large quartz phenocrysts in a matrix of devitrified shards and pumice.

Locality 22 Sample JV 126

lat. 46° 39.9′ N., long. 121° 35.6′ W.

Prominent exposures along State Route 12, 1.8 mi southwest of junction with highway 123. Basal Stevens Ridge ash flow.

White crystal-rich tuff with about 25% plagioclase, 10% quartz, pumice clasts as much as 1 cm across, and altered shards. Extensive carbonate alteration.

Locality 23 Sample JV 32

lat. 46° 42.7′ N., long. 121° 35.6′ W.

Basal Stevens Ridge ash flow along Stevens Canyon Road just east of the crest of Backbone Ridge. Slightly welded.

About 20% volcanic lithic fragments, 17% plagioclase crystals, flattened pumice lapilli (<2 cm), in a matrix of altered shards. Laumontite-prehnite alteration.

Locality 24 Sample JV 92

lat. 46° 56.7′ N., long. 121° 52.1′ W.

Prominent road cut 0.4 mi northwest of Mowich Lake. Slightly welded ash-flow tuff.

About 25% volcanic lithic clasts, 10% plagioclase crystals, and a trace of quartz. Flattened pumice fragments as much as 2 cm across in a matrix of altered ash. Laumontite alteration.

Locality 25 Sample JV 165

lat. 47° 2.4′ N., 121° 35.7′ W.

Sun Top tuff. USFS road 1849 just below Sun Top Lookout. Lithic ash-flow tuff.

About 40% volcanic lithic clasts as much as 4 cm in diameter. Crystals: approximately 4% plagioclase; 2% quartz; 2% greenish-brown hornblende; 1% biotite; trace of clinopyroxene. Small pumice fragments and an ash matrix.

Locality 26 Sample JV 36

lat. 47° 3′ N., long. 121° 33.9′ W.

Sun Top tuff. The Dalles, the lower falls on the Dalles Creek trail. From basal columnar zone of dark welded ash-flow tuff unit.

About 18% volcanic lithic fragments. Crystals include about 8% plagioclase, 3% quartz, 3% hornblende, 1% biotite, and a trace of clinopyroxene in matrix of highly compressed glass shards.

Locality 27 Sample JV 163

lat. 47° 3.8′ N., long. 121° 43.8′ W.

One mile northeast of Frog Mountain on the west branch of USFS road 1810. Clearwest Peak intrusion?

Buff rhyodacite with contorted flow lamination. About 4% plagioclase crystals in devitrified glass matrix.

Locality 28 Sample JV 354

lat. 46° 51.3′ N., long. 121° 6.9′ W.

Upper Nile Creek, about 5 km southeast of Bald Mountain. Upper part of Fifes Peak section (collected by Brad Carkin).

Tan pumiceous tuff, 8% broken plagioclase phenocrysts, and 15% andesitic lithic clasts.

APPENDIX II.–Discussion of U-Pb Data

JV 67

Zircons from JV 67 (Table 1), from the Stevens Ridge unit, show relatively minor contributions from older sources. The coarse fraction contains a small, but distinct component of "old" Pb, whereas the fine fraction, which has a large analytical uncertainty, could contain a very small older contribution, or could be considered concordant at 25.0 m.y. (Table 2, Fig. 6). Considering the possibility of a small inherited component in the fine fraction, its $^{206}Pb/^{238}U$ age of 25.0 ± 0.2 m.y. would be a maximum age of crystallization age. Using only the well-defined coarse fraction, and assuming upper intercept ages similar to those observed for the other samples with larger contributions of old zircon (JV 165 and JV 310), we can cross-check the age inferred from the fine fraction. Using the extreme range of upper intercept ages of 1,150 and 1,630 m.y., we obtain lower intercept ages of 24. 6 and 25.1 m.y., respectively. These are in excellent agreement with the age inferred from the fine fraction of zircon, and we take the crystallization age of the tuff as 24.8 ± 0.3 m.y.

JV 239

Zircons from JV 239, a sample from the Fifes Peak area, give the only completely concordant ages among the samples studied by U-Pb isotopic methods. Both the coarse and fine fractions of zircon yield $^{206}Pb/^{238}U$ ages of 25.5 m.y., and $^{207}Pb/^{206}Pb$ ages are in agreement with the U-Pb ages within analytical uncertainties (Table 2 and /fig. 6). Thus, the age is well determined at 25.5 ± 0.1 m.y.

JV 165

Zircons from JV 165, from the Sun Top area, show clear effects of inheritance of Precambrian zircons in their U-Pb systematics. However, the two fractions are precisely determined, plot close to the lower intercept, and show a modest spread on concordia (Fig. 6). This permits good control on the lower concordia intercept, and hence, on the age of crystallization of the tuff. Regression using a York-type program yields a lower intercept (crystallization) age of 22.2 +0.3/-0.5 m.y. The upper intercept is subject to very large uncertainties owing to the long projection to concordia: 1,510 +200/-180 m.y.

JV 310

Zircon systematics of JV 310, an ash-flow tuff near the base of the Tertiary section in the Summit Creek area, are dominated by an inherited component of Precambrian age. Even the finest fractions of zircon have $^{207}Pb/^{206}Pb$ ages in the range of 600 m.y. The fact that the zircon fractions do not define a single chord within analytical uncertainties (Fig. 6) indicates that the inherited component is heterogeneous; this precludes a precise determination of the crystallization age of the tuff. Nevertheless, some limits can be placed on the age

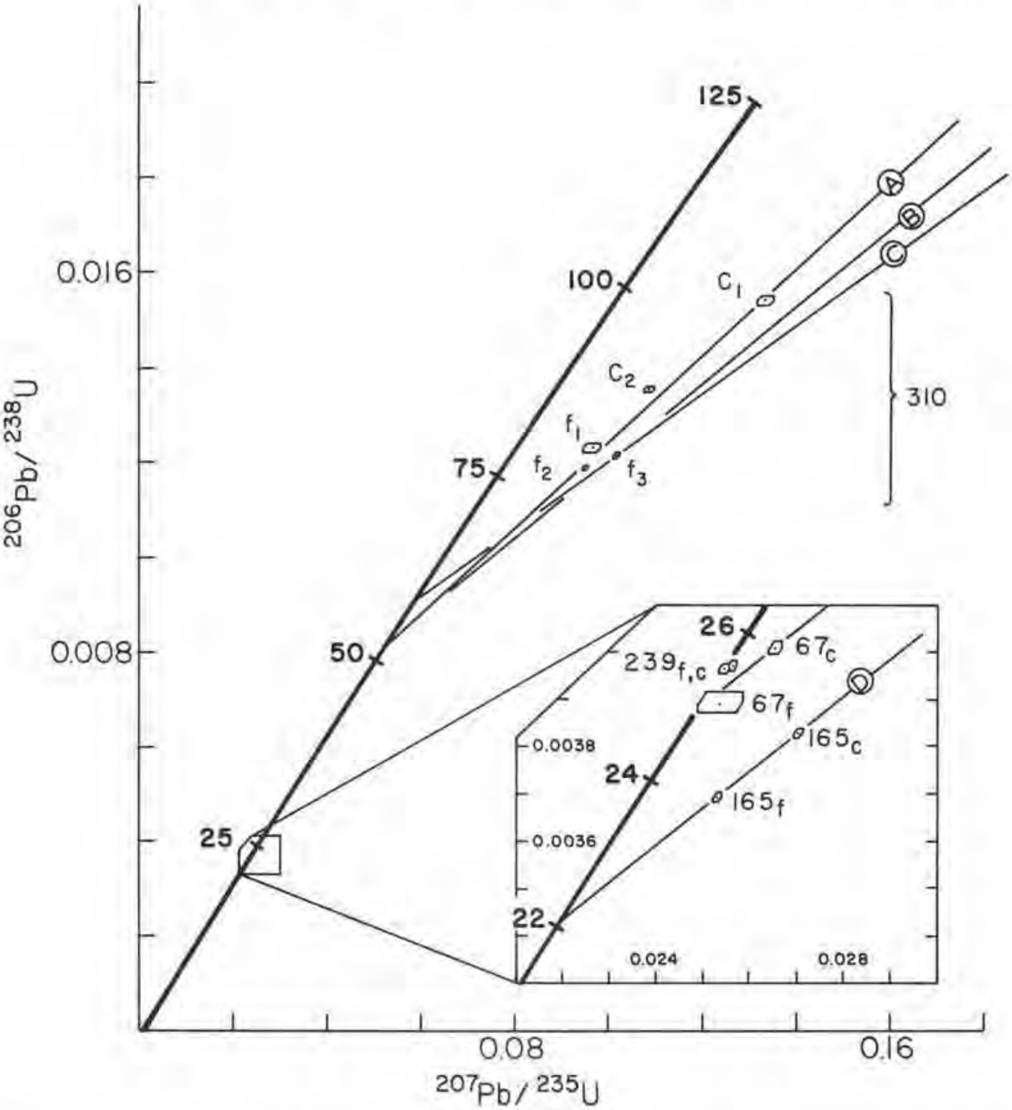

Figure 6.–Concordia diagram for zircon data. Hexagons show estimated uncertainties in the individual points at the 95% confidence level. Three reference chords (A, B, and C) are shown for the JV 310 data. Upper and lower intercept ages for (A) are 1,150 and 52 m.y., respectively; for (B), 1,390 and 53 m.y.; and for (C), 1,630 and 58 m.y. Chord (D) (inset) is fitted to the data for JV 165. It intersects the concordia curve at 1,510 and 22.2 m.y. See Appendix II for detailed discussion.

of the sample and on the nature of the inherited component. A simple linear regression for the five data points yields chord (A), with upper and lower intercepts of 1,150 and 52 m.y., respectively. Because of the geological scatter in the data, the intercept ages (especially the upper intercept age) are subject to substantial uncertainties and need not have rigorous age significance. Further constraints can be placed on the crystallization age of the tuff by making some reasonable assumptions about the age of the inherited component and "forcing" chords through these points and the observed data. The nearest known source of zircons with isotopic systematics of the type required by the observed data [that is, the upper intercept for chord (A)] is the Swakane Gneiss in the southern part of the North Cascades (Mattinson, 1972), about 110-120 km northeast of the Summit Creek area. If we assume that the gneiss or sedimentary rocks derived from the gneiss are the source of the old zircons in the tuff, its isotopic age may be further refined. Taking the most "outboard" datum point for JV 310 (f3 on Fig. 6), and forcing chords through this point and the "youngest" and "oldest" zircon data points for the Swakane Gneiss from Mattinson (1972) yields chords (B) and (C) respectively. Chord (B) yields upper and lower intercepts of 1,390 and 53 m.y., and chord (C) yields upper and lower intercepts of 1,630 and 58 m.y. If the source of inherited zircons contained a mixture of material from the Swakane Gneiss and younger sources (for example, the Mesozoic plutonic rocks of the North Cascades), then reference chords (B) and (C) would yield lower intercepts that are too old. We note, however, that the fission-track data discussed above indicate a minimum age of about 45 m.y. Thus, the ages suggested by the three reference chords (A, B, and C) in Figure 6, that is, about 55 ± 3 m.y., are not unreasonable.

Table 2.–U-Pb isotopic data from selected samples (localities 1, 18, 22, 25) of zircon for Tertiary volcanic rocks in the Mount Rainier–Tieton River area, Washington.

(1) c = >200 mesh fraction; f = <200 mesh fraction.

(2) Concentrations in ppm. * indicates radiogenic Pb. Analytical methods miniaturized from those of Krogh (1973).

(3) Pb isotopes collected simultaneously in four Faraday cups in the UCSB MAT261 multicollector mass spectrometer. Corrected for 0.10% per mass unit mass fractionation.

(4) Ages calculated using the decay constants of Jaffey and others (1971). Uncertainties in the $^{206}Pb/^{238}U$ ages are 0.5% or better, depending on the 204/206 ratios. Uncertainties in the $^{207}Pb/^{206}Pb$ ages are based on measurement uncertainties at the two-sigma level for the ratios, plus a ±0.1 uncertainty in the ratio of 207/204 (initial) used in correcting for common Pb. The latter is by far the major source of uncertainty in the $^{207}Pb/^{206}Pb$ ages.

	Concentrations(2)		Pb isotopic composition(3)			Age (in m.y.)(4)	
Sample(1)	$^{206}Pb*$	^{238}U	208/206	207/206	204/206	206*/238	207*/206*
JV 67 c	1.907	549.8	0.2151	0.08278	0.002359	25.7	100 ± 15
JV 67 f	2.124	631.0	0.4195	0.16018	0.007649	25.0	67 ± 45
JV 165 c	2.222	670.4	0.2729	0.06504	0.000938	24.6	253 ± 10
JV 165 f	2.038	637.1	0.2440	0.06348	0.000937	23.7	181 ± 10
JV 239 c	0.716	208.2	0.1844	0.05636	0.000648	25.5	38 ± 10
JV 239 f	0.653	190.3	0.2609	0.08457	0.002577	25.5	24 ± 15
JV 310 c1	1.869	140.2	0.5755	0.24220	0.01241	98.4	700 ± 55
JV 310 c2	1.796	153.1	0.3110	0.13101	0.004992	86.7	532 ± 20
JV 310 f1	2.099	197.0	0.6799	0.2738	0.01479	78.8	484 ± 80
JV 310 f2	1.820	177.0	0.1779	0.07233	0.000985	76.1	530 ± 10
JV 310 f3	2.005	190.6	0.1918	0.07980	0.001299	77.8	634 ± 10

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CENOZOIC STRATIGRAPHY, UNCONFORMITY-BOUNDED SEQUENCES, AND TECTONIC HISTORY OF SOUTHWESTERN WASHINGTON

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ABSTRACT

The Cenozoic stratigraphy of southwestern Washington consists of a complex array of sedimentary and igneous units that reflect a sequence of tectonic events including seamount accretion, arc magmatism, and forearc deformation. This paper reviews the stratigraphic nomenclature and geology of the Cenozoic units, organizes those units into five depositional sequences separated by regionally significant unconformities, and interprets the relative role of global sea-level fluctuations and regional tectonics in the development of the stratigraphic sequences. The data support an interpretation of regional tectonic events being the dominant influence on the Cenozoic depositional history of southwestern Washington. The tectonic events are: (1) middle Eocene accretion of a seamount chain; (2) early late Eocene westward relocation of subduction; (3) late Eocene onset of Cascade arc volcanism; (4) late early Miocene plate readjustment due to back-arc extension in the Columbia River Plateau and Great Basin; and (5) late Pliocene to early Pleistocene northeast compression forced by continued subduction of remnants of the Kula plate beneath North America.

INTRODUCTION

The Cenozoic stratigraphy of southwestern Washington consists of a complex array of sedimentary and igneous units formed by seamount accretion, arc magmatism, and forearc deformation. This paper reviews the stratigraphic units mapped, organizes those units into depositional sequences separated by regionally significant unconformities, and interprets the relative role of global sea-level fluctuations and regional tectonics in the development of these sequences.

The study area encompasses part of the Washington Coast Range, lying between the subduction complex of the Oregon-Washington continental shelf and the magmatic arc of the Cascade Range (Fig. 1). The Coast Range province is divisible into the Olympic Mountains of northwest Washington, the Willapa Hills and Black Hills of southwest Washington, and the Oregon Coast Range uplift south of the Columbia River (Figs. 1 and 2).

The Coast Range of southwestern Washington contains Cenozoic strata of marine sedimentary and volcanic origin that crop out in each of several present-day structural basins. These basins form the drainage areas of the Chehalis River-Grays Harbor, Willapa River-Willapa Bay, and lower Columbia River. The composite Cenozoic section, measuring nearly 23,000 feet thick, is mapped as fifteen formations and can be subdivided into five unconformity-bounded sequences (Fig. 3). Coeval units known by different names occur in and adjacent to the study area and are discussed in the section titled "Stratigraphy". Recent literature, including maps and detailed stratigraphic descriptions, are listed under "References Cited". Maps and stratigraphic studies of particular note include Henriksen (1956), Pease and Hoover (1957), Rau (1958, 1967), Snavely and others (l958), Wells (1979, 1981), and Wells and others (1983). Summaries of the Cenozoic geologic history of western Oregon and Washington are available in Snavely and Wagner (1963) and McKee (1972), and for western North America as a whole in Nilsen and McKee (1979) and Cole and Armentrout (1979). A review of current concepts of plate motion and volcanotectonic evolution of western Oregon and Washington is provided by Wells and others (1984) and Wells and Coe (1985).

STRATIGRAPHY

Cenozoic rocks of the Centralia, Chehalis, and Grays Harbor area in southwestern Washington have a measured composite thickness of from 10,000 to 23,000 feet. Figure 2 is a simplified geologic map of the study area. General features include:

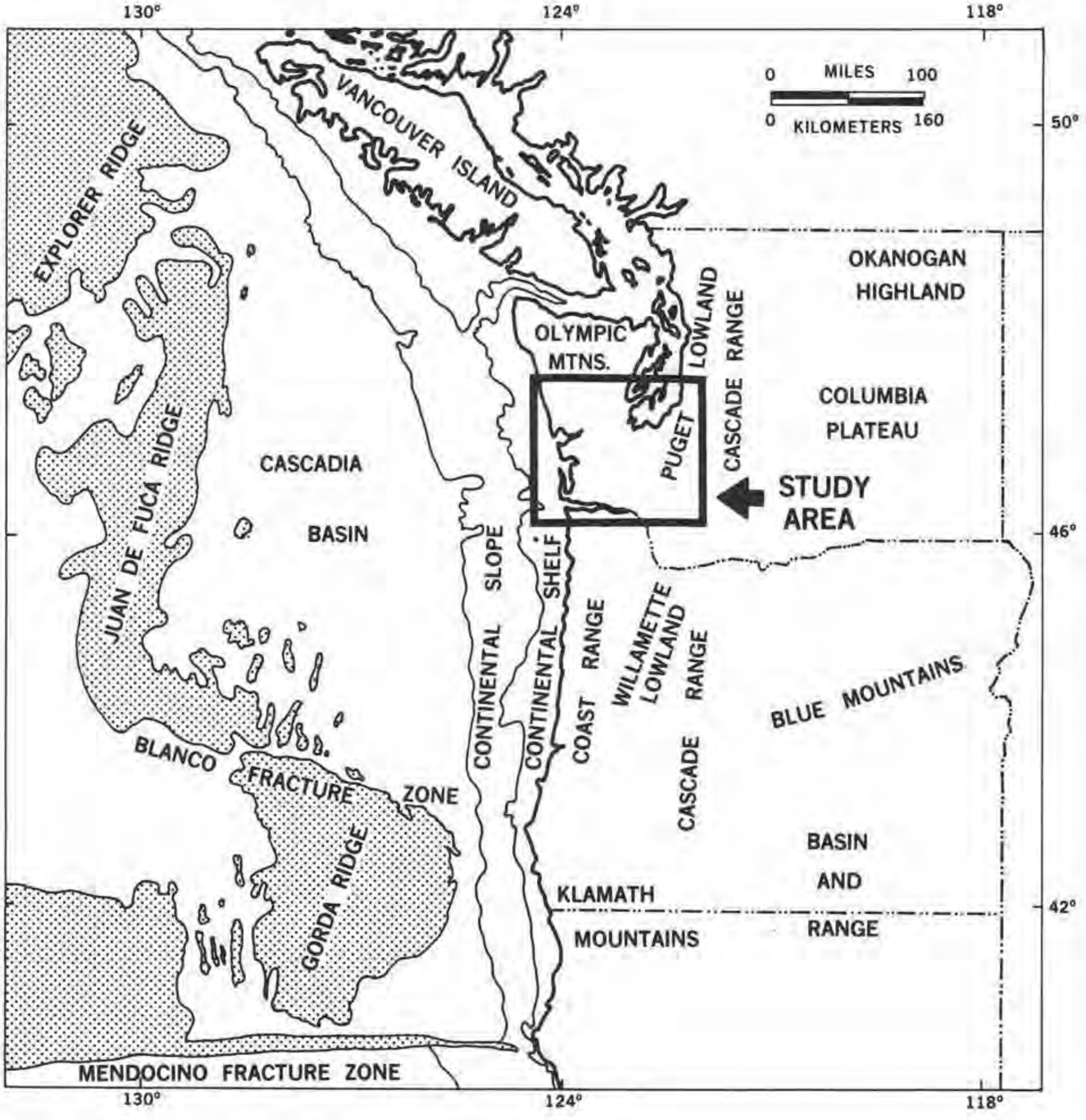

Figure 1.–Location map and geomorphic provinces of the study area (modified from McKee, 1972).

1. Upland areas cored by early to middle Eocene basalt of the Crescent Formation.
2. Lowlands underlain by middle Eocene to middle Miocene marine and nonmarine sedimentary rocks of the Cowlitz, Lincoln Creek, and Astoria (?) Formations, with associated Eocene igneous rocks of the "Cowlitz volcanics," Northcraft Formation, Goble Volcanics, and basalts of the Miocene Columbia River Basalt Group.
3. Late Miocene and Pliocene rocks restricted to structural downwarps that form the drainage pathways flowing into the Grays Harbor, Willapa Bay, and lower Columbia River embayments. Facies of the Montesano and Quinault Formations typify this embayment phase of sedimentation.

Not shown are Pliocene and Pleistocene terrace deposits formed along river valleys and coastal embayments. Principal Cenozoic rock units include: Crescent

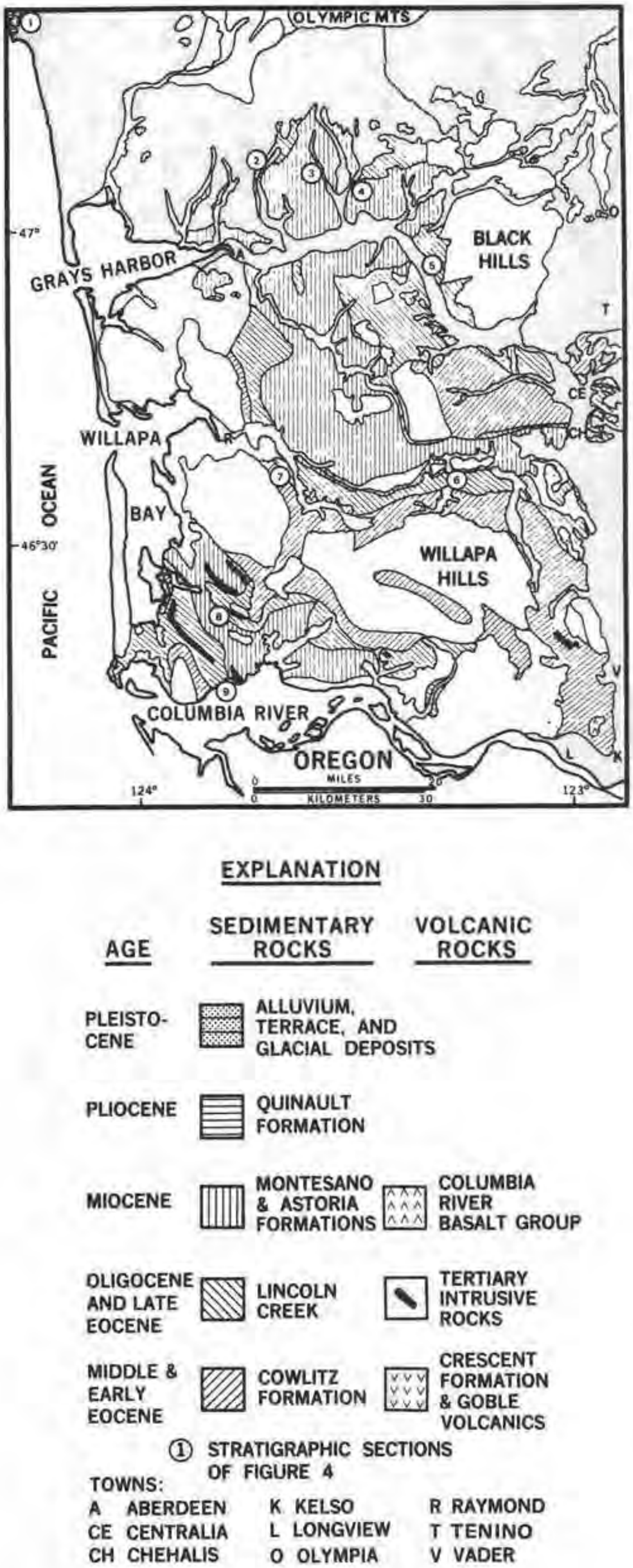

Figure 2.–Simplified geologic map of southwest Washington with location of stratigraphic sections shown on Figure 4 (modified from Huntting and others, 1961).

Formation, Cowlitz Formation, McIntosh Formation, Northcraft Formation, "Cowlitz volcanics", Skookumchuck Formation, Goble Volcanics, Lincoln Creek Formation, Astoria (?) Formation, Columbia River Basalt Group, and Montesano Formation (Fig. 3). Fluvial and glacial-fluvial deposits of the Troutdale Formation, Logan Hill Formation, terrace deposits, Vashon Drift, and alluvium are included under "alluvium and glacial drift" in Figure 3. The strata are dominantly marine sedimentary and igneous rocks that grade eastward into nonmarine sedimentary and volcanic rocks. Age relations of these and coeval units of the Oregon-Washington Coast Range are reviewed in Armentrout (1981) and Armentrout and others (1983) with slight modifications on the Eocene-Oligocene boundary as suggested by Prothero and Armentrout (1985). Wells (1979, 1981) and Wells and others (1983) mapped the area along the Columbia River in southwestern Washington and present brief descriptions of the rock units in that area. A simplified correlation chart of stratigraphic sections (Fig. 4) and a summary chart of Cenozoic sedimentation patterns and events (Fig. 5) help define stratigraphic relations. These figures are discussed in detail under "Unconformity-bounded Sequences".

Crescent Formation

The oldest rocks in the study area are assigned to the Crescent Formation of late Paleocene to middle Eocene age. Crescent Formation rocks are typically aphanitic to porphyritic augite-rich submarine tholeiitic basalts interbedded with and overlain by marine sandstone and siltstone. The type area for the formation is Crescent Bay along the northern shore of the Olympic Peninsula (Arnold, 1906). Crescent Formation basalts underlie the topographic highs of the Willapa Hills and Black Hills of southwestern Washington (Fig. 2). The basalts are commonly zeolitized (Snavely and others, 1958).

Locally marine sedimentary rocks are interbedded with and overlying volcanic rocks of the Crescent Formation. These rocks consist of interbedded sandstone and siltstone, largely of turbidite origin, and include units mapped as "rhythmically bedded sandstone and siltstone" of Wagner (1967) along the Willapa River; "Unit A" of Wolfe and McKee (1968) along the Grays River; and "sandstone of Megler" (Wells, 1979) along the southwest flank of the Willapa Hills. These units are assigned to the Ulatisian Foraminiferal Stage (W. W. Rau, personal commun., 1986).

The Crescent Formation basalts are part of a major, upper Paleocene to middle Eocene volcanic sequence that extends from Vancouver Island on the

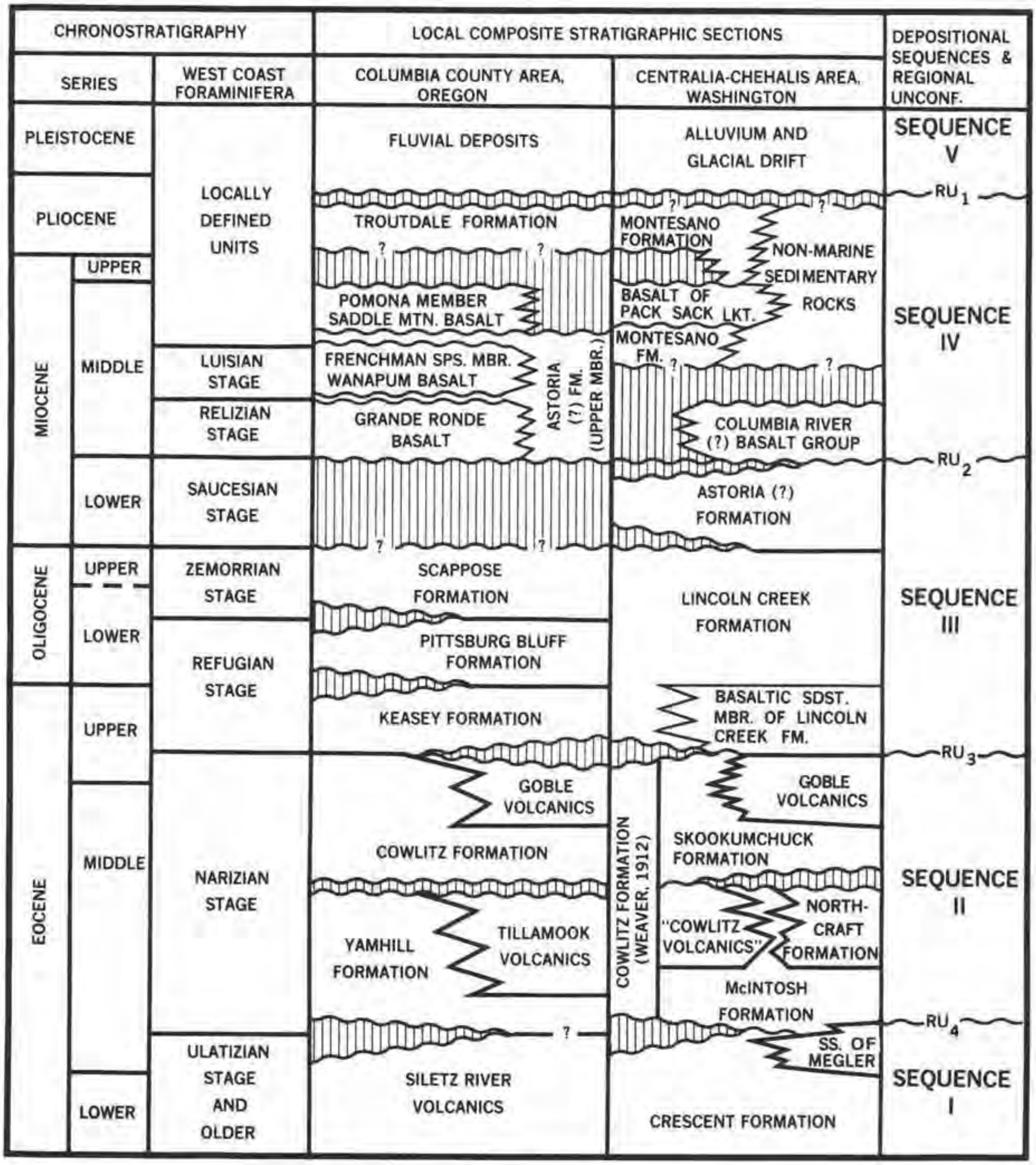

Figure 3.–Composite stratigraphic correlation chart for northwest Oregon and southwest Washington (compiled from Armentrout and others [1983] with modifications discussed in text). The two sections shown border the study area on the southwest and northeast and together include the principal stratigraphic units discussed in the text. RU, regional unconformity.

north to the flanks of the Klamath Mountains in southern Oregon (Figs. 1 and 6). This sequence includes at least 60,000 cubic miles of flood basalts that erupted onto the sea floor from fissures and vents, either as flows or extrusive breccia (Snavely and Wagner, 1963; Snavely and others, 1968; Cady, 1975). Local volcanic centers appear to have become subaerial toward the end of middle Eocene time. The base of the Crescent Formation is not exposed in southwest Washington. See Cady (1975) for a discussion of basal relationships of the Crescent Formation in the Olympic Peninsula.

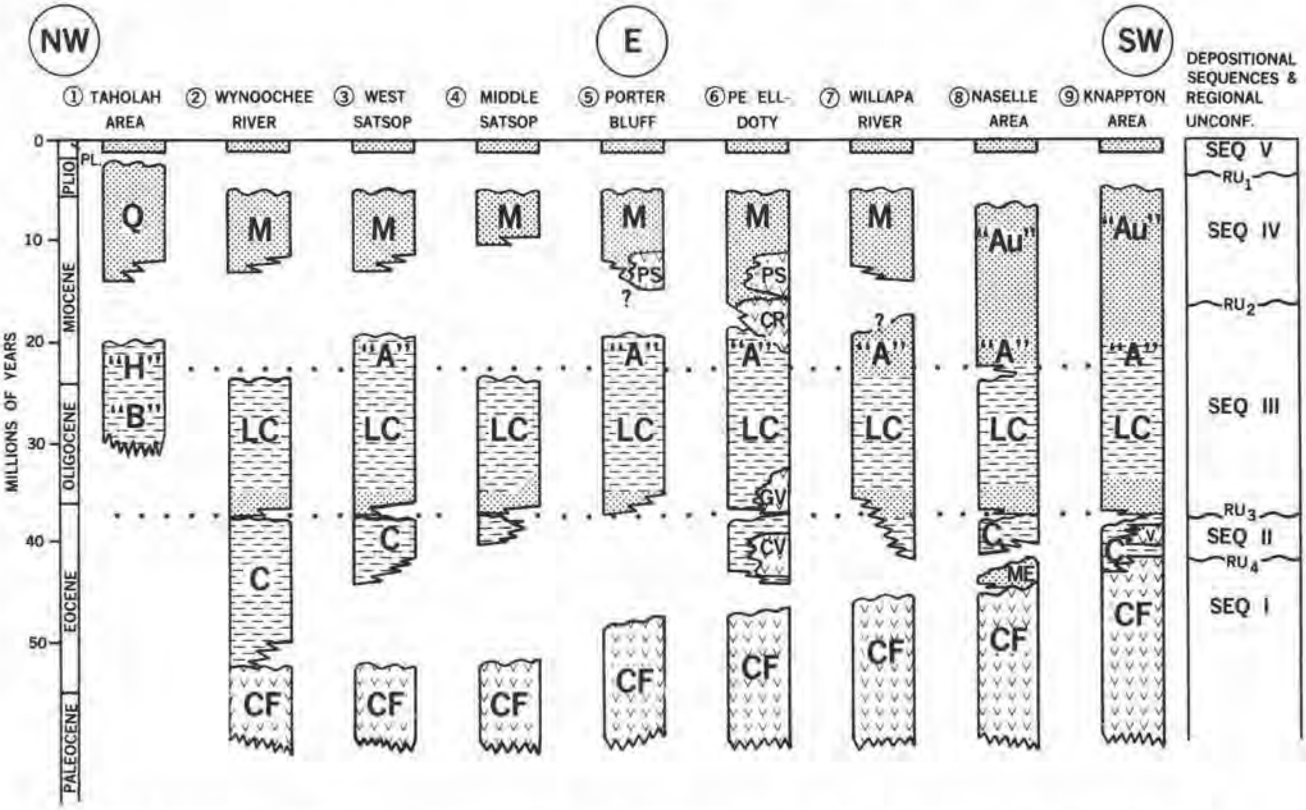

Figure 4.–Correlation of simplified stratigraphic sections in southwest Washington (compiled from Rau, 1966, 1967, 1970; Armentrout, 1973b, 1975; Wells, 1979, 1981). Section locations are shown on Figure 2. The two rows of dots are to aid visual correlation. The stratigraphy represents five depositional sequences separated by four regional unconformities. Formational units are CF, Crescent Formation; ME, sandstone of Megler; C, Cowlitz Formation; CV, "Cowlitz volcanics"; GV, Goble Volcanics; LC, Lincoln Creek Formation; "B", Blakeley equivalent; "H", Hoh complex; "A", Astoria (?) Formation; "Au", upper Astoria (?) Formation; CR, Columbia River (?) Basalt Group; Q, Quinault Formation; M, Montesano Formation; PS, basalt of Pack Sack Lookout. Patterns represent dominant rock type: dots for sandstone, dashes for mudstone, and v's for volcanic rocks.

Volcanic rocks that correlate with the Crescent Formation include the Crescent (?) Formation in Washington (Pease and Hoover, 1957; Snavely and others, 1958), the Metchosin Volcanics of Vancouver Island (Clapp, 1917), and the Siletz River Volcanic Series (Snavely and Baldwin, 1948; Snavely and others, 1968) and Roseburg Formation volcanic rocks of Oregon (Baldwin, 1976) (Fig. 6). Duncan (1982) suggests that these upper Paleocene to middle Eocene igneous rocks are an accreted seamount chain extruded from a hot spot on or near a spreading ridge. Alternatively, Wells and others (1984) suggest these rocks were episodically generated along leaking fractures and transform faults during changes in seafloor spreading directions and then accreted as part of the accretionary prism of western Oregon and Washington.

The late Paleocene to middle Eocene age of the Crescent Formation has been determined from fossils recovered from interbedded and overlying sedimentary rocks (Rau, 1964, 1966, 1981) and from radiometric ages on the basalts that range from 54 to 46 m.y.B.P. (Duncan, 1982; Armentrout and others, 1983). Paleomagnetism studies of Crescent Formation volcanic rocks in southwest Washington suggest that structurally separate domains are characterized by clockwise rotation, the oldest showing a 65° clockwise rotation diminishing to a 20° clockwise rotation for the youngest, when compared to the Eocene reference pole of North America (Wells and Coe, 1985) (See also Fig. 12.)

Cowlitz Formation

Arnold (1906) referred a molluscan fauna from the area around Vader, Washington, to the Eocene and, using California nomenclature, called it a "Tejon correlative". The strata containing that fauna were subsequently named Cowlitz Formation by Weaver (1912). The formation has never been adequately

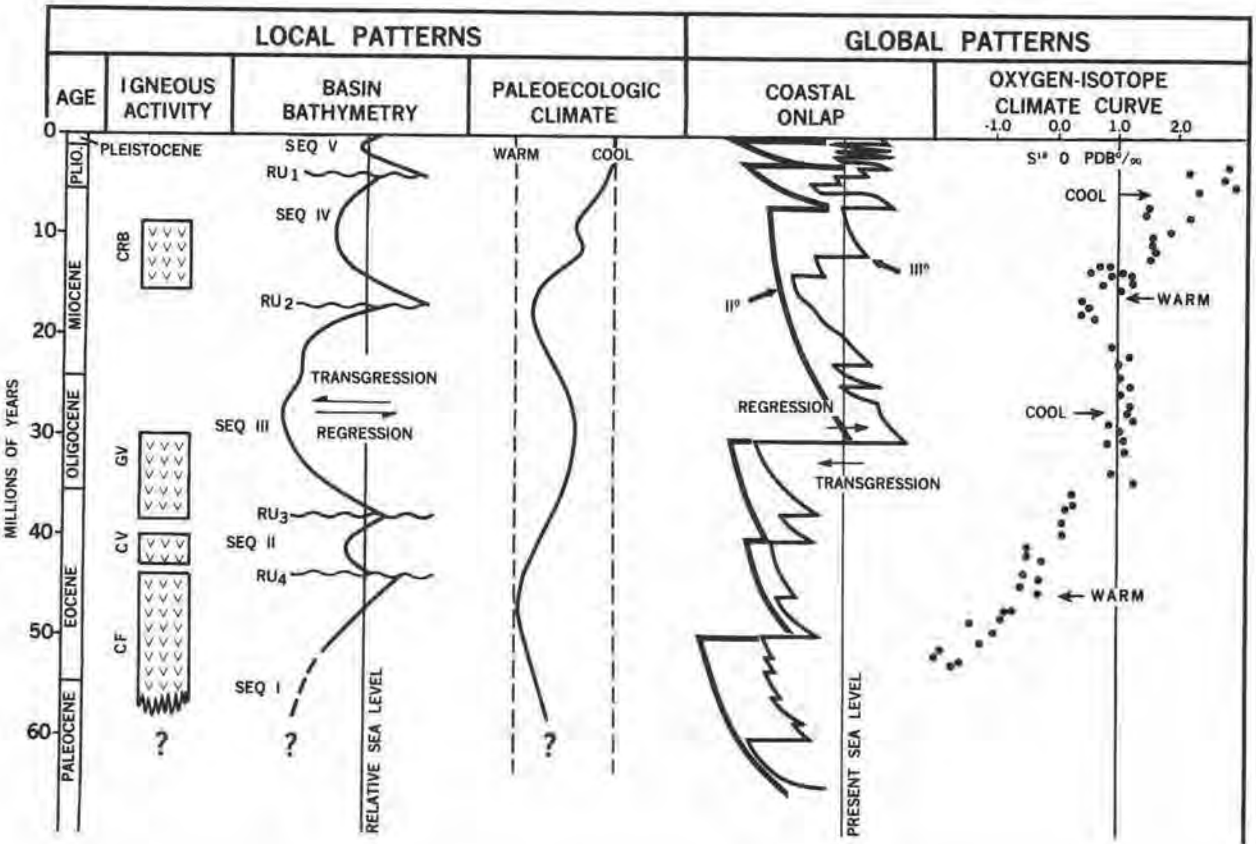

Figure 5.–Cenozoic patterns of igneous activity, bathymetry, and climate for southwest Washington compared with global coastal onlap and oxygen-isotope climate curves. Igneous activity is summarized in Figure 6. Igneous units include: CF, Crescent Formation and equivalents; CV, "Cowlitz volcanics" and equivalents; GV, Goble Volcanics and equivalents; CRB, Columbia River Basalt Group. Basin bathymetry is derived from lithofacies analysis and paleoecologic studies (See Figs. 7 and 8.). Climate is interpreted from paleoecologic studies; see text discussion, "Sequence Patterns". SEQ, unconformity-bounded sequence; RU, regional unconformity. Global patterns are from Vail and others (1977) for coastal onlap and Shakelton and Kennett (1975) for oxygen isotopes.

defined (Livingston, 1966), but current usage follows that of the geologic map of Washington (Weaver, 1937) where the post-Crescent Formation and pre-Lincoln Creek Formation unconformity-bounded sequence is referred to the Cowlitz Formation (Snavely and others, 1958).

The type area of the Cowlitz Formation is along Olequa and Stillwater Creeks and the Cowlitz River near Vader. The maximum thickness of the Cowlitz Formation is approximately 9,500 feet. The unit is predominantly siltstone with sandstone beds locally well developed; marine and nonmarine sedimentary and interbedded igneous rocks also occur in the type area (Henriksen, 1956). Molluscan fossils from the type section of the Cowlitz Formation have been studied by Weaver (1912), Van Winkle (1918), and Nesbitt (1982); Foraminifera have been studied by Beck (1943), Rau (1958), and McDougall (*in* Armentrout and others, 1980). Both faunas indicate a middle to early late Eocene age and a warm subtropical environment (Armentrout and others, 1980, 1983).

The Cowlitz Formation, as used by Weaver (1912, 1916a, 1937), includes several distinct sedimentary formations of present usage: the McIntosh and Skookumchuck Formations of the Centralia-Chehalis area (Snavely and others, 1951a,b, 1958); the "sedimentary rocks of Late Eocene age" in the Satsop River area (Rau, 1966, 1967) now named the Humptulips Formation (Rau, 1984); the McIntosh Formation as mapped along the Willapa River (Wagner, 1967); and the "siltstone of Skamokawa Creek" on the south flank of the Willapa Hills (Wells, 1982). These Eocene units represent deltaic environments to the east (Skookumchuck) and open marine environ-

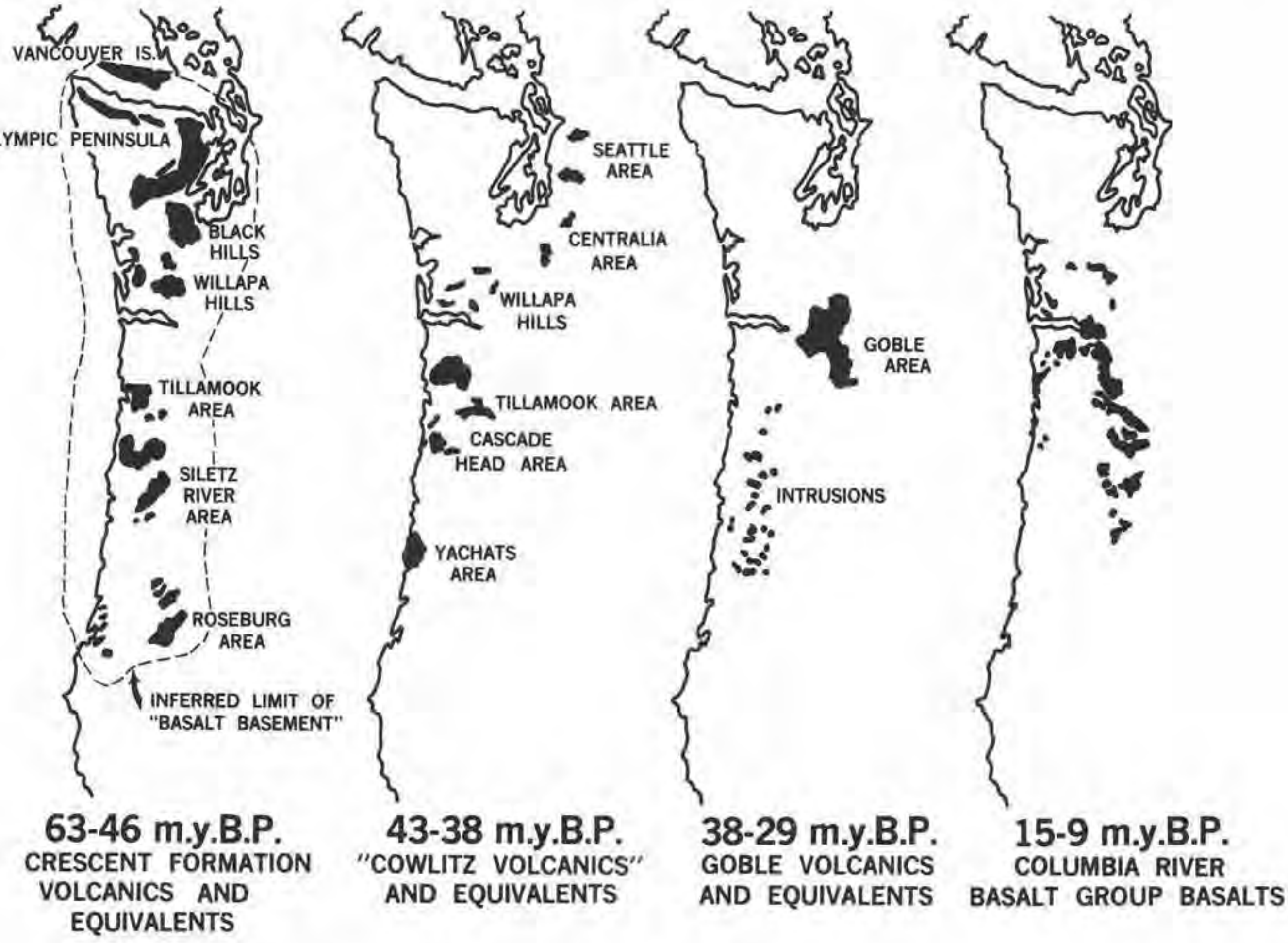

Figure 6.–Simplified map showing the distribution of Oregon and Washington Coast Range igneous outcrops and their age range (modified from Duncan, 1982, fig. 2). See text discussion, "Stratigraphy", for each igneous sequence.

ments to the northwest and west (Humptulips and McIntosh).

The following discussion of McIntosh, Northcraft, and Skookumchuck Formations and the informally named "Cowlitz volcanics" provides an introduction to the lithofacies recognized in southwestern Washington that are correlative with the Cowlitz Formation of Weaver (1912, 1916a, 1937). Regional sedimentology related to these units is described by Buckovic (1979).

McIntosh Formation

The McIntosh Formation is a middle Eocene sequence of tuffaceous marine sedimentary rocks and local porphyritic basalt flows. The formation was named by Snavely and others (1951a) from road cuts along the south shore of McIntosh Lake. The type area is the axial part of the Crawford Mountain anticline east of Tenino. Maximum measured thicknesses of the McIntosh Formation are 5,000 feet. The middle McIntosh siltstone and claystone are rich in Foraminifera that indicate a middle Eocene age (Rau, 1956; Armentrout and others, 1983). The lower and upper parts of the McIntosh Formation include nearshore sequences of basaltic and arkosic sandstone, respectively, and grade westward into offshore deep-water deposits of marine shale.

The McIntosh Formation unconformably overlies the Crescent Formation along coeval structural and paleotopographic highs and is gradationally overlain by or locally interbedded with the Northcraft and Skookumchuck Formations and various units discussed here as the "Cowlitz volcanics". The McIntosh Formation is correlative with parts of the Aldwell, Lyre, and lower Hoko Formations of the Olympic Peninsula (Rau, 1964; Snavely and others, 1977), with the middle to late Eocene units of the Humptulips Formation in the Satsop River area (Rau, 1966, 1984) and along the Willapa River (Rau, 1951; Wagner, 1967). In Oregon, the Coaledo and Yamhill Formations are correlative with the McIntosh Formation (Snavely and others, 1969; Armentrout and others, 1983).

Northcraft Formation and "Cowlitz volcanics"

Field mapping augmented by major element geochemistry permits differentiation of two suites of middle to early late Eocene volcanic rocks interbedded with the Cowlitz Formation and age-equivalent rocks in southwest Washington. Basalts that have high titanium oxide contents are present in the Willapa Hills area and here are referred to as the "Cowlitz volcanics"; andesites and basaltic andesites having low titanium oxide contents occur along the eastern margin of the study area and are mapped as the Northcraft Formation (Fig. 6). Mapped patterns and geochemistry suggest that the more western outcrops of "Cowlitz volcanics" are of oceanic affinity and the more eastern Northcraft rocks are of magmatic arc affinity.

Geochemistry reported by Wolfe and McKee (1972), Beck and Burr (1979), McElwee and others (1985), Phillips and Kaler (1985) and W. M. Phillips (personal commun., 1986) demonstrates that the basalts herein referred to as the "Cowlitz volcanics" have a high titanium oxide content and include units previously mapped as: "basaltic volcanics" of the Cowlitz Formation by Weaver (1937); "Goble volcanics member" of the Cowlitz Formation by Henriksen (1956); "Astoria basalts" immediately west of Olequa Creek by Henriksen (1956); and "Unit B volcanics" of Wolfe and McKee (1972) mapped by Wells (1981) as "Goble volcanics" of Livingston (1966). The "Cowlitz volcanics" are both geochemically distinct from and older than the Goble Volcanics as used in this study.

The Northcraft Formation includes a broad spectrum of igneous and sedimentary rock types. Preliminary major element geochemistry suggests the andesites and basalts are low in titanium oxide (W. M. Phillips, personal commun., 1986). The formation was named from typical exposures near Northcraft Mountain on the southwest flank of Crawford Mountain anticline near Tenino (Snavely and others, 1951b). The Northcraft Formation gradationally overlies or is interbedded with the McIntosh Formation and is overlain by the Skookumchuck Formation with apparent local angular unconformity. Those parts of the Northcraft Formation that are interbedded with fossiliferous sediments of the McIntosh Formation permit assignment of a middle Eocene age (Snavely and others, 1958; Rau, 1958). To the east, the Northcraft Formation appears to be as young as late Eocene where it is interbedded with sandstones of the Skookumchuck Formation and may be gradational with volcaniclastic rocks of the late Eocene lower Ohanapecosh Formation of the ancestral Cascade Mountains (W. M. Phillips, personal commun., 1986).

The Northcraft Formation consists chiefly of andesite and basalt flow rocks with pyroclastic rocks in the upper part and basaltic conglomerate, sandstone, and pyroclastic material in the lower part. In the vicinity of Northcraft Mountain, the formation is nearly 1,500 feet thick and is predominantly basalt and andesite. To the west and south of Northcraft Mountain, this formation is thinner and is mostly fine-grained pyroclastic rocks (Snavely and others, 1958).

The Northcraft Formation and "Cowlitz volcanics" are local middle to late Eocene volcanic centers and are coeval with other volcanic centers throughout the Coast Range Basin province of western Oregon and Washington (Duncan, 1982; Armentrout and others, 1983). Correlative units include the Yachats Basalt of the central Oregon coast (Snavely and MacLeod,1974), the Tillamook Volcanics of northwest Oregon (Warren and others, 1945; Magill and others, 1981), and the Tukwila (Waldron, 1962; Vine, 1969) and Northcraft (Gard, 1968) Formations in the Seattle area, Washington.

Skookumchuck Formation

The Skookumchuck Formation is typified by interbedded shallow marine and continental facies. The formation is defined by Snavely and others (1951b), and the type section is along the Skookumchuck River. The Skookumchuck Formation may be as much as 3,500 feet thick in the Centralia-Chehalis area. Lower basaltic and upper arkosic sandstone members are separated by carbonaceous siltstones that thicken westward. Tuffaceous coals are abundant in the lower and upper members. The coal ranks from lignite to subbituminous B; most of the coal is subbituminous C (Snavely and others, 1958).

The Skookumchuck Formation overlies the Northcraft Formation with local angularity. To the west, the underlying siltstones of the McIntosh Formation grade upward into siltstones of the Skookumchuck Formation. The Skookumchuck Formation is conformably overlain by the Lincoln Creek Formation in structural lows and unconformably overlain by the Lincoln Creek Formation over structural highs developed during early late Eocene deformation.

Molluscan and foraminiferal fossils of the Skookumchuck Formation indicate a middle to early late Eocene age (Snavely and others, 1958; Armentrout and others, 1983). Correlative rocks include the upper part of the type section of the Cowlitz Formation of Weaver (1912); the Hoko River Formation of the Twin River Group on the Olympic Peninsula (Rau, 1964; Snavely and others, 1977); and the "sedimentary rocks of Late Eocene age" of Rau (1966) now named the Humptulips Formation (Rau, 1984) in the Grays Harbor basin of Washington and the "siltstone of Skamokawa Creek" (Wells, 1981, 1982) along the north side of the Columbia River. In Oregon, correlative units include the Cowlitz Formation of northwestern Oregon (Niem and Van Atta, 1973; Niem and Niem, 1985), the Spencer Formation of the Willamette basin (Baldwin, 1964), the Nestucca Formation of the central Coast Range (Snavely and Vokes, 1949), and the Coaledo Formation of the Coos Bay basin (Allen and Baldwin, 1944).

Goble Volcanics

A thick sequence of late Eocene and early Oligocene basaltic andesite, andesite, and dacite flow and pyroclastic rocks is interbedded with the Paleogene marine sedimentary rocks on both sides of the Columbia River in southwestern Washington and northwestern Oregon (Fig. 6). The volcanic rocks were mapped as Goble Volcanics (Warren and others, 1945). The type area is in the vicinity of Goble, Oregon, where the unit's thickness is more than 5,000 feet (Wilkinson and others, 1946). Both the upper and lower contacts of the Goble Volcanics are transitional, interfingering with the Eocene sedimentary rocks of the uppermost Cowlitz Formation and latest Eocene to Oligocene rocks of the Lincoln Creek and Toutle Formations (Livingston, 1966; Armentrout, 1975). Radiometric ages from Goble outcrop samples in southwest Washington range from 45 to 32 m.y.B.P., which spans middle Eocene to early Oligocene (Beck and Burr, 1979; Wells and Coe, 1985). Lithostratigraphic and biostratigraphic correlations suggest that the Goble Volcanics are no older than about 40 m.y. (coeval with and younger than upper Cowlitz Formation = upper Skookumchuck Formation, which contains a tuff dated at 40 m.y.B.P. [D. M. Triplehorn, written commun., 1985]). Phillips and Kaler (1985) mapped the Goble Volcanics and demonstrated that the unit is both geographically continuous with and geochemically similar to the Hatchet Mountain Formation of Roberts (1958). The Hatchet Mountain Formation sits unconformably above the Cowlitz Formation and below the Toutle Formation (Roberts, 1958), which contains an andesite flow dated at 35 m.y.B.P. (Phillips and Kaler, 1985).

Paleomagnetic studies of Goble Volcanics indicate that the mean direction of remanent magnetism is about 23° to 25° east of the expected mid-Tertiary geomagnetic field direction for the area (Beck and Burr, 1979; Wells and Coe, 1985). Paleomagnetic rotation data are graphically displayed in Figure 10.

Lincoln Creek Formation

From 2,000 to 9,000 feet of sedimentary beds are assigned to the Lincoln Creek Formation. The type section of the formation, originally named the Lincoln Formation, was established along the Chehalis River near Lincoln Creek west of Centralia, and was considered to be of middle Oligocene age (Weaver, 1912, 1916c, 1937). Subsequently, the formation was redefined (Weaver and others, 1944) to include a composite of many sections along the Chehalis River between Centralia and Porter. This formation was further redefined and renamed the Lincoln Creek Formation by Beikman and others (1967). It is now recognized as late Eocene through Oligocene in age (Armentrout and others, 1983; Prothero and Armentrout, 1985).

Three rock types express different depositional environments in the Lincoln Creek Formation. Along the easternmost margin of the exposures of the formation are sandstone and conglomerate strandline and continental deposits of the Lincoln Creek Formation of Snavely and others (1958) and correlative beds of the Toutle Formation (Roberts, 1958). These rocks grade westward into marine glauconitic basaltic sandstone that is 1,500 feet thick just east of Centralia (Snavely and others, 1958). Farther westward, the sandstone becomes thinner but persists in all sections studied. The basaltic sandstone grades upward to fine-grained tuffaceous sandstone and siltstone. This gradation reflects both westward-deepening depositional environments and overall basin subsidence and consequent transgression.

The Lincoln Creek Formation is thickest toward the western margin of the outcrop area in basin deeps where it is conformable with underlying marine units of the Cowlitz Formation. The formation thins locally and unconformably onlaps highs of older Eocene rocks that now form the Doty Hills, Black Hills, and Minot Peak (Pease and Hoover, 1957). The unconformity at the base of the Lincoln Creek Formation is coeval with the unconformities at the base of the Cascade arc volcanic sequence and the volcanogenic John Day Formation of the Columbia River Plateau (Wells and others, 1984).

The Lincoln Creek Formation molluscan (Armentrout, 1973b, 1975) and foraminiferal faunas (Frizzell, 1937; Mumby, 1959; Rau, 1948a, 1958, 1966, 1981; McDougall, 1980) both indicate an age range from late Eocene to latest Oligocene. The Lincoln Creek strata correlate with the Blakeley Formation of the Seattle and Bainbridge Island areas (Tegland, 1933;

Weaver, 1937; Durham, 1944; Fulmer, 1975) and with the Makah Formation to the north on the Olympic Peninsula (Snavely and others, 1980b). Correlatives in northwest Oregon include the Keasey, Pittsburg Bluff, and Scappoose Formations (Armentrout and others, 1983) and the Smuggler Cove and possibly in part the Northrop Creek formations (informally named in Niem and Niem, 1985); in the central Oregon Coast Range the uppermost Nestucca, Alsea, and Yaquina Formations (Snavely and others, 1969; Goodwin, 1973); in the Willamette Valley the Eugene Formation (Hickman, 1969); and in the southern Oregon Coast Range the Tunnel Point Formation (Baldwin and others, 1973; Armentrout, 1980).

Paleomagnetic studies of the Lincoln Creek Formation provide both a magnetostratigraphic record for correlation (Prothero and Armentrout, 1985) (vertical column of Fig. 9) and suggest clockwise rotation of about 15° to 20° (D. R. Prothero, written commun., 1985).

Astoria (?) Formation

The Lincoln Creek Formation is overlain by the Astoria (?) Formation, a dark- to medium-gray micaceous and carbonaceous fine-grained sandstone that generally has a glauconitic sandstone bed(s) at its base in marine facies. Snavely and others (1958) describe the Astoria (?) Formation as including those strata referred to as "basaltic conglomerate" of Miocene (?) age along the North Fork of the Newaukum River mapped by Snavely and others (1951b), lower Astoria Formation of Pease and Hoover (1957) in the Doty-Minot Peak area, and the Astoria Formation of Etherington (1931) in Grays Harbor and Thurston Counties. The question mark after Astoria follows Snavely and others (1958) and Wells (1979, 1981), denoting probable correlation with the type Astoria Formation mapped south of the Columbia River (Niem and Niem, 1985). Astoria (?) Formation and lower Astoria (?) Formation (Wells, 1981) are older than rocks mapped as upper Astoria (?) Formation of Wells (1979, 1981; Wells and others, 1983), which are correlative with the Montesano Formation (Armentrout and others, 1983).

In the Centralia-Chehalis area, where a series of late Oligocene structural highs developed, the Astoria (?) Formation unconformably overlies the Lincoln Creek Formation (Snavely and others, 1958). Farther west, away from the structural highs, the Astoria (?) Formation conformably overlies the Lincoln Creek Formation. The Astoria (?) Formation is unconformably overlain by younger units that differ from place to place. Because of these unconformable relations, the thickness of the Astoria (?) Formation ranges from absent to 5,600 feet. The foraminiferal faunas of the Astoria (?) Formation have been studied by Rau (1948b, 1951, 1958, 1966, 1967), and the molluscan faunas have been studied by Etherington (1931), Moore (1963), Strong (1967), Armentrout (1973b), and Addicott (1976). All faunas suggest early to middle Miocene age assignments (Addicott, 1976; Armentrout and others, 1983).

The Astoria (?) Formation correlates with the Clallam Formation of the northern Olympic Peninsula (Rau, 1964), the Cape Falcon conglomerate and Astoria Formation of northwest Oregon (Niem and Niem, 1985), and the Nye Mudstone and Astoria Formation of the Newport embayment, Oregon (Snavely and others, 1969).

Columbia River Basalt Group

In southwestern Washington, tholeiitic basalts rest unconformably upon sedimentary rocks of the Astoria (?) and Lincoln Creek Formations and are overlain by beds ranging in age from late Miocene to Holocene. These basalts are correlated with members of the Columbia River Basalt Group. Flow units of the Columbia River Basalt Group were extruded in eastern and central Washington and flowed westward, where they became interbedded with marine units along the coastline (Pease and Hoover, 1957; Snavely and others, 1958; Beeson and Moran, 1979; Beeson and others, 1979). Units of the Columbia River Basalt Group in western Washington include Grande Ronde Basalt locally known as the Columbia River (?) Basalt near Doty (Snavely and others, 1958; Turner, 1970), the Frenchman Springs Member of the Wanapum Basalt along the Columbia River (Wells, 1979, 1981), and Pomona Member of the Saddle Mountains Basalt locally known as the "basalt of Pack Sack Lookout" (Snavely and others, 1973; Wells, 1981; Magill and others, 1982). Radiometric dates on these rocks indicate an age of 15 to 12 m.y.B.P. (Magill and others, 1982; Armentrout and others, 1983).

Paleomagnetic studies of the "basalt of Pack Sack Lookout" suggest clockwise rotation of about 16° with respect to stable North America (Magill and Cox, 1980; Magill and others, 1982) (Figs. 10 and 12).

Montesano Formation

A sequence of interbedded fluvial, lacustrine, brackish-water, and shallow-marine deposits overlies the Astoria (?) Formation and is interbedded with those flows of the Columbia River Basalt Group cropping out along the Columbia River between Portland, Oregon, and the mouth of the Columbia River. Contacts are conformable along synclinal axes and unconformable on structural highs where rocks of the Montesano Formation overlie rocks as old as the Lincoln Creek Formation. The continental deposits are the Wilkes Formation (Roberts, 1958, and "unnamed nonmarine" rocks of Snavely and others, 1958), whereas the marine units are mapped either as Montesano Formation (Weaver, 1912, 1937; Etherington, 1931; Rau, 1966, 1967) or as upper Astoria (?) Formation (Pease and Hoover, 1957; Wells, 1979, 1981). The Montesano Formation varies locally in thickness, but approaches a maximum thickness of 3,000 feet in the Wynoochee River area (Rau, 1967).

The marine deposits of the Montesano Formation have been dated with molluscs as middle and late Miocene (Weaver, 1912, 1937; Weaver and others, 1944; Addicott, 1976), with foraminifers as late middle and late Miocene (Rau's 1981 "undifferentiated assemblage of the Montesano Formation"; Fowler, 1965; Rau, 1966, 1967; Bergen and Bird, 1972), and with diatoms as early late Miocene (Barron, 1981). Correlative units in Washington include a lower part of the Quinault Formation of the Olympic Peninsula (Rau, 1970; Addicott, 1976) and the upper member of the Astoria (?) Formation along the Columbia River (Wells, 1981). In Oregon, rock units that correlate with the Montesano Formation include the Empire Formation of Coos Bay (Baldwin and others, 1973; Barron and Armentrout, 1980), and the Gnat Creek formation along the Columbia River in Clatsop County (Niem and Niem, 1985).

Troutdale Formation

The name Troutdale was first used by Hodge in 1933 and was formally proposed by him in 1938 to describe conglomerate and sandstone beds that crop out near Troutdale, Oregon. Semiconsolidated gravel and sand correlative with the Troutdale Formation occur in restricted areas along the lower Columbia River in the Longview and Cathlamet areas of Washington (Livingston, 1966). In that area, the Troutdale Formation unconformably overlies the late Eocene to early Oligocene Goble Volcanics. The interbedded well-rounded cobbles, lens-shaped gravel beds, and friable sandstone and mudstone indicate a fluvial origin for this formation. Thicknesses range from 60 to 900 feet. Compositions of the pebbles suggest derivation from quartzitic and metamorphic rock sources, acidic volcanic terrain such as occurs in northeastern Washington or eastern Oregon and Idaho, as well as basaltic andesites of the Cascade Range and basalts of the Coast Range. The restricted occurrence of Troutdale Formation deposits along the Columbia River and the clast types suggest that these fluvial deposits were probably associated with an ancestral Columbia River system.

Fossil leaves from the Troutdale Formation (Trimble, 1963) indicate an early Pliocene age (Livingston, 1966). Tolan and Beeson (1984) have expanded the concept of the formation to include rocks of Miocene to Pleistocene age.

Logan Hill Formation

The Logan Hill Formation is composed mainly of reddish- to yellowish-brown, iron-stained gravel and minor amounts of interbedded sand and silty clay and is as much as 200 feet thick. The formation was named by Snavely and others (1951b) in the Centralia-Chehalis area, where the deposits form flat upland surfaces. The Logan Hill Formation is interpreted as glaciofluvial in origin, the primary sediment source being alpine glaciers to the east. The depositional surface of the Logan Hill Formation has an altitude of approximately 1,000 feet in the eastern outcrop area and decreases uniformly westward to about 350 to 400 feet just west of Centralia (Snavely and others, 1958).

The Logan Hill Formation is considered to be early Pleistocene in age, a product of outwash from the alpine valley glaciers of the western Cascade Mountains (Snavely and others, 1958; Armentrout and others, 1983).

Terrace Deposits

Unconsolidated gravel and sand of glaciofluvial origin form terraces along many of the river and stream valleys of southwestern Washington. The terraces are highly dissected and in most places are difficult to map separately from alluvium (Snavely and others, 1958). The pebbles and cobbles of the terrace deposits are principally porphyritic andesite and basalt derived from the Northcraft Formation, but some cobbles of middle Eocene sedimentary rocks, andesite, and weathered rocks reworked from the Logan Hill Formation are also present. The terrace deposits are assigned to the Pleistocene, based on vertebrate fossils (Roberts, 1958; Livingston, 1966). Marine coastal terraces of Pleistocene age along the Pacific Coast of southwest Washington consist of both beach and bay deposits (Florer, 1972; Clifton and Phillips, 1980).

Vashon Drift

The Tenino Prairie, between Olympia and Centralia, is a terrace surface on which the Vashon Drift was deposited. The Vashon Drift is a glacial deposit of morainal and outwash silt, sand, and gravel laid down by the Puget Lobe of the Cordilleran Ice Sheet. The

Puget Lobe moved southward through the Puget Sound area at least four times during the late Pleistocene. The maximum advance of the Puget Lobe, between 15,000 and 13,500 years ago (Vashon Stade of Fraser Drift) reached an area just north of Centralia and rode up to a maximum altitude of about 1,100 feet along the volcanic highlands extending from Tenino eastward (Armstrong and others, 1965; Easterbrook, 1969).

The Vashon Drift is generally less than 50 feet thick (Snavely and others, 1958). Vashon till is light bluish gray and has a matrix of well-compacted silt and clay containing subangular to rounded cobbles and boulders as much as 8 to 10 feet in diameter. Outwash sand and gravel are composed of only slightly weathered igneous and metamorphic rock fragments. Outwash deposits are massive to well bedded. Cross bedding and foreset bedding are common. Vashon gravels have three sources: (1) exotic igneous and metamorphic rocks from far to the north, carried in by the Puget Lobe; (2) locally derived igneous rocks; and (3) igneous rocks from the Cascade Mountains carried westward into the Puget Lowland by alpine glaciation. Locally, the Vashon Drift has been reworked by streams and rivers into a series of terraces. Older glaciofluvial terrace gravels of the lower upper Pleistocene Wingate Hill and Hayden Creek Drifts have been highly dissected (Snavely and others, 1958; Armstrong and others, 1965). They can be distinguished from terraces in the Vashon Drift by the deeper weathering and by iron-stained rinds on the clasts. Clast types of the older terrace gravels do not include the exotic igneous or metamorphic rock types characteristic of Puget Lobe-Vashon Drift sediments. Kettle topography is locally developed, and many of the outwash prairies in the northern part of the area are mounded (Snavely and others, 1958).

Alluvium

Alluvium as used in this paper includes a variety of Holocene sands and gravels associated with river, stream, and valley fills. The maps of Pease and Hoover (1957) and Snavely and others (1958) detail these deposits.

UNCONFORMITY-BOUNDED SEQUENCES

The Cenozoic stratigraphy of southwest Washington (Fig. 4) can be subdivided into five unconformity-bounded sequences (informally termed Sequences I through V) separated by regionally significant unconformities (informally termed RU_1 through RU_4) (Figs. 4 and 5). Formal classification and naming of the sequences and unconformities awaits synthesis of the Cenozoic stratigraphic history of all of Oregon and Washington. Wheeler and Mallory (1963, 1970) provided an earlier study in which they defined regionally significant sequences in western Oregon and Washington.

Sequence I

The lowermost sequence consists of the Crescent Formation oceanic basalts and associated marine sedimentary rocks of late Paleocene and early to late middle Eocene age. Duncan (1982) has dated the igneous rocks of Sequence I at 63 to 46 m.y.B.P. (Fig. 6; also Fig. 15). Cady (1975) describes rocks of the Crescent Formation along the southeastern flank of the Olympic Mountains where deepwater marine basalts having interbeds of red limestone containing planktonic Foraminifera grade upward into shallow marine and locally subaerial volcanic rocks. The shallowing-upward pattern is regionally characteristic of Sequence I (Fig. 5, basin bathymetry curve) and is interpreted as the upbuilding of local volcanic centers, which became subaerial toward the end of middle Eocene time (Snavely and Wagner, 1963; Snavely and others, 1968; Cady, 1975), but the pattern may also include a component of shallowing due to tectonic uplift during accretion. The base of the sequence is not exposed in southwestern Washington, but it probably lies on earliest Cenozoic deep marine sedimentary rocks (Cady, 1975).

Sequence II

Unconformably overlying Sequence I is a depositional cycle of middle to early late Eocene sedimentary rocks represented by the Cowlitz Formation (including open-marine McIntosh and deltaic Skookumchuck facies). The unconformity (RU_4 of Figs. 3, 4, and 5) is most clearly demonstrated along structural uplift and grades laterally to depositional continuity in structural lows (Fig. 4, sections 2 and 9). Regional unconformity RU_4 is of middle Eocene age and separates Sequence I from Sequence II; it is interpreted to be largely the consequence of northeast-southwest compression that formed a northwest-southeast structural grain of folds and faults (Wells and others, 1984).

Sequence III

Sequence III consists of the Lincoln Creek and Astoria (?) Formations and represents a complete transgressive-regressive cycle in which nearshore and nonmarine basin-margin facies along the eastern side of the study area grade westward into deep shelf-margin basin facies. The early late Eocene Regional

Unconformity RU_3 (Fig. 5) at the base of the Lincoln Creek Formation is well developed along structural highs where basal rocks of the Lincoln Creek Formation onlap differentially eroded older Eocene rocks (Pease and Hoover, 1957). Within the mudstone and siltstone strata in basinal depocenters, this unconformity is represented by the influx of basaltic sandstones eroded from uplifted igneous rock highs. Regional Unconformity RU_3 is coeval with the unconformity at the base of the Cascade Range volcanic sequence and is interpreted as marking the onset of ancestral Cascade arc magmatism (Armentrout, 1973a; Wells and others, 1984).

Local volcanic centers of basaltic, andesitic, and dacitic lavas, such as the Goble Volcanics, formed in the late Eocene or early Oligocene forearc basins of western Oregon and Washington between 40 and 32 m.y.B.P. (Fig. 6). These igneous rocks are interbedded with and intruded into sedimentary rocks of Sequence III that contain abundant volcanic ash derived from explosive volcanism in the ancestral Cascade Mountains, and Basin and Range area farther east (Peck and others, 1964; Priest and others, 1982). Intrusive rocks dated as 38 to 29 m.y.B.P. that are present in Oregon are correlatives. (See Fig. 6.) Igneous rocks intruded into late Eocene and Oligocene marine sedimentary rocks are known as far north as Seattle and are probable correlatives of the above units.

Sequence IV

Sequence IV sedimentary rocks deposited in the middle and late Miocene and possibly early Pliocene structural downwarps consist of the Montesano, Quinault, and upper Astoria (?) Formations. The middle Miocene Regional Unconformity RU_2 (Fig. 5) separates Sequence IV rocks from those of Sequence III and can be mapped throughout Oregon and Washington from the continental shelf and Coast Range province eastward through the Cascade Mountains and into the Columbia Plateau (Snavely and others, 1980a; Priest and others, 1982). Erosion that formed this unconformity differentially truncated older sequences. The middle Miocene unconformity formed as a consequence of renewed compressional uplift of the Coast Range province of southwest Washington along northwest-southeast-trending folds, many of which reactivated older structural features (Magill and Cox, 1980). Wells (1981) suggested that this event resulted from the close approach of an ocean-spreading ridge about 16 m.y.B.P. Basalts of the Columbia River Basalt Group ranging in age from 15 to 12 m.y.B.P. (Fig. 6) flowed over this eroded surface and are interbedded with coeval sedimentary rocks of Sequence IV along old river systems and in marine embayments. The basalts consist of the older Columbia River (?) Basalt and younger "basalts of Pack Sack Lookout" now recognized as Grande Ronde Basalt and Pomona Member of the Saddle Mountains Basalt, respectively (Wells, 1981).

The sedimentary rocks of Sequence IV grade from shallow neritic deposits containing abundant molluscan faunas (Snavely and others, 1958; Addicott, 1976) westward to deep shelf-upper slope deposits with abundant foraminiferal assemblages (Fowler, 1965; Bergen and Bird, 1972; Rau, 1970). Less tuffaceous material is present in Sequence IV than in Sequence III as a consequence of the explosive volcanism in the ancestral Cascade Mountains having given way to extrusion of basaltic, andesitic, and dacitic lavas at about 18 m.y. B.P. (Peck and others, 1964; Priest and others, 1982; Wells and others, 1984).

Sequence V

Another compressional pulse (comparable to those forming RU_3 and RU_2) during latest Pliocene through Pleistocene time further deformed Sequences I through IV, displaced the marine shoreline westward to essentially the modern coastline, and resulted in erosion of structural highs forming Regional Unconformity RU_1. Sequence V deposits are restricted to coastal synclines and embayments and along river valleys. They range from marine terrace deposits (Florer, 1972; Clifton and Phillips, 1980) to river terrace and glacial deposits (Snavely and others, 1958; Wolfe and McKee, 1968, 1972), and are largely the consequence of eustatic fluctuation in sea level and flow volumes in glacially fed rivers. Coastal terrace deposits are tilted to the east as a consequence of continued subduction beneath western Oregon and Washington (Adams, 1984). Clear definition of regionally significant depositional trends in this sequence awaits detailed correlations of spatially discontinuous deposits (Wehmiller and others, 1978; Kvenvolden and others, 1979).

Sequence Patterns

The depositional patterns defined by the five unconformity-bounded sequences of southwest Washington are the consequence of interplay between relative sea level and tectonics. Sea level is reflected in patterns of transgression-regression along the basin margin. Tectonics is reflected in rates of basin subsidence or uplift and rates of sediment supply. The depositional patterns for southwestern Washington are summarized by the bathymetry curve of Figure 5.

Patterns of basin bathymetry reflect the water depth of the basin floor as a consequence of subsidence rate, sediment accumulation rate, and global sea-level fluctuations. At basin depth below the influence of eustatic sea-level fluctuations, the absolute water depth increases most significantly when subsidence exceeds sediment accumulation; absolute water depth decreases most significantly if sediment supply exceeds subsidence. The Cenozoic bathymetry curve for southwestern Washington shows basin deepening between the shoaling associated with each regional unconformity (Fig. 5). The bathymetry curve for southwest Washington is derived from lithofacies analyses, as summarized under "Unconformity-Bounded Sequences", and paleoecologic studies of marine faunas. Figures 7 and 8 illustrate the process of constructing the basin bathymetry curve. In Figure 7, the interpreted pattern of basin bathymetry is shown for six stratigraphic sections: three in the basin depocenter and three toward the basin margin. Relative water depth is estimated from the modern depth range of genera represented in the fossil assemblage and on lithofacies patterns ("traditional method" of paleobathymetric interpretations as described by Hickman, 1984). The relative water depth is tentatively calibrated to absolute water depth based on a preliminary use of the "taxonomic structure method" for mollusks (Hickman, 1984) which results in water depth interpretations in agreement with similar analyses using benthic Foraminifera (McDougall, 1980) for some of the same stratigraphic sequences. The local-section curves are synthesized to generate a basin bathymetry curve that reflects the regional pattern in southwest Washington (Fig. 8).

The basin bathymetry curve of Figures 5 and 8 is based on basinal facies in the study area (Fig. 7) that were deposited below water depths directly influenced by eustatic sea-level fluctuations as estimated by Vail and others (1977). Thus, the pattern of basin deepening and shallowing is interpreted as largely a consequence of subsidence exceeding sediment supply. Facies trends indicate initial coastal onlap by marine shelf deposits of the Cowlitz Formation (Buckovic, 1974; Nesbitt, 1982) followed by continued transgression by deeper marine basinal siltstones and shales of the Lincoln Creek Formation (Rau, 1966, 1967; Armentrout, 1973b; Armentrout and others, 1980). Local uplift and shoaling associated with structural growth of anticlines are reflected by (1) progradation and gravity-flow events of slightly coarser clastics into basinal mud-rich sequences, and (2) occurrences of faunas of shallower bathymetric zones (Fig. 7). Onset of regional regression is reflected in the coarsening-upward and shallowing-upward facies of the early to middle Miocene Astoria (?) Formation (Armentrout, 1973b; Armentrout and others, 1980).

Based on sediment thickness and age data (Prothero and Armentrout, 1985), a plot of sediment thickness versus age can be constructed for Sequence III. The plot shows an average sediment accumulation rate of approximately 1,000 feet/m.y. between 38 and 35 m.y.B.P. versus approximately 375 feet/m.y. between 34 and 27 m.y.B.P. (Fig. 9). This change in relative rate of sediment accumulation is reflected in sediment type, with siltstone and interbedded sandstone typical of the older, higher rate, and siltstone with interbedded claystone typical of the younger, lower rate. The 38-to-35-m.y.B.P. interval correlates with the initial basin-deepening phase of Sequence III, suggesting that tectonic steepening of the regional gradient caused slightly coarser sediments to be brought into the deepening basin rather than eustatic sea-level rise with coastal onlap, which would result in coastal entrapment of sediment and sediment starvation of the basin depocenter. The 34-to-27-m.y.B.P. interval correlates with the maximum basin-depth phase of Sequence III, suggesting relative coarse-sediment starvation. Lack of precise age control on the top of Sequence III sections limits confidence in calculations of sediment accumulation rates, but the increase in sand content in the Astoria (?) Formation correlates well with the suggested increase to approximately 1,400 ft/m.y. sediment accumulation rate for that interval (Fig. 9).

Sequence IV rocks are restricted to synclinal troughs formed during middle Miocene northeast-southwest compression. The basin bathymetry curve for this sequence is defined by initial deepening and marine transgression (lower Montesano Formation) followed by rapid progradation of fluvial sediments having an eastern source (Wilkes Formation and "unnamed non-marine" unit) that displaced the shallow marine facies (upper Montesano and Quinault Formations) farther to the west.

Sequence V rocks are not yet adequately correlated to define a precise depositional cycle suitable for correlation to global patterns.

The southwest Washington basin bathymetry curve reflects regionally significant trends that affected basins within the Cenozoic forearc region of Washington and Oregon. The regional curve can be compared with the global coastal onlap curves of Vail and others (1977) (Fig. 5). Global patterns of sea-level rise and fall and coastal onlap cycles have been synthesized by Vail and others (1977) using biostratigraphically calibrated seismic sections; the derived curves represent the consequence of interplay of globally significant basin subsidence, sediment supply, and sea level (Vail and others, 1977, 1984: Vail and Hardenbol, 1979; see also May and others, 1984, for a discus-

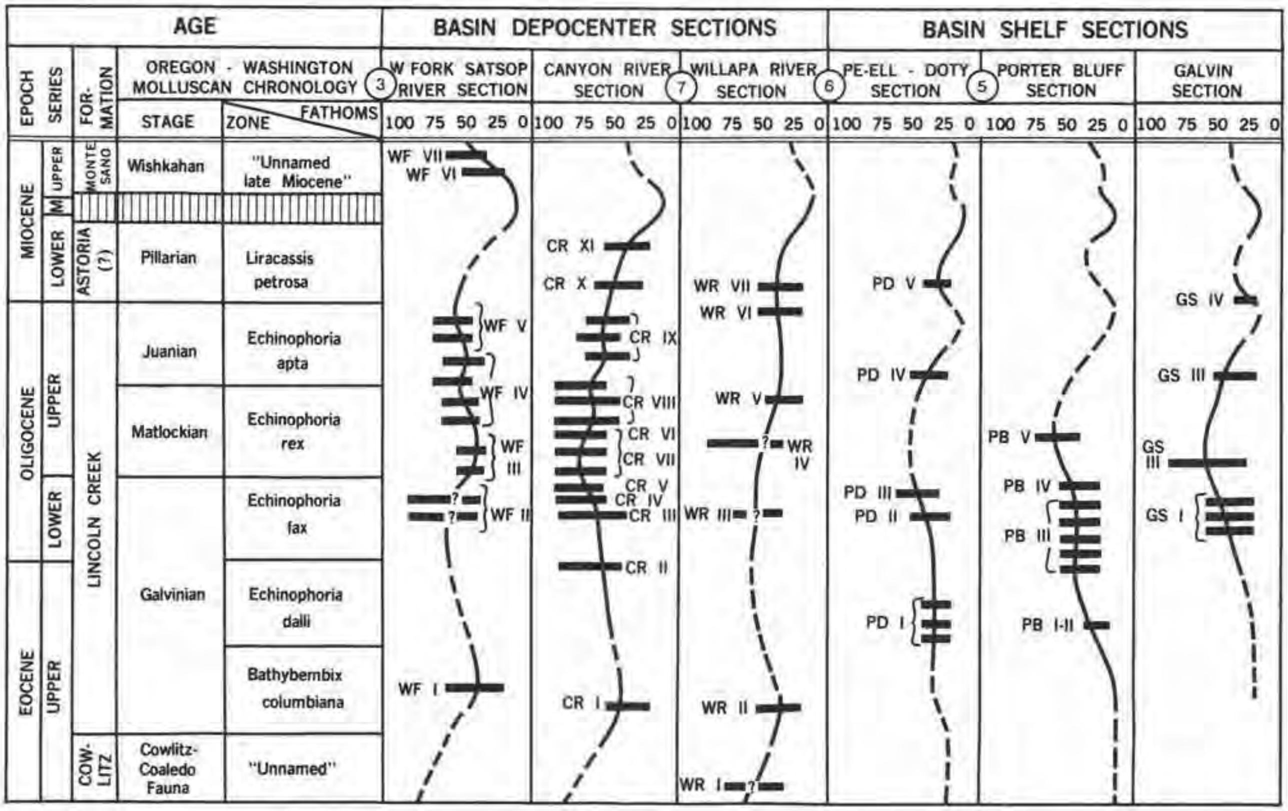

Figure 7.–Paleobathymetric curves based on molluscan faunas of the late Eocene to early Miocene Sequence III in southwest Washington. A code identifies interpreted fauna in each section, i.e., GS I, Galvin Section molluscan faunule I. Horizontal bars represent a water depth range based on the modern depth range of genera represented in the fossil assemblage. The heavy line connecting the bars represents an interpreted local bathymetric curve for the stratigraphic section and incorporates data from lithofacies patterns (modified from Armentrout, 1973b, fig. 19). See the text discussion for interpretation of water depth. Sections numbered as in Figures 2 and 4, except for Canyon River section, which is located between West Fork and Middle Fork of the Satsop River sections, and the Galvin section, which is located south of the Black Hills.

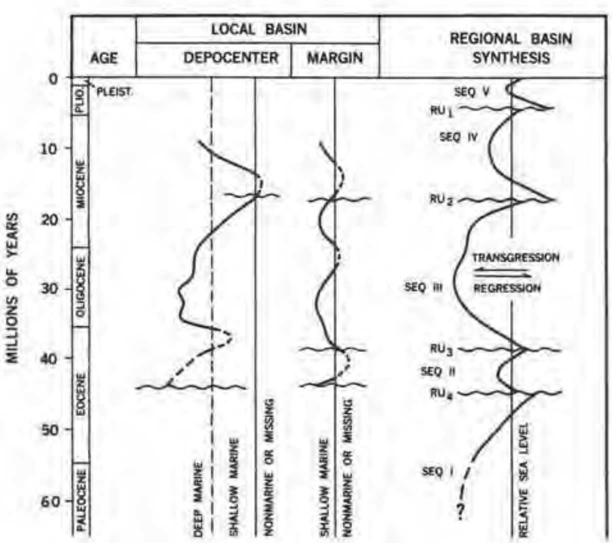

Figure 8.–Example of synthesis of local bathymetric curves into a regional basin bathymetry curve. The two local curves are redrafted from Figure 7 with the Canyon River section representing basin depocenter facies and Porter Bluff section representing basin slope facies. The basin bathymetry curve results from synthesis of local bathymetric curves supplemented with lithofacies and stratigraphic patterns. Deep marine implies water depth probably greater than 600 ft; shallow marine implies less than 600 ft. See text discussion of bathymetric curve calibration in "Unconformity-Bounded Sequences" and "Sequence Patterns". These curves are not coastal onlap curves as used by Vail and others (1977) and Vail and Hardenbol (1979).

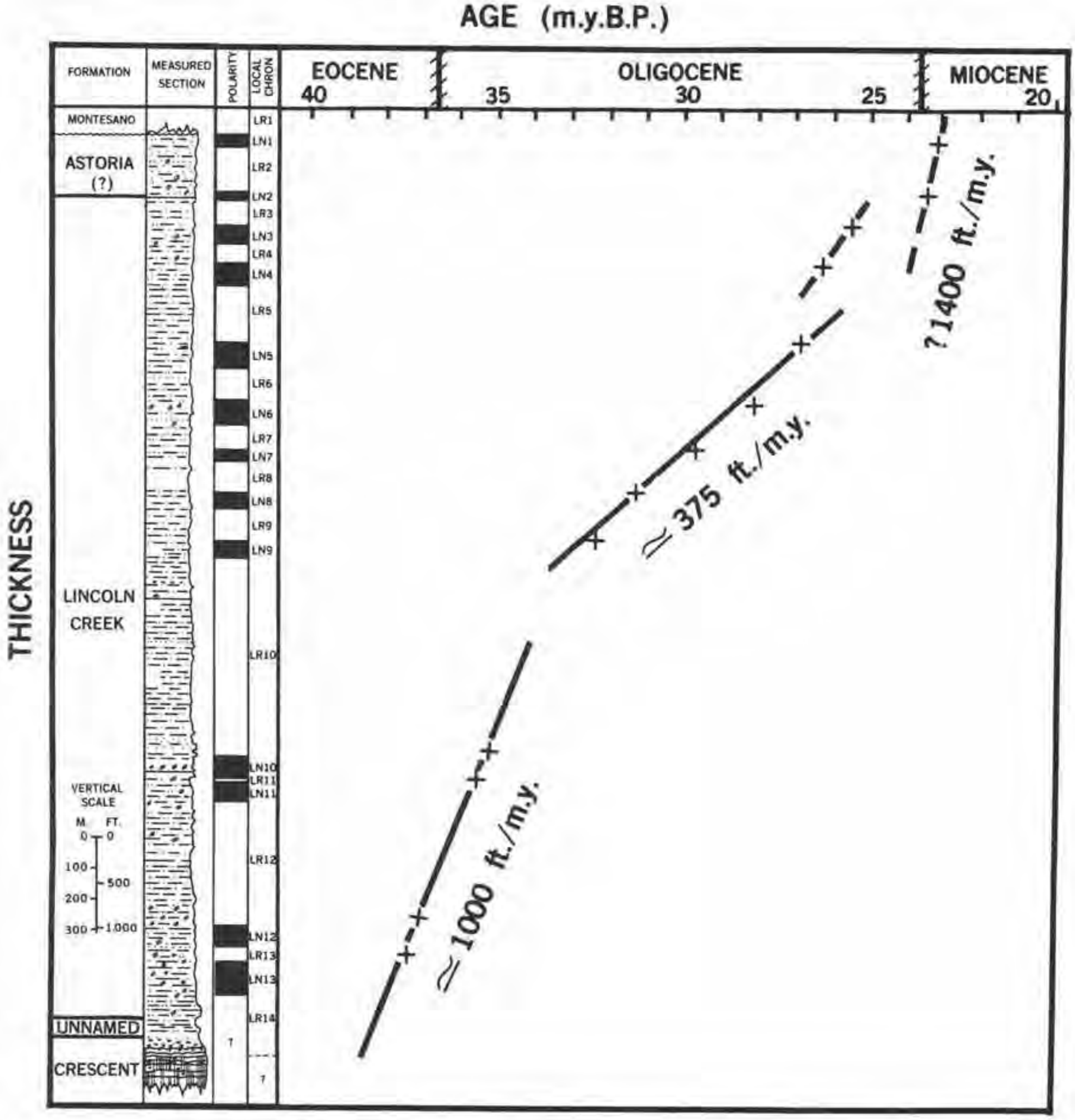

Figure 9.–Plot of rate of sediment accumulation for the Canyon River section of southwest Washington. Age control is derived from correlation of local normal (LN) and local reversed (LR) polarity events with the global magnetostratigraphic time scale following correlations of Prothero and Armentrout (1985). Plotted points represent the age/depth pair for the stratigraphic position of the top of each local polarity event.

sion of sea-level curves versus coastal onlap curves for tectonically active margins). Vail and others (1977) generated global coastal onlap curves (first called sea-level curves; Vail and Hardenbol, 1979) for the Phanerozoic showing cycles of three orders of magnitude: 2 first-order cycles of 225 and 300 m.y. duration; 14 second-order cycles of 10 to 80 m.y. duration; and more than 80 third-order cycles of 1 to 10 m.y. duration. The second- and third-order cycles for the Cenozoic are shown on Figure 5.

Comparison of the local Washington-Oregon basin bathymetry curve with the second- and third-order cycles of Vail and others (1977) shows the local curve is not in phase with the global curves; the sequence-bounded unconformities of western Oregon and Washington occur during periods of global transgression. This is not the consequence of incorrect chronostratigraphy, because the local sequence has been carefully correlated to the chronostratigraphic scale used by Vail and others (1977) (Armentrout, 1981; Armentrout and others, 1983; Prothero and Armentrout, 1985).

Vail and others (1977) considered climate to be a major factor in forcing Cenozoic global coastal onlap cycles. If global sea-level fluctuations are climatically driven, then warm periods should be represented by transgressions coeval with maximum glacial melt, and cool periods should be represented by regression coeval with maximum glacial buildup. This is in fact the pattern suggested when second-order Cenozoic coastal onlap cycles of Vail and others (1977) are compared (Fig. 5) with Cenozoic climate based on oxygen-isotope analyses (Shakelton and Kennett, 1975; Williams, 1985). Intervals of relative global transgression are correlative with warm climates during the Eocene and middle Miocene, and intervals of relative global regression are correlative with cool climates during the Oligocene to early Miocene and late Miocene to Recent. However, the transgressive-regressive events of the third-order coastal onlap curves of Vail and others (1977) clearly represent forces reflecting events not entirely attributable to global climate. These forces are most probably global tectonic events, which affect ocean basin volume, rates of basin subsidence or uplift, and rates of sediment supply to the basins (Vail and Hardenbol, 1979; Sloss, 1979; May and others, 1984).

The Cenozoic climate curve for western Washington and Oregon (Fig. 5) parallels the global climate-oxygen isotope curve. The local climate curve is based on integration of paleoecologic data from shallow water, open-marine molluscan assemblages (Dickerson, 1917; Smith, 1919; Durham, 1954; Addicott, 1976) and megafloral assemblages (Wolfe, 1981). Periods of relatively warm climate are indicated for the middle Eocene and middle Miocene; cooler climates prevailed during the Oligocene through early Miocene and late Miocene through Pleistocene. The warm intervals are coeval with major unconformities of the middle Eocene (RU_3 of Fig. 5) and the middle Miocene (RU_2 of Fig. 5). The cool intervals are coeval with periods of maximum basin deepening. Clearly, the western Washington and Oregon Cenozoic bathymetry curve is not governed by climatically controlled eustatic cycles.

Comparison of the local basin bathymetry curve with the second- and third-order global onlap curves (Vail and others, 1977) indicates that local events are not coeval with global long-term eustatic events. Therefore, regional tectonic events, as they affect the relations of subsidence and sediment supply, become likely candidates for the major factors controlling the pattern of Cenozoic deposition in western Oregon and Washington shelf-margin basins.

TECTONICS

Tectonic forces governing the transgressive-regressive cycles of western Oregon and Washington can be identified from data on basalt geochemistry, radiometric dating, remanent magnetic patterns, and sedimentary provenance studies.

Igneous Activity

Four phases of igneous activity have been identified in the Coast Range stratigraphy of Oregon and Washington (MacLeod and Snavely, 1973)(Figs. 5 and 6). They can be differentiated geochemically and age-bracketed radiometrically, and each is attributed to a different style of volcanism (Duncan, 1982; Wells and others, 1984). These styles of volcanism are directly related to the tectonic factors also affecting formation of the depositional sequences described above.

The upper Paleocene to middle Eocene (63 to 46 m.y.B.P.) basalts of Sequence I are attributed to seamount chain volcanism in an oceanic realm (Snavely and others, 1968; Duncan, 1982; Wells and others, 1984). Sequence II contains middle Eocene to upper Eocene (43 to 38 m.y.B.P.) volcanic rocks representing local accumulations around volcanic centers that were erupting in a transitional oceanic ("Cowlitz volcanics") to magmatic-arc (Northcraft Formation) system. The petrologically complex igneous assemblages of Sequence II were intruded along faults associated with northwest-southeast extension in the forearc basin (Wells and others, 1984). Sequence III contains abundant volcanic ash and interbedded Goble Volcanics of late Eocene to Oligocene age (40 to 32 m.y.B.P.) (Beck and Burr, 1979, as modified in this report). (See "Stratigraphy", Goble Volcanics.) These rocks are of andesitic composition and reflect a magmatic arc origin. Upper Eocene to Oligocene (38 to 29 m.y.B.P.) intrusions are mapped in the Oregon Coast Range where they represent a spectrum of magma types (MacLeod and Snavely, 1973). Middle Miocene (15 to 12 m.y.B.P.) volcanic rocks interbedded in Sequence IV consist of basalts of the Columbia River Basalt Group, which flowed westward into the coastal area from sources far to the east in the extensional back-arc regime of the Columbia Plateau (Snavely and others, 1973; Beeson and Moran, 1979).

Tuffaceous marine siltstones record the history of active arc magmatism in the ancestral Cascade Mountains. Such units are oldest in southern Oregon and youngest in Washington (Snavely and Wagner, 1963; Atwater, 1970). Regional development of explosive Cascade arc-magmatism occurred between about 40 and 38 m.y.B.P. when tuffaceous siltstones became pervasive in western Oregon and Washington—for example, the upper Cowlitz to lower Lincoln Creek Formations (Heller and Ryberg, 1983; Kadri and others, 1983; Prothero and Armentrout, 1985). Wells and others (1984) attribute the latest Eocene through early Miocene episodes of pyroclastic volcanism to shallow crustal emplacement of magmas and associated extension within the arc, and they note that this style of volcanism is coeval with the period during which Farallon plate-North American plate convergence rates decreased from pre-43 m.y.B.P. rates of 150 km/m.y. to post 28 m.y.B.P. rates of 40 km/m.y.

In summary, the history of volcanism defines four major phases of igneous activity in the forearc basins of western Oregon and Washington: (1) at 63 to 46 m.y.B.P., oceanic seamount extrusion; (2) at 43 to 38 m.y.B.P., arc-trench gap volcanism; (3) at 40 to 32 m.y. B.P., arc volcanism; and (4) at 15 to 12 m.y.B.P., back-arc flood basalt extrusion (Fig. 6).

Remanent Magnetism

Several research groups have measured the remanent magnetic declination of the igneous rocks interbedded with or intruded into the depositional sequences of western Oregon and Washington. Figure 10 is from Magill and Cox (1980) with the addition of one data point—the "basalt of Pack Sack Lookout" from Magill and others (1982). Although a straight-line curve suggesting a relatively constant rate of Cenozoic rotation can fit the data, Magill and Cox (1980) prefer a two-phase rotation history defined by the heavy line on Figure 10. The two phases of rotation consist of an older 60 to 42 m.y.B.P. rotational phase and a younger 20 to 0 m.y.B.P. rotational phase separated from each other by a 40 to 20 m.y.B.P. relatively nonrotating interval.

A two-phase rotational model better fits the geologic history of western Oregon and Washington. The major inflections of the Magill and Cox (1980) curve are coincident with the middle Eocene (RU_1) and middle Miocene (RU_2) regional unconformities, which also separate the principle types of igneous rocks described above (Fig. 5).

The two-phase rotation model of Magill and Cox (1980) and Magill and others (1981) suggests a 60 to 42 m.y.B.P. rotation phase of the Oregon Coast Range in a clockwise direction into the eastward-dipping Challis-Absaroka subduction zone (Fig. 11). Approximately 40° to 50° of clockwise rotation is attributed to the first phase, which was completed about 46 to 42 m.y. B.P. and was followed by a period of relative tectonic stability. The 46 to 42 m.y.B.P. timing is coincident

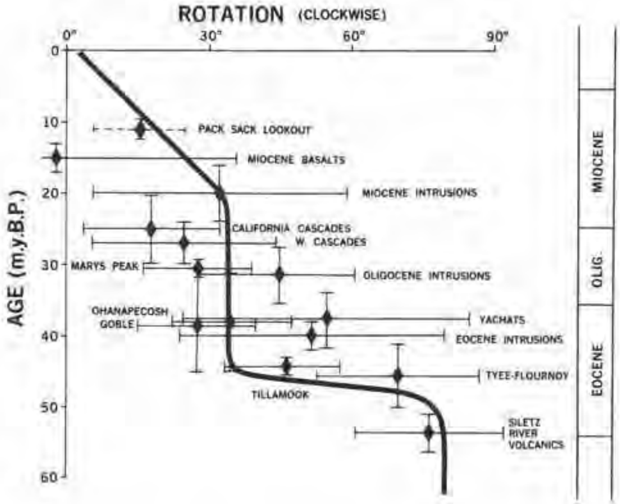

Figure 10.–Paleomagnetic rotation versus age for geologic units of the Cascades and Oregon-Washington Coast Range (modified from Magill and Cox, 1980, fig. 12). Vertical error bars for age; horizontal bars for rotation. The heavy line represents a best-fit curve for two phases of rotation, pre-42 m.y. ago and post-20 m.y. ago. The data for "basalt of Pack Sack Lookout" (Pomona flow) is from Magill and others (1982).

with the major folding and faulting, uplift, and erosion of the rocks of the lowermost unconformity-bounded sequence of southwest Washington, and it probably represents the accretion of Sequence I oceanic seamounts into the forearc accretionary prism of western Oregon and Washington (Tabor and Cady, 1978; Heller and Ryberg, 1983; Wells and Coe, 1985).

The 42 to 20 m.y.B.P. interval of relative tectonic quiescence is represented by the relatively uninterrupted deposition of the late Eocene through early Miocene unconformity-bounded Sequences II and III. This period also brackets the building of the ancestral Cascade Mountain arc.

Renewal of rotation occurred about 20 m.y.B.P., coincident with middle Miocene deformation of the lower Tertiary unconformity-bounded Sequences I, II, and III. Approximately 20° (Magill and Cox, 1980) to 35° (Wells and Coe, 1985) of clockwise rotation occurred during this second phase. Magill and Cox (1980) and Magill and others (1981) attribute this rotation to the extensional opening of the Great Basin.

The Magill and Cox (1980) model used a rotation of coherent terrains: the Oregon Coast Range during phase one, and the Oregon Coast Range and Cascade terrains together during phase two (Fig. 11). Others workers (Beck and Burr, 1979; Globerman and others,

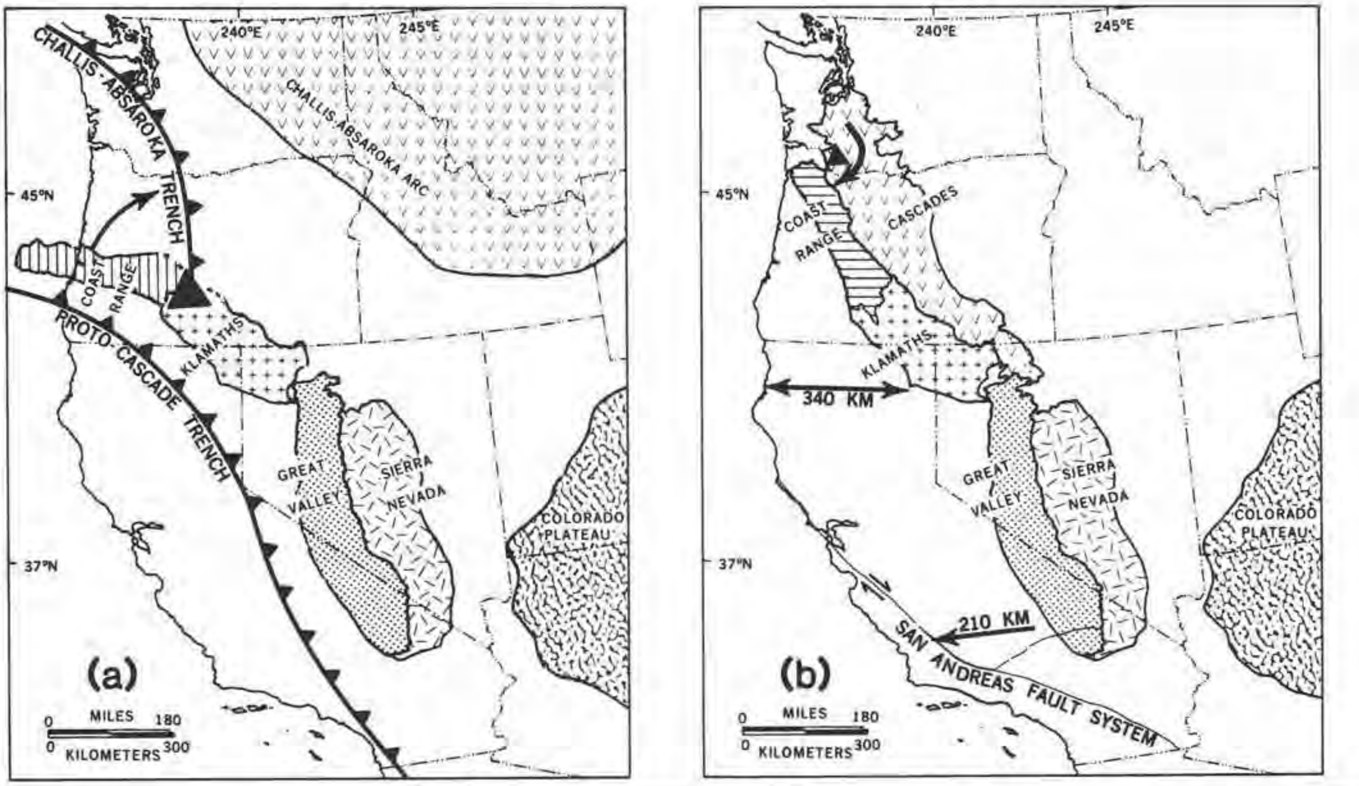

Figure 11.–Palinspastic maps for western North America showing a two-phase rotational model suggested by Magill and Cox (1980). (a) Phase I: 50 to 42 m.y. ago—Rotation occurred about a northern Klamath Mountain pivot- point (triangle), moving the Oregon Coast Range terrain clockwise toward the Challis-Absaroka arc subduction zone (heavy line with triangles on the upper plate (Magill and Cox, 1980, figure 3b). (b) Phase II: 20 to 0 m.y. ago—Rotation about a southern Washington pivot-point occurred as a consequence of Basin and Range extension and the westward displacement of several terrains as shown by arrows with extensional displacement in kilometers (Magill and Cox, 1980, fig. 13).

1982; Wells and Coe, 1985; Wells and others, 1984) have found that basalts and sedimentary rocks of southwest Washington have rotational histories different from that of the Oregon Coast Range (Simpson and Cox, 1977; Magill and others, 1981). Wells (1982) mapped three distinct structural blocks of Crescent Formation basalts in southwest Washington, each with a different clockwise rotational history (Fig. 12). Wells (1982) and Wells and Coe (1985) suggest that the data show differential rotation of the Crescent Formation basalt basement of southwest Washington. This early phase of differential rotation ceased about 46 to 42 m.y.B.P., prior to the extrusion of upper middle Eocene "Cowlitz volcanics" (mapped as Goble Volcanics by Wells, 1981), which show a pattern of coherent rotation. Coherent rotation is suggested by "Cowlitz volcanics" from several distinct structural blocks having the same degree of clockwise rotational history.

Regional northwest-southeast lineations are well known in the northwest United States and have been interpreted as right-lateral faults (Lawrence, 1976) (Fig. 13). The subordinate left-lateral fault systems represent a second-order failure between clockwise rotational blocks (Fig. 14). It is possible that the rotation of Coast Range structural blocks is, for the most part, a consequence of this shear system, with only minor components of microplate movement (Beck, 1976; Wells and Coe, 1985). Magill and Cox (1980) and Magill and others (1982) have suggested that the driving force for this shear system is the opening of the Great Basin; Wells and Coe (1985) suggest the driving force results from shear along the Pacific-North American plate boundary.

Interpretation of rotational patterns of Eocene rocks in western Oregon and Washington requires incorporation of Duncan's (1982) seamount chain model. Duncan (1982) has radiometrically dated the oceanic tholeiitic basalts of the Coast Range (Fig. 15). He has documented a geographically symmetrical pattern of ages for these basalts and suggests they represent a seamount chain extruded from an oceanic rift

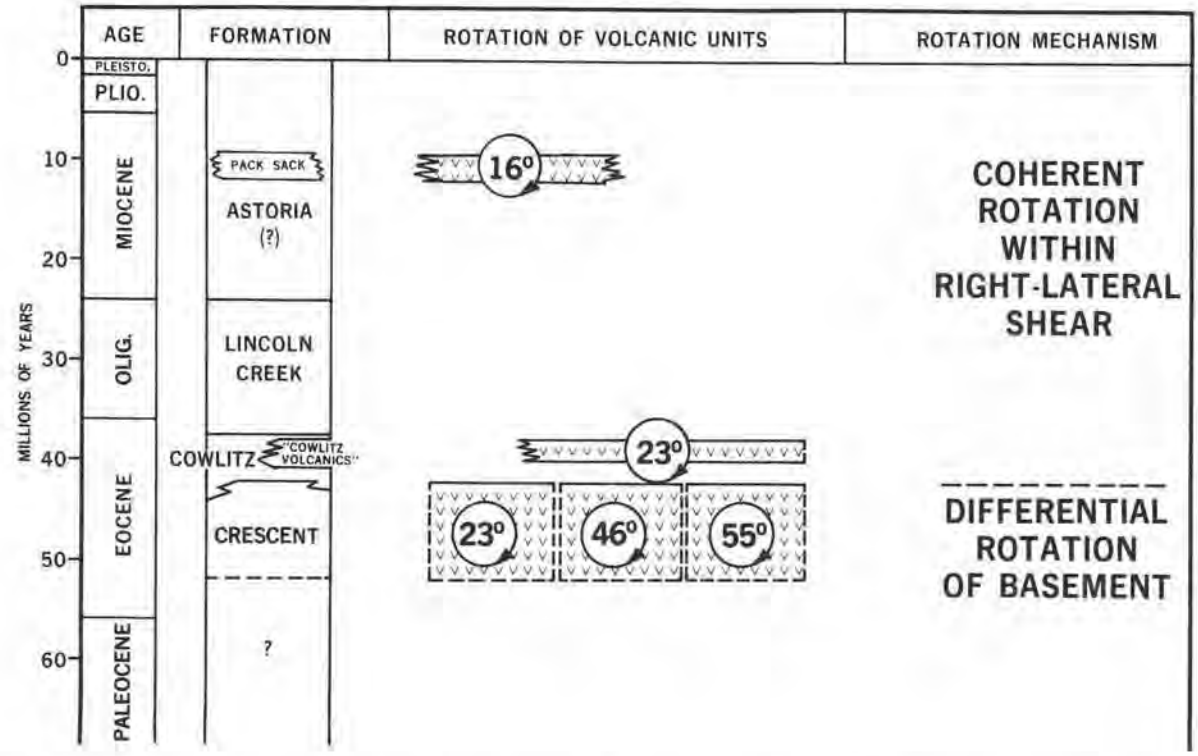

Figure 12.–Summary chart for paleomagnetic rotation of volcanic rocks in southwest Washington. Wells (1982) and Wells and Coe (1985) mapped several fault-bounded domains of Crescent Formation volcanic rocks characterized by different amounts of clockwise rotation, suggesting differential rotation of Coast Range basement. Magill and others (1982) and Wells and Coe (1985) mapped rocks of the informally named Cowlitz volcanics and basalt of Pack Sack Lookout and found that all outcrop areas studied for each stratigraphic unit had essentially the same amount of rotation despite extending across two or more of the differentially rotated domains in the Crescent Formation. This suggests regionally coherent rotation of the Cowlitz volcanics and the basalt of Pack Sack Lookout.

system onto the Kula and Farallon plates. The extrusions must have occurred prior to accretion of this seamount chain to western Oregon and Washington. Some component of rotation may have been imposed on this seamount chain prior to accretion by (1) microplate deformation such as proposed by Carlson (1981) for the Juan de Fuca and Gorda plates, or (2) impingement rotation similar to that suggested by Simpson and Cox (1977). The accretion of a single Paleogene seamount chain (Duncan, 1982) may not be entirely consistent with models for Kula-Farallon-North American plate interactions and may be the consequence of a much more complex history (Wells and others, 1984).

Sediment Provenance

Whatever the style of emplacement of the seamount terrain and consequent rotations, stratigraphic relations of Sequence I rocks define relative geographic proximity of a seamount terrain to the North American continent by 65 to 50 m.y.B.P. and cessation of oceanic seamount extrusion by 46 m.y.B.P. Arkosic sandstone of cratonic derivation is associated with ocean ridge tholeiitic basalt of the Roseburg Formation in southern Oregon (Heller and Ryberg, 1983). Basalts of the Crescent Formation on the Olympic Peninsula have interbedded sedimentary rocks containing clasts attributed to the Mesozoic and perhaps Paleozoic terrains north of the Straits of Juan de Fuca and east of Puget Sound (Cady, 1975). The seamount chain must have been in proximity to the craton by 62 to 58 m.y.B.P. in the southern Oregon area, and by 52 to 50 m.y.B.P. in the Olympic Peninsula area, the age of the basalts in each area (Duncan, 1982).

The youngest seamount volcanic rocks are enveloped by arkosic sandstones of middle to early late Eocene (Sequence II, Cowlitz Formation. about 44 to 40 m.y.B.P.; Snavely and others, 1951a; Armentrout and others, 1983). Based on geochemistry of the mica, the source area for the arkosic sands was the Idaho

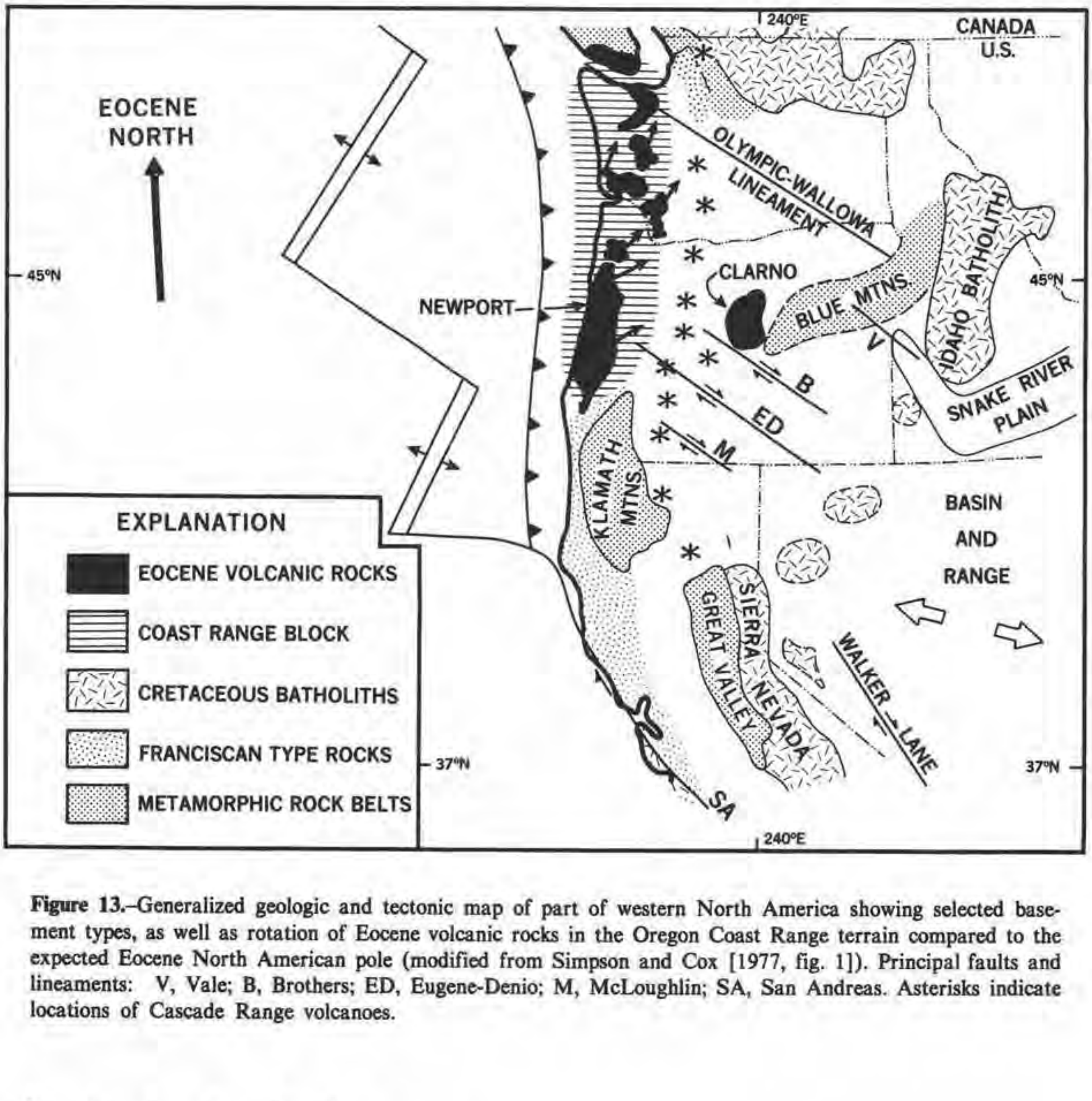

Figure 13.–Generalized geologic and tectonic map of part of western North America showing selected basement types, as well as rotation of Eocene volcanic rocks in the Oregon Coast Range terrain compared to the expected Eocene North American pole (modified from Simpson and Cox [1977, fig. 1]). Principal faults and lineaments: V, Vale; B, Brothers; ED, Eugene-Denio; M, McLoughlin; SA, San Andreas. Asterisks indicate locations of Cascade Range volcanoes.

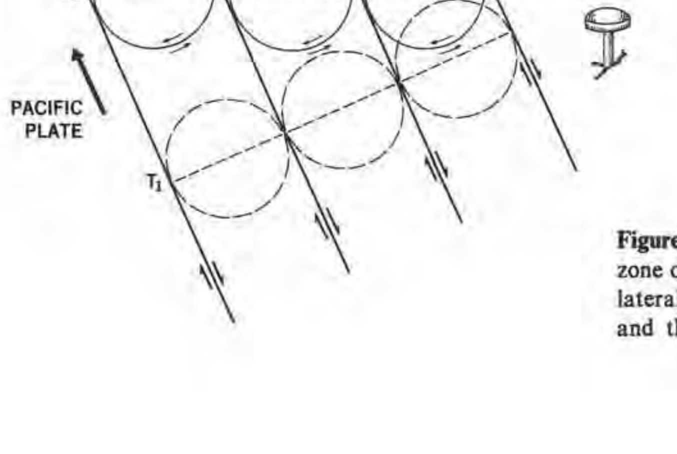

Figure 14.–Ball-bearing model for rotation of continental fragments in a zone of right-lateral shear (modified from Beck, 1976, fig. 4). Note left-lateral rotation of "ball bearings" from time-one (T_1) to time-two (T_2) and the northward displacement of each.

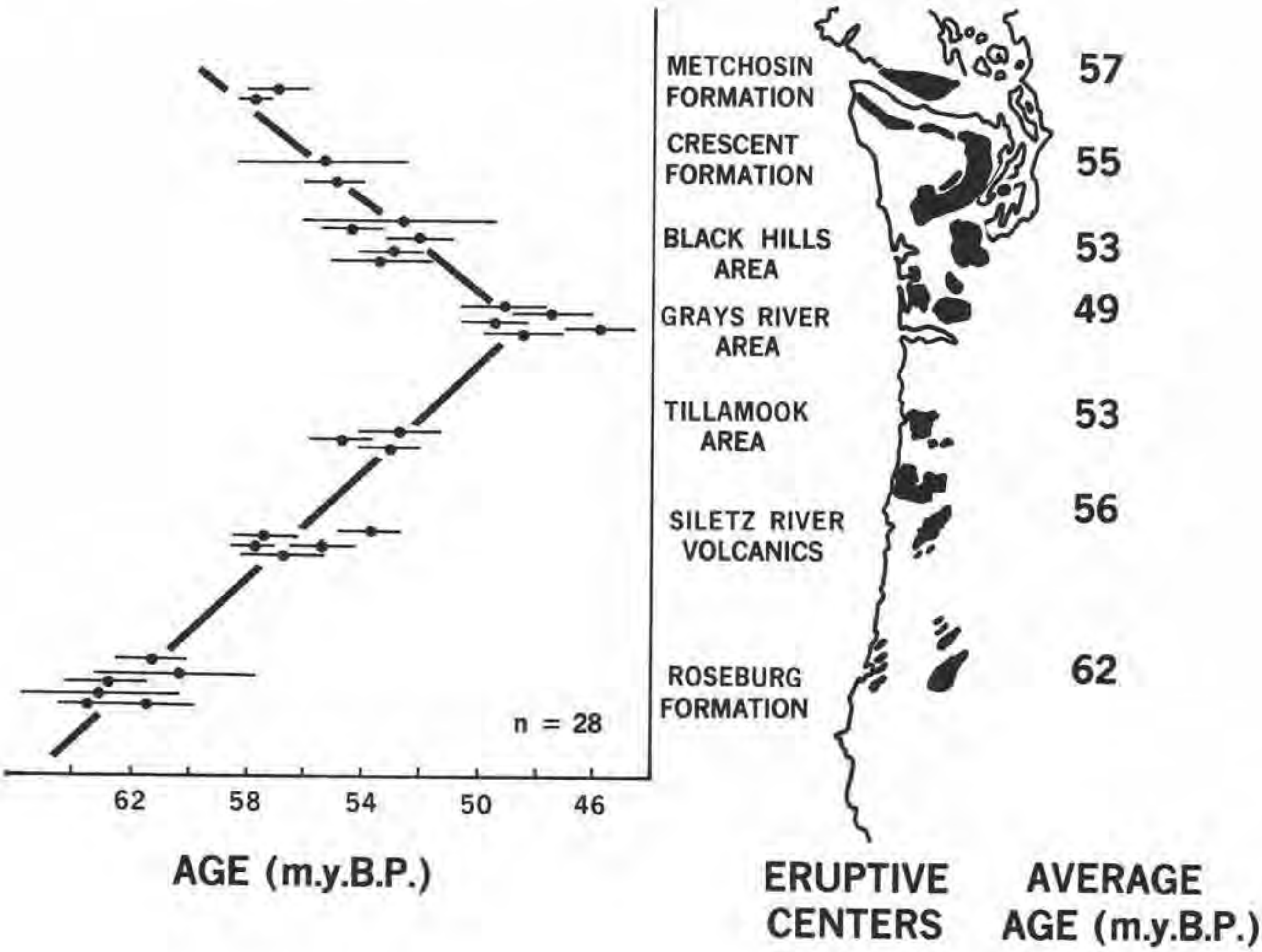

Figure 15.–Ages of Paleocene and Eocene Coast Range basalts (Duncan, 1982). Radiometric dates (with error bars) are plotted opposite the geographic position of the outcrop area shown in the map. The average age for each outcrop area is shown on the right and suggests a V-pattern wherein the youngest unit is at Grays River and successively older units are both north and south. Duncan (1982) interprets this pattern as suggesting extrusion of seamounts at a hotspot on a spreading ridge with progressive displacement of coeval rocks north and south by sea-floor spreading. See Figure 17 for an oblique view depicting this relationship.

Batholith; the sands were deposited as part of a broad fluvial plain/coastal plain/shelf-margin basin that buried the old subduction-accretion complex, including the accreted seamounts (Kadri and others, 1983; Heller and Ryberg, 1983). About 40 to 38 m.y.B.P., a new subduction system became established, and the ancestral Cascade Mountain arc began developing; this event formed a major sequence boundary in the Cascade Mountains and Columbia River Plateau provinces and is expressed along structural highs in southwest Washington (RU_3 of Fig. 5; coeval with the base of the Lincoln Creek Formation). Arc-derived pyroclastic debris was mixed with the arkosic sediments brought in from farther east by west-flowing rivers and deposited as forearc tuffaceous sediments, which dominate the latest Eocene, Oligocene, and early Miocene stratigraphy of Sequence III in western Oregon and Washington (Snavely and Wagner, 1963,1982; Snavely and others, 1980a; Armentrout and Franz, 1983; Heller and Ryberg, 1983; Wells and others, 1984).

Sequences IV and V consist of sediments derived from (1) local reworking of older Tertiary rocks within southwest Washington, (2) volcaniclastic material from the active Cascade arc, and (3) the continued input of sediments via the ancestral Columbia River drainage system.

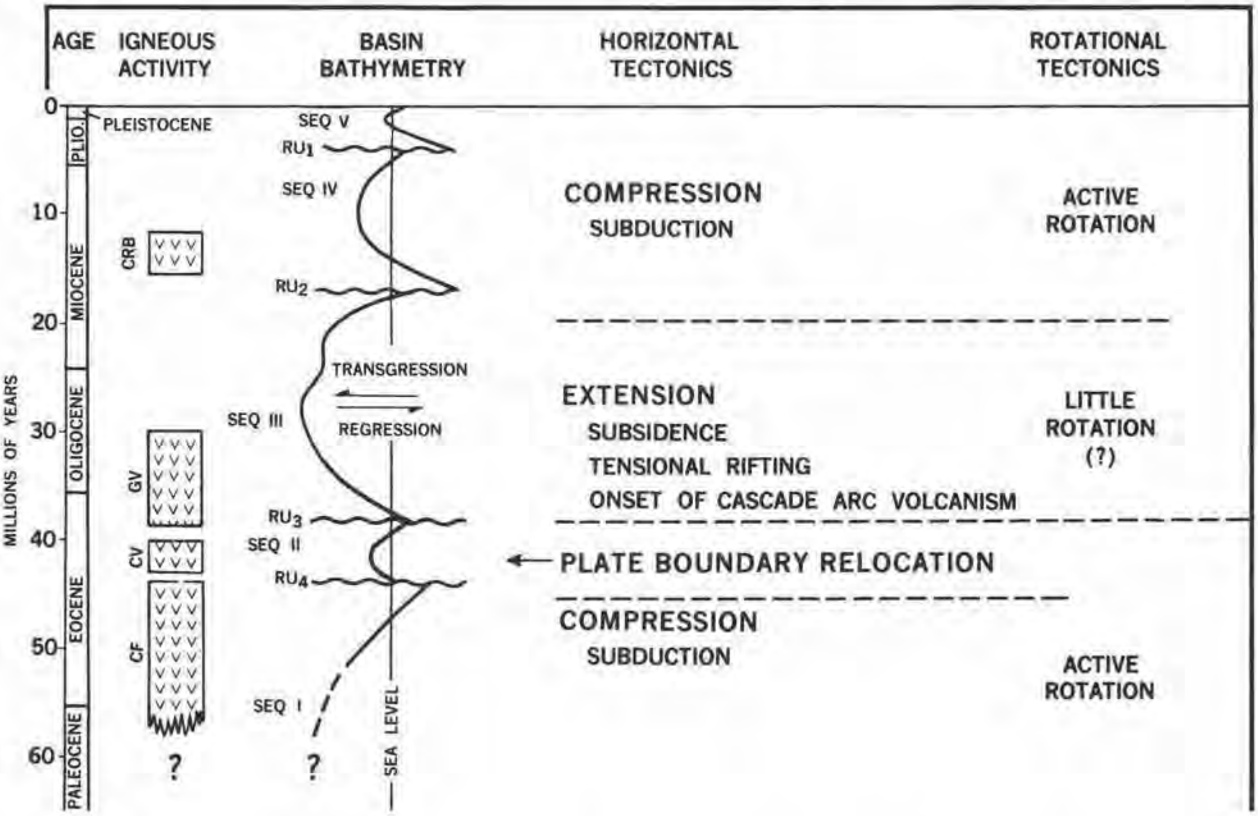

Figure 16.–Summary of Cenozoic tectonic events and depositional patterns for southwestern Washington, suggesting the horizontal and vertical rotational tectonic regimes that controlled the development of the Cenozoic depositional sequences of southwest Washington. Abbreviations as in caption for Figure 5.

SUMMARY

The Eocene tectonics of western Washington and Oregon are summarized by the maps of Figure 11, the chart of Figure 16, and the schematic block diagrams of Figures 17, 18, 19 and 20.

During early Eocene time, from 64 to 46 m.y.B.P., a northwest-southeast seamount chain was extruded on the Kula and Farallon plates west of the eastern eastward-dipping Challis-Absaroka subduction zone (Fig. 16). Subduction of the oceanic plate moved the seamount chain obliquely toward the subduction zone, possibly with some right-lateral small-plate shear resulting in clockwise rotation of from 20° to about 50°. The seamount chain was adjacent to the early Eocene forearc basin of western North America by 50 m.y.B.P. when continentally derived sedimentary rocks became interbedded with the oceanic basalts. During middle Eocene time, 46 to 42 m.y.B.P., the seamount chain (Crescent Formation) reached the Challis-Absaroka subduction zone, became accreted, and forced a relocation of subduction to the west, forming the Cascade subduction zone. As the subducting crust within the Challis-Absaroka system was consumed, the Challis-Absaroka arc decreased its activity. Continued northeastward, oblique subduction of the Kula and Farallon plates beneath North America resulted in northeast-southwest compression and formation of a northwest-southeast fault and fold trend, with consequent uplift and erosion that formed Regional Unconformity RU_4.

Coincident with and continuing after the subduction zone relocation and seamount chain accretion, Sequence II arkosic sandstones (Cowlitz Formation) prograded westward and, with associated marine shales, enveloped the basaltic seamounts in the newly formed forearc basin (Fig. 17). Local intrusion of the forearc basin ("Cowlitz volcanics") occurred along northeast-southwest tensional fault systems in the accretionary prism (Fig. 18). This fault system is interpreted as the consequence of dextral shear along the Oregon-Washington coast driven by northeast-directed oblique subduction of the Farallon plate beneath North America during the Eocene (Wells and

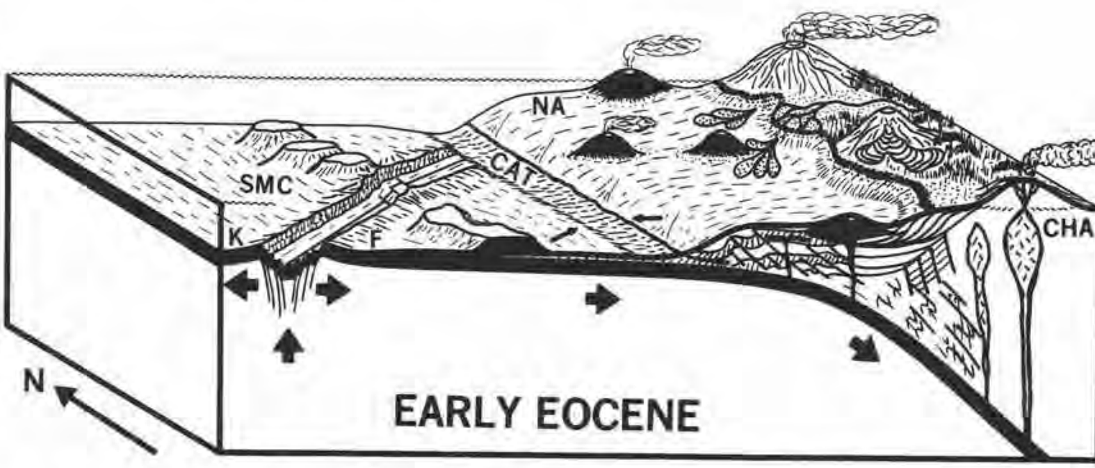

Figure 17.–Early Eocene (about 48 m.y. ago) tectonic framework for part of northwest Oregon and southwest Washington, showing a seamount chain (SMC) formed along the Kula (K)–Farallon (F) spreading ridge and Challis-Absaroka arc (CHA) volcanism on the North American plate (NA) above the eastward-dipping Challis-Absaroka trench (CAT) marking the site of subduction of the Kula-Farallon plates. Deltas and submarine fan systems developed along the margin of the forearc basin. Arrows represent relative motion of plates or faults. Modified from Armentrout and Suek (1985, fig. 19).

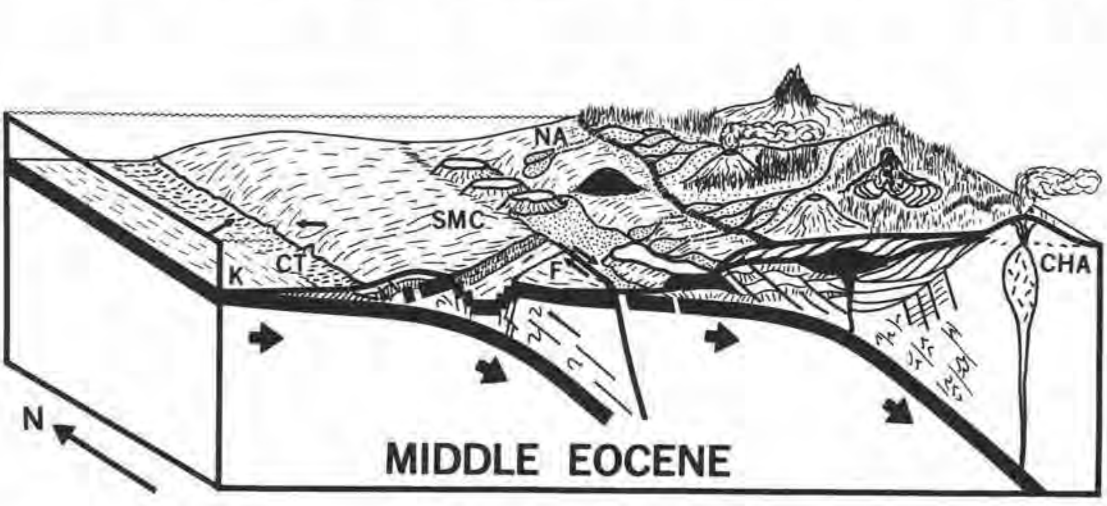

Figure 18.–Middle Eocene (about 42 m.y. ago) tectonic framework for part of northwest Oregon and southwest Washington, showing the relocation westward of subduction to the Cascade trench (CT), the emplacement of the Kula-Farallon seamount chain (SMC) into the old Challis-Absaroka trench and the subsequent progradation of the continental margin westward enveloping the seamounts in sediments of the new forearc basin. Modified from Armentrout and Suek (1985, fig. 20); other symbols as in Figure 17.

Coe, 1985). Plate-boundary readjustment associated with formation of the Cascade trench-arc system resulted in uplift and erosion that formed Regional Unconformity RU_3.

As the Cascade subduction zone developed, the arc-trench gap was under tension due to the frictional forces of the subducting crust. This tensional system resulted in (1) subsidence exceeding sediment supply in the shelf-margin basin, and (2) in the deep-water deposits of Sequence III transgressing the basin-margin coastal plain deposits (Fig. 19). As the newly formed subducting slab reached sufficient depth, arc magmatism, characterized by ash-flow volcanism, began along the axis of the ancestral Cascade Mountains at about 42 to 38 m.y.B.P. (Goble Volcanics; Wells and others, 1984; W. M. Phillips, personal

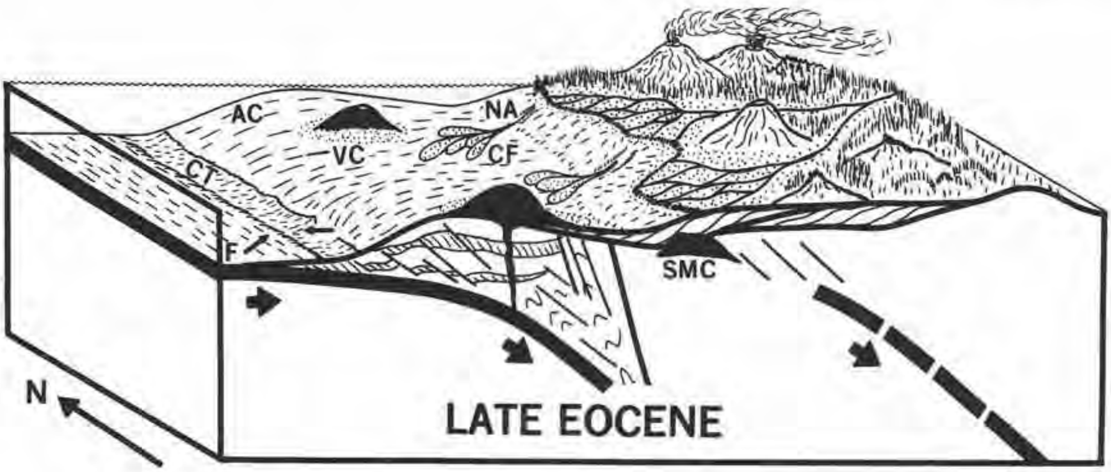

Figure 19.–Early late Eocene (about 40 m.y. ago) tectonic framework for part of northwest Oregon and southwest Washington, showing subduction of the Farallon plate (F) into the Cascade trench (CT) and beneath the newly accreted prism (AC) of western North America (NA). The Kula-Farallon seamount chain (SMC) has been accreted and buried by prograding sediments of a river-dominated coastal plain. Submarine clastic fans (CF) were shed into and local volcanic centers developed within the forearc basin. Modified from Armentrout and Suek (1985, fig. 21).

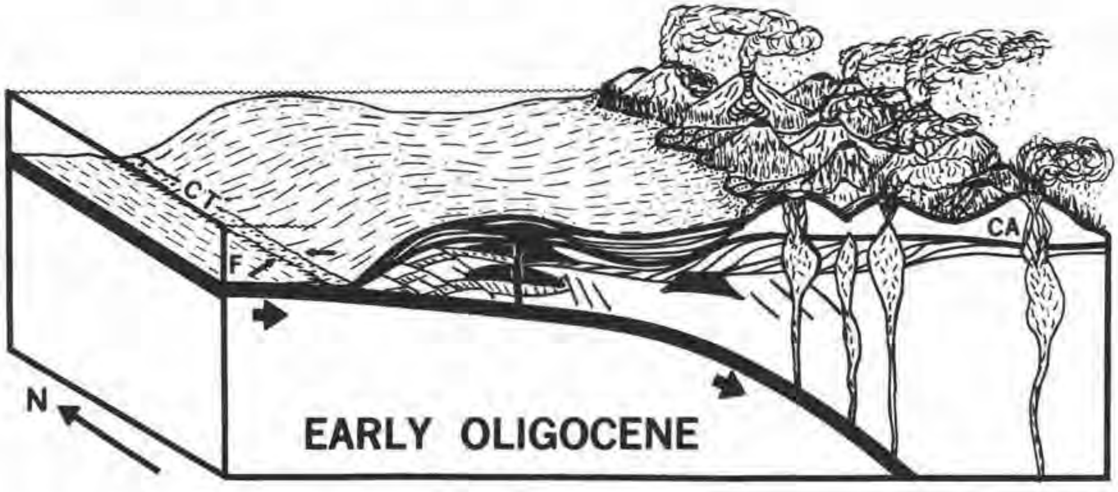

Figure 20.–Early Oligocene (about 35 m.y. ago) tectonic framework for part of northwest Oregon and southwest Washington, showing Cascade trench (CT) subduction of the Farallon plate (F) and ancestral Cascade Mountains arc development (CA). Basin subsidence resulted in the regional transgression along margins of the self-margin forearc basin.

commun., 1986). The abundant arc-derived ash was washed westward and mixed with marine sand and mud (Lincoln Creek and Astoria (?) Formations).

Late early Miocene readjustment of plate configuration deformed Sequences I, II, and III, forming Regional Unconformity RU_2. Subsequent events included backarc extension of the Great Basin, accompanied by extrusion of the 17 to 12 m.y.B.P. basalts of the Columbia River Basalt Group, replacement of Cascade ash flow volcanism by andesitic, basaltic, and dacitic lava flows, and renewed movement along northwest-southeast fault systems. This late Cenozoic tectonism is represented in the 20 to 0 m.y.B.P. rotational phase of Coast Range and Cascade terrains during opening of the Basin and Range province. Consequent northwest-southeast-aligned basins are occupied by the Columbia, Willapa, and Chehalis Rivers. These basins were filled from the east by nonmarine to marine progradational Sequence IV (Montesano and Quinault Formations).

The late Pliocene to early Pleistocene Regional Unconformity RU_1 developed by the erosion of areas uplifted by continued northeast compression. The compression was forced by subduction of Kula plate remnants into the Cascade trench and beneath North America.

Late Cenozoic global cooling resulted in major ice sheet development and consequent sea-level fluctuations. Sequence V deposits are largely the result of glacially forced fluctuations of stream gradient and flow volume, as well as shoreline position along a continental margin undergoing continued compression due to plate subduction.

Interpretation of the geologic history of western Washington and Oregon has undergone significant revision over the last several years as detailed mapping and sample analysis have been interpreted in the context of plate tectonic models. It is reasonable to expect similar dramatic reinterpretations in the future as detailed field mapping and stratigraphic studies provide a foundation for high resolution sample analysis and synthesis.

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CHRONOLOGY OF PLEISTOCENE SEDIMENTS IN THE PUGET LOWLAND, WASHINGTON

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ABSTRACT

The early Pleistocene history of the Puget Lowland is marked by repeated advances of the Cordilleran ice sheet into the southern Puget Lowland where deposits of at least three glaciations older than 0.8 million years are recognized: the Orting, Stuck, and Salmon Springs, separated by the interglacial Alderton and Puyallup Formations. Until 1981, the chronology of these stratigraphic units was unknown, and correlations were based entirely on relative age considerations. Paleomagnetic, fission-track, and tephra analyses now provide a basis for establishing the chronology of the type sections of these stratigraphic units and to develop a standard for correlations throughout the Puget Lowland. Reversed magnetic polarity and fission-track ages of 800,000-900,000 years from Lake Tapps tephra at three localities indicate that the Orting Drift, Alderton Formation, Stuck Drift, Puyallup Formation, and Salmon Springs Drift were all deposited during the Matuyama Reversed Polarity Epoch, which began about 2.48 million years ago and ended about 0.73 million years ago. Many of the correlations previously made with Salmon Springs Drift and older units in the Puget Lowland are no longer viable in view of the antiquity of the Salmon Springs, Puyallup, Stuck, Alderton, and Orting sediments. A chronologic gap between 200,000 and 800,000 years ago is now apparent in the Pleistocene stratigraphic record of the Puget Lowland.

The chronology of Double Bluff Drift, Whidbey Formation, and Possession Drift has been determined by amino acid analyses of shells and wood. Amino acid age estimates for marine shells in Double Bluff glaciomarine drift suggest an age of 110,000-178,000 years, but the ages could be as old as about 250,000 years. The age of shells in sediments of the Whidbey Formation is calculated as 96,000 ± 35,000 years by leucine and 107,000 ± 9,000 years by alloisoleucine/isoleucine D/L ratios. Marine shells in Possession glaciomarine drift give estimated amino acid ages of 80,000 ± 22,000 years.

Amino acid analyses of wood associated with shells in Double Bluff Drift, Whidbey sediments, and Possession Drift and of younger wood dated by radiocarbon were made to determine if wood could be used for amino acid age determinations. Most of the wood analyses gave consistent results, but some samples gave widely disparate results for reasons which were not readily apparent. Age calculations for wood were not made because kinetic models of racemization in wood are not yet well understood. The results of amino acid analyses in wood look encouraging, but additional data are needed before the method can be used with confidence.

The chronology of late Pleistocene sediments in the Puget Lowland is well established with many radiocarbon dates. An early advance of ice into the Fraser Lowland of British Columbia between 18,000 and 21,500 years ago that may be equivalent to the Evans Creek Drift in the Cascade Mountains apparently did not reach the Puget Lowland of Washington. The Cordilleran ice sheet advanced across the international boundary shortly after 18,000 years ago. The Puget lobe and the Juan de Fuca lobe apparently advanced synchronously. The Juan de Fuca lobe retreated from the western part of the strait shortly before 14,500 years ago, and the Puget lobe retreated from its terminus to the vicinity of Seattle by 14,000 years ago. Shortly thereafter, the ice sheet thinned sufficiently to allow marine water into the Puget Lowland, and floating of the remaining ice and its progessive melting resulted in deposition of Everson glaciomarine drift over an area of about 18,000 square kilometers. More than 80 radiocarbon dates from shells and wood in Everson glaciomarine drift show that the drift was deposited nearly contemporaneously from berg ice over the whole region, rather than transgressively from a retreating, calving ice front.

Cordilleran ice readvanced a short distance into the northern Puget Lowland during the Sumas Stade about 11,500 years ago and disappeared by 10,000 years ago.

INTRODUCTION

Construction of the stratigraphic framework of Pleistocene sediments in Washington was initiated by Willis (1898) and Bretz (1913). Willis recognized two glaciations in the Puget Lowland, which he named the Vashon and Admiralty, separated by the Puyallup Interglaciation. Additional pre-Vashon glacial sediments were included by Bretz in the Admiralty Glaciation. Hansen and Mackin (1949) were the first to document more than one pre-Vashon glaciation by identifying two tills, separated by interglacial sediments, beneath Vashon till north of Seattle. Evidence for four glaciations in the southern Puget Lowland was documented by Crandell and others (1958), and recognition of some of the stratigraphic units was extended throughout the southern and central Puget Lowland (Fig. 1). Pleistocene deposits beyond the range of radiocarbon dating were correlated throughout the region entirely on the basis of relative stratigraphic position and comparison with previously defined stratigraphic units.

This framework was later redefined and expanded by Armstrong and others (1965) and Easterbrook and others (1967) and served as the basis for interpretation of Pleistocene stratigraphy and chronology in the Puget Lowland until 1981. A radiocarbon date of 71,500 years ago from the type locality of the Salmon Springs glaciation (Stuiver and others, 1978) suggested that it was early Wisconsin in age, and deposits of the next-to-last glaciation elsewhere in the Puget Lowland, which had previously been widely correlated with the Salmon Springs Drift, were also assumed to be the same age. However, the demonstration by Easterbrook and others (1981) that the Salmon Springs Drift at its type locality was about 850,000 years old made most previous correlations with the Salmon Springs Drift invalid and left the regional chronology of pre-Fraser deposits in need of extensive revision. This conclusion followed the discovery of reversely magnetized silt at Auburn (Othberg, 1973) and at the Salmon Springs type locality (Easterbrook and Othberg, 1976) and a zircon fission-track age of 0.84 ± 0.21 m.y. from tephra immediately below the silt (Easterbrook and Briggs, 1979; Easterbrook and others, 1981). The reversed polarity and fission-track age at the type locality place the Salmon Springs Drift within the Matuyama Reversed Polarity Chron. The revised stratigraphic sequence in the Puget Lowland is shown in Table 1.

Recent paleomagnetic investigations have shown that the pre-Salmon Springs stratigraphic units are also reversely magnetized at their type localities and are thus between about 800,000 and 2.4 m.y. old (Roland, 1984; Easterbrook, in press). The paleomagnetic and fission-track studies have thus shown a gap in the Pleistocene chronology of the Puget Lowland between about 250,000 and 850,000 years ago (Easterbrook, in press).

In order to better establish correlation and chronology of pre-Fraser–Salmon Springs sediments, amino acids in shells and wood from a number of critical localities in the northern and central Puget Lowland were analyzed (Fig. 2). Enantiomeric ratios of amino acids in fossil mollusk shells from the west coast of North America have been utilized by many workers in the past few years as an aid in correlation of geological units and to determine ages (Karrow and Bada, 1980; Kennedy, 1978; Kennedy and others, 1982; Kvenvolden and Blunt, 1980; Kvenvolden and others, 1979a, b, 1980; Masters and Bada, 1977; Miller and Hare, 1980; Rutter and others, 1979; Schroeder and Bada, 1976; Atwater and others, 1981; Wehmiller and others, 1977; Williams and Smith, 1977). In most instances, correlation using amino acids has been successful, whereas dating has been hampered by difficulty in obtaining the average temperature history of the specimen, a necessary requisite for accurate results.

Amino acid geochemistry of fossil wood has potential applicability to problems of geochronology, but details of amino acid reactions in wood are not well known. Although much less work has been done on wood than on shells, results obtained so far on wood look encouraging for correlation purposes (Easterbrook and others, 1982; Engel and others, 1977, 1979; Lee and others, 1976; Rutter and others, 1980; Rutter and Crawford, 1983). However, more data on kinetic modeling and paleotemperature effects are necessary before dates on wood become consistent and reliable.

Shells and wood were analyzed by gas chromatography (Fig. 3) at the laboratory of the U.S. Geological Survey; many of the wood samples were analyzed at the amino acid laboratory of the University of Alberta. Both laboratories followed essentially the same analyt-

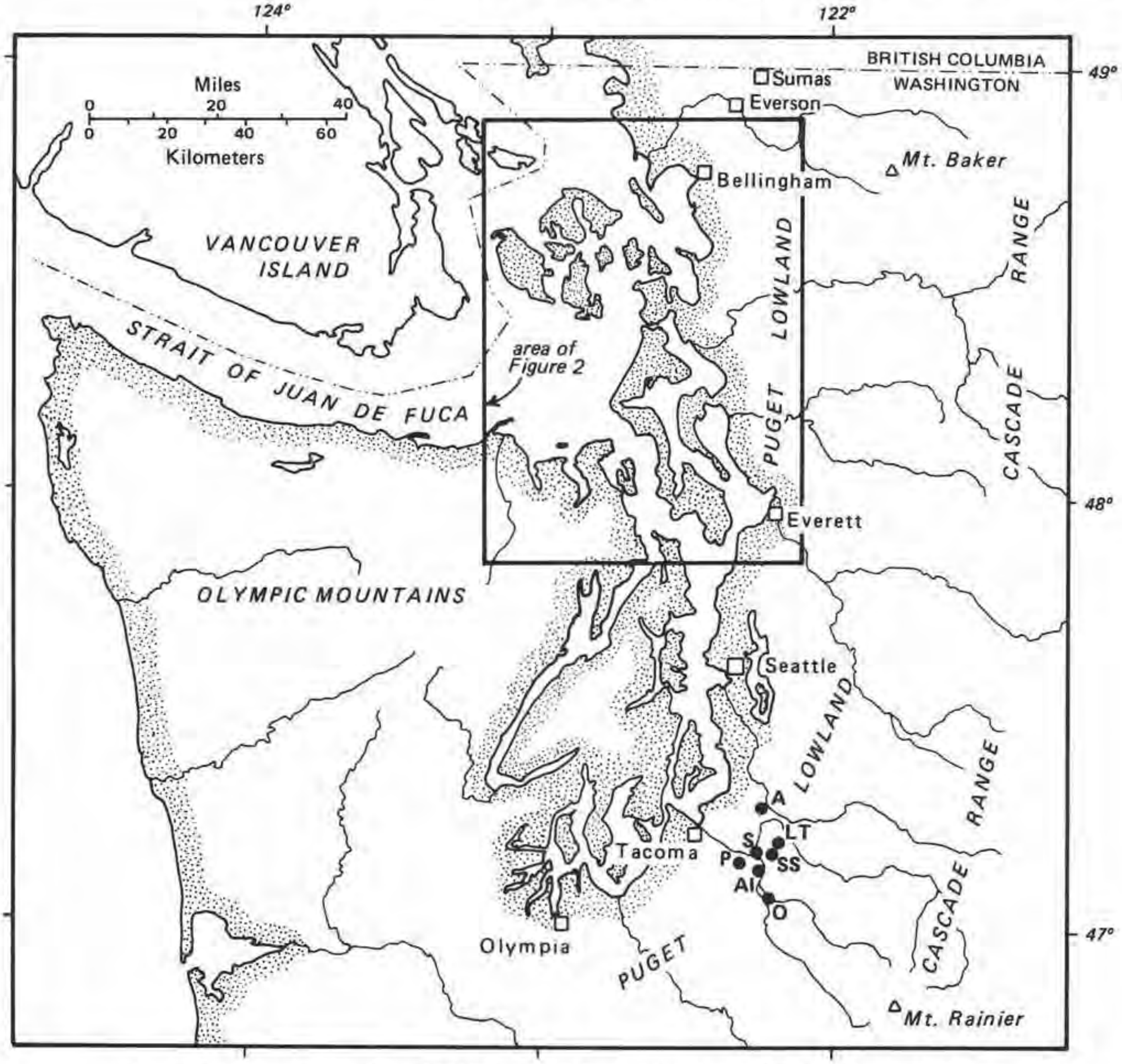


Figure 1.–Index map of the Puget Lowland. A, Auburn; Al, Alderton; LT, Lake Tapps; O, Orting; P, Puyallup; S, Sumner; SS, Salmon Springs.

ical procedures outlined by Hare (1969) and Kvenvolden (1975). Some shells were also analyzed by the University of Colorado laboratory. Details of analytical procedures are given in the Appendix.

STRATIGRAPHY AND CHRONOLOGY

Early Pleistocene Sediments

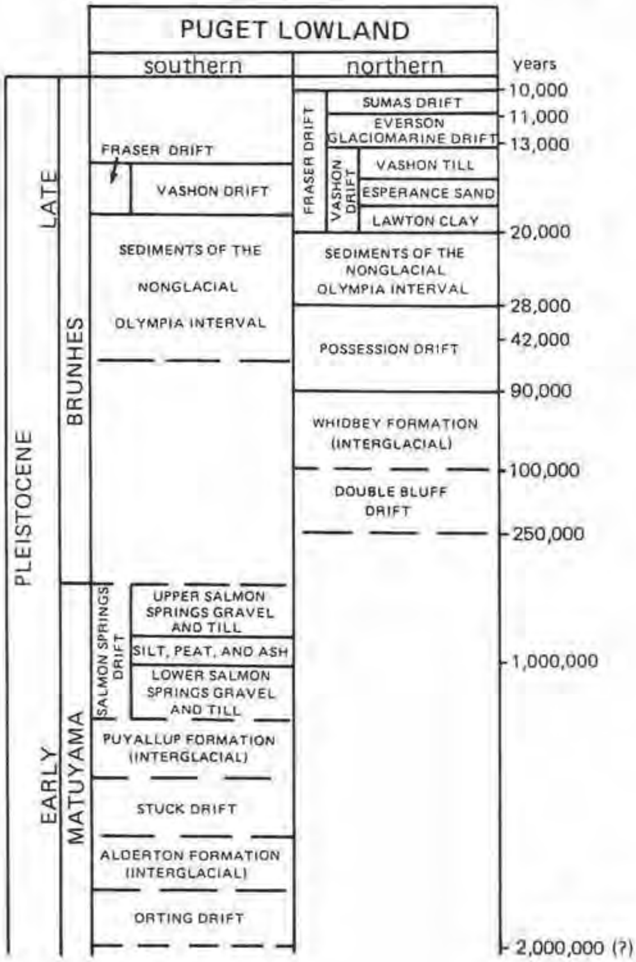
Introduction

Early Pleistocene deposits in the Puget Lowland are characterized by interbedded glacial and nonglacial sediments. Glacial lobes of the Cordilleran ice sheet extended southward from the Canadian border, splitting into two main lobes. The Juan de Fuca lobe flowed westward out the Strait of Juan de Fuca, while the Puget Lobe advanced southward across the central and southern Puget Lowland. The maximum extent of pre-Wisconsin glaciations is difficult to determine because those deposits were buried by younger sediments. No early Pleistocene deposits have been documented from the Juan de Fuca lobe.

Orting Drift

The oldest glacial unit recognized in the Puget Lowland, the Orting Drift, occurs in a few outcrops in

Table 1.–Stratigraphic sequence in the Puget Lowland



the Puyallup valley south of Seattle. Willis (1898) first used the name Orting for gravel exposed near the town of Orting. Later, Crandell and others (1958) identified at least three tills within the Orting gravel and redefined it as Orting Drift, consisting of a thick basal gravel deposited by westward-flowing streams from the Cascade Range prior to the beginning of volcanism at Mount Rainier and overlain by till and gravel of the Puget lobe. The tills are compact, unstratified, poorly sorted diamictons containing 10 to 15 percent clasts of northern provenance and 5 to 15 percent garnet in the heavy-mineral fraction. Stratified gravel and sand not closely associated with till are usually of central Cascade origin but lack clasts from Mount Rainier, presumably because the deposit predates Mount Rainier.

Orting Drift commonly occurs as continuous outcrops as much as 60 m thick and reaches a maximum thickness of 79 m. Tills less than 8 m thick occur at more than one horizon but are interpreted to be oscillations of a single major glaciation because the tills are not laterally continuous and no significant unconformities have been found (Crandell and others, 1958; Crandell, 1963).

Orting drift lies on upper Tertiary rocks and is overlain unconformably by sediments of the Alderton Formation. The drift has long been considered the oldest Pleistocene glacial deposit in the Puget Lowland on the basis of its intense oxidation, but no numerical ages have been available until recently.

Alderton Formation

The Alderton Formation lies unconformably above the Orting Drift and is named for exposures along the west wall of the Puyallup valley near the town of Alderton (Crandell and others, 1958). It consists of interbedded mudflows, lahars, alluvium, lake sediments, peat, and volcanic ash derived from an active ancestral Mount Rainier volcano that predated the modern cone. Mudflows consist of very compact, angular to subrounded pebbles and boulders of hornblende-hypersthene andesite in a sandy clay matrix. Thicknesses of individual mudflows vary from 1 to 8 m.

The predominance of Mount Rainier lithologies in the Alderton sediments and the absence of northern-provenance components are interpreted to indicate that the Alderton is interglacial. Englemann spruce pollen from the lower part of a peat bed in the unit represents a climate somewhat cooler than at present, whereas the upper part of the same peat contains large amounts of Douglas fir and alder pollen, indicative of a climate similar to that of the present (Leopold and Crandell, 1957; Crandell and others, 1958). Pollen from other peat and silt beds suggest climatic conditions ranging from cool and moist to warmer and drier than the present (Heusser, 1977), but most of the Alderton exposures are now covered with vegetation, limiting opportunities for reconstruction of the climate for the Alderton as a whole.

No numerical dates have been obtained for the Alderton Formation, but recent paleomagnetic studies have shown that the geomagnetic polarity of the Alderton at its type locality is reversed with an average declination of 176° and an average inclination of -50° (Roland, 1984; Easterbrook and others, unpub. data, 1986). The reversed polarity has been overprinted with a viscous normal remanent magnetism that can be removed with high alternating field demagnetization. Additional paleomagnetic measurements of samples at an exposure 645 m from the type section gave an average declination of 170° and inclination of -65° at the 500-oersted level of demagnetization.

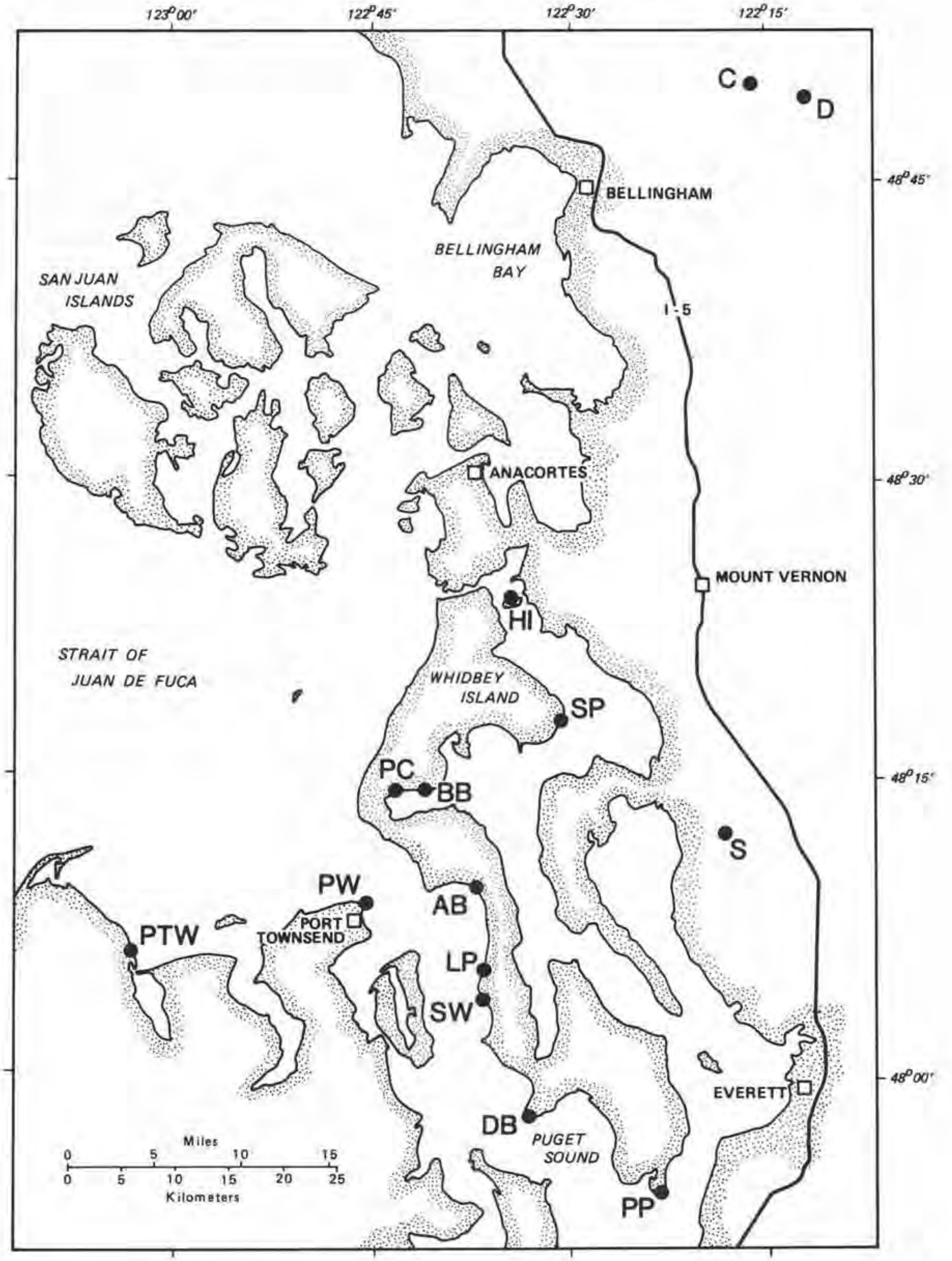


Figure 2.–Index map of the northern and central Puget Lowland. D, Deming; C, Cedarville; HI, Hope Island; SP, Strawberry Point; PC, Penn Cove; BB, Blowers Bluff; AB, Admiralty Bay; SW, South Whidbey; LP, Lagoon Point; DB, Double Bluff; PP, Point Possession; S, Stillaguamish; PW, Point Wilson; PTW, Port Williams. (Modified from USGS 1:250,000 topographic base map).

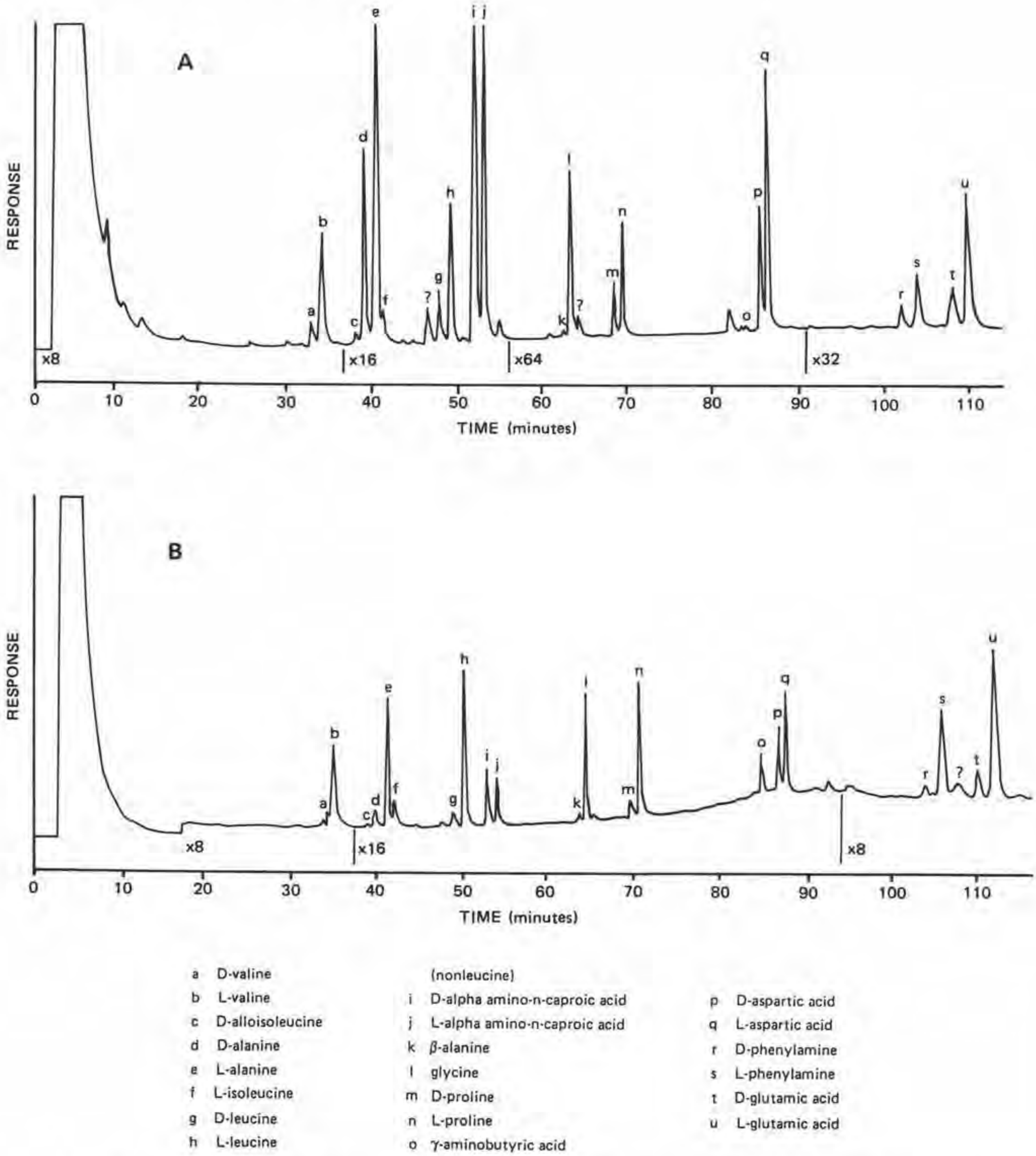


Figure 3.–A. Gas chromatographic trace of amino acids in fossil *Saxidomus giganteus*, sample 80-83c from interglacial deposits at Admiralty Bay, Washington. B. Gas chromatographic trace of amino acids in fossil wood sample 80-22 from interglacial clay deposits at South Whidbey State Park, Washington. See key for amino acid peak identifications.

The reversed remanent magnetism of the Alderton Formation and its stratigraphic position below the Salmon Springs Drift and its ≈800,000-yr fission-track dates places its age within the Matuyama Reversed Polarity Chron between 0.8 and 2.4 m.y. Volcanic ash in the Alderton sediments is presently being examined for possible fission-track dating.

Stuck Drift

Stuck Drift is defined on the basis of oxidized till, containing 10 to 15 percent clasts of rock types transported from the north by the Cordilleran ice sheet, interbedded with oxidized sand and gravel which overlies the Alderton Formation southwest of the town of Alderton (Crandell and others, 1958). In the type area, Stuck Drift consists of a single till between outwash sand and gravel, but elsewhere, two tills are separated by 15 to 50 m of lacustrine and fluvial silt, sand, and gravel (Crandell, 1963; Mullineaux, 1970). Clasts in the alluvium are derived from the Cascade Mountains to the east, indicating that the Cordilleran ice sheet had retreated sufficiently to permit reestablishment of westward-flowing drainage from the mountains. Exposures of Stuck Drift are found only along the walls of the Puyallup valley, so the southern extent of Stuck Glaciation is not known and correlation with deposits elsewhere in the Puget Lowland has not been made. Poor exposures in the area prevent lateral tracing of deposits to determine whether the Stuck Drift includes two tills separated by a nonglacial interval or whether one of the tills belongs to a different glaciation.

No numerical dates have been obtained from the Stuck Drift, but recent paleomagnetic studies have shown that silt in the Stuck near the type locality has reversed geomagnetic polarity. The mean magnetic declination calculated at the 700-oersted demagnetization level is 183°, and the mean inclination is -19°. Based on its pre-Salmon Springs stratigraphic position and reversed remanent magnetism, the Stuck Drift was deposited during the Matuyama Reversed Polarity Chron (Roland, 1984; Easterbrook and others, unpub. data, 1986).

Puyallup Formation

Nonglacial sediments in the southeastern Puget Lowland, derived mostly from ancestral Mount Rainier, were named the Puyallup Sand by Willis (1898). Crandell and others (1958) later expanded the unit to include fluvial sand and gravel, lacustrine silt and clay, peat, lahars, and mudflows which they designated the Puyallup Formation, which includes, but is not restricted to, the Puyallup Sand of Willis (1898). At its type locality in the Puyallup valley, the Puyallup Formation consists of about 41 m of lacustrine silt and sand, alluvial sand and gravel, mudflows, tephra, and peat. The mudflows, many of which contain as much as 95 percent Rainier-provenance clasts, were derived from Mount Rainier. Most gray, unoxidized sand beds in the unit are also of Rainier provenance, whereas most oxidized fluvial sand and gravel typically consist of mixed lithologies of Rainier and central Cascade provenance. Like the Alderton Formation, Puyallup sediments represent deposition along an ancestral river flowing from Mount Rainier across the southeastern Puget Lowland.

Pollen from the lower Puyallup Formation indicates an early postglacial climate followed by gradual warming. Other pollen profiles higher in the section suggest a warm climate followed by a cooling trend, but poor exposures prevent lateral tracing of sampled units, and an unknown amount of the upper part of the sediments, represented by a weathering profile and unconformity, may be missing. A major erosional unconformity, with a strong weathering profile characterized by kaolin, was recognized by Crandell and others (1958) at the top of the Puyallup Formation and interpreted to represent a long period of erosion and weathering prior to deposition of the overlying Salmon Springs Drift (Crandell, 1963; Mullineaux, 1970).

No numerical dates have yet been obtained from the Puyallup Formation, but paleomagnetic measurements from part of the original type section and from nearby exposures indicates reversed magnetic polarity with a complex magnetic overprint (Roland, 1984; Easterbrook and others, unpub. data, 1986). The remanent magnetism of sediments at another nearby section, correlated with the Puyallup on the basis of lithology and stratigraphic position, was clearly reversed. The mean magnetic declination was 183°, and mean inclination was -42° with an alpha-95 of 15° at the 700-oersted demagnetization level. The reversed magnetic polarity, together with its stratigraphic position beneath the Lake Tapps tephra, places the Puyallup Formation within the Matuyama Reversed Polarity Chron (Roland, 1984; Easterbrook and others, unpub. data, 1986).

Previous tentative correlation of the Puyallup Formation with the interglacial Whidbey Formation of the central Puget Lowland (Crandell, 1965; Easterbrook, 1969, 1976a,b) is now known to be impossible because of the reversed magnetic polarity of the Puyallup and the ≈800,000-yr fission-track age of the Lake Tapps tephra in the overlying Salmon Springs Drift (Easterbrook and others, 1981), together with the ≈100,000-yr ages estimated from amino acid analyses for the Whidbey Formation and its normal magnetic polarity (Easterbrook, 1976a).

Salmon Springs Drift

At its type locality in the Puyallup valley, the Salmon Springs Drift unconformably overlies the Puyallup Formation and is unconformably overlain by Vashon Drift (Crandell and others, 1958). Prior to 1979, the age of the Salmon Springs Drift was thought to be early Wisconsin on the basis of radiocarbon enrichment dates of 50,100 ± 400 years (GrN-4116c) (Armstrong and others, 1965) and 71,500 ± 1,700 years (Stuiver and others, 1978) from peat at the type locality, and Salmon Springs Drift was widely correlated throughout Washington on the belief that it was the next oldest drift beneath Vashon Drift. The first indication that these ages might be in error was the discovery of reversed magnetic polarity in a silt bed directly beneath the radiocarbon-dated peat bed at the Salmon Springs Drift type locality (Easterbrook and Othberg, 1976). A paleomagnetic profile from the silt included both normal and reversely polarized samples, as did two later magnetic profiles (Easterbrook and others, 1981). Because some of the samples showed high angular standard deviations for individual spin measurements and low sample stability, the validity of the normal polarities is difficult to evaluate. Strong normal magnetic overprinting has been found elsewhere in other profiles of Salmon Springs and older sediments, suggesting that the normal polarities in the Salmon Springs silt may be caused by postdepositional overprinting.

A fission-track date of 0.84 m.y. on zircon from volcanic ash that directly underlies the reversely magnetized silt at the type locality provided the conclusive evidence for the antiquity of the Salmon Springs Drift. The ash, which lies directly on lower Salmon Springs outwash gravel, grades upward into the overlying silt, which gradually increases in organic content and becomes the radiocarbon-dated peat. No break in pollen spectra is apparent at the contact between the silt and peat (Stuiver and others, 1978), further supporting the conclusion that no unconformity exists between the peat and ash. Thus, the ash-silt-peat sequence represents continuous deposition, and the radiocarbon dates are considered invalid (Easterbrook and others, 1981).

The tephra at Salmon Springs was named the Lake Tapps tephra by Easterbrook and others (1981). It has now been identified at five localities in the Puget Lowland, and additional fission-track dates of 0.87 ± 0.27, 0.84 ± 0.21, and 0.89 ± 0.29 m.y. have been obtained from it (Easterbrook and others, 1981; Naeser and others, 1984; J. A. Westgate, written commun., 1986). Reversed magnetic polarities have been measured in silt associated with the Lake Tapps tephra at all five localities, and the magnetic remanence at two of these localities is exceptionally stable and clearly reversed throughout the profiles (Easterbrook and others, unpub. data., 1986).

Thus, the Salmon Springs Drift is now firmly dated at about 800,000 years, in the latter part of the Matuyama Polarity Chron of the early Pleistocene. Because the age is established on the tephra-silt unit between the upper and lower drift units of the Salmon Springs Drift as defined by Crandell and others (1958), the age of the upper drift unit can only be stated as younger than about 800,000 years, but how much younger remains uncertain.

The discovery of the antiquity of the Salmon Springs Drift has far-reaching implications. The term Salmon Springs Drift has been widely used in the literature for pre-Vashon drifts in the southern and central Puget Lowland and the Olympic Peninsula, and correlations have been made over wide areas in the Pacific Northwest. In many instances, these interpretations have been based solely on stratigraphic position, under the assumption that the Salmon Springs Drift was the next-to-youngest glaciation in the region. Distinguishing those drift units that are truly correlative with Salmon Springs Drift from younger drifts is very difficult without dating control, and the designation of "Salmon Springs Drift" in many pre-1981 publications is very likely erroneous.

Middle Pleistocene (800,000 to 300,000 years ago)

At present, no exposures of Pleistocene sediments between 300,000 and 800,000 years old are known in the Puget Lowland. Deposits of this interval may be present below sea level in the Puget Lowland, but they are not accessible and have not been identified in cores.

Late Pleistocene

The late Pleistocene stratigraphy and chronology of the Puget Lowland is broken down into several phases:

Pre-late Wisconsin

Double Bluff Drift (about 100,000-250,000 years ago)

Whidbey Formation (about 90,000-100,000 years ago)

Possession Drift (an early phase at about 70,000-90,000 years ago and a late phase about 35,000-50,000 years ago)

Late Wisconsin

Olympia nonglacial sediments (about 20,000-30,000 years ago)

Fraser Drift (about 10,000-20,000 years ago)

The ages of pre-late Wisconsin sediments have been determined largely by amino acid analysis, whereas the Olympia nonglacial sediments and Fraser Drift have been dated by radiocarbon. The amino acid geochronology of the Puget Lowland has not been previously published, so it is treated in somewhat more detail.

AMINO ACID GEOCHRONOLOGY

Amino acid racemization geochronology relies on the fact that almost all living organisms contain amino acids which are configured as the L-amino acid stereoisomer. Upon the death of the organism, the L-amino acid stereoisomer (enantiomer) undergoes a reversible interconversion forming the mirror image D-amino acid enantiomer by the racemization reaction:

$$\text{L-amino acid} \underset{k_D}{\overset{k_L}{\rightleftharpoons}} \text{D-amino acid}$$

where k_L and k_D are the relative rates of racemization of the L- and D-amino acid enantiomers. The sensitivity of the reaction rate to temperature and the reliability of the reaction in different species of mollusks are limiting factors in calculations of absolute ages by the method of amino acid racemization geochronology.

The amino acid racemization reaction in fossil mollusks has been applied to marine terrace, estuarine, and archaeological midden deposits along the west coast of North America. For example, Wehmiller and others (1977) have utilized the technique for correlation and determination of the chronology of Pleistocene marine terrace localities along California, Oregon, and Washington. Karrow and Bada (1980) applied the technique to marine terraces in San Diego County. Kvenvolden and others (1979a) used the method for correlation and determination of the chronology of Pleistocene estuarine assemblages at Willapa Bay, Washington. Atwater and others (1981) applied the method to buried Pleistocene estuarine assemblages in San Fransisco Bay, California. Miller and Hopkins (1980) used this method to study fossiliferous deposits in Alaska. Archaeological middens have been studied by analyzing amino acid racemization in mollusks in southern California (Masters and Bada, 1977) and in Alaska (Kvenvolden and others, 1979b). These investigations suggest that the amino acid racemization reaction in mollusks can provide meaningful information regarding the fossiliferous deposits in the Puget Sound Lowland.

Fossils and Localities

Fossil mollusks were collected in the Puget Sound Lowland during the summers of 1979 and 1980. Localities and specimens analyzed are described in Table 2. The pelecypods studied represent one genus from each of six families.

Family	Genus	Species
Veneridae	*Saxidomus*	*gigantea* (Deshayes, 1839)
Tellinidae	*Macoma*	*calcarea* (Gmelin, 1791)
Myacidae	*Mya*	*truncata* (Linné, 1758)
Nuculanidae	*Nuculana*	*fossa* (Baird, 1863)
Cardiidae	*Clinocardium*	*ciliatum* (Fabricius, 1780)
	Clinocardium	*nuttalli* (Conrad, 1837)
Hiatellidae	*Hiatella*	*artica* (Linné, 1758)

In this study, comparisons of amino acid geochemistry will be restricted to the genus level. The species reported represent only a few mollusks from glacial-marine deposits of the Fraser Glaciation, and more complete faunal lists have been reported elsewhere (Easterbrook, 1963, 1969). The localities at South Whidbey State Park, Port Williams, and Admiralty Bay do not have published descriptions of their molluskan fauna. Only a few isolated mollusks have been found in the Double Bluff drift (Easterbrook, 1968).

The application of amino acid geochemistry in fossil mollusks to problems of correlation and geochronology depends on a predictable rate of protein diagenesis that is not significantly influenced by environmental factors. Estimates of the effective temperature at which in-situ amino acid reactions occur are imprecise, and this limits the accuracy of age calculations. Determination of the kinetics of amino acid racemization in different species of mollusks is important if relative correlations and age estimates are to be reliable.

Duplicate analyses of amino acid D/L ratios for *Saxidomus*, *Macoma*, *Clinocardium*, and *Nuculana* indicate that the reproducibility of amino acid D/L ratios is variable, depending on genera and extent of individual amino acid racemization. Figure 3A is an example of the gas chromatographic trace of amino acid D- and L- abundances in fossil *Saxidomus giganteus* from intraglacial deposits at Admiralty Bay. An example of amino acid D- and L- abundances in fossil wood is discussed later in the text and is shown in Figure 3B. Laboratory procedures are described in the Appendix.

Intrageneric Relationships

The apparent amino acid racemization kinetics in fossil mollusks can be studied by comparing D/L ratios

Table 2.–Amino acid geochemistry (D/L ratios) of Pleistocene mollusks, Puget Lowland

Leu, leucine; allo/iso, alloisoleucine/isoleucine; glu, glutamic acid; asp, aspartic acid; pro, proline; nd, no data

GEOLOGIC CLIMATE UNIT GEOLOGIC UNIT LOCATION GENERA	LEU	ALLO/ ISO	GLU	ASP	PRO	C-14 AGE
FRASER GLACIATION						
(Everson Interstade)						
Bellingham glaciomarine drift						
Cedarville (80-37)						11,800±400
Macoma calcarea	0.13	0.18	0.12	0.30	0.26	
Nuculana fossa	0.11	nd	0.16	0.34	0.14	
Nuculana fossa	0.11	nd	0.12	0.36	0.15	
Nuculana						
(AAL-3070A)	nd	0.086	nd	nd	nd	
(AAL-3070B)	nd	0.092	nd	nd	nd	
(AAL-3070C)	nd	0.102	nd	nd	nd	
Kulshan glaciomarine drift						
Deming (80-38)						12,970±280
Macoma calcarea	0.12	0.15	0.14	0.33	0.20	
Nuculana fossa	0.12	nd	0.11	0.34	0.14	
Everson glaciomarine drift						
Hope Island (79-11)						12,400±190
Saxidomus giganteus	0.11	0.08	0.12	0.29	0.23	
Clinocardium ciliatum	0.12	0.10	0.14	0.37	0.12	
Penn Cove (80-24)						13,010±170
Macoma sp.	0.12	0.08	0.13	0.34	0.18	
Mya truncata	0.11	0.08	0.11	0.31	0.14	
Nuculana fossa	0.14	nd	0.16	0.36	0.22	
Clinocardium nutalli	0.11	0.07	0.14	0.36	0.16	
Hiatella arctica	nd	0.08				
Hope Island						
Mya truncata						
(3071A)	nd	0.080	nd	nd	nd	
(3071B)	nd	0.083	nd	nd	nd	
(3071C)	nd	0.076	nd	nd	nd	
POSSESSION GLACIATION						
Possession glaciomarine drift						
Port Williams						
Clinocardium ciliatum						
(79-1A)	0.15	0.13	0.19	0.41	0.16	
(79-1B)	0.21	0.15	0.21	0.36	0.29	
(79-1D)	0.10	nd	0.13	0.33	0.11	
(79-1D)	0.13	0.12	0.16	0.37	0.13	
(79-1E)	0.08	0.07	0.13	0.35	0.09	
(79-1E)	0.09	0.09	0.13	0.35	0.10	
(AAL-1076A)	nd	0.063	nd	nd	nd	
(AAL-1076B)	nd	0.068	nd	nd	nd	
(AAL-1076C)	nd	0.055	nd	nd	nd	

Table 2.–Amino acid geochemistry, molluscs (continued)

GEOLOGIC CLIMATE UNIT GEOLOGIC UNIT LOCATION GENERA	LEU	ALLO/ ISO	GLU	ASP	PRO	C-14 AGE
POSSESSION GLACIATION (continued)						
Possession glaciomarine drift						
Port Williams (continued)						
Nuculana sp.						
(AAL-2939)	nd	0.20±.08				
Blowers Bluff						
Clinocardium sp.						
(AAL-1882)	nd	0.23	nd	nd	nd	
(AAL-1883)	nd	0.25	nd	nd	nd	
(AAL-1884)	nd	0.29	nd	nd	nd	
(AAL-1885)	nd	0.28	nd	nd	nd	
(AAL-1886)	nd	0.26	nd	nd	nd	
(AAL-3072A)	nd	0.138	nd	nd	nd	
(AAL-3072B)	nd	0.197	nd	nd	nd	
(AAL-3072C)	nd	0.110	nd	nd	nd	
Nuculana sp.						
(AAL-1881)	nd	0.30	nd	nd	nd	
Nuculana sp.						
(AAL-2938)	nd	0.18±.01				
Stillaguamish						
Saxidomus sp.						
(AAL-1876)	nd	0.19				
Clinocardium sp.						
(AAL-1878A)	nd	0.23				
(AAL-1878B)	nd	0.16				
Nuculana sp.						
(AAL-1877A)	nd	0.33				
(AAL-1877B)	nd	0.37				
(AAL-1877C)	nd	0.38				
Hiatella sp.	nd	0.16				
Point Wilson						
Clinocardium						
(AAL-3067A)	nd	0.094				
(AAL-3067B)	nd	0.106				
(AAL-3067C)	nd	0.067				
(AAL-3068A)	nd	0.122				
(AAL-3068B)	nd	0.106				
(AAL-3068C)	nd	0.096				
Balanus						
(AAL-3069A)	nd	0.124				
(AAL-3069B)	nd	0.124				
(AAL-3069C)	nd	0.028				
Admiralty Bay						
Saxidomus giganteus						
(79-9)	0.29	0.24	0.24	0.44	0.40	
(80-26B)	0.36	0.27	0.30	0.45	0.52	
(80-30A)	0.40	0.31	0.31	0.47	0.52	
(80-30B)	0.34	0.21	0.28	0.46	0.47	

Table 2.–Amino acid geochemistry, molluscs (continued)

GEOLOGIC CLIMATE UNIT GEOLOGIC UNIT LOCATION GENERA	LEU	ALLO/ ISO	GLU	ASP	PRO	C-14 AGE
POSSESSION GLACIATION (continued)						
Admiralty Bay (continued)						
(80-30B)	0.36	0.25	0.30	0.50	0.46	
(80-30C)	0.37	0.27	0.30	0.47	0.47	
Mya truncata	0.41	0.44	0.30	0.59	0.47	
Macoma						
(80-26)	0.37	0.42	0.38	0.57	0.51	
(30-30)	0.47	0.48	0.39	0.65	0.58	
WHIDBEY INTERGLACIATION						
Whidbey Formation						
South Whidbey State Park						
Macoma						
(80-22A)	0.41	nd	0.45	0.52	0.46	
(80-22B)	0.24	nd	0.15	0.29	0.33	
(80-22C)	0.42	nd	0.33	0.54	0.51	
DOUBLE BLUFF GLACIATION						
Double Bluff glaciomarine drift						
Double Bluff						
Nuculana						
(80-27)	0.47	nd				
Foulweather Bluff						
Macoma	nd	0.34				
Mya	nd	0.34				

of different amino acids. Lajoie and others (1980) reported that intrageneric relationships (comparisons of two different amino acid D/L ratios in a single genus of mollusk) in *Saxidomus* and other fossil mollusks follow linear trends. The intrageneric relationships of alloisoleucine/isoleucine (hereafter, allo/iso) vs. D/L leucine for *Saxidomus, Macoma, Clinocardium, Nuculana* and *Mya* are shown on Figure 4. In general, the ratios for *Saxidomus, Clinocardium,* and *Nuculana* follow the same linear path, whereas data for *Macoma* show higher D/L ratios. The curve for *Mya* is intermediate between those for *Macoma* and the other genera.

Differences in amino acid D/L ratios among different genera are more obvious in samples from older sediments. For example, at Admiralty Bay, the leucine D/L ratio is about 0.37 in *Saxidomus* 0.47 in *Macoma,* and 0.41 in *Mya.* These data indicate that leucine racemization in *Macoma* tends to be slightly faster than in *Saxidomus,* as reported earlier by Wehmiller and others (1977).

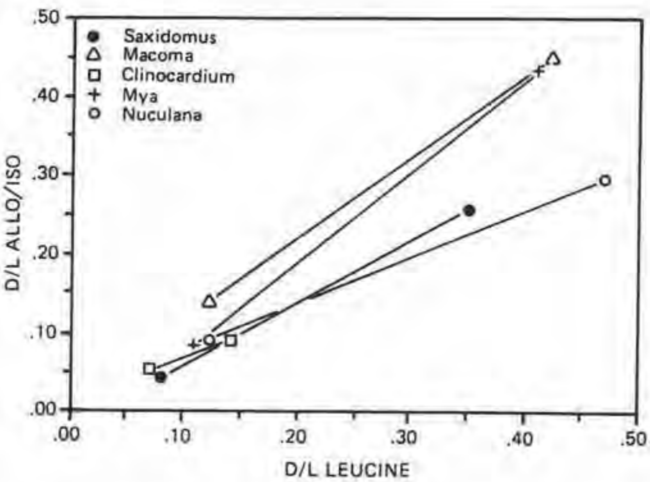


Figure 4.–Intrageneric relationships for selected mollusk genera based on leucine and allo/isoleucine D/L ratios.

Amino Acid Geochemistry of Fossil Wood

The reliable application of the amino acid racemization reaction in fossil wood to correlation and geochronology depends on the predictability of the reaction. If too many environmentally sensitive variables are involved or need to be evaluated with each sample, then the usefulness of the technique is greatly diminished. Factors that affect the racemization of amino acids in wood include temperature, moisture content, terpene abundance, formation of amino sugars, and formation of melanoidin-type substances (Lee, 1975; Zumberge, 1979; Zumberge and others, 1980). In the present study, the main factor considered was whether the racemization followed a predictable trend. Table 3 lists the amino acid D/L ratios in wood recovered from localities in the Puget Lowland.

Table 3.–Amino acid geochemistry (D/L ratios) of Pleistocene wood, Puget Lowland

GEOLOGIC CLIMATE UNIT GEOLOGIC UNIT LOCALITY SAMPLE NUMBER	ASP	PRO	GLU	LEU	C-14 AGE
FRASER GLACIATION					
(Everson Interstade)					
Bellingham glaciomarine drift					
Cedarville					
UA-1200	0.29	0.016	0.068	0.018	11,800±400
Deming Sand					
UA-1206	0.17	nd	0.054	0.017	11,640±275
UA-1207	0.21	nd	0.030	0.016	11,640±275
80-36	0.21	0.05	0.08	0.04	
OLYMPIA NONGLACIAL INTERVAL					
Strawberry Point					
UA-1195 (mid peat 6)	0.20	nd	0.059	0.018	22,700
UA-1197 (silt E, 90 cm)	0.18	nd	0.067	0.020	24,800
UA-1196 (silt E)	0.19	nd	0.067	0.020	24,800
UA-1198 (silt D, 70 cm)	0.22	nd	0.060	0.018	27,600
UA-1191 (peat 5c)	0.28	nd	0.070	0.025	27,650
POSSESSION GLACIATION					
Possession Drift					
Strawberry Point DP-37					
UA-1194 (top peat 3)	0.22	nd	0.062	0.018	34,900
UA-1192 (middle peat)	0.24	nd	0.067	0.021	35,400
UA-1190 (top peat 2)	0.25	nd	0.089	0.025	43,600
UA-1189 (peat 2)	0.30	0.029	0.072	0.029	43,900
UA-1193 (lowest peat)	0.29	nd	0.061	0.030	47,600
80-25a	0.18	0.05	0.13	0.03	
80-25b	0.25	0.07	0.11	0.06	
80-25c	0.18	0.04	0.15	0.04	
WHIDBEY INTERGLACIATION					
Whidbey Formation					
Double Bluff					
80-29	0.28	0.08	0.13	0.06	
80-28	0.25	0.09	0.12	0.07	
80-28a	0.22	0.07	0.10	0.04	
80-27b	0.37	0.06	0.09	0.04	
UA-929 (lowest peat)	0.26	0.08	0.115	0.048	
Maxwelton, Mx-10					
UA-926 (peat 5F)	0.26	0.056	0.117	0.047	
UA-927 (peat 5L)	0.24	0.032	0.156	0.040	
Strawberry Point (beach)					
UA-931	0.23	0.063	0.110	0.066	

Table 3.–Amino acid geochemistry, wood (continued)

GEOLOGIC CLIMATE UNIT GEOLOGIC UNIT LOCALITY SAMPLE NUMBER	ASP	PRO	GLU	LEU	C-14 AGE
WHIDBEY INTERGLACIATION (continued)					
Whidbey Formation (continued)					
Blowers Bluff					
UA-928 (below pumice)	0.26	0.052	0.115	0.034	
Swantown					
UA-933 (DP-25)	0.29	nd	0.104	0.042	
UA-932 (DP-29)	0.14	nd	nd	0.029	
Polnell Point					
UA-934 (DP-23)	0.28	0.041	0.137	0.041	
Freeland					
UA-935 (F-35)	0.26	0.072	0.107	0.042	
Lagoon Point					
UA-1205 (F-20)	0.42	0.097	0.174	0.050	
South Whidbey State Park					
UA-1209 (F-26)	0.13	nd	0.099	0.036	
80-22	0.65	0.12	0.19	0.07	
Camano Island					
UA-924 (L-14)	0.42	0.162	0.197	0.078	
UA-925 (L-12b)	0.22	0.197	0.208	0.090	
80-41 Pebble Beach	0.16	0.04	0.10	0.04	
Point Wilson					
80-19	0.06	0.03	0.05	0.04	
80-20	0.32	0.05	0.08	0.04	
UA-1199	0.34	nd	0.082	0.032	
Point Roberts					
UA-937	0.28	nd	0.094	0.041	
F-14					
UA-1204	0.18	0.021	0.097	0.024	
West Beach					
80-32	0.12	0.04	0.08	0.04	
80-33	0.12	0.04	0.14	0.08	
80-34	0.13	0.03	0.21	0.03	
DOUBLE BLUFF GLACIATION					
Double Bluff glaciomarine drift					
Double Bluff					
80-27a	0.30	0.13	0.22	0.07	
80-35	0.29	0.13	0.15	0.08	
80-35	0.28	0.11	0.14	0.08	
Possession Point (beneath glaciomarine drift)					
UA-1293 (Mx-20)	0.33	0.040	0.156	0.039	
UA-930 (Mx-20)	0.20	nd	0.092	0.050	
Camano Island					
80-41	0.16	0.04	0.10	0.04	
UA-1201	0.13	nd	0.257	0.021	

Aspartic Acid and Proline D/L Ratios in Wood

The trend of aspartic acid vs. proline D/L ratios in wood samples is shown on Figure 5. These two amino acids are selected because they have been used by previous workers to age-date samples of fossil wood. To be useful the trend should be linear, as demonstrated by the mollusk data. The general trend of these data points shows an increase with age for both amino acid ratios; a few data points scatter above the majority of the points. This comparison indicates that the relative kinetics of racemization between aspartic acid and proline are complex and do not follow a simple relationship. This condition has also been noted in wood samples from Willapa Bay (Blunt, 1982), where aspartic acid racemization in wood suggests non-linear kinetics at D/L values of about 0.3.

Environment of Preservation

The explanation for anomalous chemical behavior of amino acids in wood specimens includes factors of in-situ preservation. Sample W80-22 is composed of wood fragments 1 to 3 mm thick collected in silty clay of the Whidbey Formation at South Whidbey State Park. This sample has the highest measured aspartic acid D/L ratio (0.65) and the highest relative content of γ-aminobutyric acid. The non-protein γ-aminobutyric acid is common in clay, and its presence in a sample may indicate the need for additional sample cleaning (Schroeder, 1975). Non-protein amino acids are found in the mureide complex of bacterial cell walls and are considered to indicate a bacterial process in the formation of peat (Casagrande and Given, 1980).

Other wood specimens also show anomalous chemical behavior (Fig. 5). Sample W80-27b was collected in sandy silt of the Whidbey Formation at Double Bluff. This sample is from the interior of a 2-3-cm-thick wood specimen. Non-protein amino acids were not detected. Sample W80-20 was collected in sandy silt of the Whidbey Formation at Point Wilson. This sample is from the interior of a wood specimen 4 to 7 cm thick and does not have measurable quantities of γ-aminobutyric acid. Of three samples (W80-22, W80-27b, and W80-20), none have anomalously high D/L ratios of alanine or glutamic acid, a condition that could be an indication of micro-organism sources in wood (Zumberge and others, 1980). Other factors, such as the composition of the organic matrix of individual wood samples and the position of aspartic acid within peptides, need to be evaluated (Engel and others, 1979).

Identification of wood samples contaminated by micro-organisms might be accomplished through the

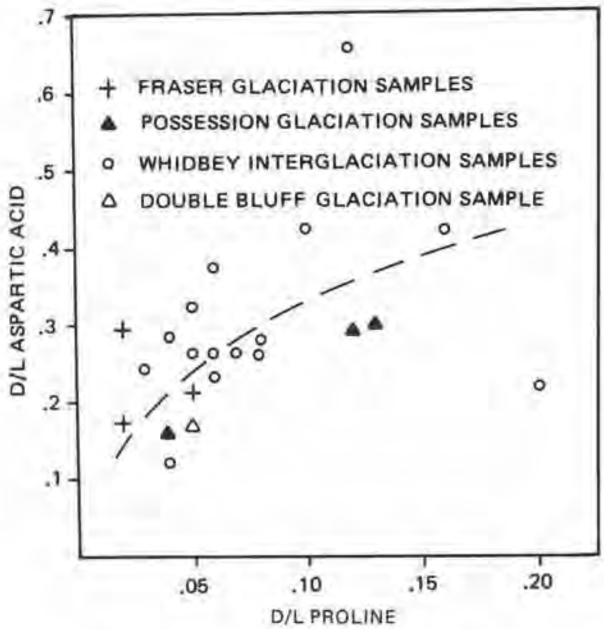


Figure 5.–Relationships of aspartic acid and proline D/L ratios in wood from Pleistocene deposits, Puget Lowland.

stereochemistry of amino acids. Differences in the amino acid stereochemistry of fresh and rotten bristlecone pine have been reported by Zumberge and others (1980). According to their results, glutamic acid D/L ratios are relatively higher in the bound fraction of rotten bristlecone pine than in the bound fraction of relatively fresh wood in the same sample. Samples W80-33 and W80-34 from consolidated sand in the Whidbey Formation at West Beach have anomalously high glutamic acid D/L ratios (0.13) with respect to the D/L ratio of aspartic acid (0.12). These two samples probably have been contaminated to some extent by micro-organisms.

Amino Acid Chronology and Correlation of pre-Wisconsin Deposits

Both isoleucine and leucine have been used to correlate samples from different localities, as well as to calculate ages. The D/L ratios of leucine in *Saxidomus* have been used with some success to correlate localities in the Pacific Northwest (Wehmiller and others, 1977; Kennedy, 1978; Kvenvolden and others, 1979a; Kvenvolden and Blunt, 1980). Leucine D/L ratios of five mollusk species from the Puget Lowland (Fig. 6) cluster for species of the similar radiocarbon age. The leucine D/L ratios of *Macoma* (80-22b) and *Saxidomus* (79-1) are relatively low compared with ratios for other samples from the same locality.

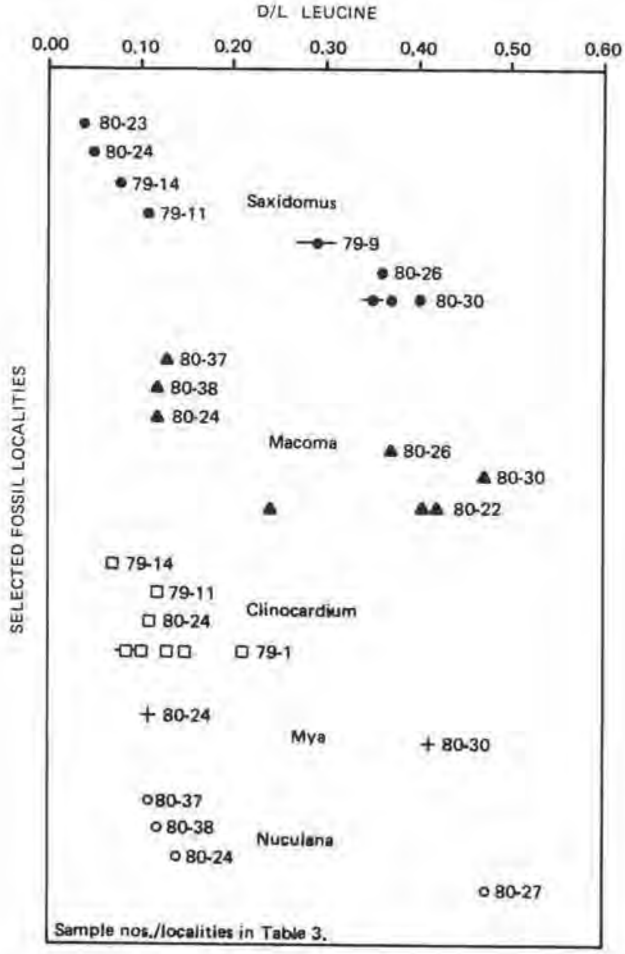


Figure 6.–Relative correlation of leucine D/L ratios in five genera of mollusk at fossil localities in the Puget Lowland.

The allo/iso D/L ratios in fossil mollusks from localities in the Puget Lowland cluster for genera of comparable age (Fig. 4 and Table 2). The allo/iso ratios in *Clinocardium* from Port Williams are not as consistent, and *Macoma* (80-22b) from South Whidbey State Park and *Saxidomus* (79-1) from Admiralty Bay have relatively low allo/iso ratios compared to other samples from the same localities.

Temperature Considerations

The rate of racemization is a function of temperature. The average effective in-situ temperature at which a fossil has been insulated during its burial history is the least reliable factor in age calculations using amino acid racemization (Wehmiller and others, 1976; Wehmiller and others, 1977; Kvenvolden and others, 1979a, b; Kvenvolden and others, 1980; Miller and Hare, 1980). Nevertheless, rate constants can be adjusted to the estimated effective diagenetic temperature conditions of a buried fossil. Adjustments of rate constants require estimating glacial temperature reductions and estimating effective temperature histories. Kvenvolden and others (1980) have employed the method of adjusting calibrated rate constants to the effective temperatures of fossil mollusks in the Puget Lowland.

Temperature reductions of as much as 4° to 6°C due to Pleistocene climatic deterioration in the Pacific Northwest have been reported from palynologic records in sediments (Heusser, 1977; Heusser and Shackleton, 1979; Heusser and Heusser, 1981) and from oxygen isotope records in calcite speleothems (Gascoyne and others, 1980). Due to the nature of palynological studies, the variations in July temperatures are usually reported. In the method established by Wehmiller and others (1976) and used by Kvenvolden and others (1980), 29 years of modern monthly air temperature variations are taken as a model of effective temperature history. The effective temperature must be adjusted downward to account for cooler Pleistocene climates. The adjusted temperature is assumed to have been the average diagenetic temperature at which the Pleistocene fossil was insulated. An average Pleistocene temperature reduction of 3.6°C for pre-Wisconsin sediments has been calculated from palynological records by Kvenvolden and others (1980).

Temperature Calibration

Temperature calibration sites listed on Table 4 were selected on the basis of proximity to radiocarbon-dated localities from which fossil mollusks were collected for analysis of leucine and glutamic acid D/L ratios. The mean annual air temperature and the effective annual air temperature were calculated from monthly air temperature data which have been averaged over the long-term recording period from 1941 to 1970 (U.S. Department of Commerce, 1975). Sediment insulation properties predict that the mean annual temperature approximates the in-situ temperature of a sample if it is buried by more than about one meter of sediment (Wehmiller and others, 1976). As the sediment cover decreases, the in-situ temperature increases and approximates the effective temperature just below the sediment surface. The effective temperatures listed in Table 4 are about 1° or 2°C higher than the mean annual temperatures because of the logarithmic increase in the calculation for warmer months (Wehmiller and others, 1976).

Table 4.–Temperature calibration localities in the Puget Lowland, Washington.

Average of monthly air temperatures (in °C) over the recording period 1941-1970 from U.S. Department of Commerce Climatological Data Annual Summary, 1975. Eff T, average effective temperature calculated using log k = 15.77 - 5939/T, following the method of Wehmiller and others (1977)

Locality	J	F	M	A	M	J	J	A	S	O	N	D	Mean	Eff T
Port Angeles	3.6	5.2	6.1	8.4	11.2	13.5	15.2	15.0	13.7	10.1	6.6	4.8	9.4	10.8
Coupeville	3.4	5.2	6.2	9.0	11.9	14.3	16.1	16.1	13.9	10.2	6.6	4.7	9.8	11.3
Anacortes	4.0	5.9	7.0	9.7	12.6	15.1	16.7	16.4	14.7	11.1	7.4	5.3	10.5	11.9
Bellingham	2.9	4.9	6.1	8.7	11.9	14.7	18.2	16.2	13.9	10.1	6.3	3.9	9.7	11.9
Sedro Woolley	3.1	5.4	6.8	9.5	12.7	15.3	17.1	16.9	14.7	10.8	6.7	4.3	10.3	12.2
Concrete	2.4	5.1	6.8	10.1	13.8	16.3	18.7	18.4	16.2	11.6	6.6	3.8	10.8	13.3

Adjustment of Racemization Rates

Leucine and glutamic acid racemization rates are adjusted to account for temperature variability using the Arrhenius relationship:

$$\ell n\ (k_2/k_1) = Ea\ (T_2 - T_1)/R(T_2T_1)$$

where k_2 and k_1 are the rates of amino acid racemization at the respective temperatures of T_2 and T_1 in degrees Kelvin; Ea is the activation energy of racemization in calories per mole; and R is the gas constant 1.987. Activation energies of amino acid racemization in fossil mollusks have been reported as 29.4 kcal $mole^{-1}$ for isoleucine in *Mercenaria* (Mitterer, 1975); 29.0 kcal $mole^{-1}$ for isoleucine in *Hiatella arctica* (Miller and Hare, 1980); and 29.4 kcal $mole^{-1}$ for an average of eight amino acids in *Saxidomus* (Blunt, 1982). The activation energy of 29.4 kcal $mole^{-1}$ is applied to leucine and isoleucine racemization in the present study.

Uncertainty in using the mean or effective temperature from Table 4 arises from the unknown insulation history of the fossils collected at each calibration locality. At Penn Cove (loc. 80-24), samples were collected within one meter of the soil surface. Near Cedarville (loc. 80-37), fossils were collected on the surface from deposits recently exposed by river erosion. The fossil mollusks collected near Deming (loc. 80-38) were from a shallow roadcut. Although these samples may have been only recently exposed, in-situ temperatures were likely warmer about 8,000 years ago during the Holocene temperature maximum (Hansen and Easterbrook, 1974; Heusser and others, 1980). Because of these uncertainties in estimating in-situ temperature histories, the average between mean annual and effective annual air temperatures is used in calculations. Samples from Admiralty Bay, South Whidbey State Park, Port Williams, and Double Bluff had more than one meter of overburden, allowing reliable use of mean annual air temperatures in calculations. (See Table 5 for method of calculation.)

Double Bluff Drift

Double Bluff Drift is named for till, glaciomarine drift, glaciofluvial sand and gravel, and glaciolacustrine silt at Double Bluff on Whidbey Island (Easterbrook and others, 1967). It occurs at or near sea level in sea cliff exposures in the central Puget Lowland, and underlying sediments are not exposed (Easterbrook, 1968). Consequently, no deposits spanning the approximately 500,000-yr gap between the Salmon Springs Drift and the Double Bluff Drift are known.

Amino acid measurements of wood and shells in Double Bluff Drift suggest an age of about 145,000 years. Three wood samples from glaciomarine drift at the type locality gave aspartic acid D/L ratios of 0.30, 0.29, and 0.28, slightly higher than typical ratios for the overlying Whidbey Formation and consistent with relative age. Proline values for the same samples were 0.13, 0.13, and 0.11, all considerably higher than those of the Whidbey Formation. Wood in silt at the base of Double Bluff glaciomarine drift at Possession Point on Whidbey Island yielded a high aspartic acid D/L ratio for one sample (0.33), but the D/L ratio for a second sample from the same site analyzed later was only 0.20, indicating that further study of species differentiation and other factors is needed.

Calculated age estimates of the Double Bluff Drift at Double Bluff were based on leucine and glutamic acid racemization in *Nuculana*. The calibration of

Table 5.–Example of time calculation

The extent of time required to achieve a given D/L ratio is a function of the specific kinetics of the amino-acid racemization (or epimerization) reaction. Linear first-order kinetics are expressed in equation (1):

$$\ln\left(\frac{1 + D/L}{1 - k'\ D/L}\right) - \ln\left(\frac{1 + D/L}{1 - k'\ D/L}\right)_{t=0} = (1 + k')kt \qquad (1)$$

where D/L is the ratio of the D- and L- amino acids

k' is inverse of the ratio of the forward and reverse rate constants at equilibrium; k' for D/L Leu = 1; K' for D-allo/L-iso at equilibrium = 0.714

k is the rate of racemization or epimerization

t is time (age of the sample)

Both the rate constant and average temperature for the "calibration site" are calculated by solving the Arrhenius expression in equation (2):

$$\ln\left(\frac{k_2}{k_1}\right) = \left(\frac{Ea}{1.987}\right)\left(\frac{T_2 - T_1}{T_2 T_1}\right) \qquad (2)$$

where k_2 is calculated in equation from a site having a known radiocarbon date, measured D/L ratio, and incubation temperature (T_2)

T_2 is the average "in situ" temperature in degrees kelvin at the calibration site

T_1 is the estimated temperature at the site that includes a glacial reduction

Ea is the racemization activation energy of 29,400 calories per mole

k_1 is the rate constant of the site of unknown age.

The steps to calculate the age of fossil Nuculana (AAL-1877) in Possession Drift at Stillaguamish using Nuculana from Cedarville (C/4 = 11,640 years) are as follows:

(A) Calculate k_2 at Cedarville:

$$\frac{\ln\left(\frac{1 + 0.11}{1 - (0.714)\ 0.11}\right) - \ln\left(\frac{1 + 0.02}{1 - (0.714)\ 0.02}\right)_{t=0}}{1.714\ (11,640)} = k_2 = 7.6 \times 10^{-6}$$

(B) Estimate T_2 at Cedarville and T_1 at Stillaguamish using methods in Table 4:

Cedarville = 10.9°C; Stillaguamish = 9.8° - 3.6° glacial reduction = 6.2°C

(C) Calculate k_1 using equation (2):

$$\ln\left(\frac{7.6 \times 10^{-6}}{k_1}\right) = \left(\frac{29,400}{1.987}\right)\left(\frac{4.7}{284.05 \times 279.35}\right)$$

$$k_1 = 3.2 \times 10^{-6}$$

(D) Calculate the age of Nuculana (D-allo/L-iso = 0.36 ± 0.03) at Stillaguamish by equation (1):

$$\frac{\ln\left(\frac{1 + 0.36}{1 - (0.714)\ 0.36}\right) - \ln\left(\frac{1 + 0.02}{1 - (0.714)\ 0.02}\right)_{t=0}}{1.714\ (3.2 \times 10^{-6})} = 104,000 \text{ years}$$

these racemization rate constants follows the previously described procedures of Kvenvolden and others (1980). These rate constants were calibrated at radiocarbon-dated localities containing *Nuculana* at Penn Cove, Deming, and Cedarville with the temperature information from Coupeville, Bellingham, Sedro Woolley, and Concrete on Table 4. The calibrated rate constants were adjusted with a reduction of 3.6°C to the effective glacial diagenetic temperature of 6.2°C at Double Bluff. The calibrated rate constants coupled with the range of possible temperatures give a variety of calibrated rate constants at 6.2°C and a range of calculated ages. The estimated ages of the Double Bluff Drift, using leucine D/L ratios, range from 111,000 to 178,000 years. The overall age range reflects the uncertainties in the calibration method. Caution must be given to use of this tentative age estimate because only a single sample of *Nuculana* was analyzed from the Double Bluff Drift at its type locality. The age thus calculated for Double Bluff Drift may be too young because of the few samples available from the drift from throughout the area that were used for amino acid determination and the possibility of incorrect assumptions upon which computations are based. Additional samples are needed to confirm the age, but fossils are rare in this unit. D/L ratios from shells from glaciomarine drift at Foulweather Bluff also suggest relatively great antiquity for the Double Bluff Drift. Independent confirmation of the amino acid age estimates presently awaits thermoluminescence dating of clay in the Double Bluff Drift.

The remanent magnetism of Double Bluff glaciomarine drift is normal. An average declination of 1° and an average inclination of 49° were measured (Easterbrook, 1976a, 1983), providing additional evidence that the Double Bluff Drift cannot be correlative with Salmon Springs Drift in the southern Puget Lowland.

Glacial deposits stratigraphically beneath Vashon Drift (Sceva, 1957) or drift beyond the Vashon limit was commonly mapped as Salmon Springs Drift in the south-central Puget Lowland and Olympic Peninsula (Molenaar, 1965; Frisken, 1965; Noble and Wallace, 1966; Molenaar and Noble, 1970; Carson, 1970; Heusser, 1974, 1977, 1978; Gayer, 1977; Grimstad and Carson, 1981). Colman and Pierce (1981) proposed the name McCleary Drift to replace the "Salmon Springs" Drift of Carson (1970), but that drift is probably equivalent to the Double Bluff Drift. Deeter (1979) found that some of the "Salmon Springs" drifts in the southwestern Puget Lowland were in fact Double Bluff glaciomarine drift containing shells that gave amino acid racemization age estimates of about 250,000 years (Easterbrook and others, 1982). Glacial sediments extending beyond the margin of the Fraser glaciation have been tentatively correlated with the Double Bluff Drift on the basis of weathering and soil characteristics (Lea, 1984).

Whidbey Formation

Floodplain silt, sand, and peat of the Whidbey Formation accumulated during the last major interglaciation in the Puget Lowland (Easterbrook and others, 1967; Easterbrook, 1968, 1969; Hansen and Easterbrook, 1974). Pollen analyses from peat and silt indicate that Whidbey sediments were deposited during an interglacial period characterized by a warm climate, but with cooler intervals at its beginning and end (Hansen, 1947; Hansen and Mackin, 1949; Hansen and Easterbrook, 1974; Heusser and Heusser, 1981).

Twenty radiocarbon dates on wood and peat from the Whidbey Formation (Table 6) have yielded only infinite ages (Easterbrook, 1969, 1976a, in press).

Table 6.–Radiocarbon dates from the Whidbey Formation

>33,200	(I-1446)	>40,000	(W-1523)
>35,000	(I-1385)	>40,000	(UW-447)
>35,000	(I-1194)	>40,000	(UW-39)
>35,700	(I-1528)	>40,000	(UW-450)
>39,900	(I-2283)	>40,000	(I-10375)
>40,000	(I-974)	>42,000	(I-722)
>40,000	(I-975)	>42,000	(I-723)
>40,000	(I-1203)	>43,000	(W-1578)
>40,000	(W-1446)	>47,000	(GS-2131)
>40,000	(W-1516)	>49,400	(GRN-4971)

Six anomalous finite radiocarbon dates, ranging from 28,910 to 43,250 years reported from the Whidbey Formation (Johnson and others, 1980; Stoffel, 1980), are all considered invalid because these samples were subsequently rerun by other laboratories and found to be more than 40,000 years old. Another half dozen finite radiocarbon dates (Dorn and others, 1962), measured prior to the definition of the Whidbey Formation (but from Whidbey sediments), were also later re-analyzed and dated as older than 40,000 years. Thus, no confirmed finite radiocarbon dates are known from the Whidbey Formation.

Amino acid analyses of wood and shells in nonglacial deposits of the Whidbey Formation suggest an age of about 100,000 ± 20,000 years, well beyond the range of radiocarbon dating. Age calculations based on leucine and isoleucine racemization in *Saxidomus* were made for sediments at Admiralty Bay, Whidbey Island. Calibration of leucine and isoleucine racemization rate constants followed the method of Kvenvol-

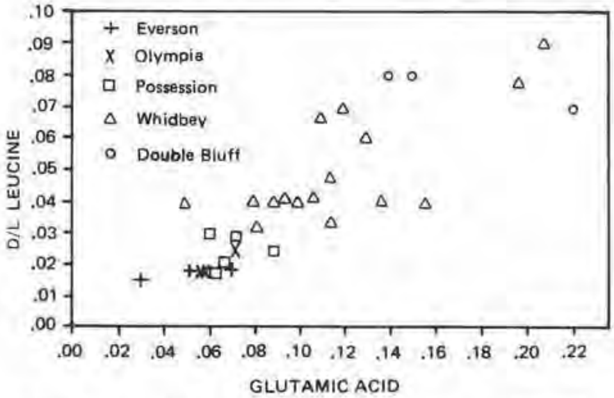


Figure 7.–Leucine and glutamic acid D/L ratios in wood from the Puget Lowland.

den and others (1980), who used Anacortes and Coupeville as temperature calibration sites and calculated an age of 77,000 years for the Admiralty Bay locality.

The calibrated rate constant (k_2) from *Saxidomus* at Hope Island and the estimated diagenetic temperature at Anacortes were used in the Arrhenius expression to calculate the rate constant (k_1) at the estimated diagenetic temperature (T_1) at Admiralty Bay. The calculated ages range from 61,000 to 131,000 years. If temperature assumptions are accurate within 0.5°C, the uncertainty in calculations is about 20 percent. The age of these deposits is here estimated as 96,000 ± 35,000 years by leucine and 107,000 ± 9,000 years by allo/iso.

The range of estimated ages for the Admiralty Bay sediments is significant because the sediments at South Whidbey State Park are correlated with those at Admiralty Bay on the basis of similar amino acid D/L ratios in *Macoma*. Fossiliferous blue clay at South Whidbey State Park was correlated on the basis of its lithology and stratigraphic position relative to the Whidbey Formation by Easterbrook (1968). The estimated amino acid age of 97,000 ± 35,000 years for these sediments appears to confirm earlier correlation of the clay with the Whidbey Formation.

Results of aspartic acid and proline measurements made on 20 wood samples from the Whidbey Formation at 11 localities (Table 3 and Fig. 5) are not as consistent as those for leucine and glutamic acid. Twelve of 19 aspartic acid measurements fall within a narrow range of values (0.20 to 0.29), but some D/L ratios of wood yielded unexpectedly higher values. Wood from shell-bearing silt and clay at South Whidbey State Park has D/L ratios of 0.65 for aspartic acid and 0.12 for proline, and wood samples from Lagoon Point and Camano Island also gave anomalous values, higher than those of Whidbey sediments elsewhere. Leucine and glutamic acid D/L ratios may be more reliable for chronostratigraphic correlation (Fig. 7). Because kinetic models are not well understood for these amino acids in wood, no ages are calculated.

Correlation of the Whidbey Formation with the Puyallup Formation in the southern Puget Lowland can be shown to be impossible because measurements of remanent magnetism of Whidbey sediments show normal polarity with declinations within 10° of north and inclinations ranging from 46° to 64° (Easterbrook, 1976b). The Puyallup Formation is reversely magnetized.

Possession Drift

Possession Drift (Easterbrook and others, 1967) occurs as discontinuous lenses in the central and southwestern Puget Lowland (Easterbrook, 1968, 1969, 1976a; Deeter, 1979). It consists of a single till sheet as much as 25 m thick at its type locality, Possession Point, but elsewhere in the lowland, Possession Drift includes glaciomarine sediments and outwash sand and gravel.

Calculations based on amino acid analyses of marine shells in Possession glaciomarine drift at several localities suggest an age of about 75,000-90,000 years (Easterbrook and Rutter, 1981, 1982). Amino acid D/L ratios were measured in shells of six mollusk genera in Possession glaciomarine drift at three localities in the Puget Lowland: Port Williams, Stillaguamish, and Blowers Bluff. Only two genera, *Nuculana* and *Clinocardium*, were found in Possession glaciomarine drift at Blowers Bluff on Whidbey Island (Fig. 8) and at Port Williams on the Olympic Peninsula (Fig. 9).

Saxidomus, *Nuculana*, *Clinocardium*, *Hiatella*, *Mya*, and *Macoma* were all present in pre-Fraser glaciomarine drift at the Stillaguamish locality (Fig. 10). The shells were radiocarbon-dated at 46,500 ± 1,100 years, but peat immediately overlying the shells was dated as older than 49,000 years (Minard, 1980). Because radiocarbon dates of shells older than about 30,000 years are much more likely to give dates to young, the peat date is believed to be more reliable, and the shells are considered older than 49,000 years. Amino acid age calculations gave a mean age of 80,000 ± 22,000 years.

Three analyses of *Nuculana* at Stillaguamish gave quite similar allo/iso ratios (0.33, 0.37, 0.38). The allo/iso ratio from the first sample analyzed from Blowers Bluff was 0.30, but ratios from seven analyses on two samples run 2 years later averaged 0.18. Analytical values for three samples from Port Williams range from 0.12 to 0.28.

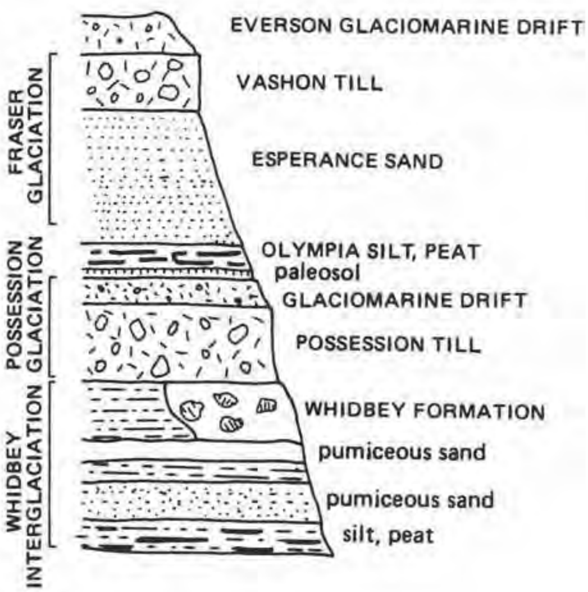


Figure 8.–Geologic section, Blowers Bluff, Whidbey Island.

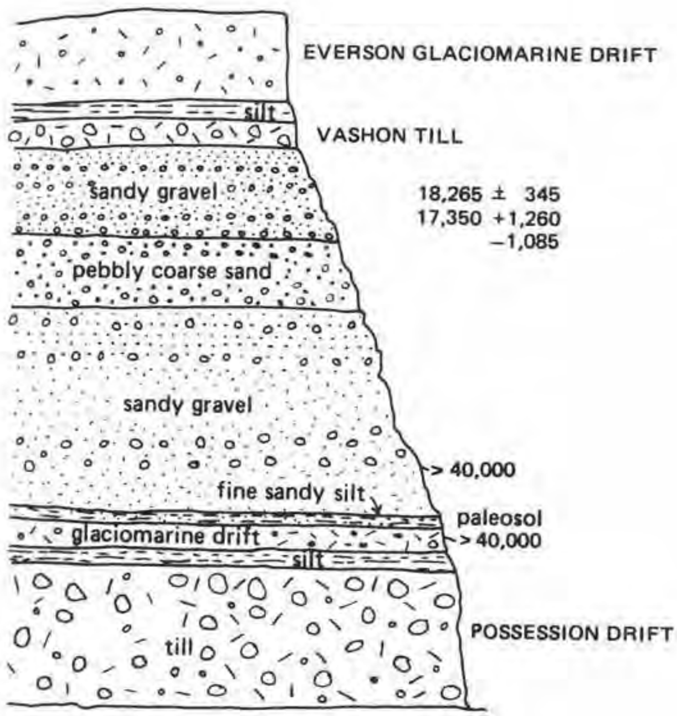


Figure 9.–Geologic section, Port Williams.

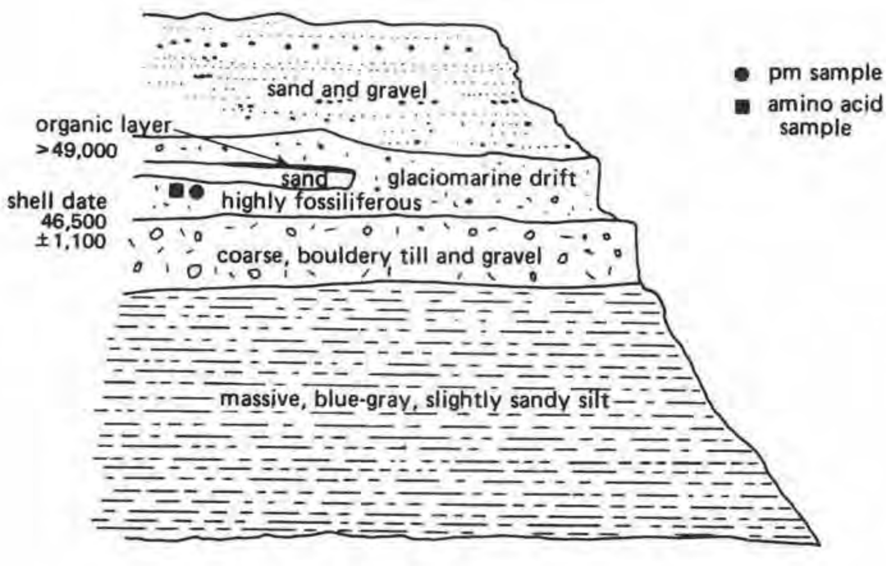


Figure 10.–Geologic section, Stillaguamish River.

At Port Williams, shell-bearing, pre-Fraser glaciomarine drift (Fig. 9) is overlain by gravel radiocarbon-dated as older than 40,000 years old. Although no direct evidence precludes that glaciomarine drift from being slightly younger than the glaciomarine drift at Stillaguamish, both occupy similar stratigraphic positions and are believed to correlate with one another. Some of the leucine and allo/iso ratios in a few *Clinocardium* from Port Williams (loc. 79-1) are similar to D/L ratios of *Clinocardium* from localities that have radiocarbon ages of about 12,000 years. However, because the fossil mollusks in the Port Williams glaciomarine drift and in the overlying gravel are radiocarbon-dated as older than 40,000 years, the Port Williams deposit cannot be correlative with such young sediments.

All radiocarbon dates on shells and wood in glaciomarine drift included in the Deception Pass Stade of the Possession Glaciation by Easterbrook (1976a) have been infinite. However, at Strawberry Point on Whidbey Island, outwash sand and gravel containing two peat beds are present between two tills. Wood in the upper of the two peats has been radiocarbon-dated at 34,900 ± 3,000 (I-1880), 35,400 ± 200 (QL-148A), and 35,600 ± 300 (QL-148B) years, and wood in the lower peat has been dated at 47,600 ± 3,300 (GrN-5257), 43,900 ± 1,000 (QL-149), and 43,600 ± 1,000 (QL-151) years (Hansen and Easterbrook, 1974; Stuiver and others, 1978). Dates of 27,600 ± 1,000 (WW-11), 27,200 ± 1,000 (I-2285), and 26,850 ± 1,700 (I-1111) years from basal peat on the upper till provide a minimum age for the upper till (Easterbrook, 1976a).

The glacial origin of these deposits is confirmed by the following evidence: (1) the upper till is compact and shows deformation structures; (2) both the till and gravel contain clasts of rock types that crop out only in Canada; (3) the gravel and till contain particles much too coarse to have been deposited in the center of the lowland by any process other than glaciation; and (4) pollen from the peat in the outwash gravel indicates a cool, tundra-like environment.

These glacial deposits provide the physical evidence for the Oak Harbor Stade of the Possession Glaciation (Easterbrook, 1976a). Pollen from the Olympic Peninsula indicates a short, cold, climatic phase from about 34,000 to 40,000 years ago (Heusser, 1977), but no evidence of glaciation during this time has been found to the north in Canada (Fulton and others, 1976) where radiocarbon dates have been measured for all but a few thousand years of the time recorded in the glacial deposits at Strawberry Point. Easterbrook (1976b,c) suggested that the Oak Harbor glacial advance occurred during a short period not covered by radiocarbon dates in Canada, but a number of recent dates from Canada coincide with those in the outwash gravel at Strawberry Point. Because the lithology of the Strawberry Point deposits precludes a glacial source in the Cascades, a very limited ice advance from British Columbia must have brought the erratic clasts.

Amino acid ratios from wood samples in late Possession and Olympia peats were measured, but no wood samples were associated with shells. All the Possession wood samples came from Strawberry Point on Whidbey Island, where radiocarbon dates were available for age comparison. The oldest wood samples at Strawberry Point were from the lowest peat in outwash gravel, dated at 47,600 years by Vogel (Hansen and Easterbrook, 1974) and 43,000 years by Stuiver, and the youngest samples were from 22,000-28,000-yr-old wood in overlying peat of the Olympia nonglacial interval (Hansen and Easterbrook, 1974). New radiocarbon dates were made from splits of the samples used for amino acid analyses and are discussed in the following section.

Aspartic acid D/L values for wood in the lowest peat, radiocarbon-dated at 47,600 years, are 0.29, 0.30, and 0.25, as compared to 0.24 for the overlying peat and 0.22 for the highest peat in the gravel (radiocarbon-dated at 34,900 years). Ratios for wood in the overlying Olympia nonglacial peat beds cluster around 0.20, although one value of 0.28 was measured. From the data collected so far, D/L ratios appear to increase with the age of the wood, but individual measurements overlap, making precise age derivations difficult.

Olympia Nonglacial Interval

Floodplain and lacustrine silt, clay, and peat of the Olympia nonglacial interval were deposited during "the climatic episode immediately preceding the last major glaciation, and represented by nonglacial strata lying beneath Vashon Drift" (Armstrong and others, 1965). Chronology of the Olympia sediments is based largely on radiocarbon dating.

Radiocarbon dates from the type locality at Fort Lawton range from 18,100 to 22,400 years (Mullineaux and others, 1965), and dates in the central Puget Lowland extend to 28,000 years (Hansen and Easterbrook, 1974; Easterbrook, 1976a). Radiocarbon dates as young as 15,000 years have been obtained from deposits immediately beneath Vashon till in the Seattle area and on the Kitsap Peninsula (Mullineaux and others, 1965; Deeter, 1979). Pollen from peat suggests that the climate during the Olympia nonglacial interval was somewhat cooler than that of the present (Hansen and Easterbrook, 1974; Alley, 1979; Clague, 1978).

RADIOCARBON CHRONOLOGY AND CORRELATION OF THE LATE WISCONSIN DEPOSITS

Coquitlam Drift

Piedmont ice which advanced into the Fraser Lowland of British Columbia late in the Olympia interval, but which apparently did not extend into Washington, deposited the Coquitlam Drift (Hicock, 1976, 1980; Armstrong, 1977; Armstrong and Hicock, 1976; Clague and others, 1980; Hicock and Armstrong, 1981; Clague and Luternauer, 1982, 1983; Armstrong and others, 1985). Wood from Coquitlam Drift in the Fraser Lowland has been radiocarbon dated at:

25,800; ± 310 (GSC-2273)
21,700 ± 240 (GSC-2235)
21,700 ± 130 (GSC-2416)
21,600 ± 200 (GSC-2203)
21,500 ± 240 (GSC-2536)
21,300 ± 250 (GSC-3305)

Mammoth tusks from Coquitlam outwash have been radiocarbon dated at:

21,400 ± 240 (SFU-65)
21,600 ± 240 (SFU-66)

Dates from overlying organic material limit the age of the Coquitlam Drift to about 18,000 years (Clague, 1980, 1981; Armstrong and others, 1985). The dates are:

18,700 ± 170 (GSC-2344)
18,600 ± 190 (GSC-2194)
18,300 ± 170 (GSC-2322)
18,700 ± 150 (GSC-2371)
17,800 ± 150 (GSC-2297)

On the basis of the range in ages, the Coquitlam Drift appears to be correlative with the Evans Creek Drift in the Cascade Mountains (deposition of which immediately preceded that of Vashon Drift), and with late Olympia nonglacial sediments in the Puget Lowland, which were deposited prior to the extension of the Cordilleran ice sheet into Washington during the Fraser glaciation (Armstrong and Clague, 1977; Easterbrook, in press).

Fulton and others (1976) have contended that the Olympia is a true interglacial period, extending from about 20,000 to beyond 59,000 years in British Columbia, but till and outwash of the Oak Harbor Stade in Washington, radiocarbon-dated between 28,000 and 35,000 years, indicate that the Olympia nonglacial interval was relatively short. Further, pollen from Olympia sediments suggests that climatic conditions were not quite as mild as at present (Hansen and Easterbrook, 1974).

Paleomagnetic data from Olympia silt indicate normal polarity with oscillatory declinations (Easterbrook, 1976a).

Fraser Drift

Fraser Drift was deposited during the last major glaciation of the Pacific Northwest by the Cordilleran ice sheet (Armstrong and others, 1965). It has been subdivided into four stades: (1) Evans Creek Drift, deposited during an early alpine phase; (2) Vashon Drift, deposited during the maximum advance of the Cordilleran ice sheet; (3) Everson glaciomarine drift, deposited from floating ice during deglaciation of the lowland; and (4) Sumas Drift, deposited during a short readvance of the ice before complete deglaciation.

Vashon Drift

Vashon Drift includes (1) the Esperance Sand Member in the Puget Lowland and Quadra Sand in southern British Columbia, both deposited by meltwater streams from the advancing Cordilleran ice sheet (Newcomb, 1952; Mullineaux and others, 1965; Clague, 1976, 1977); (2) Vashon till (Willis, 1898; Bretz, 1913), which overlies the Esperance and Quadra units; and (3) recessional outwash sand and gravel and ice-contact deposits.

The time of advance of the Cordilleran ice sheet is limited by glaciolacustrine deposits that lie beneath Vashon till in the Fraser Lowland. The glaciolacustrine sediments have been radiocarbon dated at 18,700 ± 170 (GSC-2344) and 17,800 ± 150 (GSC-2297) years (Clague, 1980; Armstrong and others, 1985; Hicock and others, 1982; Hicock and Armstrong, 1985). Other radiocarbon dates that limit the date of the Vashon advance include: (1) 18,000 ± 400 years (I-2282) (Easterbrook, 1969) from peat in outwash sand of the Esperance Sand Member at the east end of the Strait of Juan de Fuca; (2) 16,510 ± 320 (UW-445), 15,450 ± 450 (UW-448), and 15,350 ± 210 (10374) years from floodplain sediments in the south-central Puget Lowland (Deeter, 1979); and (3) 15,000 ± 400 (W-1227), 15,100 ± 300 (W-1305), and 16,070 ± 600 (W-2125) years from organic material in the Seattle area (Mullineaux and others, 1965; Yount and others, 1980).

Waitt and Thorson (1983) proposed that the Juan de Fuca lobe (Bretz, 1920), which flowed westward out the strait to the ocean, may have been out of phase with the Puget lobe, which flowed southward down the Puget

Lowland (Fig. 11). They suggested that the Juan de Fuca lobe may have reached the western end of the strait by about 17,000 years ago, several thousand years earlier than the Puget lobe reached its maximum southern extent. However, radiocarbon dates of 17,000 ± 240 (GSC-2829) years from southeastern Vancouver Island (Keddie, 1979), 17,250 ± 1,000 from the central Strait of Juan de Fuca (Anderson, 1968); 17,350 ± 1,260 (B-1062), 18,265 ± 345 (B-1063), and 18,000 ± 400 years (I-2282) (Fig. 19) from the southeastern Strait of Juan de Fuca (Easterbrook, 1969; G. W. Thorsen, personal commun., 1981) demonstrate that the Juan de Fuca lobe did not pass the eastern part of the strait until after 17,000 years ago. Both lobes seem to have advanced synchronously.

The terminus of the Puget lobe in the southern Puget Lowland lacks a conspicuous end moraine and is marked only by patchy segments of moraines, extensive outwash deposits, and kame-kettle complexes (Bretz, 1913; Mackin, 1941; Crandell, 1963; Carson, 1970; Porter and Carson, 1971; Lea, 1983, 1984). During the interval between 13,000 and 14,500 years ago, the Puget Lobe retreated from its terminal position by backwasting, depositing recessional outwash and proglacial lacustrine sediments (Bretz, 1913; Curran, 1965; Deeter, 1979; Waitt and Thorson, 1983). The Puget Lobe had melted back from the latitude of Seattle before 14,000 ± 900 (1-330), 13,650 ± 550 (1-346A), and 13,430 ± 200 years (Rigg and Gould, 1957; Leopold and others, 1982). Wood in ice-dammed lake deposits east of Seattle was radiocarbon-dated at 13,570 ± 130 (UW-35) years, suggesting that the Puget lobe may still have occupied the Puget Lowland then.

The age of the recession of the Juan de Fuca lobe from its terminal zone is limited by radiocarbon dates of 14,460 ± 200 (Y-2452) years from a bog on Vashon Drift near the western margin of the Strait of Juan de Fuca and by the dates listed below from wood in till at five localities near the terminus of the Juan de Fuca lobe (Heusser, 1973a,b, 1982).

13,380 ± 250 (RL-140)
13,100 ± 180 (Y-2449)
13,080 ± 260 (Y-2450)
13,010 ± 240 (RL-139
12,020 ± 210 (RL-138)

Other radiocarbon dates that limit the age of retreat of the Juan de Fuca lobe include 14,400 ± 400 and 13,150 ± 400 years from bottom sediments in the Strait of Juan de Fuca (Anderson, 1968) and 12,100 ± 310 (WSU-1866) years from basal peat in a bog at a mastodon site along the southern side of the strait (Petersen and others, 1983).

Everson Glaciomarine Drift

As the Cordilleran ice sheet retreated northward from its terminus, it also thinned rapidly. Soon after it had receded northward from the Seattle area, the ice had thinned sufficiently to allow marine water to enter the Puget Lowland through the Strait of Juan de Fuca. The remaining ice floated, and progressive melting deposited Everson glaciomarine drift over a large area in the central and northern Puget Lowland (Easterbrook, 1963, 1966a, 1968, 1969, 1970, 1971, 1976a, 1979). Unbroken, articulated marine shells preserved in growth positions demonstrate that the glaciomarine drift represents in-situ deposition. More than 80 radiocarbon dates from shells and wood in Washington and British Columbia establish the age of Everson glaciomarine drift between 11,000 and 13,500 years.

Everson glaciomarine drift has been found in an area of approximately 18,000 km^2. It is believed to have been deposited largely from berg ice, more or less simultaneously over the central and northern Puget Lowland and southwestern British Columbia (Armstrong and Brown, 1954; Easterbrook, 1963, 1969). Pessl and others (1981) and Domack (1983) proposed the contrasting view that the Everson glaciomarine drift was deposited by calving of ice from a northward-retreating, backwasting terminus and thus is time-transgressive. Domack (1983) proposed that Everson glaciomarine drift was deposited from an adjacent, progressively retreating, ice terminus during deposition of the Everson glaciomarine drift. This type of origin requires transgressive deposition of the glaciomarine drift over a distance of more than 170 km and would mean that the glaciomarine drift in the northern part of the lowland should be younger than glaciomarine drift farther south. However, the many radiocarbon dates from Everson glaciomarine drift are not progressively younger northward. Similar ages are found for all of the Everson glaciomarine sediments, regardless of their geographic location. The southernmost radiocarbon date from Everson glaciomarine drift is 12,670 ± 90 years (USGS-64) from the southern end of Whidbey Island. Other radiocarbon dates on shells from Everson glaciomarine drift on Whidbey Island include:

11,850 ± 240 (I-1448)
12,300 ± 180 (I-2154)
12,400 ± 190 (I-2286)
12,535 ± 300 (I-1079)
13,010 ± 170 (UW-32)

Radiocarbon dates on Everson glaciomarine drift from the San Juan Islands, about in the geographic middle of the area covered by Everson glaciomarine drift are:

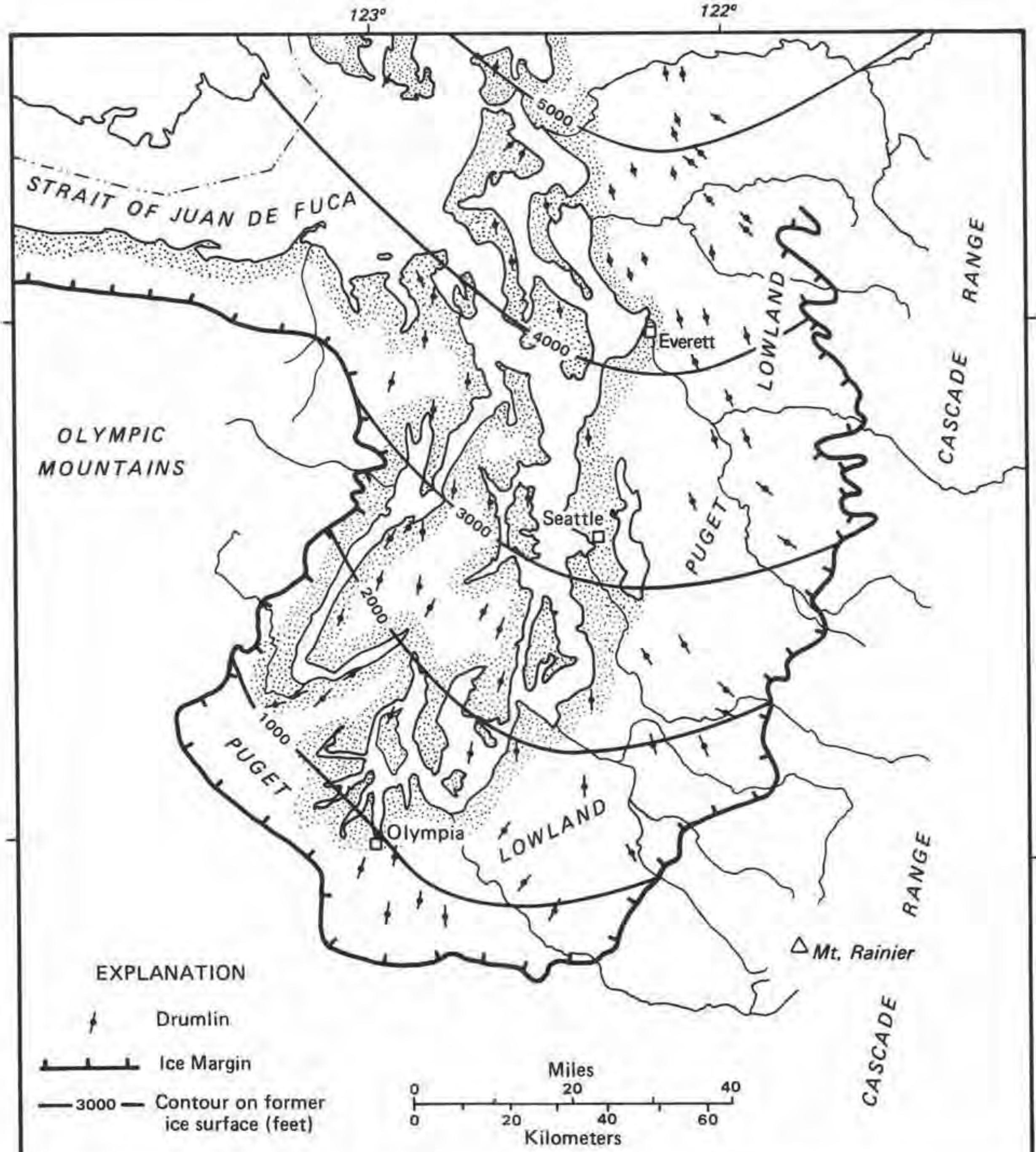


Figure 11.–The Cordilleran ice sheet in the Puget Lowland during the maximum extent of the Vashon Stade of the Fraser Glaciation (after Easterbrook, 1979).

11,900 ± 170 (I-2156)
12,000 ± 450 (I-1471)
12,160 ± 160 (I-1470)
12,350 ± 330 (I-1469)
12,350 ± 400 (I-969)
12,600 ± 190 (I-1881)

Radiocarbon dates from the northernmost Everson glaciomarine drift in Washington include 12,970 ± 280 (I-1447) and 12,090 ± 350 (W-984) years (Easterbrook, 1963, 1969) and in British Columbia, more than 20 dates range from 12,900 ± 170 to 12,000 ± 100 (GSC-2177) (Armstrong and others, 1985; Clague, 1980). The overlapping of such a large number of radiocarbon dates distributed over such a wide area and the occurrence of many old dates in the northern outcrops and young dates in the southern outcrops invalidates the Domack transgressive, calving-ice-front model for deposition of the Everson glaciomarine drift. Deposition from shelf ice is considered unlikely in view of paleoecological conditions suggested by diatoms and foraminifera in the glaciomarine drift (Crandall, 1979). Deposition largely from abundant berg ice is thus the most likely mode of origin for the glaciomarine drift.

Sumas Drift

Following deposition of Everson glaciomarine in the lowland, Cordilleran ice readvanced a short distance from British Columbia into northern Washington and deposited Sumas Drift near the international boundary (Easterbrook, 1963, 1966a,b, 1969, 1971, 1974, 1976d; Armstrong, 1977, 1981; Armstrong and others, 1965). Sumas till and ice contact deposits lie directly on post-Everson outwash, demonstrating that deposition of outwash preceded the advance of Sumas ice and that the Sumas was a subaerial advance, not just grounding of floating ice (Easterbrook, 1963). Radiocarbon dates from wood in Sumas till in southwestern British Columbia (Armstrong and others, 1965; Armstrong, 1981) include:

11,600 ± 280 (GSC-1675)
11,500 ± 1,100 (GSC-L-221D)
11,400 ± 170 (GSC-1695)

Erratics from the Fraser Canyon are frequently found associated with Sumas Drift. Radiocarbon dates (Mathews and others, 1972; Mathews and Heusser, 1981; Mathews and Rouse, 1975) from bogs in the Fraser Canyon, indicating that the canyon was ice-free at that time, include:

11,430 ± 150 (I-6057)
11,140 ± 260 (I-6058)
11,000 ± 170 (I-53460

The dates from the Fraser Canyon limit the possible thickness of ice in the canyon to 250 m at 11,430 ± 150 and 150 m at 11,140 ± 260 years ago (Clague, 1981). Sumas ice filled the Columbia Valley of Washington and the contiguous lower Chilliwack Valley of British Columbia in the Cascade Mountains with as much as to 250 m of outwash (Easterbrook, 1971). Wood in ice-contact sediments interbedded with the outwash has been radiocarbon-dated at 11,300 ± 100 (GSC-2523) (Armstrong, 1981; Clague, 1980). The time of the end of the Sumas Stade is limited by radiocarbon dates of 9,920 ± 760 (I-2280), 9,750 ± 150 (WW-6), 9,500 ± 200 (WW-8), and 9,300 ± 250 (I-2281) from the bases of peat bogs in abandoned outwash channels a few miles south of the international boundary (Easterbrook, 1969, 1971). Figure 12 summarizes the radiocarbon chronology of the late Pleistocene in the Puget Sound.

CONCLUSIONS

The chronology of early Pleistocene sediments in the Puget Lowland is based on identification and fission-track dating of the Lake Tapps tephra and paleomagnetic measurements of associated deposits. The Lake Tapps tephra, defined on the basis of outcrops at the Salmon Springs Drift type locality, is especially useful as a time-stratigraphic marker.

The ages of the Orting Drift, Alderton Formation, Stuck Drift, and Puyallup Formation are now confirmed as early Pleistocene. All lie stratigraphically beneath the Lake Tapps tephra, whose age is established as 0.8-0.9 m.y. by fission-track dating and paleomagnetic measurements. Reverse magnetization of all these stratigraphic units and the fission-track dates establish their deposition during the Matuyama Reversed Polarity Chron between 0.73 and 2.4 m.y. A numerical age for the lowermost units awaits additional dating of interbedded tephra.

Many correlations previously made with the Salmon Springs and older units in the Puget Lowland are invalidated by the antiquity of the Salmon Springs, Puyallup, Stuck, Alderton, and Orting outwash sediments. None of the Salmon Springs and older deposits can be equivalent to the Double Bluff Drift, Whidbey Formation, or Possession Drift, all of which are normally magnetized. A chronologic gap in the Pleistocene stratigraphy of the Puget Lowland is now apparent for the interval between 200,000 and 800,000 years ago.

The ages of Double Bluff Drift, the Whidbey Formation, and Possession Drift are based on amino acid analyses of shells and wood. Calculation of amino acid age estimates for marine shells in Double Bluff glaciomarine drift suggest an age of 110,000-178,000 years, but the unit could be as old as about 250,000 years. The age of shells in Whidbey sediments is calculated as 96,000 ± 35,000 years by leucine and 107,000 ± 9,000 years by allo/iso D/L ratios. Marine shells in Possession glaciomarine drift give amino-acid-estimated ages of 80,000 ± 22,000 years.

Amino acid analyses were also made on wood associated with shells in Double Bluff Drift, Whidbey sediments, and Possession Drift, as well as on younger wood dated by radiocarbon. The purpose of these analyses was to determine if wood could be used for

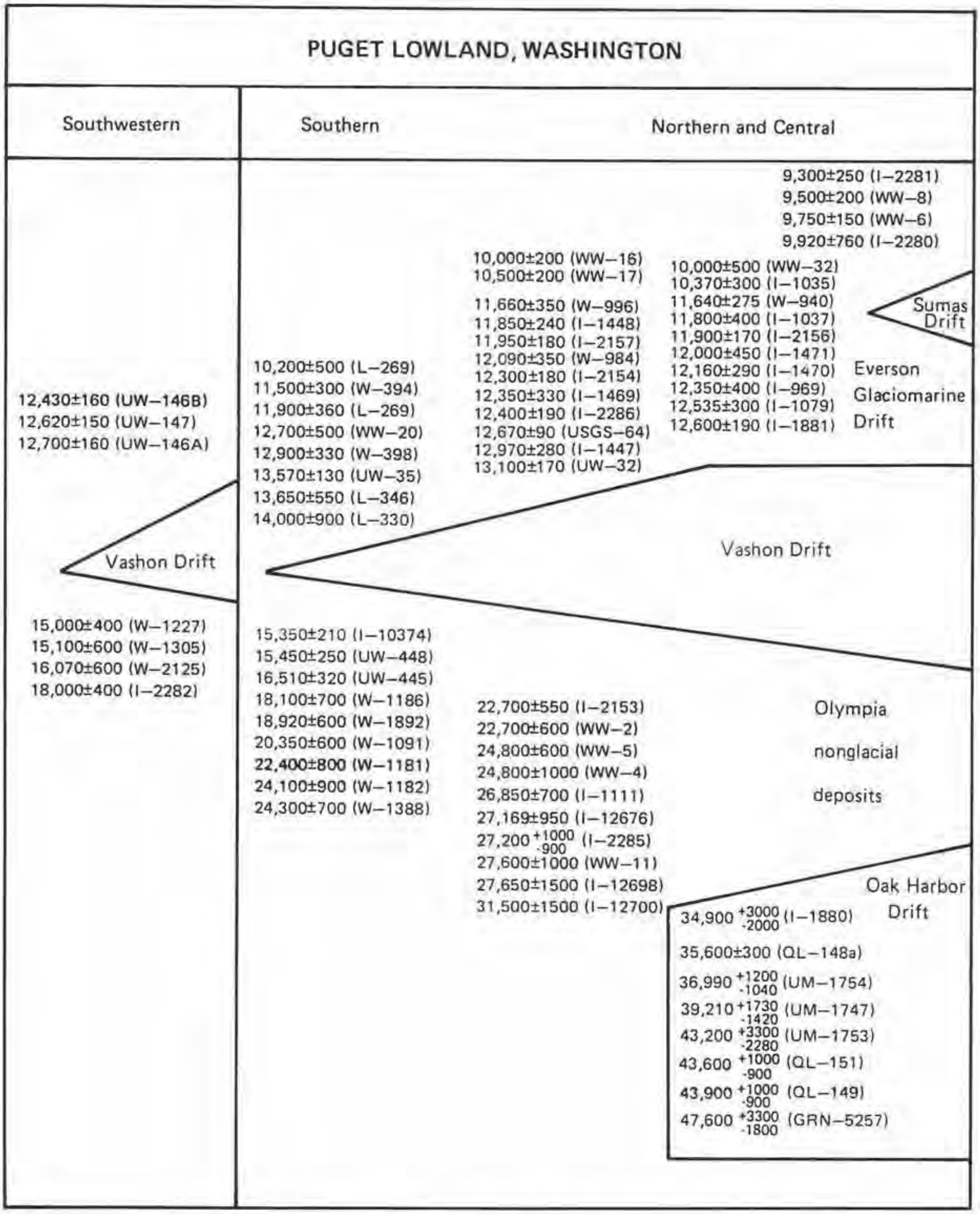


Figure 12.–Radiocarbon chronology of the late Pleistocene in the Puget Sound.

amino acid age determinations. Most of the wood analyzed gave consistent results, but some samples gave widely varying results for reasons that were not readily apparent. Age calculations for wood were not made because kinetic models of racemization in wood are not yet well understood. At this early stage in the development of amino acid age estimates in wood, results look encouraging, but additional data on variation between species and the effects of the degree of preservation of wood are needed.

The chronology of late Pleistocene sediments in the Puget Lowland has been established with numerous radiocarbon dates (Fig. 12). Advance of the Cordilleran ice sheet during the Fraser Glaciation is well documented by radiocarbon dating. An early advance of ice into the Fraser Lowland of British Columbia between 18,000 and 21,500 years ago apparently did not reach the Puget Lowland of Washington. It may be equivalent to the Evans Creek Drift in the Cascade Mountains, deposition of which also immediately preceded the main Fraser advance. The Cordilleran ice sheet advanced across the international boundary shortly after 18,000 years ago and split into two lobes at the junction of the Puget Lowland with the Strait of Juan de Fuca. Both lobes apparently advanced synchronously. The Juan de Fuca lobe retreated from the western part of the strait shortly before 14,500 years ago, and the Puget lobe retreated from its terminus to the vicinity of Seattle by 14,000 years ago. By 13,000 years ago the ice sheet had thinned sufficiently to allow incursion of marine water into the Puget Lowland, and floating of the remaining ice resulted in deposition of Everson glaciomarine drift over an area of about 18,000 km^2. More than 80 radiocarbon dates from shells and wood in Everson glaciomarine drift show that the drift was deposited nearly contemporaneously from berg ice over the whole region, rather than transgressively from a retreating, calving ice front.

Cordilleran ice readvanced a short distance into the northern Puget Lowland during the Sumas Stade about 11,500 years ago. Radiocarbon dates from basal peat in meltwater channels indicate that the Sumas ice disappeared by 10,000 years ago.

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APPENDIX

Amino Acid Geochemistry Procedures

U.S. Geological Survey Laboratory

Wood shavings were cut from the interior of each wood sample (frozen since collection) using a clean razor knife. Each shaving was immersed in 1 N HCl and sonicated for 30 sec, followed by a distilled water rinse. The pretreated shavings were dried in a vacuum oven and stored in a desiccator. Sample weights after cleaning ranged from 0.15 g to 0.70 g. Shell samples (frozen since collection) were slightly dissolved in 1 N HCl for 30 sec with sonication and rinsed in distilled water. The pretreated shell samples were dried in a vacuum oven and stored in a desiccator. Sample weights after cleaning were about 0.2 g.

Amino acids (both free and bound) were extracted from wood and shells by hydrolyzing the material in 6 N HCl at 110° C for 20 hours under nitrogen in screw-cap hydrolysis tubes. The resulting hydrolysate was evaporated to dryness, taken up in a norleucine standard at pH 1 and applied to the top of a 14 ml column of AG 50-X8 (H+) cation exchange resin. Amino acids were eluted with 2 N NH_4OH and evaporated to dryness. The amino acids were taken up in 2 ml of distilled water and split; half was acidified with 0.1 N HCl and evaporated as a salt in preparation for derivatization; the other half was evaporated to dryness and taken up on 1 ml pH 2.2 sodium citrate dilution buffer in preparation for ion-exchange chromatography. The derivatized portion is discussed in this paper.

The N-pentafluoropropionyl-amino acid-(+)-2-butyl ester derivative was resolved by gas chromatography on two different stainless steel capillary columns: 60 m x 0.08 cm coated with UCON 75-H 90,000 and 60 m x 0.12 cm coated with Carbowax 20M. Quantification of the D- and L-amino acid derivatives was made by peak-height measurement on chromatograms.

University of Alberta Laboratory

Each wood sample was thoroughly cleaned with distilled water, air dried on a plastic weighing dish, broken into small fragments, and crushed using an IKA analytical mill. The wood particles were washed twice in a plastic disposable centrifuge tube using 2 N HCl and twice with double distilled water. Between washings, the sample was sonicated, centrifuged, and decanted. The cleaned particles were transferred to a Buchner funnel connected to a water-vacuum tap and fitted with Whatman glass fiber paper (GF/A-4.25 cm) and washed several times with double-distilled water. The filtrate was discarded and the washed particles collected in small plastic vials where they were dried.

About 100 mg of washed, dried sample was placed in a glass screw-top culture tube (13 x 100 mm). About 6 to 8 ml 5.5 N HCl (constant boiling) was added and the mixture allowed to reflux at 108° C for 24 hours in a heating block. Once cooled, the supernatant liquid was evaporated to dryness in a Speed Vac Concentrator. The residue was dissolved in 1 to 2 ml double distilled water and added to freshly regenerated cation exchange resin (Dowex Ag 50W-X8, 50-100 mesh). Two bed volumes of 3 N NH_4OH were added to elute the amino acids. About 10 ml of amino acid eluate were collected in a clean 13 x 100 mm screw-top culture tube when the solvent front was about 1.5 to 2. 0 cm from the bottom of the column. Esterification was carried out by adding 0.1 ml isopropanol/3.5 N HCl to the dried eluate. This was sonicated until homogeneous, then heated at 100°C for 15 minutes in an oil bath. After evaporation to dryness, the sample was acylated by adding 0.1 ml PFPA (pentafluoropropionic anhydride) and 0.3 ml distilled CH_2Cl_2 (methylene chloride). The sample was sonicated until dissolved and then heated in an oil bath for 100 °C for 5 minutes. The excess PFPA and CH_2C_{12} were cold evaporated on a Buchi rotary evaporator using liquid N_2. Next, the sample was washed with 0.5 to 0.1 ml CH_2C_{12} and after allowing the residue to dissolve completely, it was cold evaporated to dryness using a rotary evaporator. The sample was diluted to 0.5 ml CH_2C_{12} and filtered through a Gelman alpha-200, 0.20 μ metricel filter. About 0.2 to 1.0 μl was injected onto a Hewlett-Packard Model 5840A gas chromatograph equipped with FID Detector and Chirasil-val capillary column (25 ml). The system is controlled by a digital micro-processor terminal which reports amino acid peak areas by automatic integration.

D/L ratios of alanine, glutamic acid, valine, leucine, phenylalanine, proline, aspartic acid were routinely determined in both laboratories. Aspartic acid, glutamic acid, and leucine proved to be the most useful because of their relatively fast rate of racemization and reliability. Examples of gas-chromatograms and ion-exchange chromatograms of mollusks and wood are shown on Figure 3. The leucine and alloisoleucine/isoleucine D/L ratios of fossil mollusks are reported in Table 2. The aspartic acid, glutamic acid, and leucine D/L ratios in fossil wood are reported in Table 3.

Age of Midden Deposits at Penn Cove

The age of an archeological midden at Penn Cove on Whidbey Island was calculated using the racemization of amino acids in *Saxidomus.* Leucine and glutamic acid racemization rate constants were calibrated using data from *Saxidomus* from Bainbridge Island. The leucine D/L ratio of 0.08, the leucine time-zero correction of 0.02 and the radiocarbon age of 3,260 ± 80 years give Kleu = (1.85 ± 0.05) x 10^{-5} yr^{-1}. The glutamic acid D/L ratio of 0.09, the glutamic acid time-zero correction of 0.03 and the radiocarbon age of 3,260 ± 80 years give Kglu = (1.85 ± 0.05) x 10^{-5} yr^{-1}. These rate constants and the leucine and glutamic acid D/L ratios measured in *Saxidomus* 80-24m are substituted into equation (1) to give an age of 1,600 years by leucine and 1,600 years by glutamic acid. In a similar manner an age of about 1,000 years is calculated for *Saxidomus* (80-23). Subsequently, a radiocarbon age of ~ 845 years has been obtained from *Saxidomus* shells at Penn Cove, confirming the amino acid calculation.

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MINOR EXPLOSIVE ERUPTIONS AT MOUNT ST. HELENS DRAMATICALLY INTERACTING WITH WINTER SNOWPACK IN MARCH-APRIL 1982

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ABSTRACT

An eruption at Mount St. Helens in March and April 1982 comprised two distinct extrusive phases, each beginning with explosive activity. Stratigraphic relations of erosional and depositional effects in the crater and breach area reveal numerous separate but related events of the explosive phase beginning on 19 March. A pre-eruption temperature rise starting at about 19:05 PST at a tiltmeter in the breach is attributed to steam generated from meltwater that cut into hot 1980 deposits; resulting steam explosions produced a plume from which brown ash settled on the east flank of the mountain. An explosion at 19:27 ejected gas, pumice, and gray-dacite blocks from the south-southeast side of the dome. The blast dislodged snow from the crater walls; the snow and injected rock debris avalanched into and through the crater and down the north flank of the volcano. A vertical eruption column rose to an altitude of 14 kilometers; as the ash cloud drifted downwind, juvenile pumice fell for tens of kilometers to the southeast. Hot juvenile pumice that accumulated on the crater floor melted snow and rapidly formed a lake from which phreatic explosions deposited fans of pumice on adjacent snow surfaces. The rapidly deepening, hydraulically dammed lake discharged simultaneously around both sides of the dome as a flood of pumice-laden water, which swept through the breach and down the north flank. Successively lower strandlines and terraces of water-laid pumice give evidence of irregular lowering and eventual disappearance of the crater lake. A second eruption column rose at 01:37 on 20 March. Viscous dacite extruded on the southeast sector of the dome between 20 and 23 March.

On 4 April at 20:52 PST a vertical eruption column rose to 8.5 kilometers, from which pumiceous ash was carried by winds to the northeast. The northern brow of the dome collapsed to form a rock avalanche that moved as far as ½ kilometer from the base of the dome. The hot blocks melted snow, producing a small flood that swept through the breach and down channels on the north flank. From 5 to 9 April new lava extruded on the north brow of the dome.

In the future, deposits of the March-April 1982 eruption probably will be poorly preserved and their origin difficult to interpret. Although the 1982 airfall deposits were conspicuous when they overlay snow, after the snow melted they blended into the similar, underlying 1980 airfall deposits. The snow-avalanche deposit became thin once its constituent snow melted. If it becomes considerably dissected or buried, the 1982 snow-avalanche deposit on the north flank could be overlooked or mistaken for other deposits. Much of the extensive laharic deposits has been reworked since 1982. The conspicuous avalanche, lake, and flood deposits in the crater and breach became largely buried in 1982-1987 by the growing dome and by talus and snow from the crater walls.

Processes apparently similar to those at Mount St. Helens in 1982 have occurred during historic eruptions of several other snow-clad volcanoes when minor but hot erupted products interacted with snow and ice. Deposits of these events are fleeting, and with time they tend to become obscure or difficult to interpret. The more than 20,000 fatalities at Nevado del Ruiz volcano, Colombia, in November 1985 starkly illustrate the need to understand these processes.

INTRODUCTION

Following the six mainly explosive eruptions of May to October 1980 at Mount St. Helens (Lipman and Mullineaux, 1981; Foxworthy and Hill, 1982), 11 generally nonexplosive dome-building eruptions occurred between December 1980 and February 1983 (Swanson and others, 1983). Like other eruptions since 1980, the March-April 1982 eruption added new lobes of viscous dacitic lava to the dome, but it differed in three ways: (1) it was preceded by an unusually long

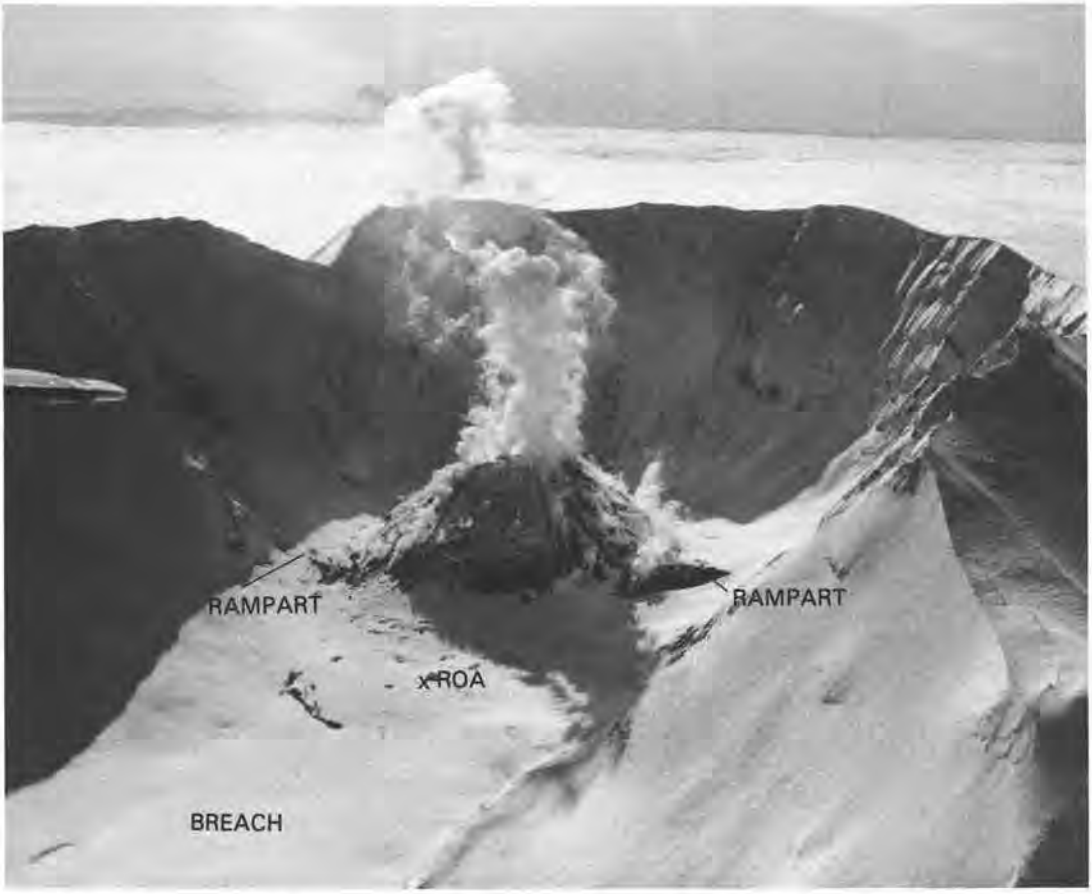

Figure 1.–View southward of the snowy crater of Mount St. Helens on 5 March 1982. The crater is about 2 km wide at the rim.

period of seismicity that included many earthquakes located at depths between 4 and 10 km—the first since October 1980 that were deeper than 3 km (Weaver and others, 1983); (2) it comprised two separate extrusive phases of a few days each (mid-March, early April) separated by a period of dome inflation without extrusion; and (3) each extrusive phase began with explosive activity. Owing to interaction between hot volcanic products and a winter snowpack, effects of the minor explosive activity were dramatic and complex, and they reached far beyond the crater area.

Destructive effects of snow and laharic flows of the 19 March 1982 phase have been summarized (Waitt and others, 1983) and the lahars downstream analyzed (Pierson and Scott, 1985). The present report reconstructs in detail the sequence of explosive events and effects of 19 March 1982 in the crater area, presents some of the photogenic record, summarizes the April 1982 eruptive phase, and shows the meager geologic record left by such eruptions. Helicopter access to the crater during unseasonably fair weather just after the 19 March phase allowed us to acquire much stratigraphic data before the evidence was disrupted or lost by melting or buried by new snow. Because most of the deposits contained large proportions of snow or were deposited on snow and therefore changed as the snow melted during spring and summer, we describe effects mainly as they appeared within a few days of the events.

EXPLOSIVE PHASE OF 19-20 MARCH

Summary of Main Events and Effects

Following several weeks of seismic activity and a gradually accelerating swelling of the dome (Malone and others, 1983; Swanson and others, 1983), the explosive eruptive phase of 19 March 1982 occurred when snow thickly mantled the crater floor and wall (Fig. 1). At about 19:27 PST (00:27 UT, 20 March) a lateral blast carrying hot juvenile pumice, dacite blocks, and gas was directed southward and eastward from the south side of the dome. Snow dislodged by the blast from the south and east crater wall avalanched into the crater. The avalanche split and flowed

Figure 2.–View south-southeastward of avalanche and flood deposits in the breach.

around both sides of the dome, joined 400 m farther north in the breach, and descended the north flank of the volcano; it gradually slowed as it spread across the pumice plain to Spirit Lake and the North Fork Toutle River some 8.4 km from the south crater wall (Figs. 2-4) (Waitt and others, 1983).

Shortly after the avalanche, a crescent-shaped lake formed between the dome and crater wall as hot eruptive products melted crater-floor snow and avalanched snow that lodged behind the dome. A flood of water and pumice discharged simultaneously through outlets both east and west of the dome, swept down the crater breach, and cascaded down the steep north flank of the volcano. Entraining much rock debris, the flood became a lahar by the time it reached the pumice plain. The lahar spread as a broad sheet across the pumice plain; it discharged into Spirit Lake as well as down the North Fork Toutle River (Figs. 3 and 4). (For details see Pierson and Scott, 1985.)

Detailed Sequence in Crater and Breach Area

Stratigraphic relations between deposits and erosional features reveal a complex sequence of explosive and flow events in the crater and breach areas. The sequence apparently began before 19:10 PST, but most events occurred in rapid succession beginning at 19:27. Snow had already avalanched from the crater walls and the flood had passed beyond the volcano flanks before the first night-flight observations at 21:30. The events are thus inferred mainly from the deposits but are calibrated by instrumental data. They are enumerated in a composite stratigraphic order assembled from field observations (Table 1).

Principal events are labeled 1, 2, 3, etc., subdivisions of them A, B, C. Many of these events occurred swiftly and simultaneously, and the relative stratigraphic positions of some continuous deposits changed with distance from the vent. Thus the number-letter sequence of Table 1 is not everywhere strictly consecutive.

1. Small Flood

Seismic signals with gradual onsets at 19:04 and 19:07 PST, recorded by seismometers on the outer volcano flank (C. Boyko, oral commun., 1982), suggest

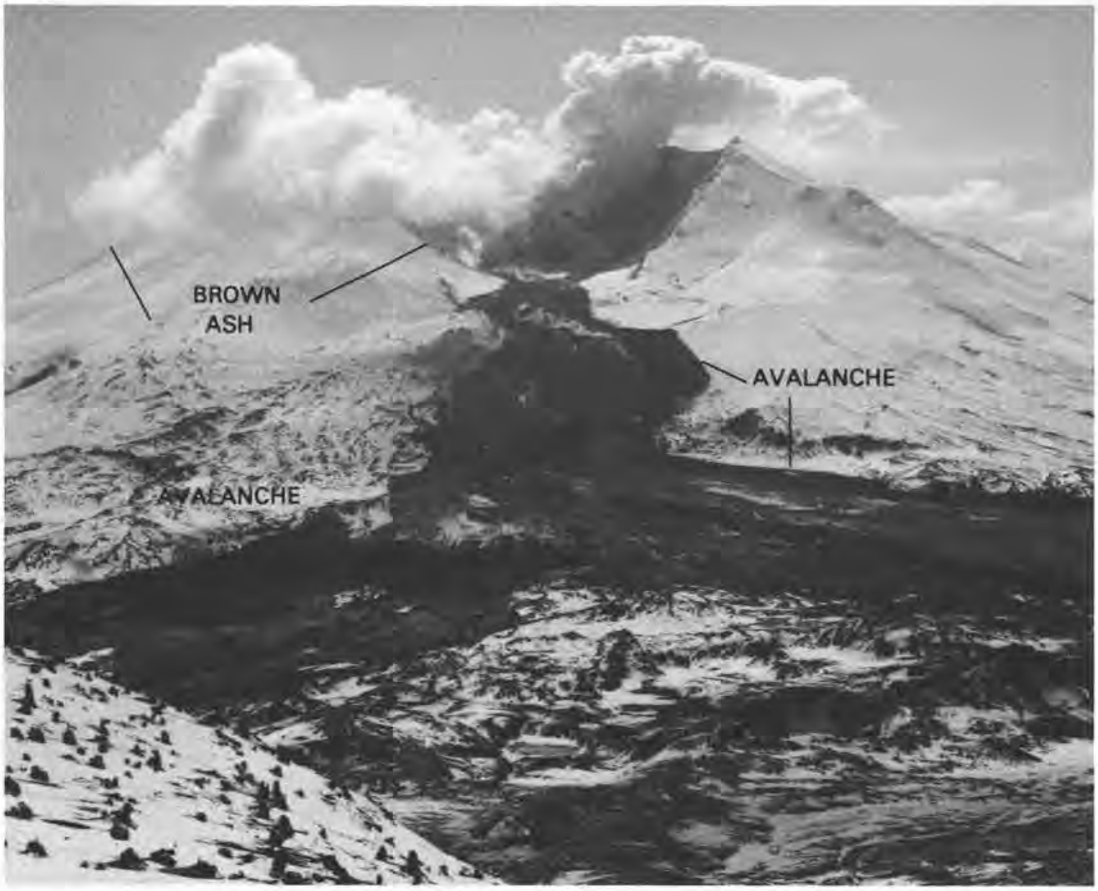

Figure 3.–View southward from Harrys Ridge of avalanche and flood deposits of 19 March 1982. Hummocky topography in middle ground is part of the deposit of the great landslide that created the crater void on 18 May 1980. Splintered stumps in the foreground are remains of old-growth coniferous forest leveled by the initial pyroclastic density current on 18 May 1980.

minor explosive activity. Between 19:00 and 19:10 PST, ambient temperature rose sharply from 0°C to 8.5°C at the YEL tilt station beneath the snow in the breach (Figs. 2 and 4) and rose to 22.2°C by 19:30, when telemetry was lost (Daniel Dzurisin, personal commun., April 1982). The recording thermister lay undisturbed beneath the snowpack as it had been installed before winter. Telemetry ceased because the transmitting antenna some distance away was damaged by the outer edge of the snow avalanche (event 3B). Although isolated from above-snow events, the barrel did lie within a laterally extensive snow cavern that had been gradually melted by fumaroles during winter (Daniel Dzurisin, personal commun., 1983).

In the breach a distinctive pale-brown ash stratigraphically underlay all other deposits of 19 March, including those of the snow avalanche from the crater wall. The brown ash mantled the snow only on the east side of the breach and on the east and northeast flanks of the mountain—the approximate downwind direction on 19 March. The distribution of ash shows that it was derived from the breach well north of the dome. The brown ash was ejected, carried downwind, and deposited before arrival of the swift-moving snow avalanche, which apparently began at 19:27. Observers in aircraft reported a red glow at Mount St. Helens as early as 6 min before the vertical eruption column. We suppose this to be evidence of incandescence in the crater before the onset of the main eruption. We infer that well before the explosive activity at 19:27 a thermal event in the crater melted enough snow that water flowed down the breach, eroding through the snowpack and into underlying 1980 deposits. Interaction of this water with the still-hot 1980 ashflow deposits generated phreatic explosions; a brown plume must have convected upward hundreds of meters and deposited the sand- and silt-sized ash downwind to the east. The anomalous, sharp rise in temperature at the YEL tilt station prior to the main eruption probably resulted from steam of the phreatic explosions migrating

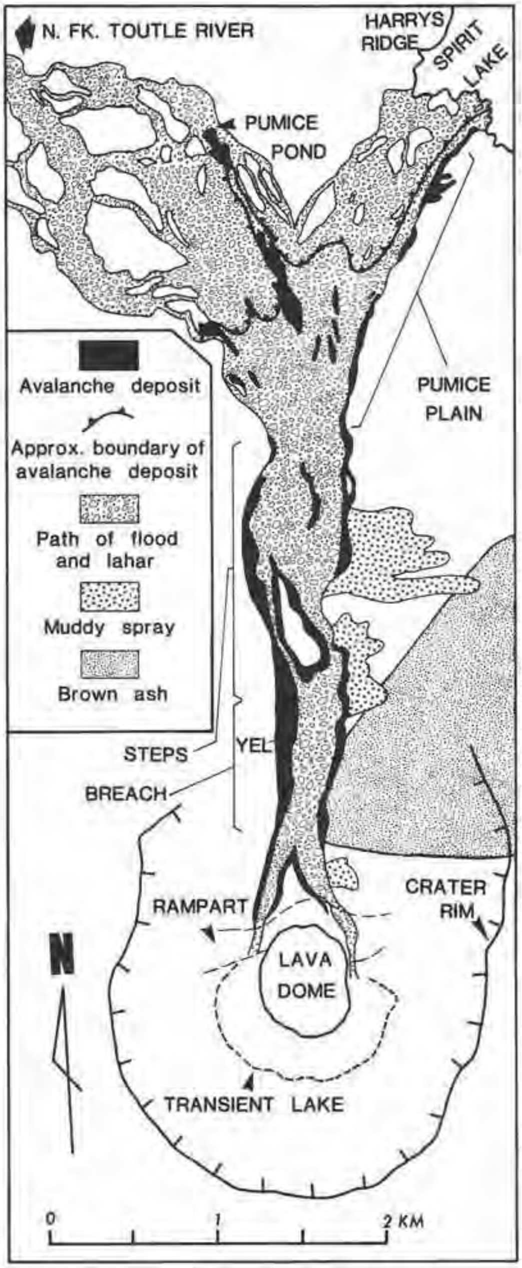

Figure 4.–Map of the avalanche and flood of 19 March 1982, drawn by T. C. Pierson from vertical aerial photographs taken on 20 March 1982. Slightly modified from Waitt and others (1983, fig. 2); see also Pierson and Scott (1985, fig. 4).

through the existing caverns at the base of the snowpack.

2A. Initial Blast

At 19:27 PST an aircraft crew and ground observers 40 to 50 km to the south reported glowing projectiles apparently being lobbed onto the upper south flank of the volcano. We infer that these projectiles were thrown out by a lateral blast. This hot blast propelled gas, pumice, and dome blocks from the south-southeast side of the dome with enough momentum to dislodge most of the snow from the steep crater walls in a sharply defined 120° sector from east to south-southwest. Snow that remained high on the steep crater walls just beyond these limits was densely studded with rock fragments that had been injected laterally (Fig. 5).

3A. Snow Avalanche

The dislodged snow and injected rock debris avalanched into and flowed through the crater. The source area of the avalanche was clearly marked by the bare crater walls (Fig. 6) and at the base of the walls by the snow fan that had become deeply grooved radially downslope.

2B. Further Blasts

Angular blocks of gray breadcrusted dacite as large as 2 m lay in deep cylindrical melt pits, which pocked the grooved snow surfaces east of the dome (Fig. 7). The blocks indicate that an explosion—perhaps a slightly later phase of a continuing eruption that began with the lateral blast (event 2A)—hurled large hot projectiles from the dome. Because there was no evidence that the avalanche had piled behind these blocks, and because many of them lay in craters that truncated the grooves etched by the avalanche (Fig. 7), most of the projectiles must have landed after the snow avalanche had passed through the crater.

Many dacite blocks also impacted and melted into snow in the breach north and northeast of the dome. There they postdated the brown ash (event 1) and were far more abundant outside the margin of the snow-avalanche deposit (event 3B) than inside it. Apparently most of the projectiles fell just before the avalanche arrived from the crater.

4A. Pumiceous Explosion

Gray lithic-crystal ash containing angular pumice lapilli and small blocks formed a mantle 1 to 3 cm thick on the fluted snow surfaces in the crater and around the base of both the eastern and western walls. This mantle became thinner and finer northward, just beyond the crater. This deposit must have resulted from a ground- and wall-hugging pyroclastic surge of ash and pumice lapilli erupted from a vent on the south part of the crater floor or south flank of the dome. The surge divided into separate north-curving flows, one banking along the east crater wall, the other

Table 1.–Sequence of events and effects of 19 March 1982 eruption phase
Note: Timing is off for some events in more distal areas

Event	Activity (approximate time, PST)	Evidence (erosional effect or deposit)
1	Small flood (19:05-19:27)	Brown ash east of breach
2A	Initial blast (19:27)	Removal of crater-wall snow
3A	Snow avalanche off wall	Radial fluting of snow at base of crater wall
2B	Further blasts from dome	Ballistic blocks of dome dacite
4A	Pumiceous explosion	Pyroclastic-surge deposit around crater
4B	Pyroclastic surge passes avalanche	Pyroclastic-surge deposit in breach
3B	Snow avalanche down breach	Avalanche-margin deposits
5	Vertical eruption column (19:30-19:33)	Crater-rim deposit; crater-floor pumice
6A	Lake forms, phreatic explosions	Pumice fans centrifugal from lake
6B	Lake rapidly rises and crests	Highest strandline in crater
6C	Lake drains	Lower strand lines; scour at outlets
7	Flood (19:35-19:40?)	Maximum flood lines; gorges; deposits
----------(time break)----------		
8	Further blasts from dome	Dome-rock projectiles in impact craters
9	Second eruption column (01:37, 20 March)	Clots of airfall ash in SE part of crater
10	Collapse of crater snow	Large circular pits and concentric scarps in lake deposit
11	Lava extrusion (20-23 March)	Lobe added to SE side of dome; rockfalls

along the west. The deposit overlay and thus postdated the dacite blocks (event 2B) and the fluted snow surface (event 3A) (Fig. 7).

4B. Pyroclastic Surge Passes Avalanche

In the breach just north of the dome, the sharp, lobate outer margin of the pyroclastic-surge deposit overlay the brown ash (event 1), but it underlay and was truncated by the snow-avalanche deposit (event 3B), which it therefore predated (Fig. 8). The avalanche took more than 2 min to descend the crater wall and flow down the breach. The pyroclastic surge followed the early phase of the avalanche (event 3A) near the dome, but preceded a slightly later phase of the same avalanche (event 3B) somewhat farther from the

Figure 5.–Snow on the steep upper crater wall east-northeast of dome, studded with angular fragments of gray dacite hurled laterally from the dome. Largest visible block is about 2.5 m in diameter.

dome. Therefore the pumiceous pyroclastic surge flowed swiftly enough to overtake the avalanche as the two fast-moving flows left the crater. The pyroclastic surge either was caused by a discrete eruption event or was the last stage of a sustained blast whose initial stage triggered the snow avalanche.

3B. Snow Avalanche Down Breach

The snow-avalanche deposit headed on the north-sloping rampart both east and west of the dome. The two paths united 1 km north of the dome (Figs. 2 and 4). From there the avalanche flowed along the axis of the breach, descended the north flank, and spread across the pumice plain (Figs. 3 and 4).

As it moved through the breach, the avalanche left a continuous deposit. After an ensuing flood (event 7), the deposit remained as a broad belt along each margin of the avalanche path, overlying undisturbed snow. The deposit was typically thinner than 25 cm and was very dark because of admixed rock fragments and ash. It consisted half of recrystallized snow and half of pumice and rock fragments, with an estimated porosity of 40 percent. Juvenile pumice in the deposit is light gray and of low density, identical to angular pumice that fell to the southeast from the eruption column (event 5). In places the avalanche flowed into dome-rock impact craters (event 2B) in the snow (Fig. 9).

The distinctive brown ash east of the avalanche margin in the breach (event 1) was not present on the avalanche deposit, and therefore the avalanche arrived after the ashfall had ceased. At its margin in the breach, the snow-avalanche deposit overlay and truncated the gray pyroclastic-surge deposit (event 4B), whereas on the southeast side of the dome the snow grooved by the avalanche underlay the pyroclastic-surge deposit.

5. Vertical Eruption Column

A near-vertical eruption column rose 2 to 3 min after the ballistic projectiles sprayed the crater rim, according to distant eyewitnesses. The column started at 19:30 PST and rose to an altitude of 14 km by 19:33 PST. According to U.S. National Weather Service (NWS) radar at Portland (R. A. Nordberg, personal

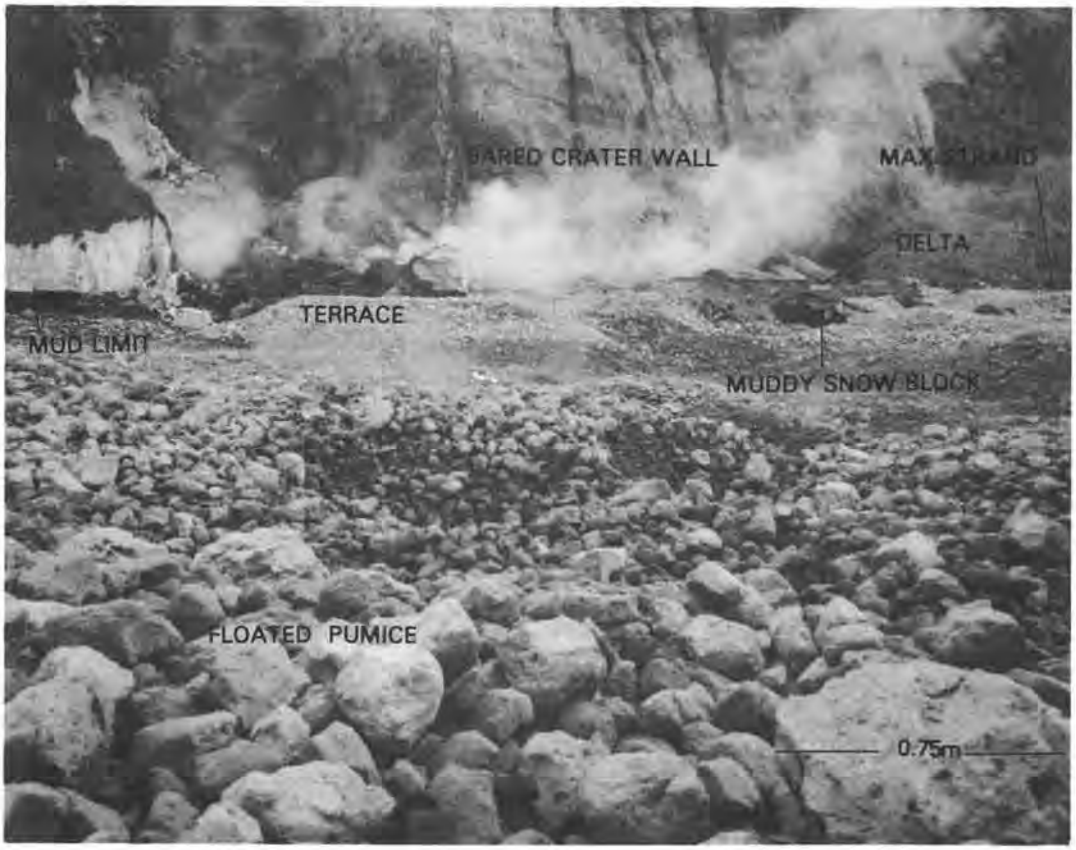

Figure 6.–View southwestward across the crater floor, showing effects of directed blast and deposits of the subsequent transient lake.

commun., 1982), a high-density radar reflection lasted only about 5 min. After the column reached its maximum altitude, the plume drifted southeast at about 15 km/hr; it deposited coarse pumice blocks on the southeast and south mountain flanks and lapilli and ash beyond (Fig. 10). Stratigraphic relations of this airfall material to other 19 March deposits were unclear inside the crater, but because airfall material was not recognized on top of pumice deposited in the crater-floor lake (events 6B and 6C), it probably is older; it may have been contemporaneous with some or all of events 2A through 6A.

6A. Lake Forms Phreatic Explosions

Large volumes of white pumice blocks as large as 2 m accumulated on the crater floor from the initial blast and from the rapid venting that fed the vertical eruption column. The absence of coarse juvenile pumice on the dome suggests that this column and any attendant explosions were directed slightly away from the dome, perhaps from a vent on the south crater floor adjacent to the dome. The hot pumice and probably gas swiftly melted several meters of crater-floor snow, probably including some from the avalanche trapped behind the dome, to form a lake south of the dome.

On the east, northeast, and west-northwest sides of the lake, elongate lobes as thick as 2 m of angular small blocks and lapilli of pumice spread from the lake tens of meters up the avalanche-scoured snow surfaces of event 3A adjacent to the lake. These lobes overlay both the gray-dacite blocks in melt pits (event 2A) and the pyroclastic-surge deposit (event 4A); one pumice lobe also splayed up onto the west-northwest sector of the dome. The pumice lobes apparently resulted when phreatic explosions threw water and pumice from a lake that was deepening in the crescent-shaped moat between the dome and the crater walls.

6B. Lake Rises and Crests

The lake continued to deepen as hot pumice and erupting gas melted snow. The lake became partly clogged with pumice. Subangular pumice blocks formed discontinuous, horizontal strandline benches as high as 15 m above the irregular crater floor along the fluted snow surfaces of event 3A at the base of the

Figure 7.–View upslope of snow surface fluted by overriding snow avalanche, which was impacted by a breadcrusted dacite block, which later melted the cavity. The dark mantle is the gray pyroclastic-surge deposit, which generally overlies the inmelted blocks.

crater walls (Fig. 6). These deposits overlay the more angular and generally smaller pumice blocks of the pyroclastic-surge deposit (event 4A) as well as the angular pumice of the upslope-directed lobes from phreatic explosions (event 6A). Local deltas at and just below the uppermost pumice strandline indicate contemporaneous rapid flow of water from melting snow on the crater walls. Below a sharp, horizontal upper limit of mud on some freshly truncated snow fans (Fig. 6), the snow was intricately scalloped by melting.

Below the uppermost strandline, the moat between the dome and the crater walls was underlain by pumice (blocks, lapilli, ash), sparse lithic debris, and melt-scalloped, mud-stained blocks of granular snow as large as 4 m. Deposits as thick as 3 m in low parts of the moat consist of lithic-rich blocks and lapilli at the base, grading upward to coarse crystal-vitric ash and pumice lapilli at midsection and to fine vitric ash near the top. This graded deposit records a single short-lived lake. It is overlain by openwork, low-density, floated pumice blocks, some piled precariously one on another (Fig. 11). The pumice blocks floating on the eruption-agitated lake surface evidently chafed against each other and became somewhat rounded; the ash produced by abrasion settled in the lake. Most pumice

Figure 8.–View east-northeastward showing the abrupt outer edge of the gray pyroclastic-surge deposit northeast of dome. The deposit is overlain by deposits of the great snow-debris avalanche and small white-snow avalanches from the crater wall. Downslope from the small avalanche on the right, impact craters are far less dense than immediately to the north. Thus, earlier avalanches, probably contemporaneous with event 3A, swept down this slope after the dacite-block projectiles fell (event 2A or 2B) but before the surge (event 4B) and main snow avalanche (event 3B).

blocks deposited at or near the lake surface are ash free. Many blocks deposited well below the lake surface became coated by the fine ash in the muddy lake water. Gases emitted from the cores of the pumice blocks eroded small conical pits through these ashy coatings. Thus although the pumice blocks accumulated in water, their interiors remained hot.

There was no evidence of a material dam at the shallow, relatively narrow thresholds east and west of the dome that became lake outlets. Had the lake been held in by dams of pumice, snow-avalanche deposit, or other material, overflow would have rapidly incised the low point on the crest of such a dam, and remnants should have been prominent on both sides of spillways. Such material was absent despite the gentle underlying slopes that would have encouraged accumulation and preservation. This and the unusual circumstances of more than one outlet indicate that the water was dammed hydraulically at the outlets. The presence of large volumes of pumice blocks and ash probably increased the effective viscosity of the escaping fluid, but the fundamental cause of ponding was that snowmelt produced water quicker than it could be discharged through the undammed outlets.

6C. Lake Drains

In the crater, several lower-level pumice strandlines and terraces consisting of openwork, somewhat rounded pumice lay 2 to 8 m below the uppermost pumice strand (Fig. 6). From its maximum level the lake apparently lowered rapidly by discharging through its two outlets. Minor benches along the west outlet mark successively lower surfaces of the subsiding lake and flood (Fig. 12).

Water-laid, sand-sized ash along the lowest level of the west outlet is dimpled by horseshoe-shaped scours ("horns" directed northward) around protrud-

Figure 9.–Lateral margin of avalanche deposit, spilled into a bomb-impact crater in snow. Plowed-up snow blocks constitute part of avalanche-margin deposits. Ice-axe handle is 80 cm long.

ing boulders, and pumice blocks are jammed on the south side of such boulders (Fig. 13). This and other evidence shows that the last flow through both outlets was northward from the crater; the water at this stage apparently carried only incidental pumice rather than being a water-pumice slurry. Water from the lake doubtless flowed last through the slightly lower east outlet.

7. Flood

A pumice-laden flood as deep as 10 m flowed from each outlet of the lake and northward down the rampart; the two arms converged in the breach 1 km north of the dome. From the level of the highest strandline at the west outlet, a pumice levee descending northward down the west side of the rampart marks the maxi-

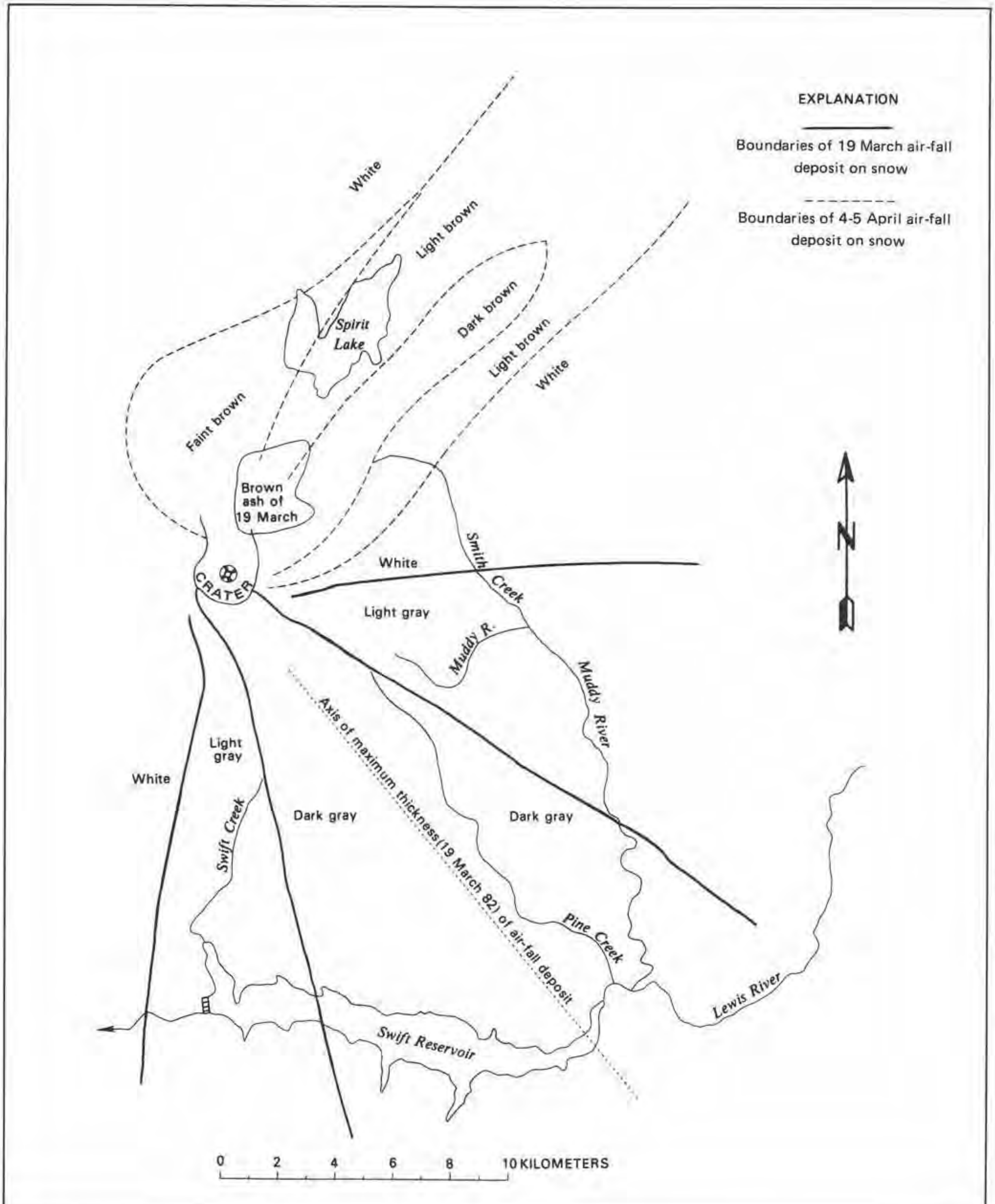

Figure 10.–Map of proximal airfall deposits of March and April 1982. Darkness is an index of relative thickness within each deposit over the white snow.

Figure 11.–Pumice blocks piled precariously one on another in the area of the transient crater lake. Ice-axe handle is about 60 cm long.

Figure 12.–View westward across area of the west outlet of the transient crater lake. Several minor pumice strands and terraces in pumiceous waterlaid ash mark successive water levels during the rapid subsidence of the lake and wane of the flood. Floating pumice blocks were stranded atop grounded, ash-veneered snow blocks that had floated earlier. Exposed part of ice-axe handle (in front of snow block) is about 70 cm long.

Figure 13.–Pumice boulder at west outlet of the lake has scour depression around it and smaller pumice blocks piled against upcurrent side. Water flow was to the right and into the plane of the photograph. Ice-axe handle is about 80 cm long.

mum flow of water from the west outlet (Fig. 14). The exact time of the flood is not known, but it must have begun within a few minutes of the eruption column; the water supply would have been exhausted within a few minutes.

As it flowed through the breach, the flood followed the same course taken by the avalanche (event 3B); the flood effects were entirely nested within the avalanche deposits (Figs. 2 and 4). The flood cascaded down the steps and spread across the pumice plain (Figs. 3 and 4). In the breach the flood was mostly eroding; it left only scattered deposits of juvenile pumice, dome dacite, locally derived angular lithic clasts as large as 80 cm, and 1980 pumice.

The flood cut two steep-walled canyons as deep as 30 m into the steep north slope of the rampart by cataract retreat. The canyons become shallower northward to the junction of the two flood streams, where they merge into an erosional channel 1 to 4 m deep and about 200 m broad. The flood left a variety of erosional and depositional landforms analogous to features formed by the great Lake Missoula floods in the Channeled Scabland of eastern Washington but orders of magnitude smaller (Bretz and others, 1956; Baker and Nummedal, 1978). These features include recessional gorges, cataract alcoves, expansion bars, anastomosing channels, overtopped minor divides, and boulder fields.

8. Further Blasts From Dome

Sparse, angular, breadcrusted dacite blocks fell on the west part of the area that had been briefly occupied by the lake. Each block lay within a shallow crater surrounded by an ejecta rim of pulverized pumice. These craters and the unmodified ejecta rims showed that the blocks were hurled from the dome after the lake had entirely emptied and flow through the west spillway had ceased. This explosion either occurred late on 19 March or was related to an explosion at 01:37 PST on 20 March (event 9). The blocks were one or two orders of magnitude less abundant

Figure 14.–View northward down levee of pumice marking limit of maximum flood from the west outlet. Exposed part of ice-axe handle is about 45 cm long.

within the lake boundary than beyond its margins, which shows that most of the dacite blocks had been ejected fairly early in the sequence (event 2).

9. Second Eruption Column

Brown clots of silt-sized ash dotted the openwork, subrounded pumice blocks in the south-through-east sector of the crater. (Wind directions were southward and eastward.) This ash could have landed only after the lake had entirely drained and the blocks had stabilized. This ash probably fell from the ash plume erupted at 01:37 on 20 March.

10. Collapse of Crater Snow

Collapse pits as large as 100 m in diameter by 10 m deep, outlined by concentric infacing scarps, formed in the thick water-laid pumice-and-ash deposits on parts of the crater floor. After the lake had drained, the still-warm pumice deposits evidently continued to melt snow buried beneath the deposits. This process may have started before the lake had entirely drained, but some of the fractures cut surface ash in the lowest channels of the east outlet. Thus most of the collapse occurred after the lake drained.

11. Lava Extrusion

A new extrusion of viscous dacitic lava was first seen at 19:30 PST on 20 March during a night observation flight. It grew for several days until 23 or 24 March from the southeast of the dome brow, draping over the southeast sector of the dome (Fig. 15). As the lobe advanced, hot dacite blocks detached and tumbled

Figure 15.–Northwestward view of growing dome lobe of dacitic lava on 24 March 1982. Composite dome is about 215 m high. Pumice and mud-saturated snow blocks in foreground are stranded on avalanche-fluted snow surface overlain by pyroclastic-surge deposit.

onto the 19 March water-laid pumice adjacent to the dome.

Correlation of Deposits, Seismic Record, Other Observations

The pilot and copilot of an aircraft some 40 km south-southeast of the mountain reported stationary "bright-red balls" above and east of the mountain roughly 3 min before a blast that sprayed glowing projectiles on the mountain and roughly 6 min before a vertical eruption column began to rise. The "balls" probably were from incandescence in the crater that reflected off clouds or an ash plume that was not directly visible at night. Such a plume must have been responsible for the pale-brown ash on the east flank of the mountain. These observations are thus consistent with the inferred precursory thermal event and ash plume (event 1) before the main eruption.

The initiation of the explosive eruption at 19:27 PST on 19 March was recorded on the regional seismic network, according to S. D. Malone and others (personal commun., April 1982). Stations on the volcano flank were "saturated" for more than 3 min 20 sec. The YEL station (Fig. 4) in the breach failed 2 min 50 sec after the initial seismic burst. The seismic signal from the SOS station 6 km northeast of the vent had two main pulses, the second and stronger at 19:30 PST. This signal continued strongly for 4 min and then gradually died down during the next 30 min. An ensuing seismically "quiet" period lasting 23 hr was broken only by occasional seismic signals of a type that typically accompanies gas emissions. The strongest of these events began at 01:35 PST on 20 March and lasted 2 min 30 sec.

The quick succession of events inferred from geologic evidence in the crater and breach (described above and listed in Table 1) correlates well with this seismic record. Small seismic bursts at 19:04 and 19:07 probably record phreatic explosions responsible for the precursory temperature rise and brown-ash deposit (event 1). The strong seismic burst at 19:27 PST

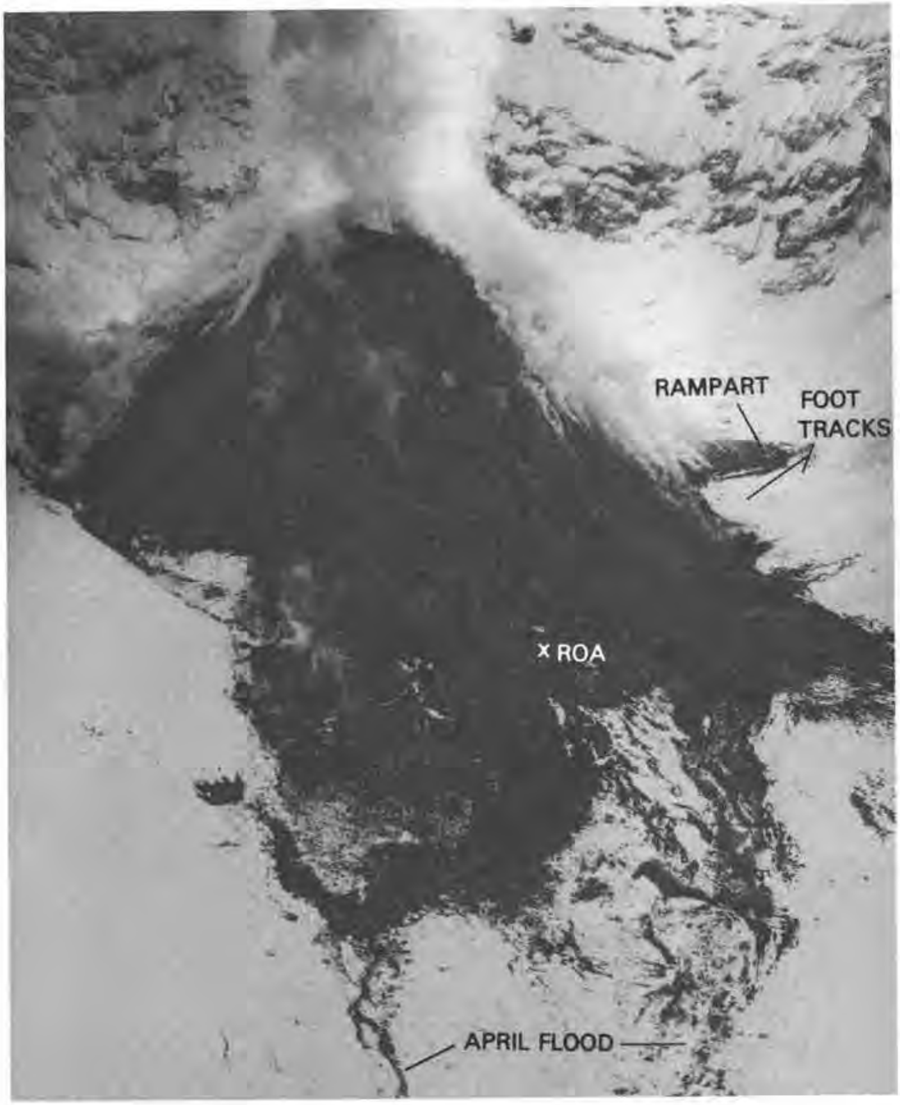

Figure 16.–View of the north side of the dome, showing rock-avalanche due to dome collapse of 4 April. Largest boulders in the talus are about 8 m in diameter. New dome lobe is being emplaced on upper brow of the composite dome. Fresh snow covers all effects of 19 March avalanche and flood and largely obscures flood features formed on 4 April. ROA is the site of destroyed seismic station. Snowy boulder just upslope from ROA is about 5 m in diameter.

corresponds to the initial blast (event 2) that initiated the snow avalanche (event 3). If the avalanche started immediately and was a single event, it took 2 min 50 sec to descend from the crater wall to the YEL seismic station in the breach 2.6 km to the north—a mean velocity of 55 km/hr (15 m/sec). The strong but diminishing signal from SOS between 19:27 and 19:30 PST corresponds to the initial blast (event 2A), further blasts from the dome (event 2B), the pyroclastic-surge explosion (event 4), and shaking due to the moving avalanche (event 3) and to impacts of ejected blocks (event 2B).

The very strong 4-min seismic burst beginning at 19:30 PST probably records the most voluminous venting of gas and juvenile pumice, responsible for the eruption column (event 5). This seismic evidence corresponds closely to the eyewitness observation that the column shot up 2 to 3 min after the observed ballistic shower (event 2A). It also corresponds to the NWS-Portland radar report that the column rose swiftly to 14 km between 19:30 and 19:33 PST; the maximum radar reflection (greatest concentration of particles in column and plume) lasted 5 min or less. The gradual decline in the seismic signal for 30 min after 19:34 PST reflects the inferred decline and cessation of venting (event 6B). The seismicity was caused partly by shaking due to the flood (event 7) and perhaps by continued gas venting and later blasts from the dome (event 8).

The seismically "quiet" period that followed corresponds to the lack of any geologic, radar, or eyewitness evidence of explosive venting. A sharp seismic burst at 01:35 PST on 20 March corresponds to an eruption column observed by Portland radar (event 9), which is probably responsible for the brown clots of ash in the crater southeast of the dome. A modest increase in seismicity starting on 20 March corresponds to the main period of dome growth and rockfalls (event 11) and to later bulging of the dome due to intrusion. These nonexplosive events were closely monitored from the crater. (See, for example, Swanson and others, 1983.)

EXPLOSIVE PHASE OF 4-5 APRIL

After two weeks of continuous swelling of the north brow of the dome (Swanson and others, 1983), the second phase of the spring 1982 eruption began on 4 April. The north side of the dome began to collapse in the late afternoon and evening of 4 April (unpub. photograph by B. K. Furukawa; Daniel Dzurisin, written commun., May 1982). The main eruption began at 20:52 PST when an eruption column (observed by NWS-Portland radar) rose to an altitude of 8.5 km. Pumiceous ash from this column drifted to the northeast (Fig. 10). The associated seismic signal had three main pulses, the third and largest lasting 3 min 30 sec. The ROA seismic station just north of the dome (Figs. 2 and 16) stopped transmitting abruptly at 20:54 PST (S. D. Malone and others, written commun., May 1982). The eruption occurred at night during cloudy weather, but we were able to glimpse the crater and north flank by 23:50 PST from a fixed-wing aircraft. The north flank of the dome had a glowing talus pile. From the northwest side of the pile a narrow dark swath trailed across the crater breach and down the north flank to the North Fork Toutle River. The feature did not change during or after eruption of the second ash column of this phase, which we witnessed between 00:35 and 00:42 PST on 5 April. This second column convected to an altitude of 10 km, and from it slight ash fell to the northeast as far as 40 km.

Ground observations on 5 April showed that a large part of the oversteepened north brow of the dome had collapsed as a rock avalanche, which left lobate deposits of large blocks as far as 1/2 km from the base of the dome (Fig. 16). An associated thin layer of gray ash mantled the snow in the breach as far as 2 km north of the dome. This ash thinned from about 5 cm at the rock avalanche to 4 mm at its sharp, lobate outer edge. The grain size also decreased outward from coarse to fine ash. The ash did not melt underlying snow that had accumulated after 19 March. The ash evidently was deposited from a cool, ground-hugging cloud of ash winnowed from the avalanche. This cloud eventually lost enough load and lateral momentum to abruptly rise, aided by being warmer than the subfreezing ambient air.

Associated with the avalanche was a flood of water, slush, and entrained rock debris, whose deposits formed the dark band seen on the night observation flight. We examined the deposit briefly on the upper part of the pumice plain. It consisted of a deposit 1 to 30 cm thick of angular fragments of gray dome rock, locally derived lithic fragments, pumice from 1980 and March 1982 deposits, and snow blocks—all in a dark matrix of lithic granular sand (Fig. 17). The flow had been wet enough to saturate snow along the margins, but in many places it failed to melt even a thin layer of snow beneath its deposits. Thus, by the time the flow reached the pumice plain, it could not have been much above freezing. The thin, discontinuous deposit it left was highly conspicuous on snow, which had covered the flood deposits of 19 March.

By early morning of 6 April, a new viscous dacite lobe had begun to grow on the northern brow of the dome. Over the next 3 days it swelled, draped northward, and produced rockfalls that formed new talus aprons (Fig. 16).

EXPLOSIONS IN 1983 AND 1984

Dome-building eruptions in February 1983 and June 1984 were each preceded by two or more explosions that laid hot volcanic debris against snow in the crater. Surficial processes included rock avalanches, snow avalanches, small pyroclastic surges, slushflows, and watery floods that downchannel bulked with debris to form lahars. The deposits, to be described and analyzed in forthcoming reports, were generally analogous to deposits of March-April 1982, although no two sequences were entirely the same. The fact that deposits analogous to those of March-April 1982 resulted from several subsequent volcanic explosions that interacted with existing snow illustrates the importance of these processes on snow-clad volcanoes.

EFFECTS OF WINTER SNOWPACK ON SURFICIAL ERUPTIVE PROCESSES

Most erosional features and deposits of the spring 1982 eruption were confined to the crater, breach, and upper flanks of the volcano. But the snow-avalanche deposit of 19 March, much of which later melted, extended as far as 8.4 km to Spirit Lake (Figs. 2 and 18). The flood deposits of 19 March and 4 April also accumulated on the lower north flank (Figs. 18 and 19) and down the North Fork Toutle River valley (Waitt and others, 1983). Neither the 19 March nor 4 April deposits would have existed had the eruptions occurred when there was little or no snow in the crater. Effects of the 19 March and 4 April eruptive phases

Figure 17.–Flood deposit of 4 April where it overflowed from the main gully onto the head of the pumice plain. Fine-grained deposits are only a few centimeters thick and overlie scarcely melted snow. Ice-axe handle is about 80 cm long.

extended far beyond the vent because they caused the snow to mass waste and melt.

SPARSE LASTING EVIDENCE OF THE SPRING 1982 ERUPTION

Airfall Deposits

Airfall materials of the spring 1982 eruptions fell in areas that received similar coarse airfall materials during the six explosive magmatic eruptions in 1980. Of the earlier deposits only those of 18 May, capped by finer ash, are stratigraphically segregated from the 1982 materials (Waitt and Dzurisin, 1981; Waitt and others, 1981). The airfall of 19 March 1982 overlies 1980 airfall deposits of 25 May, 12 June, and 16-18 October. Some coarse pumice of March 1982 fell in the same area as did coarse pumice from the June and October 1980 eruptions (Waitt and others, 1981, figs. 366 and 369; Sarna-Wojcicki and others, 1981, fig. 349). In most of its proximal area the March 1982 airfall material is coarser and more pumiceous than the underlying June 1980 deposit. But once the snow melted, the 1982 deposit became scarcely distinguishable even along its axis of maximum thickness.

On the southeast flank the stratigraphy is a thin and fine layer from 25 May 1980, overlain by a coarse and thick layer from 12 June 1980 (Waitt and others, 1981, fig. 374) and by a thinner but coarser layer from 19 March 1982 (Fig. 10). All of these deposits could be interpreted as a single, reversely graded layer resulting from a single eruption. If such a layer were seen within a stratigraphic section of ancient airfall deposits, it would be difficult or impossible to infer that three eruptions were represented. Reversely graded airfall deposits are common to pumiceous eruptions of many volcanoes (Williams, 1942; Wentworth and MacDonald, 1953; Lirer and others, 1973; Sparks and Wilson, 1976). At Mount St. Helens two superposed, reversely graded pumiceous sequences accumulated along the main airfall lobe during the single eruption of 18 May 1980 (Waitt and Dzurisin, 1981, figs. 354 and 357).

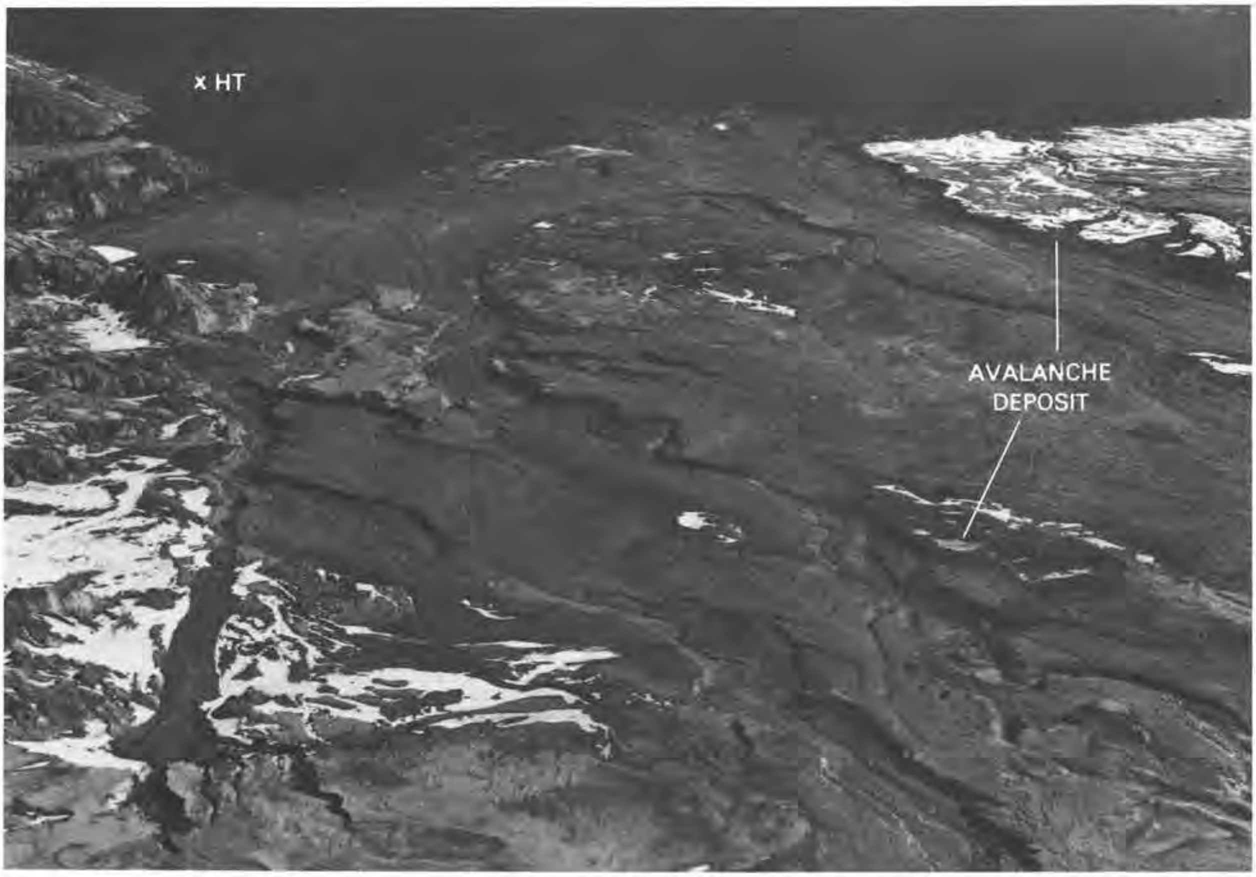

Figure 18.–Snow-avalanche and flood deposits of 19 March 1982 on pumice plain. Hummocks on left are deposits of the great landslide of 18 May 1980. HT marks the approximate site of Harry Truman's lodge before the 1980 eruptions.

The sparse pumiceous tephra of 4-5 April 1982 fell in the area of coarse airfall material from the 7 August 1980 eruption, which overlapped southward the similar 22 July 1980 airfall deposit (Waitt and others, 1981, figs. 366, 371, and 373). All of these deposits consist of loose, discontinuous fragments overlying a coherent surface of fine ash deposited late on 18 May 1980. Even though the original distribution of the April 1982 airfall deposit is known, after one season it became indistinguishable from those of July and August 1980.

Crater and Breach Deposits

The voluminous deposit in the crater and breach recorded many details of the March and April phases of the spring 1982 eruption. But several rock avalanches and explosion-generated snow avalanches, slushflows, and floods in 1983, 1984, and 1986 have obliterated most effects of 1982 in the crater and part of the breach. Between mid-1982 and late 1986 several new dome lobes and their talus aprons have encroached tens to hundreds of meters over the spring 1982 deposits on much of the crater floor and upper part of the breach. If the dome continues to grow even for a few more years, the dome and its talus will bury the remaining crater floor. Great fans of rock debris and snow have also grown tens to hundreds of meters out onto the crater floor and the sides of the breach from the steep crater walls hundreds of meters high (Figs. 1, 2 and 3). By 1987, accretion of these debris and snow fans have buried most of the former crater floor and some of the breach. Continuation of this activity may eventually bury most effects of the spring 1982 eruption that remain in the breach.

Snow-Avalanche and Flood Deposits on the North Flank

The 4 April 1982 flood was very small and mostly confined to a narrow gully system (Fig. 19); its deposits were thin and composed mostly of reworked materials. Once the adjacent and underlying snow melted, the deposits could not be distinguished from

Figure 19.–Flood deposit of 4 April 1982 where it descended the steps and fanned out on the upper part of the pumice plain.

those of 19 March 1982. Small eruption-caused snow avalanches, slushflows, and floods in February 1983, May-June 1984, and May 1986 had similar extents and distributions below the breach. Erosional gullies were deepened and greatly widened between 1982 and 1987 by noneruptive streamflow, and most 1982 deposits were removed from the gullies.

The snow avalanche and large flood of 19 March each left extensive deposits on the pumice plain (Figs. 18 and 20) that were clearly recognizable after the constituent and adjacent snow had melted (Figs. 21 and 22). The March 1982 avalanche and flood extensively modified the surface of 1980 pumiceous pyroclastic flows that had formed the 1980 pumice plain (Rowley and others, 1981; Wilson and Head, 1981). These 1982 erosional effects and some of the deposits will remain distinct until the area is far more dissected by winter runoff, swept by a larger flood, or covered by new pyroclastic deposits. The 1982 record on the pumice plain would then be patchy and difficult to interpret. The deposits initially lay along and adjacent to a broad shallow channel system, but this distribution could become obscured by dissection or burial. The flood deposits and the unconformity beneath them would survive in places, and the rounded pumice, angular lithics, and the sorted, sandy matrix may suggest an origin by water flow. But deposits of other origins—for instance, parts of the devastating 18 May 1980 pyroclastic density current (Waitt, 1981) and some pyroclastic flows—have some similar properties.

The snow-avalanche deposit, if buried, would be even more enigmatic, for it is a poorly sorted diamict containing generally angular clasts and is lying on a scarcely eroded surface. Once the surface form and areal distribution were obscured, remnant deposits could be misinterpreted. If such a deposit were encountered in a stratigraphic section, probably few investigators would think of snow avalanche as a likely emplacement process. Yet, if eruptive history of a volcano is to be inferred from its deposits, it is important to correctly ascertain the origin of such deposits.

Decades from now will it be possible without *a priori* knowledge to distinguish any effects of 1982 from those of 1980? The diverse deposits of the 1980 eruptions are voluminous, and they broadly overlie soil, root mats, leveled trees, manmade roads, and other unambiguously pre-1980 surfaces. But the identity of individual deposits and the internal stratigraphy of post-May-18 1980 deposits are understood with the assistance of contemporary field observations and photographs. The spring 1982 eruption of Mount St. Helens may eventually become unrecognizable in the preserved geologic record.

IMPLICATIONS OF 1982 VOLCANO-SNOW INTERACTIONS FOR OTHER ERUPTIONS AND VOLCANOES

Details of the flows during the March-April 1982 eruption at Mount St. Helens are known because they occurred on a volcano that was being closely monitored, because of access to the crater by helicopter, and because weather was unseasonably clear just after the March episode. Such eruption effects are probably fairly common on snow-clad volcanoes. Avalanches and floods remarkably similar to those of 19 March 1982 were indeed produced by much smaller explosions in the snowy Mount St. Helens crater in 1983, 1984, and 1986.

Dark, snowy flows have occurred as a consequence of minor explosive activity at many snow-clad volcanoes. A minor steam explosion at Mount Wrangell volcano, Alaska, on 3 September 1899 produced a dark flow that descended the mountain for many kilometers and melted gorges into the snow and ice but was cold by the next day and buried by new snowfall

Figure 20.–View northward on 24 March 1982 down right-lateral levee of snow-avalanche deposit of 19 March on pumice plain. Deposits of trailing flood (left) were confined by the avalanche levee.

(Benson and Motyka, 1979, p. 24). A similar dark, cold flow resulted from a minor explosion at Chokai volcano, Japan, in 1974 (Ui and others, 1976-77). An eruption of Grímsvötn, Iceland, in May 1983 produced a black snow avalanche strikingly similar to the proximal part of the March 1982 avalanche at Mount St. Helens (Jökull, Ar. 34 [1984] cover photography). A lava flow interacting with snow and ice generated rock-snow-ice avalanches and lahars at Villarica volcano, Chile, in 1984 (Gonzalez-Ferran and others, 1984). In November 1985, small pyroclastic surges and pyroclastic flows swiftly melted only a few percent of the snow and ice on the upper flanks of Nevado del Ruiz volcano, Colombia (Naranjo and others, 1986; Comite de Estudios Volcanologicos, 1986, unpub. report), to generate floods that claimed some 25,000 human lives. On glacier-clad volcanoes like Mount Wrangell and Nevado del Ruiz, avalanches and mudflows become part of the glacier once they are buried by the next snowfall, and thus on the volcanic cone depositional record of the event is soon lost. Minor activity between major explosive eruptions is probably more common than is commonly recognized, for flows must be large enough to reach beyond the base of the volcano for their effects to be noted.

Investigators of deposits at all snow-clad volcanoes should recognize the importance of loose snow to volcanic-flow phenomena. Potential hazards include large snow avalanches and slushflows triggered by an eruption. The snow avalanche at Mount St. Helens on 19 March 1982 was as mobile and moved as far from the vent as had pumiceous pyroclastic flows during the 1980 and several earlier, "major" eruptive episodes (Crandell and Mullineaux, 1978; Rowley and others, 1981; Crandell, in press). The devastation near Ruiz volcano on 13 November 1985 poignantly illustrates the importance of understanding volcano-snow phenomena. This minor eruption was vastly magnified because hot materials interacted with surficial snow and ice.

Figure 21.–View westward in October 1982 across hummocky, unsorted deposit of 19 March 1982 snow-avalanche. Openwork pumice blocks are concentrated along the outer margin. The distinct levee of Figure 20 has largely collapsed, owing to loss of the dominant constituent, snow. The flood deposit in the upper part of photograph contains larger and more scattered lithic blocks. Shovel handle is 70 cm long.

ACKNOWLEDGMENTS

The explosive events of 19-20 March and 4-5 April 1982 occurred at night, and most of the affected area was accessible only by helicopter. Some pertinent information from this and later eruptions originated from night-flight observers and volcano-monitoring crews of the USGS Cascades Volcano Observatory. Helicopter pilots made skillful landings under sometimes marginal conditions. Radio dispatchers, electronics technicians, and many others all together make possible a winter field operation at an active volcano in mountainous terrain. Our inferred sequence of events for 19 March has been influenced by discussions with colleagues, particularly Daniel Dzurisin, R. T. Holcomb, R. J. Janda, T. C. Pierson, D. A. Swanson, and Barry Voight (The Pennsylvania State University). The manuscript has benefitted from reviews by D. R. Crandell and C. G. Newhall.

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Figure 22.–View southward in October 1982 across surface swept by flood on 19 March 1982. The outer left-lateral margin of the flood deposit is marked by a levee 10 to 50 cm high of openwork pumice blocks comparable to those shown in Figures 13 and 21.

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ABBREVIATED BIOGRAPHIES[1]

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The five retired professors to whom the symposium on the geology of Washington is dedicated wish to express their appreciation for the honor that is done them by the Cordilleran Section and the conveners of the meeting.

A. LINCOLN WASHBURN

A. Lincoln Washburn, an honors graduate of Dartmouth College, majored in geology and skiing—he made the Olympic ski team in 1936. After entering Yale graduate school, he emerged in 1942 with a Ph. D. earned under the direction of Richard Foster Flint. His problem was the geology of parts of Victoria Island in arctic Canada.

Washburn then spent some 10 years in cold-regions research and exploration in northern Canada, Alaska, Greenland, and Antarctica before becoming executive director of the Polar Institute of North America (1945-1951) and assuming the same post for the Snow, Ice, and Permafrost Research Establishment of the Army Corps of Engineers (1951-1952). In 1953 he became professor of northern geology at Dartmouth.

He moved to Yale in 1960 as a professor and later as director of graduate studies in geology, until he came to the University of Washington in 1966 to establish the Quaternary Research Center, a multi-departmental, multi-college laboratory for the study of the Ice Age and related phenomena. He established the journal *Quaternary Research*, attracting highly qualified staff and generous grants. Simultaneously he finished and saw published a ponderous tome on periglacial effects. He turned the Center over to Estella Leopold and took early retirement in 1976. Washburn's interest in cold-regions research has not abated. He has built a summer home at Resolute Bay in northern Canada, with a fine view of the Arctic Ocean.

There is no finer friend of the Pleistocene than A. Lincoln Washburn.

PETER MISCH[3]

And then there is Peter Misch, a native of Germany, where his father was a distinguished professor at the University of Göttingen. Peter began to collect fossils and minerals at an early age, and his father, a philosopher by training and practice, drew books on the subject of Peter's interest from the university library to aid him. The books were pre-Wernerian and in Latin. The going was tough, but fortunately Peter chanced to meet some geology students and their professor, Hans Stiller, in the field. When Peter, age 12, was pushed forward to answer the professor's questions about the geology that the students could not answer, a long-lasting association began.

Peter eventually took his Ph.D. degree under Professor Stiller's direction at Göttingen, working on a structural problem in the Spanish Pyrenees. The year following his graduation was spent at Göttingen as a post-doctoral assistant, before his selection as the youngest member and scientist on Willi Merkel's second expedition (1934) to Nanga Parbat in the Himalayas. Peter had been an excellent skier and climber since youth, so when the expedition ran into bad weather and six of the party were lost, Peter did heroic rescue work.

He then returned to Germany to begin study of the collected rocks and to prepare the scientific report of the expedition. Warned that the Nazis were planning to remove him from the project and confiscate his materials and writings, he and his wife escaped from Germany.

The Misches went to Canton, China, in 1936, where Peter became a professor in the geology depart-

[1] Written for the occasion of the symposium on the geology of Washington at the Cordilleran Section Meeting of the Geological Society of America, April 1982.

[2] Barksdale's autobiographical sketch was rewritten after his death by his colleague E. S. Cheney; it is presented with the memorials.

[3] Peter Misch died July 23, 1987.

ment of Sun Yat-Sen University. When the Japanese invaded China, Peter assisted in moving Sun Yat-Sen University overland in 1938 to a site near Kunming in Yunnan Province. Kunming was the city to which National Peking University and several others were relocated during the war. Peter transferred to Peking University in 1940 and taught there until war's end. Contributing to the war effort, he did extensive field work in Yunnan from 1940 to 1946.

He left China and flew to Calcutta, where he spent a month studying some of the Triassic collections of the Geological Society of India. He had been invited to continue his Triassic studies at Stanford, and he was visiting his sister and pursuing the identity of his Yunnan ammonites when in the spring of 1947 the Cordilleran Section of the Geological Society of America held its annual meeting on the Stanford campus. There he met Professor G. E. Goodspeed, and they discovered a strong mutual interest in granitization.

This interest led to Peter's being offered an appointment to the geology faculty at the University of Washington. He began his teaching and research program in metamorphic petrology and structural geology in the fall quarter of 1947. He quickly made the northern Cascades of Washington his own, and a research area for the many graduate students whose work he has supervised. Never parochial, his structural interests took him for a time to eastern Nevada for research and thesis supervision. Peter retired in 1980, but is very active, coming to his office almost every day.

HOWARD A. COOMBS

On seeing Howard A. Coombs, such a soft-spoken, mild-mannered individual, one would never suspect that he was born in Dallas. However, he was schooled in Toronto, Canada, and in the Chicago of Al Capone and Dion O'Bannion. He fled that environment to enter the University of Washington as a pre-law major.

A chance encounter with a beginning geology course disrupted his plans forever. He became the undergraduate summer field assistant to Dean Henry Landes, who was seeking to identify potential hydroelectric dam sites along the Columbia River. Coombs did all of his university work at Washington and was added to the staff in 1934 while finishing his dissertation, a study of Mt. Rainier, which he climbed dozens of times while serving as national park ranger during the study.

Many of you will remember that hardly any major dam could be built in the late 1930s and early 1940s without the approval of Charles P. Berkey, consulting engineering geologist of Columbia University. As Berkey became overtaxed, Coombs became his Washington-Oregon legs and eyes on several dam site jobs, his engineering geology reputation building on Berkey's approval.

During and after World War II, Coombs became the dam site expert of the Northwest. At the same time he was a superb teacher at both the graduate and undergraduate levels. He succeeded Goodspeed as the department chairman in 1952 and served the department in that capacity for 17 years, during the period of its greatest growth. He returned to full-time teaching in 1969, until he fully retired in 1976. It was his proud boast during his very full university career that he never missed a class because of administrative or consulting duties. Now he is free to give full attention to problems of geologic hazards affecting atomic-power-plant sites, and more recently to volcanic hazards.

MEMORIALS

MEMORIAL TO JULIAN DEVEREAUX BARKSDALE
1904-1983[1]

by
Eric S. Cheney

Julian Devereaux Barksdale died of a heart attack on 20 December 1983. "Barky," as he was known by all, will be remembered for his service of more than four decades to the University of Washington and his simultaneous geological research in the Methow Valley on the eastern flank of the Cascade Range of Washington. More important, he will be remembered as a man who befriended and remembered everybody, both geologist and non-geologist, the young and the old, the chiefs and the Indians. Any people-oriented cause would have him pulling like a Clydesdale. Barky considered himself a historian of rocks, geological science, and people. His history *Geology at the University of Washington, 1895-1973* (1974) is a masterpiece, especially for those who can read between the lines.

Barky graduated from high school in Beaumont, Texas, in 1920, the same year that his father died. He spent most of the next six years as a roustabout in the local oil fields. From 1926 to 1928 he attended the University of Texas. Then began his love affair with Stanford University, from which he received an A.B. degree in 1930, as well as a special interest in sedimentary rocks. After working for Cities Service in Mexico, New York, and Pennsylvania, he returned to Stanford in 1932 and began doctorate work on the Shonkin Sag laccolith of Montana. Between 1933 and 1937 he rose from camp cook to director of the Stanford Geological Survey, the university's summer field mapping course. Here he formed friendships with S. W. Muller, W. C. Smith, and W. C. Putnam. In 1934, however, Aaron Waters, his advisor at Stanford, shipped him off to Yale to complete his Ph.D. under Adolph Knopf.

Barky arrived at the University of Washington in 1936. This was the beginning of the Yale-Stanford geoscience cadre at U.W. Barky published on the petrology of the Shonkin Sag (1937), but taught the history of geology, nonmetallic resources, structure, field methods, and seismology. George E. Goodspeed, the first of the American granitizers, was beginning his 16-year term as chairman of the Department of Geology. Perhaps this soft-spoken Bostonian was the one who taught Barky that the oilcan is mightier than the sword. In any event, Barky greatly admired him (1977).

In Barky's own words, "In 1938 I took on two great encumbrances..." In June he married Marajane Burns Warren. At the end of the field season, Waters invited his mentors, Professor and Mrs. Knopf, and the Barksdales on a geological tour of the Chelan, Okanogan, and Methow valleys of north-central Washington. The upper part of the Methow was almost geologically unknown, but the thick Mesozoic units seemed to go forever. Barky recalled (1974, p. 38-39):

> As the party stood at Harts Pass and marvelled at the beautifully displayed arkoses, Mrs. Knopf took Marajane Barksdale aside and exhorted her not to let Barksdale begin on so vast an area in which there were no maps... The advice was sound, but the temptation was too great.

He could not have managed the Methow without the other "encumbrance." For years Marajane drove him to the end of logging roads, and while waiting for him to finish traverses, she patiently read thick volumes or gathered bark and flowers for the art classes she taught at elementary schools. At times, she and son Tucker would accompany Barky and the pack train of horses into wilderness along the Canadian border. Sometimes Barky's only field assistant was his dog of mixed ancestry, Migma. Their summer camp in the Methow always seemed to be abuzz with impromptu visitors (with their geologic maps or histories flapping). Unlike the other city folk "from the Coast," the Barksdales were almost accepted as Methow natives.

One paper on the Methow did emerge before World War II. Much to the surprise of a later generation of Pleistocene geologists, Barky noted that Canadian continental ice had overridden Harts Pass and had extended far down the Methow Valley. His mapping of the extent of erratics indicated that only the peaks above 7,200 feet had been spared (1941). Of course Barky learned much about the petrology, stra-

[1] Reprinted verbatim from the memorial published by The Geological Society of America Memorials, Volume XV, 1985.

tigraphy, and structure of the bedrock of the Methow, but World War II intervened, and after the war just one short paper (1948) and half a dozen abstracts appeared. As Elenora Knopf had predicted, it would take at least a lifetime to map the rugged 2,000 square miles of Okanogan County that friends and colleagues know as Barksdalia.

Although he was old enough to avoid military service when World War II came, Barky was among the first to go. Having volunteered to work on the naval petroleum reserves, he found himself a lieutenant commander without wings in naval aviation. While a supply officer in the South Pacific, he found that a well-placed bottle of bourbon was even mightier than the oilcan. His interest in sedimentary geology was enhanced by the military problems with coral reefs. His real reward, however, was the ability to relate to GIs returning to the University of Washington after World War II and the Korean and Vietnam wars.

Before and after World War II, other interests competed with the Methow. He taught seismology and with H. A. Coombs published (1942, 1946) on the two largest earthquakes in Puget Sound. He revisited and revised his work on the Shonkin Sag (1950, 1952). Ever the utility infielder, he handled introductory geology, physical stratigraphy, sedimentary geology, and geology in world affairs for non-science majors. Incidentally, his teaching and conversation were richly embellished with anecdotes, history, and the latest news on each geologic subject; students and colleagues were either delighted or exasperated as he threaded his way back to the original topic.

The real reason that the geology of the Methow advanced slowly was that Barky devoted his talents to the University of Washington. He was noted for his mnemonic aptitude; he never forgot a name or a face, whether faculty, student, or staff. He could recognize the face of a brother whose sister he had in class several years before. After a second meeting with you, Barky could practically recite your genealogy. This great interest in people served the University of Washington and the geologic profession well. He was chairman of the Faculty Senate in the late 1950s. In the early 1970s at the height of the black students' "unreconcilable" dissatisfaction with the university's athletic program, the quiet man from Beaumont was a major force in solving the problem. Usually it was impossible to walk across the campus with him, for he would stop to chat with everybody; after lunch, he would frequently visit the offices of the secretaries in the Administration Building, charming information out of them. From 1964 to 1970 he was the director of the Honors Program for undergraduates in the College of Arts and Sciences; of course he knew every student. From 1970 until his retirement in 1973, Barky was the undergraduate advisor for the Department of Geological Sciences; he was one of the reasons that the number of majors soared to two hundred. As Grand Marshall and bearer of the university mace at commencement and other official ceremonies, he was for many years the virtual symbol of the university.

Until the early 1960s the geology department was small and tightly knit, so the Barksdales became close friends with H. E. Wheeler, V. S. Mallory, H. A. Coombs, J. H. Mackin, E. B. McKee, and their respective wives. Mallory and McKee cooperated by holding summer field camps in the Methow. Most of the faculty members hired in the 1960s were not from the Pacific Northwest. Barky and Marajane became the surrogate grandparents of a least four families, participating in birthdays, graduations, and even some vacations. The youngsters (and their parents) received the good advice that only grandparents can offer.

Barky's love for the university cost him dearly. The student unrest and disorders of the late 1960s that struck so many universities were an attack on his university. Fully a tenth of the pages of his history of geology at the University of Washington (1974) described these confrontations, and "Somehow the fun seemed to go out of teaching..." This anxiety was partially responsible for his first serious heart attack in 1969.

Geologic friends shuddered at the prospect of Barky's knowledge of the geology of the Methow going to the grave with him. One evening while returning from Harts Pass, Barky was admonishing a young colleague about "publish or perish." By the time the party returned to the Barksdale's camp, Barky was convinced that he might perish before he published. By forsaking the detail that he had cherished, he thwarted his greatest fear that his report on the Methow (1975) would be published posthumously. Barky also lived to see his other two fondest hopes realized: Tucker married and eventually produced a granddaughter.

Despite his fragile health in recent years, Barky had the patience of Job. Soon he knew doctors and hospitals the way he did geologists and the university. In 1980 he announced, "Hooray, I have another tax deduction." When asked why, he admitted that he had just been declared legally blind. Thereafter, his great frustration was that he could only distinguish his many acquaintances by their voices. Marajane became his eyes. With "three typewriter educated fingers," he wrote a popularized version of the geology of the Methow (1983). He also worked diligently to raise funds from the alumni for the Goodspeed and other departmental scholarships, and spent considerable time cataloging the mineral specimens of the university's museum. At a Geological Society of America symposium on the geology of Washington at the Cordilleran Section Meeting of 1982, he was called upon to give the biographies of those who had recently retired from the University of Washington. When he left virtually no time for his autobiography, J. T. Whetten filled the gap; the standing ovation for the extraordinary man brought tears to the most ductless of eyes.

Barky is survived by Marajane who continues to live in Seattle, by Tucker and his family who have become stalwarts of the Methow Valley, and by many students and extended families everywhere.

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MEMORIAL TO RANDALL LEE GRESENS
1935-1982[1]

by
Eric S. Cheney

Randall Lee Gresens, his wife, Miriam Turley Gresens, and their friend E. Bates McKee were killed in a light plane crash near Wenatchee, Washington, on July 17, 1982. Randy was 47 years old. He will be remembered for his innovative and controversial contributions to metasomatism and to stratigraphic and structural interpretations in the Wenatchee area and in the Precambrian rocks of New Mexico. He was a quiet, informal, open man with an overwhelming penchant for fairness.

Randy was born on May 11, 1935, in Harvey, Illinois, the son of Harold F. G. and Helen E. Gresens. The family lived in Midlothian, Illinois, where his father owned and operated a neighborhood drug store and where Randy attended grade school and high school. Although he was an honor student at Northern Illinois University, he volunteered to be drafted in the Army. His experience as an Army clerk-typist (1954-1956) taught him (1) typing (which was the source of voluminous single-spaced memos throughout his life), (2) not to be intimidated by authority, and (3) that education is the key to escaping mindless drudgery.

His interest in geology followed his discharge from the Army when he and a friend visited some gold mines in Ontario. This must have been a continuing fascination, because some of his last papers described the setting of the epithermal gold deposit near Wenatchee (1980) and the geochemistry of the greenstone gold deposits in the nearby Blewett district (1982).

To return to college he trekked to New Mexico Institute of Mining and Technology in Socorro. This ignited his lifelong love of the American Southwest. He transferred to the University of New Mexico in 1958, where he earned his B.S. degree in geology in 1960. While at UNM he met Mimi; they were married in November of 1960.

At Florida State University Randy worked on his doctorate under George W. Devore, whom he greatly admired. His dissertation was on the genesis of zoned pegmatites in the Petaca district of northern New Mexico. Using emission spectrographic analyses and adapting the concepts of Hemley and Jones on silicate equilibria in alkali chloride solutions, Randy concluded (1967b, 1967c) that the pegmatites probably had formed hydrothermally in low-pressure zones in the schists. This was not a popular concept at a time when most petrologists favored crystallization from a silicate melt.

Leaving Florida State with the first of three daughters, the Gresens spent a postdoctoral year (1964-65) at the University of Southern California. Here their second daughter was born. Los Angeles seemed great until those nights they put their deck chairs on the roof and watched riot-torn Watts burn. The Gresens were ecstatic to move in 1965 to the climatic antithesis of the Southwest—Seattle. There, Randy became the geochemist in a rapidly expanding department at the University of Washington that was well known for its "hard-rock" petrology. The Gresens family also expanded to include a third daughter.

At the University of Washington, Randy became adept at X-ray diffraction and emission spectroscopy, atomic absorption, neutron activation, mass spectrometry, and the electron microprobe. For several years he supervised the departmental X-ray facilities; he established the atomic absorption facilities and, in the year before his death, helped to acquire and bring into operation an inductively coupled-plasma-atomic emission-spectrometer (ICP).

His experience with pegmatites suggested that mineral equilibria in aqueous chloride solutions could explain several other enigmatic rock types. He used geochemical data and his geological observations and those of others to propose the metasomatic origin of some blueschists (1969), kyanite deposits (1971), and massif-type anorthosites (1978). He was delighted to have his paper on blueschist alteration during serpentinization selected by W. G. Ernst for inclusion in *Metamorphism and Plate Tectonics*, a volume of Benchmark Papers in Geology published in 1975. His most important paper during this time probably was on the composition-volume relationships of metasomatism (1967a).

[1] Text reprinted verbatim from the memorial published by The Geological Society of America, Memorials, Volume XIV, 1984. The references have been expanded.

Meanwhile, he continued work in New Mexico with H. L. Stensrud on the geochemistry of micas (1974b), the metamorphic stratigraphy (1974a), and geochronology (1975) of the pre-Phanerozoic rocks that contain the pegmatites. His hotly debated conclusion that the metamorphic stratigraphy is inverted seems to have been substantiated by recent publications emanating from UNM.

Randy had an overwhelming sense of fairness and balance. He deluged departmental and other chairmen with memos, suggestions, and ballot issues. Many colleagues considered him an obstructionist; others regarded him as the conscience of the department; none doubted that he would champion an unpopular cause. He served on the faculty senate. He also spent countless hours on the university grievance committee, until he concluded that the university administration was both judge and jury. In the early 1970s he organized a series of seminars for state legislators on a variety of geologic issues. In 1974 he learned French, took a leave of absence for two years, became a Peace Corps volunteer, and moved his family to the Ivory Coast; however, he resigned after five months when bureaucratic infighting precluded meaningful geologic mapping.

In the early 1970s the department established a three-week field course for non-majors and pre-majors on the eastern slope of the Cascade Range. Some believe that the geochemist's place is solely in the laboratory—not Randy. When no others volunteered to continue the course, he did, and made it an outstanding success. No one was more organized or more convinced that this should be a meaningful but enjoyable geological experience. He became interested in the geology of the nearby Wenatchee area when he noted a previously unmapped unconformity in the Tertiary strata (1980). He spent several summers (employed by the Washington Division of Geology and Earth Resources) in the Wenatchee sun that shines with a southwestern intensity. He produced several abstracts and papers on the geology of Wenatchee and the enclosing Chiwaukum graben. Just before his death, he completed the first draft of a bulletin of the Wenatchee area that will be published posthumously by the state [Gresens, 1983]. He enjoyed his Wenatchee research more than any of his other professional activities.

Lately, he had redirected his research toward exploration geochemistry. This was caused by his observations on the attempted development of the epithermal gold deposits near Wenatchee and, especially, by his expertise in aqueous geochemistry and analytical techniques. During the summers of 1980 and 1981, he consulted for Chevron Resources and for Los Alamos Scientific Laboratory. His goal was to make geological and geochemical sense of the vast amounts of data reported by the NURE program for stream sediment and water samples. Rather than focus on the obvious few high values, he correlated and displayed relationships in the multi-elemental sets of data.

Although appalled by nuclear weapons, he maintained a scientific interest about them and human health. In 1982 he was investigating the merits of a major grant proposal on neutron activation due to nuclear explosions, specifically whether areas underlain by particular rocks or soils might be safer than others for humans to reoccupy.

Because of his insights into the geology of the Wenatchee area and the renewed interest by several companies in oil and gas exploration beneath the Columbia River Basalt Group, he co-authored (1981) a paper speculating on the stratigraphic and structural relationships beneath the Columbia Plateau. In 1981 he and several of his colleagues (including McKee) founded a geologic consulting firm, Cambria Corporation, based in Seattle; the firm's first major contract was to expand and attempt to document the sub-basalt relationships. As vice-president of the firm, Randy bought and proudly wore his first three-piece suit.

Establishment of the consulting firm promised a better financial future and focused his creative energies on things he could affect, rather than on the department or the university. Accordingly, he found more time for his own scientific pursuits. The geology of the Wentachee area may be the key to what lies below the Columbia Plateau, so on the beautifully cloudless day of July 17, Randy, Mimi, and Bates decided to do aerial reconnaissance of the poorly known Tertiary stratigraphy along the rim of the Columbia Plateau south of Wenatchee. He and Bates had never been happier professionally; they were doing what they loved best to do.

Randy usually had a broad smile on his oval face, especially when "singing" geological ditties while strumming the guitar that he made. He enjoyed remodeling his house, folk dancing, group singing, and bicycling. He was the master of the surprise party and friendly prank. Mimi was equally informal but even more outspoken and more hilarious. They were devoted parents. Randy's proudest trip was as a chaperone for one of his daughters who was a finalist in the Miss Teen All-American contest in Miami. For many years Mimi operated a neighborhood child-care center. She taught at the Shoreline Community College, and in August of 1982 was to begin teaching in a preschool at the University of Washington. Mimi and Randy cheered for their daughters' soccer-playing, artistic talent, and high school cheerleading. Randy and Mimi always shared neighborhood problems and triumphs; they were co-presidents of the local PSTA. The reception following the funeral was held in the local elementary school and designated a celebration of the lives of Randy and Mimi. Over 400 people joined the celebration. They are survived by their daughters, Kelli, Hayley, and Amy, and by Randy's parents who retired to the Seattle area several years ago.

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MEMORIAL TO ELLIOTT BATES McKEE, JR.
1934-1982[1]

by
Julian D. Barksdale

Elliott Bates McKee, Jr., geologist, yachtsman, and pilot, and two companions died in a plane crash near Wenatchee, Washington, July 17, 1982, while making a geological reconnaissance in the Cascade foothills.

Bates was the son of E. Bates McKee, Sr., and Katharine Pillsbury McKee; he was born January 10, 1934, at Mount Kisco, New York. He attended St. Paul's School in Concord, New Hampshire, graduating in the Form of 1951, and received his bachelor's degree in geology from Yale University in 1955.

Bates did his graduate study in geology at Stanford University and was awarded the degree of Doctor of Philosophy in 1959. He came to the University of Washington in the spring of 1958 as assistant professor with special interest in structural and engineering geology. He plunged immediately into a creative teaching program at both the graduate and undergraduate levels while continuing his research begun in California. On the basis of his teaching and research he was promoted to associate professor in 1964.

McKee's unique contribution during this early period was his recognition of the significance of the widespread occurrence in surface exposure (more than 140 square miles) of the mineral jadeite replacing sodic feldspar in clastic sedimentary and volcanic rocks of the Franciscan Formation in the central California Coast Range south of San Francisco. Because jadeite had long been thought to form only at high pressures (above 10,000 bars), McKee suggested that the Franciscan now seen in outcrop must have been at or near the Mohorovičić discontinuity at one time and that it subsequently faulted up to its present position in the area studied. He bolstered the case with papers on the association of lawsonite and glaucophane with the jadeite in the sedimentary rocks, and he pointed out that the white veins occurring throughout the 140-square-mile area near Pacheco Pass contained aragonite, the high-pressure form of $CaCO_3$ altering to calcite.

[1] Text reprinted verbatim from the memorial published by The Geological Society of America, Memorials, Volume XIV, 1984; references expanded from original.

Bates had an almost unique rapport with people of all ages. Even as a graduate student, senior professors talked to and confided in him as an equal. Conversely, he never talked down to the greenest freshman. These qualities made him a most effective teacher at all levels. He was a patient and successful administrator. Even at a junior rank he lobbied through a number of changes in long-standing, conservative curricula of geologic and supporting courses.

During the late 1960s, Bates gathered material for a forthcoming book and also found time to act as Associate Curator at the Burke Memorial Washington State Museum, giving special attention to its mineral collection.

Bates took sabbatical leave during the 1970-71 year and with his wife, Pamela, their twin boys, David and John, and his only daughter and eldest child, Katherine, visited Japan and the Asian coast, settling down in Australia and New Zealand to complete a book manuscript. *Cascadia, the Geologic Evolution of the Pacific Northwest* was published in 1972 and was an instant success. The book met a real need for students, the intelligent layman, and the amateur geologist. He began immediately to accumulate material for the book's eventual revision. There can be no better memorial to Bates than the completion of the revision on which he was working at the time of his death.

Bates returned to the campus from sabbatical in time to take charge of the mini-field course and was dragooned into the onerous task of acting departmental chairman for the spring quarter of 1972. He retired December 15, 1972, at the age of 39 to begin a career in business. The department voted and the regents appointed him Affiliate Professor of Geological Sciences, the rank he held for ten years prior to his death.

To understand the sudden career change, one need only consider the history of McKee, the sportsman. From boyhood he had two consuming passions: hockey in the winter and sailing the rest of the year. He played varsity hockey at St. Paul's and at Yale, and even participated in pick-up competition on European vacations. There was a break during his Stanford years, but he resumed the "madness" temporarily in Seattle. The sailing he never gave up. When he graduated from the small-boat class, his father, a perennial Atlantic, Bermuda-to-Bergen racer, included him several times in his crew, remarking only recently "If I had only taken Bates's advice we might have won that last race."

McKee brought his competence to the Pacific, crewing in Victoria to Maui races, and participated in Puget Sound racing at all levels. The 1982 sailing season in the Northwest bid fair to become Bates's most successful when, in his own boat with his three older sons, Bates III, Jonathan, and Charles as crew, they won their class and were placed second overall in the Swiftsure race.

Bates, with his multitude of talents, became one of the most successful yacht brokers in the Northwest and cultivated and maintained personal contacts with brokers in all the boating centers in the United States. He recently turned over the brokerage part of his business to an associate when he discovered the joys and convenience of flying his own plane. Bates always maintained his interest in geology. He was actively working on the revision of his book, *Cascadia*, and was raising money for the Corporation Fund of Washington's Department of Geological Sciences. In 1981 he joined with university colleagues in starting a promising geological consulting firm. He was truly a man for all seasons.

Bates was co-sponsor and co-organizer of the symposium at the Cordilleran Section of the Geological Society of America at Anaheim, California, in May 1982. Twenty-two papers were solicited and twenty-two presented as "The Regional Geology of the State of Washington." Bates was a Fellow of the Geological Society of America and was for eight years (1965-1973) the secretary of the Cordilleran Section of the Society; he was a member of the Geochemical Society, the Cordilleran Section of the Geological Association of Canada, the Society of the Sigma Xi, the cruising Club of America, and the Seattle Yacht Club.

In addition to the members of his immediate family previously mentioned in this Memorial, Bates is survived by his father, E. Bates McKee of Annapolis, Maryland, and two brothers, Phillip Winston McKee of Washington, D.C., and Charles Dunn McKee of Portland, Maine.

Bates McKee was a quiet but multitalented man with a fine command of the written and spoken language and a finely developed sense of good humor, who will be sorely missed by his family, geological and boating friends, and the several generations of students whose lives he touched.

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MEMORIAL TO HARRY EUGENE WHEELER
1907-1987[1]

by

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Harry Eugene Wheeler died on January 26, 1987. Although much of his early work was paleontological, he attained international prominence as a physical stratigrapher. He pioneered the distinction between time-stratigraphy (shown on area-time diagrams) and lithostratigraphy (shown on area-depth, or fence, diagrams). Before the dawn of plate tectonics, he used unconformity-bounded sequences for regional, intracontinental, and even intercontinental correlations. In his view, contemporary stratigraphers were overly preoccupied with the depositional record, sedimentology, and paleontology. Not only should volcanic rocks be included, but more especially, when the effects of erosion and non-deposition were considered, the stratigraphic patterns were clearly preservational, not depositional. Thus, unconformity-bounded sequences and their associated periods of erosion and non-deposition demonstrated (1963b):

> ". . . a striking episodicity which is unrelated to and obscured by arbitrary time-stratigraphic subdivisions (systems, etc.) . . . Moreover, both physical and biostratigraphic patterns comprise . . . a simple order, which . . . tends to negate most of the previously envisioned 'persistent'. . . intracratonic tectonic features (such as domes and basins)."

Instead of stratigraphy, Harry practiced what he called "stratology", which stressed stratigraphic principles and concepts and their implications in regional and interregional analysis and historical interpretation.

Although he was born in Pipestone, Minnesota, most of Harry's boyhood was spent in Eugene, Oregon. Later he considered himself a virtual juvenile delinquent; then he discovered geology at the University of Oregon. While an undergraduate, he had the good fortune of being a summer field assistant for Professors Earl Packard and Arthur F. Buddington. He went to Stanford in 1930 for graduate work, where, in a competitive examination, he won a Jordan Fellowship. A fellow student and life-long friend was Lawrence L. Sloss; the two of them seemed to have been imbued with somewhat the same radical brand of stratigraphy by Hubert G. Schenk, Siemon E. Mueller, and Eliot Blackwelder. Harry's A.M. and Ph. D. (1934) theses were on the McCloud limestone of northern California.

In 1935 Harry became an assistant professor at the Mackay School of Mines, University of Nevada, Reno. He spent the next 13 years at Reno working on a wide variety of paleontologic and stratigraphic problems in the Paleozoic rocks of Nevada, eastern California, and northern Arizona. He was particularly interested in Cambrian stratigraphy (1940, 1942, 1943) and the Precambrian-Cambrian boundary problem (1947, 1958b). Yet, he and his colleagues (1940,1950) were the first to recognize that the Permian Tethyan fusulinid fauna in Cordilleran strata, which are now characterized as allochthonous or suspect terranes, are unlike the North American fauna to the east.

Eventually this competent paleontologist asked a faculty member in botany in Reno to identify some fossils for him. He married the botanist, Loretta Rose Miller, in 1938. Their children, Eugene Anthony Wheeler (1939), Carolyn Wheeler Van Wyck (1940), and David Beebe Wheeler (1942) still live in the Seattle area. Until the 1960s Harry was in the field every summer, and when he did return home, he usually presented Loretta another hungry geologist or two. His avocation, like his vocation, clearly was geology.

Due to his age and weak eyesight, Harry could have avoided service in WWII, but he finally convinced the U.S. Navy to accept him. He taught in the V-12 program at Nebraska and then worked in the Hydrographic Office in Washington, D.C.

In 1948 Harry was happy to leave Reno for the greener pastures of the Pacific Northwest. George E. Goodspeed recruited him for the University of Washington to fill the gap caused by the impending retirement of Charles E. Weaver, who had specialized in the Tertiary paleontology and stratigraphy of Washington for 43 years. Until the middle 1960s the Geology Department was small and tightly knit. Harry and Loretta became close friends with Howard A. Coombs,

[1] Completed by Eric S. Cheney from materials originally prepared by Julian D. Barksdale for the symposium.

Julian D. Barksdale, J. Hoover Mackin, V. Standish Mallory, and their respective wives. A good deal of the Department's business was settled at football games or around the bridge table.

In honoring Harry at the 1982 symposium discussed in the preface to this volume, Barksdale observed:

> "Harry is a mild-mannered, soft-spoken person with a very rough pen; so rough, in fact, that a close friend and fellow stratigrapher has been known to publish what he said was a paraphrase of an old Magyar proverb: 'With Wheeler as a friend, who needs enemies?' Rough, but without malice. . . . Wheeler's stratigraphy is not always orthodox, but it is provocative."

Provocative indeed! While on the American Stratigraphic Commission from 1957 to 1960, Harry commonly found himself in the minority. His papers on the principles of stratology included the classification of stratigraphic units (1953, 1959b) into lithostratigraphic (1956), biostratigraphic (1958a), and time-stratigraphic components (1958c), as well as cyclothems (1957), unconformity-bounded units (1959a), and the true nature of base-level (1964). The applications of these concepts were even more provocative: Whereas Sloss showed in 1963 that pre-Tertiary unconformity-bounded sequences could be traced across the North American craton, Harry (1960, 1963a) extended them into the miogeoclines. With Mallory (1963c) he defined a number of Tertiary sequences in the Cordillera that workers in the Pacific Northwest only began to rediscover two decades later. He and his colleagues asserted that both the middle Devonian Catskill delta (1963a, b) and the Illinois and other intracratonic basins and domes (1960, 1963b, 1965) are erroneous constructs. Likewise, deformation in the Pacific Northwest is younger than commonly supposed (1954, 1963c), and the Columbia River basalts once extended over (not below) the Cascade Range of Oregon (1970). He stirred other controversies (1965, 1967) as well. In each case, a tall and popular tree was felled to see the forest.

Harry excelled in graduate-level courses, seminars, and brainstorming with students and colleagues. His loquacious nature and the seemingly obvious logic, but abstract nature, of stratology tended to polarize undergraduates. A former liberal arts major reflected that in his subsequent business profession it was Harry's course that taught him the discipline and ability of recognizing and evaluating missing evidence.

Harry became an honored peripatetic. During the 1950s he was a research consultant for Gulf Oil Corporation throughout the western United States. He was a visiting professor at Indiana University (1956-1957), University of Texas (1961), and Southern Methodist University (1966). In 1957 he was a guest of the French National Center for Scientific Research and provided his insight on PreCambrian-Cambrian relations. In 1960 the Wheelers traveled extensively and examined field relations throughout Europe. His most memorable trip was from April to June 1963: he was selected to participate in an exchange between the National Academy of Science and the Soviet Academy of Science. The base was Moscow, but Harry was also able to visit Georgia, Armenia, and the Crimea and to discuss remarkably similar stratigraphic problems and patterns with the leading Soviet geologists. This trip also convinced him that the Russian people were more enlightened than the American government.

Harry retired in 1976. His concepts of time-stratigraphy, unconformity-bounded sequences, long distance correlation, and global changes of sealevel were not widely accepted until the work of Vail and others at Exxon was published from 1977 to 1987. Of course, the Exxon work was based largely upon world-wide (and proprietary) seismic reflection profiles; it also benefited from pre-existing plate tectonic concepts. Neither were available during most of Harry's career, but he enthusiastically utilized both in the North Sea while consulting for Mobil in 1973. The renewed interest in unconformity-bounded units prompted the International Subcommission on Stratigraphic Classification to define them in 1987 in the Geological Society of America Bulletin; a belated recognition of Harry's contributions is in the Subcommission's report (p. 233):

> "Wheeler (1958[c], 1959a, 1959b, 1960, 1963[b]) was probably the first to recognize unconformity-bounded units as clearly distinct from other kinds of stratigraphic units."

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