Eclogitization and exhumation of Caledonian continental basement in Lofoten, North Norway

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ABSTRACT

U-Pb and $^{40}Ar/^{39}Ar$ isotopic data combine with structural and petrological information to allow insights into the timing of Caledonian tectonic burial and exhumation of lower crustal rocks now exposed in the Lofoten Islands of North Norway (latitude 68° N). Severely retrogressed eclogites occur in rare lenses or, even more rarely, hydrated shear zones within Archaean and Proterozoic granulite-facies Baltic basement gneisses. The timing of high-pressure metamorphism in Lofoten has been difficult to determine because retrogression has disturbed mineral isotopic systems and zircon apparently was not generated, leaving its tectonic significance uncertain. Recently discovered pre- and post-kinematic felsic injections provide the opportunity to bracket the age of eclogitization. U/Pb analyses of zircon and xenotime from a prekinematic syenogranite dyke that cuts the mafic host to one of the retro-eclogite lenses intruded at 1800 ± 5 Ma, and we interpret the strong disturbance of its U-Pb system at 478 ± 41 Ma to approximate the age of eclogitization. Omphacite breakdown textures imply rapid isothermal decompression during initial uplift, followed by slow uplift. Syn-upper-amphibolite-facies thrust emplacement of the overlying Leknes Group at ca. 464 Ma implies that the Lofoten basement resided in the lower crust for ca. 14 m.y. $^{40}Ar/^{39}Ar$ cooling dates from the retro-eclogites trace their exhumation path into the middle crust (hornblende ca. 433 Ma) where they resided before being slowly elevated to levels at the base of the ductile-brittle transition in the early Carboniferous (muscovite ca. 343 Ma). The Middle Ordovician Lofoten eclogites likely formed at a time similar to the "early" group of Eocaledonian eclogites (ca. 505–450 Ma) found in the Tromsø, Seve, and Bergen Arcs allochthons that were later thrust onto Baltica during the main Scandian (Siluro–Devonian) collision. Lofoten eclogites appear to be ca. 50 m.y. older than the "late" group of autochthonous, Scandian (ca. 425–400 Ma) high-pressure and ultrahigh-pressure (UHP) eclogites of the Western Gneiss Region (WGR), and the former preserve a much longer (ca. 100 m.y.) exhumation history. To date, there is no evidence to indicate either UHP or Scandian eclogite-facies metamorphism in Baltic basement along coastal Norway north of latitude 64° N. Middle Ordovician eclogites of Scandinavia are contemporaneous with Taconic eclogites found in thrust nappes of the Appalachian Orogen. If the Lofoten eclogites correlate to those in the Tromsø Nappe Complex (Uppermost Allochthon), then both terranes might represent Laurentian relics emplaced during the Scandian and left orphaned on the conjugate side of the orogen when the North Atlantic began to open in the Eocene. Lithologic, petrologic, kinematic, provenance, and palinspastic information favor, however, correlation with eclogites in transitional Baltic-Iapetan crust of the Seve Nappe Complex (Upper Allochthon), which provides a piercing point linking eclogites in autochthonous Baltic crust now exposed in the most internal parts of the orogen with those transported and preserved in the Swedish foreland. We suggest a broad twofold subdivision for the Eocaledonian eclogite provinces into those that are (1) Baltic derived, and (2) exotic with respect to Baltica. Middle Ordovician eclogites falling under category (1) can further be subdivided into those occurring in autochthonous Baltic basement (i.e., Lofoten), and those in thrust translated terranes. Apparently, continental crust in a collisional setting can be subducted to mantle depths and show only very sparse evidence of this tectonic history.

INTRODUCTION

Eclogites are important indicators of plate tectonic processes but constitute only a very minor volume of continental basement exposed today at the Earth's surface. The Western Gneiss Region (WGR) of west Norway (Fig. 1) is arguably the largest, best exposed, most accessible, and now best studied eclogitized continental basement terrane in the world (Dobrzhinetskaya et al., 1995; Hacker et al., 2003; Hacker, 2007; Walsh et al., 2007; Hacker et al., 2010). High-pressure (HP) and ultrahigh-pressure (UHP) rocks of the WGR record deep-level subduction of the ancient Baltic margin beneath Laurentia during the Scandian (Siluro–Devonian) phase of the Caledonian orogeny (Roberts and Gee, 1985; Andersen and Jamtveit, 1990). Following UHP metamorphism, the WGR was educted out or exhumed from underneath Laurentia along the system of Early Devonian extensional faults associated with the Nordfjord-Sogn detachment (Norton, 1986; Seranne and Seguret, 1987; Andersen and Jamtveit, 1990; Osmundsen and Andersen, 1994; Walsh et al., 2007). Coupled with the lack of a subsequent contractional orogenic overprint, such as that which typifies most other mid-Paleozoic orogens (e.g., the Taconic, Acadian, and Alleghanian phases of the Appalachian orogeny; see references in Miller et al. [2010]), the Caledonian lithospheric root was rather simply exhumed to emerge as the present-day WGR. The pristine preservation in the WGR of rocks and their structural geometries serve as a world-class example of deep-continental subduction and very rapid (<10 m.y.) and large magnitude (≥120 km throw) tectonic unroofing.

Scandinavian eclogite terranes are also found at a variety of tectonostratigraphic levels within the stack of Caledonian allochthons that were thrust eastwardly upon the Baltic basement. These older, Cambro–Ordovician eclogite terranes occur at various tectonostratigraphic levels

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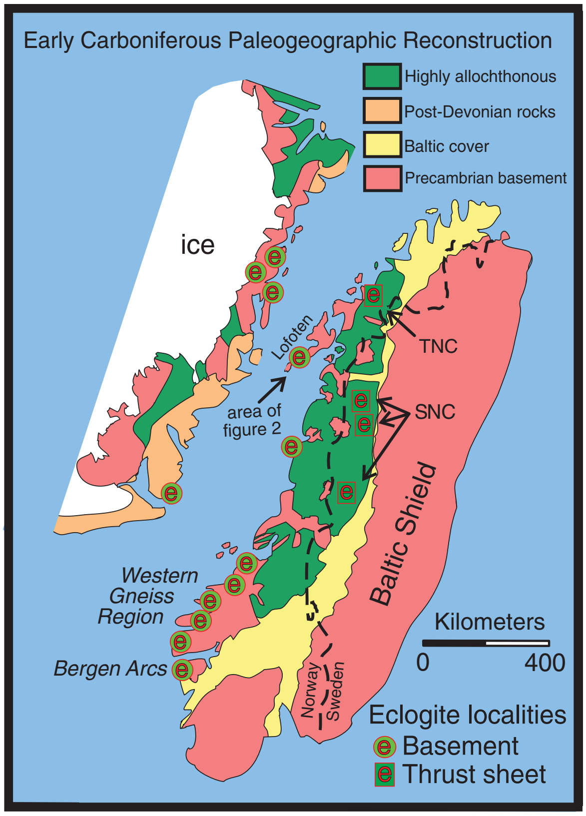

Figure 1. Carboniferous paleogeographic reconstruction of the Norway-Greenland Caledonides, depicting general distribution of eclogites (modified from Steltenpohl et al., 2006). Abbreviations: SNC—Seve Nappe Complex, TNC—Tromsø Nappe Complex.

within the Caledonian nappe stack (Fig. 1). Middle Ordovician eclogites in the Tromsø Nappe Complex (Corfu et al., 2003) occur at the highest preserved structural levels, within the exotic (Laurentian) Uppermost Allochthon (Fig. 1). The eclogitic Seve Nappe Complex (Fig. 1) constitutes the base of the Upper Allochthon in northern Sweden, and is interpreted to have been derived from the area of extended transitional crust between western Baltica and the Iapetus Ocean (Gee et al., 1985; Mørk et al., 1988). Near the base of the nappe stack, the Bergen Arcs eclogite terrane is interpreted to be a slice of Baltic basement within the Middle Allochthon (Fig. 1). Taken together, the Scandinavian eclogites thus appear to fall broadly into two temporal and kinematic groups, an "early" (Eocaledonian) allochthonous group with ages between roughly 505 and 430 Ma (Mørk et al., 1988; Griffin and Brueckner, 1985; Krogh et al., 1990; Dallmeyer and Andresen, 1992; Brueckner and van Roermund, 2001; Corfu et al., 2003) and a "late" (Scandian) autochthonous group with ages between ca. 425 and 400 Ma (Griffin and Brueckner, 1980, 1985; Berry et al., 1993; Brueckner et al., 1996; Terry et al., 2000; Tucker et al., 2001; Root et al., 2005; Hacker, 2007; Hacker et al., 2010).

The Lofoten Islands in north Norway (latitude 68° N), which are underlain by Baltic basement gneisses that project as the northern continuation of the WGR (Fig. 1), also contain eclogites, but they have not yet been dated (Wade, 1985; Kullerud, 1992, 1996; Markl and Bucher, 1997; Steltenpohl et al., 2003b, 2006; Kassos, 2008). As was earlier presumed for the WGR eclogites before more robust isotopic methods were available to date them (Krogh, 1977), the Lofoten eclogites were thought to be pre-Caledonian (e.g., Markl and Bucher, 1997) since they occur in Proterozoic basement gneisses. An earlier attempt to date the timing of eclogitization in Lofoten was unsuccessful because retrogression had disturbed the Rb-Sr and Sm-Nd mineral isotopic systems (Kullerud, 1992). Our petrographic studies indicate that the eclogite-facies rocks are barren of zircon and titanite, leaving little hope of using U-Pb methods to directly date the high-pressure event to establish Lofoten's place among the other Scandinavian eclogite terranes.

The geological history of the Lofoten islands has also been historically debated because Caledonian fabrics and structures are remarkably sparse, making it difficult to imagine how it remained so after being overridden and buried beneath an estimated ~50-km-thick pile of allochthons during the climactic continent-continent collision (Hakkinen, 1977; Tull, 1977; Griffin et al., 1978; Hodges et al., 1982). Only through targeted isotopic dating have we been able to confidently distinguish between which granite-on-granite shear zone is Proterozoic and which is Paleozoic and/or Caledonian (Hames and Andresen, 1996; Steltenpohl et al., 2003b, 2004, 2006). The paucity of Caledonian structures is thought to be the result of the limited availability of fluids to facilitate plastic deformation of the anhydrous, rigid granulites (Bartley, 1982, 1984; Hodges et al., 1982; Steltenpohl et al., 2004). The eclogites in Lofoten, likewise, are volumetrically miniscule, and are restricted in their occurrence to rare, thin (meter scale) ductile shear zones (Wade, 1985; Kullerud, 1992, 1996; Markl and Bucher, 1997; Kassos, 2008). Eclogite shear zones in continental basement rocks are exceedingly rare, and, as with those in Lofoten, are argued to have formed by eclogitizing fluids that weakened the rocks and promoted crystal-plastic flow and shear zone development (Austrheim, 1987; Austrheim and Griffin, 1985; Boundy et al., 1992; Kullerud, 1992, 1996, 2000; Markl and Bucher, 1998; Markl et al., 1997, 1998a, 1998b; Steltenpohl et al., 2006).

Herein we report field, structural, and isotopic results on a crystal-plastic shear zone on the Lofoten island of Flakstadøy (Figs. 1 and 2), which contains variably retrogressed lenses of eclogite, documenting it to be a Caledonian shear zone. During our mapping, we discovered felsic injections that both predate and postdate eclogitization, providing an opportunity to attempt isotopically bracketing the age of eclogitization. We also present U-Pb isotopic dates on minerals from one of the felsic injections that combine with $^{40}Ar/^{39}Ar$ mineral cooling dates, and regional geological information, to help place constraints on the timing of eclogitization and exhumation. In addition to clarifying how continental crust in such a collisional setting can be subducted to mantle depths and show only very little evidence for it, we believe our results are the first to link eclogitization event(s) within the basement to that recorded in one of the allochthonous terranes.

GEOLOGIC CONTEXT

The Lofoten archipelago is the deep part of a continuous crustal column exposed roughly parallel to latitude 67.5° N across northern Scandinavian Caledonides (Steltenpohl et al., 2004) (Fig. 1). Basement rocks along this transect, from east to west, range from unmetamorphosed in the Swedish foreland, through the Barrovian series, to eclogite-facies metamorphism on the island of Flakstadøy (Figs. 1 and 2). The geologic evolution of Lofoten began before 2.7 Ga, when Archean plutonic rocks

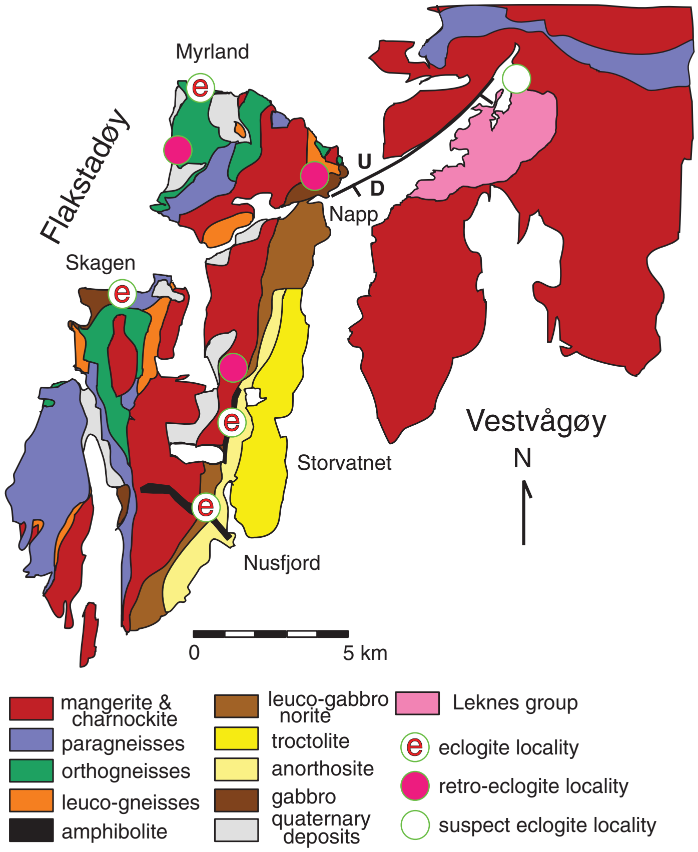

Figure 2. Geologic map of Flakstadøy and part of adjacent Vestvagøy illustrating eclogite localities (modified from Markl and Bucher, 1997; Steltenpohl et al., 2006).

intruded a supracrustal sequence preserved as rafts and xenoliths. These were migmatized at 2.3 Ga. More supracrustals were deposited at 2.1 Ga and later mangeritic and charnockitic plutons were emplaced in two pulses at 1.86–1.79 Ga under granulite-facies conditions (Griffin et al., 1978; Corfu, 2004b). Allochthonous amphibolite-facies metasedimentary rocks of the Leknes Group (Fig. 2) were thrust upon the Lofoten basement contemporaneous with metamorphism dated between 469 ± 3 Ma and 461 ± 1 Ma (Corfu, 2004a).

Our work has focused on Flakstadøy, which is underlain primarily by basement rocks of the anorthosite-mangerite-charnockite-granite (AMCG) suite (Fig. 2; Griffin et al., 1978; Markl and Bucher, 1997). The eastern third of the island is a series of north-northeast–striking and eastward-dipping mafic igneous units, the "Flakstadøy Basic Complex" of Romey (1971). From the top down, this complex comprises anorthosite, troctolite, and gabbronorite (Romey, 1971; Markl and Bucher, 1997). Both gradational and sharp contacts have been observed between these units. Pegmatites within gabbro units in the basic complex yield upper intercept ages, interpreted to be the ages of crystallization, of 1793 ± 4 Ma and 1789 ± 2 Ma (Corfu, 2004b). To the west of the basic complex is a large mangerite pluton with steeply dipping to vertical contacts (Romey, 1971). The outcrop width of the mangerite ranges from <1 km wide at the southern tip of the island to >4 km wide in the center of the island, and it is the single most abundant unit. Mangerite was emplaced at 1800 ± 2 Ma (Corfu, 2004b). The western third of Flakstadøy is predominately ortho- and paragneisses that are sheared and folded and dip subvertically (Romey, 1971).

Eclogite-facies rocks are observed in all three major rock subdivisions on Flakstadøy (Wade, 1985; Kullerud, 1992; Markl and Bucher, 1997; Kassos, 2008). The eclogites were severely to completely retrograded under amphibolite-facies conditions, but four localities, at Nusfjord, Myrland, Skagen, and Storvatnet (Fig. 2), contain the best preserved relics documenting this event. The eclogites are known to occur within ductile shear zones (Wade, 1985; Kullerud, 1992, 1996; Markl and Bucher, 1997), but we have recognized two distinct structural styles that are rheological inverses of one another. The first style occurs as narrow (<2 m thick), discrete ductile shear zones where eclogitizing fluids have mediated plastic failure within strong and rigid anhydrous granulites (Kullerud, 1992, 1996; Steltenpohl et al., 2006). The second style contains competent lenses, or knockers, of relatively massive eclogite floating within a wider (tens of meters thick), more highly distributed shear zone matrix composed of rheologically weaker, more strongly foliated and lineated schists and gneisses. Style one eclogites are well exposed at Nusfjord where Kullerud (1992, 1996) and Steltenpohl et al. (2006) have described them; we have also found this type at Skagen (Fig. 2). The second style, eclogite lenses within a wider shear zone, is critical for the context for our U-Pb and $^{40}Ar/^{39}Ar$ isotopic age studies, and the Myrland locality provides a spectacular example of this style (Figs. 2 and 3; Kassos, 2008).

ECLOGITE ASSOCIATED GRANITOIDAL VEINS AND DIKES

Pre-, syn-, and/or post-kinematic, muscovite-bearing granitoidal injections are associated with the eclogites exposed at the Myrland, Storvatnet, and Skagen localities on Flakstadøy (Fig. 2; Kassos, 2008). Prekinematic leucocratic injections are especially well exposed within and outside of eclogite lenses preserved in a >30-m-wide shear zone, the northeastern boundary of which is not exposed, at Myrland (Fig. 3). Recognizing the importance that the prekinematic veins at Myrland hold for potentially dating the timing of eclogitization, Kassos (2008) painstakingly documented their field and structural relations by gridding off the exposure, drawing and photographing 168 two-square-meter individual areas, and then digitally stitching them together to produce Figure 3. Since field relations between the prekinematic veins and the eclogite lenses at Myrland are key to how we interpret our U-Pb results, presented further below, a detailed characterization of those relations follow.

Paleoproterozoic monzodiorite and monzodioritic orthogneiss shoulder rock are progressively comminuted, foliated, and lineated toward the northeast as the Myrland shear zone is approached (Fig. 3). Eclogite lenses are encapsulated by sheared amphibolite that forms the shear zone matrix (Fig. 4A). Mafic intrusions and enclaves observed within undeformed monzonite outside of the shear zone clearly served as the host material for the formation of eclogite. The least retrogressed eclogites, containing only rare omphacite and primary garnet

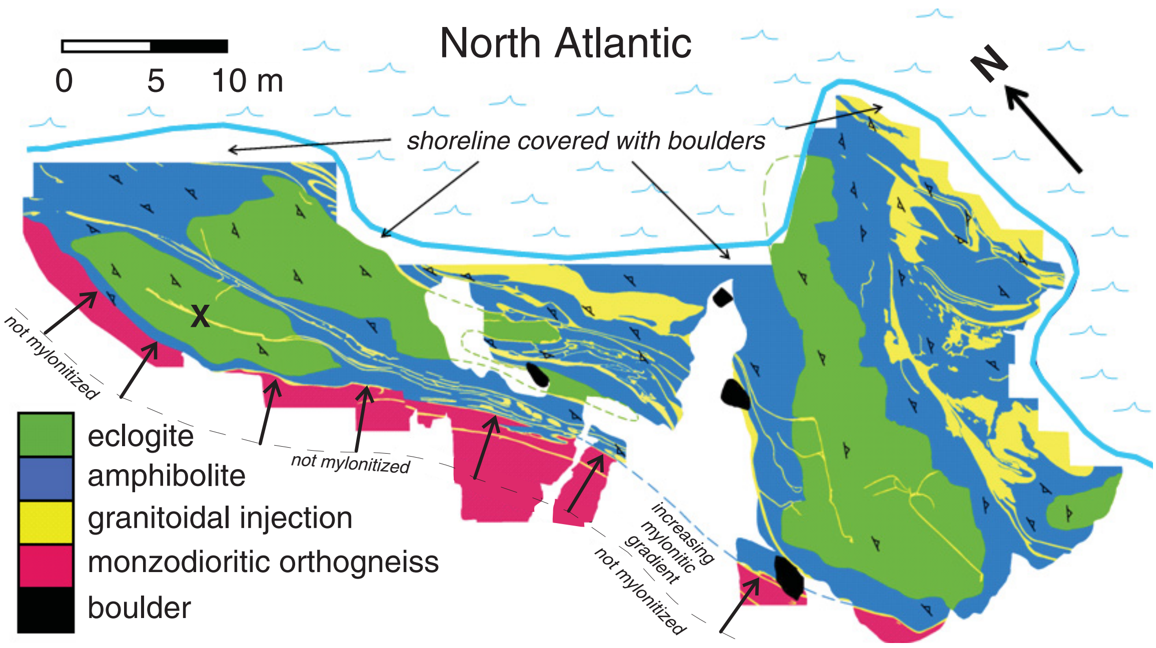

Figure 3. Detailed geologic map of the Myrland locality developed from digitally stitching together 168 maps based on a two-square-meter grid (Kassos, 2008). Strike and dip symbols reflect the attitude of the predominant amphibolite-facies schistosity and/or gneissosity, with generally moderate to steep dips. Dashed black line marks the southwest boundary of the Myrland shear zone, approximated in areas (white) where cliffs made access impossible; black arrows point in the direction of progressively increasing degrees of mylonitization of the monzodiorite. X marks location from which leucocratic vein sample AA03–3 was collected for U-Pb isotopic analysis.

relics altered variably to plagioclase, pyroxene, and amphibole, occur in the cores of the lenses where the fabric is relatively massive. Foliation becomes progressively developed toward the rims of the lenses to eventually merge with the encapsulating and anastomosing shear zone fabric of the matrix amphibolite (Fig. 4A). Rocks forming the rims of the retro-eclogite lenses and the shear zone matrix are completely devoid of any relict of primary eclogite-facies minerals, which have been wholly reconstituted to amphibolite-facies assemblages; shear zone rock microstructure, therefore, shows no sign of nonrecovered strain, meaning that the fabric now is metamorphic rather than mylonitic. Schistosity and gneissosity measured within the eclogite lenses is coplanar with that of the encapsulating amphibolite matrix, both striking northwest and dipping moderately to steeply toward the northeast or southwest (Figs. 5A and 5C). Likewise, a collinear mineral-stretching lineation is weakly developed within the lenses but is quite strong in the matrix amphibolites, lying within the foliation plane and plunging shallowly to moderately toward the east and southeast (Figs. 5B and 5D). The long axes of the eclogite lenses (e1>e2>e3) are collinear with the stretching lineation. Although asymmetric grain shapes are locally preserved within shear zone rocks when viewed perpendicular to the foliation and parallel to the lineation, no consistent sense of shear was discerned from them. Along eclogite lens boundaries, the felsic veins are consistently deflected in a dextral sense (see Fig. 3), however, which combines with the roughly strike-parallel stretching lineation to suggest primarily right-slip, with minor reverse-slip, movement. Post–shear zone folding, about an axis roughly coincident to the stretching lineation, resulted in the partial-girdle spreads of foliation pole measurements seen in Figures 5A and 5C. This late-stage folding caused the contractional imbrication of the eclogite pods apparent in Figure 3.

From Figure 3, the prekinematic granitic veins can be seen to represent as much as 15% of the total rock area exposed at Myrland (Fig. 3). The prekinematic veins occur within both the competent eclogite lenses and in undeformed monzonite outside of the Myrland shear zone, where they are tabular and planar (Figs. 4B and 4C, respectively), generally less than 20 cm thick, and have sharp, high-angle bends, clearly indicating their injection into a brittle fracture network (see Fig. 3, especially the easternmost lens). The volume of felsic material increases dramatically within the highly sheared amphibolite surrounding the eclogite pods (Fig. 3). Within the amphibolites the veins are highly contorted, foliated, isoclinally folded, and plastically necked and boudinaged within the foliation and stretched parallel to the shallow-southwest–plunging amphibolite-facies stretching lineation (Figs. 3 and 4D). Felsic dikes within the eclogite pods that meet the pod boundaries at a high angle (e.g., the southernmost pod) are either dragged immediately into parallelism with the shear zone fabric or are truncated by it. Combined with their identical characteristics in both the competent eclogite pods and in pristine monzonite outside of the shear zone (cf. Figs. 4B and 4C), we interpret the dikes to have predated eclogitization.

Bimodal grain size delineates two types of injections at Myrland, either thin (<20 cm) fine-grained veins or thicker (<2 m) pegmatites (Kassos, 2008). Pegmatites are restricted to amphibolite within the shear zone matrix whereas the finer-grained injections occur in all of the country rock units. Like the fine-grained veins within the amphibolite matrix, the pegmatites are highly contorted into flow folds and boudins (Figs. 3 and 4D). Typical mineralogy of fine-grained veins is quartz, plagioclase ($An_{15.6}$, CIPW norm value), K-feldspar, muscovite, and biotite, with accessory clinozoisite, epidote, rutile, opaque minerals, zircon, and xenotime (Kassos, 2008). Some plagioclase grains are

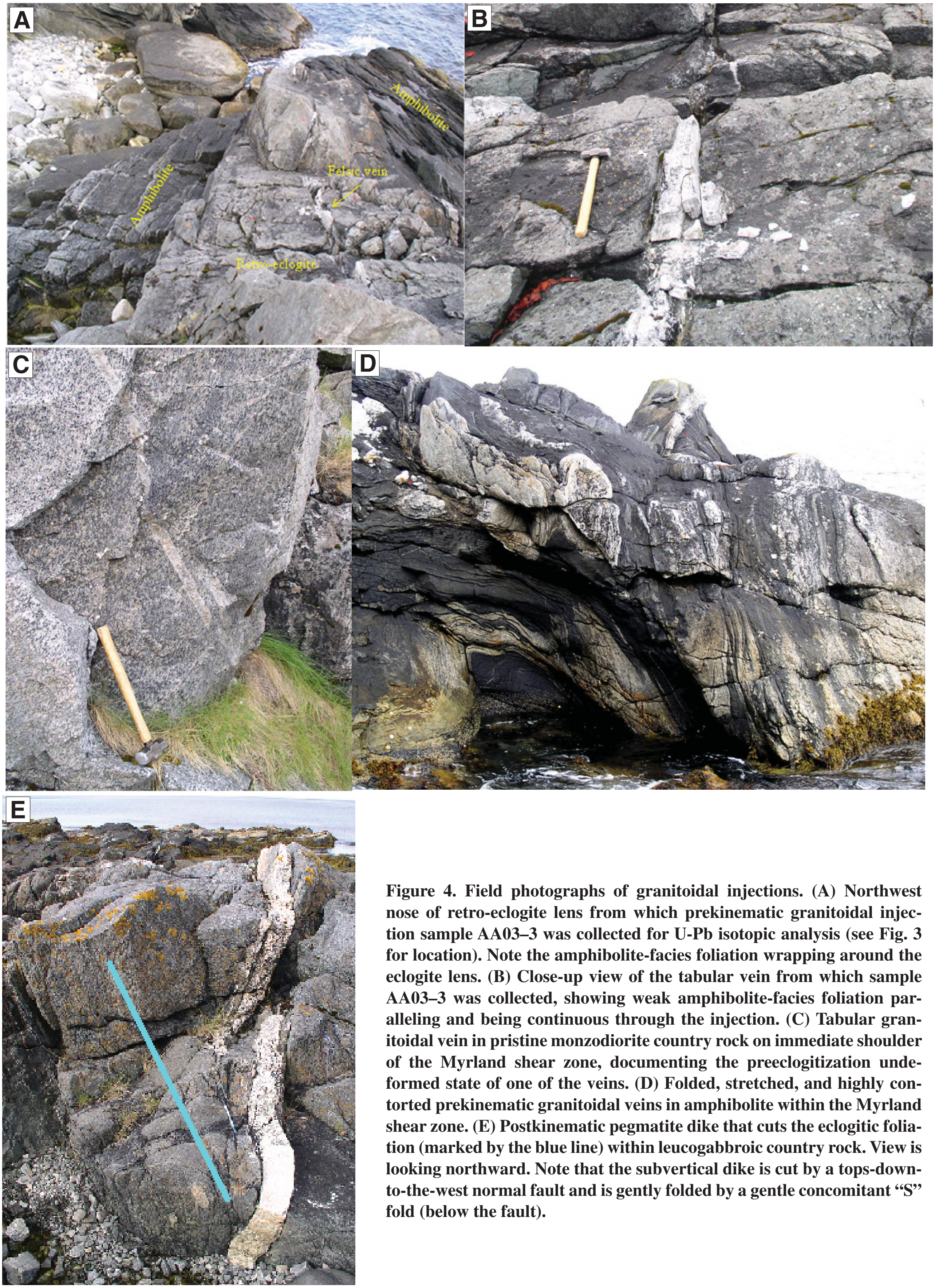

Figure 4. Field photographs of granitoidal injections. (A) Northwest nose of retro-eclogite lens from which prekinematic granitoidal injection sample AA03–3 was collected for U-Pb isotopic analysis (see Fig. 3 for location). Note the amphibolite-facies foliation wrapping around the eclogite lens. (B) Close-up view of the tabular vein from which sample AA03–3 was collected, showing weak amphibolite-facies foliation paralleling and being continuous through the injection. (C) Tabular granitoidal vein in pristine monzodiorite country rock on immediate shoulder of the Myrland shear zone, documenting the preeclogitization undeformed state of one of the veins. (D) Folded, stretched, and highly contorted prekinematic granitoidal veins in amphibolite within the Myrland shear zone. (E) Postkinematic pegmatite dike that cuts the eclogitic foliation (marked by the blue line) within leucogabbroic country rock. View is looking northward. Note that the subvertical dike is cut by a tops-down-to-the-west normal fault and is gently folded by a gentle concomitant "S" fold (below the fault).

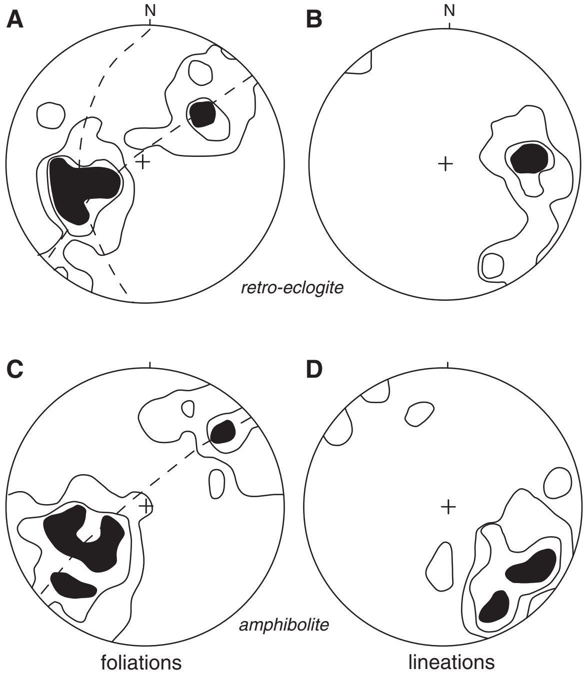

Figure 5. Equal-area, lower-hemisphere stereographic projections of fabrics measured in the Myrland shear zone. Poles to metamorphic foliation (left-hand column) and mineral stretching lineations (right-hand column) are for retro-eclogite lenses (upper row) and their encapsulating amphibolites (lower row). (A) and (C) contoured poles to metamorphic foliation; *n* = 70 and 150, respectively; contour interval (CI) for A = 0–3, 3–6, and >6% per 1% area, and for C = 0–2, 2–4, and >4% per 1% area. (B) and (D) mineral stretching lineations; *n* = 31 and 63, respectively; contour interval for B = 0–7, 7–14, and >14% per 1% area, and for D = 0–4, 4–8, and >8% per 1% area. Dashed curves in A and C are visually deduced best-fit partial girdles (see text).

twinned and/or zoned. Muscovite grains are silver to yellow-gold, up to several millimeters thick, rarely bent (as evidenced by undulatory extinction), and concentrated in 5-mm-thick bands visible in hand samples. Biotite gives the rock a speckled appearance. Fine-grained veins have a strong foliation defined by white mica that generally parallels the long dimension of the eclogite lens (Figs. 4A and 4B). CIPW normative values obtained from whole-rock geochemical data indicate a syenogranitic composition (Kassos, 2008). Pegmatite mineralogy is highly variable, ranging from ~100% quartz to plagioclase- and mica-rich. Clear quartz grains in the pegmatites are up to several millimeters in diameter, equidimensional, and exhibit no discernible signs of deformation or preferred growth direction, implying crystallization from a molten state or a high degree of static annealing following deformation. Large (up to several cm), euhedral K-feldspar and mica grains dominate pegmatite boudins.

Figure 6 summarizes our interpretation of the sequence of events that led to the field and fabric relations observed at the Myrland eclogite locality. The original monzonite pluton that forms the country rock contained enclaves of basalt (Fig. 6A). Following crystallization of the monzonitic magma, syenogranitic dikes and veins intruded both it and its enclaves (Fig. 6B). Later, the package was subducted to upper mantle levels where the basalts were locally transformed into eclogites due to fluid-mitigated eclogitization, as is well documented for other Flakstadøy eclogites, weakening the rocks and initiating crystal-plastic shear zone development (Fig. 6C; Kullerud, 1992, 1996, 2000; Markl and Bucher, 1998; Markl et al., 1997, 1998a, 1998b; Steltenpohl et al., 2006). Following exhumation to higher crustal levels, amphibolite-facies necking and dismemberment of eclogitized domains formed the eclogite lenses that were rotated as shearing within the Myrland shear zone intensified causing the zone to widen into the monzonitic shoulder rocks (Fig. 6D). Pegmatite veins were injected into the shear zone at this stage further promoting plastic flow. Late-stage folding and imbrication of the eclogite pods accompanied the final developmental stages of the Myrland shear zone. The style (i.e., its width, meso- and micro-structures, and rheological development) of mylonitic progression within the Myrland shear zone is virtually identical to that observed in shear zones throughout Lofoten where undeformed leucocratic basement rocks are sheared against supracrustal enclaves or allochthonous cover units (Klein, 1997; Klein and Steltenpohl, 1999; Klein et al., 1999; Steltenpohl et al., 2004, 2006). Focused isotopic dating on many of those shear zones has recently revealed them to be Caledonian, rather than Precambrian structures (Corfu, 2004a; Steltenpohl et al., 2003b, 2004; Ball et al., 2008; Key et al., 2008). Hence, it was the field relations that first led us to suspect that the Myrland shear zone is a Caledonian structure, which the $^{40}Ar/^{39}Ar$ study below will bear out, and if that was the case then we reasoned that eclogite formation might also be Caledonian.

In contrast to the prekinematic injections, an unusual occurrence of synkinematic granitoids with nebulus, mobilized boundaries between eclogite and amphibolite was found at the Storvatnet locality, where exposures are spotty due to high degrees of vegetative cover (Fig. 2). These apparent in situ partial melts are commonplace marginal to the eclogite lenses, but were only seen at Storvatnet. CIPW normative values indicate monzodioritic to anorthositic compositions, which combined with the field relations suggest that they are eclogite-derived "sweat-outs," interpreted to have formed due to (near) isothermal decompression melting (Vold, 1997; Watt et al., 2000; White and Hodges, 2003; Kassos, 2008). Unfortunately, petrography revealed that zircon and titanite are not present in these granitoids.

Postkinematic dikes at Skagen are nonfoliated granitic pegmatites that clearly cut across the dominant amphibolite-facies foliation that has overprinted the eclogites (Fig. 4E). Mineralogy of the pegmatites is simple, with large (up to 2 cm) K-feldspar, plagioclase, and muscovite grains, and smaller (up to 7.5 mm) quartz grains. The dikes strike north-northeast, have subvertical dips (Fig. 4E), are not foliated, and they have randomly oriented "books" of muscovite up to 1 cm thick. They are locally cut by moderately westward-dipping normal faults and shears, and are gently folded within the same kinematic plan (Fig. 4E). Based on their proximity and identical orientations and mineralogy, we suggest that the postkinematic dikes at Skagen correlate with a swarm of larger pegmatite dikes exposed on southern Flakstadøy that are dated at 410 ± 3 Ma (Corfu, 2004a).

Similar types of pre- and post-eclogite facies granitoidal injections are reported from eclogite terranes in China (Wallis et al., 2005) and East Greenland (Gilotti and McClelland, 2005) and have been used to help constrain processes and timing of eclogite formation and their exhumation. The postkinematic granitoid veins at the Skagen locality are barren of zircons, a problem also experienced for the postkinematic granitoids investigated by Wallis et al. (2005). Below we report our attempt to date one of the prekinematic injections from the Myrland locality.

U-Pb GEOCHRONOLOGY

Sample AA03–3 was taken from a prekinematic felsic dyke that occurs within an eclogite lens at the Myrland locality (location X in Fig. 3; see also Figs. 4A and 4B). This sample is a fine-grained white syenogranite with a

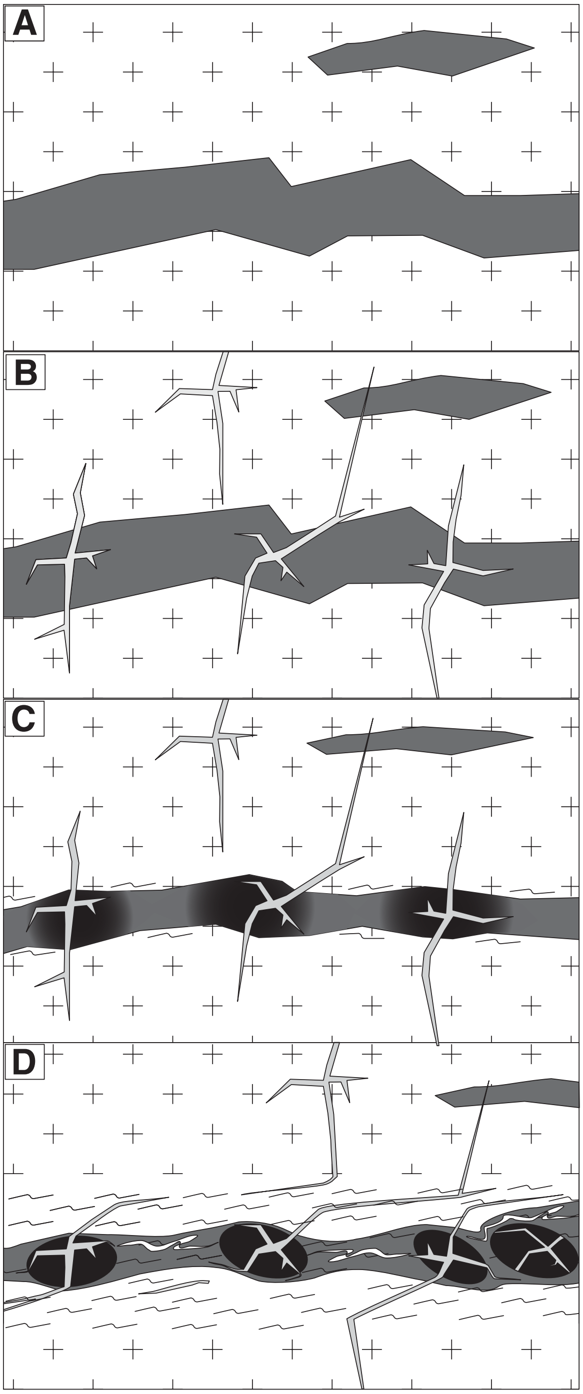

Figure 6. Sequence of events leading to the observed field and fabric relations exposed at the Myrland eclogite locality. (A) Pristine monzonite (+ pattern) with basalt enclaves (dark gray). (B) Injection of prekinematic granitic veins (light gray). (C) Localized fluid-mitigated eclogitization (nebulus blackish areas) of basalt and initiation of crystal-plastic flow (short squiggly lines). (D) Amphibolite-facies necking (dark gray is amphibolite) and dismemberment of eclogitized domains to form eclogite lenses (black ovals) that are rolled, folded, and imbricated. Short squiggly lines mark the Myrland shear zone. Contorted pegmatite veins within the shear zone are indicated by white wormy features.

moderately well-developed foliation defined by mica grains concentrated in layers. Muscovite is coarser and more abundant than brownish-green biotite. Quartz is recrystallized into millimeter thick bands. Feldspar occurs as newly crystallized tiny grains of both K-feldspar and plagioclase. Felty green patches of feldspar alteration products are common. These patches contain muscovite as well as both clinozoisite laths and to a minor degree epidote. Rutile, opaque minerals, zircon, and very rare xenotime constitute the accessory phases.

The zircon population of the sample is rather homogeneous and most grains are fractured but transparent. Size varies widely and color ranges from colorless to pink and purplish-pink. Almost all grains are prismatic, with differing aspect ratios, although a few multifaceted grains occur. No obvious rims were observed. Xenotime is very rare, but the grains present are euhedral and have a greyish pink color. Ten zircon fractions, which ranged from single grains to multigrain fractions and fragments, and one xenotime crystal were analyzed. In general the analyzed grains were free of inclusions, fractures, and turbidity (further descriptive details in Table 1).

Zircon and xenotime grains were separated and analyzed using isotope dilution thermal ionization mass spectrometry (ID-TIMS; Fig. 7) following Krogh (1973) at the Department of Geosciences at the University of Oslo; a description of the methods is in Appendix 1. The country rock has an age of ca. 1800 Ma (Corfu, 2004b). The sampled pod of melt associated with the eclogites contains a population of zircon grains, which all are discordant, variably between 1.5% and 22%. A discordia calculated for all ten analyses have an upper intercept at 1799 ± 11 Ma and a lower intercept at 485 ± 90 Ma (mean square of weighted deviates [MSWD] = 18) (Fig. 7B). Due to the scatter of $^{207}Pb/^{206}Pb$ ages, three analyses were discarded and another discordia was calculated with an upper intercept at 1800 ± 5 Ma and a lower intercept at 478 ± 41 Ma (MSWD = 4) (Fig. 7C). For this calculation analysis C was discarded due to a rather high common lead content (Table 1), and analysis B was discarded due to the low-U content, which may cause erratic behavior. Analysis J was discarded due to the older $^{207}Pb/^{206}Pb$ age, which may indicate xenocrystic zircon. The lower intercept of the resulting discordia is strongly guided by analysis F, which is the most discordant (22%). The analyzed fraction is a single colorless tip with average contents of U and Pb. However, in comparison with the other analyzed fractions, it has a conspicuously low Th/U. Varying Th/U ratios are often attributed to varying magmatic or metamorphic conditions during crystallization of different phases. Zircon formed under eclogite-facies conditions commonly has remarkably low Th/U ratios (e.g., Rubatto et al., 1999). The lower Th/U of analysis F may, therefore, indicate that a proportion of the analyzed zircon crystallized during eclogite-facies conditions. The age of 1800 ± 5 Ma is considered the best estimate of the crystallization age of the dyke.

The xenotime analysis (Fig. 7A) is very strongly reversely discordant due to anomalous behavior of the uranium. The cause of this is not known. This makes the data difficult to interpret and possibly the $^{207}Pb/^{206}Pb$ age is the most reliable estimate of the crystallization of the xenotime. Xenocrystic xenotime has been documented by Viskupic and Hodges (2001), but this was in low-temperature anatectic melts and it is not certain that these conditions are comparable to the situation studied here. Furthermore,

TABLE 1. U-Pb ISOTOPIC DATA

Fraction*	Weight (µg)	U (ppm)	Th/U†	Pb_{com}§ (pg)	$^{206}Pb/^{204}Pb$#	$^{207}Pb/^{235}U$**	2σ (abs)	$^{206}Pb/^{238}U$**	2σ (abs)	ρ	$^{207}Pb/^{206}Pb$**	2σ (Ma)	Disc.†† (%)
A—clabfr	7	763	0,29	5,2	18343	4,27916	0,02401	0,28690	0,00159	0,987	1768,9	1,6	9,1
B—lpabfr	32	45	0,30	8,1	3493	4,72100	0,01821	0,31409	0,00125	0,855	1783,0	3,8	1,4
C—sprclabfrS	8	365	0,24	52,5	915	3,73397	0,01699	0,25816	0,00103	0,858	1712,6	4,3	15,2
D—lpabfr	19	284	0,27	8,0	12460	4,43309	0,01079	0,29613	0,00065	0,959	1775,6	1,3	6,6
E—sclmfS	1	492	0,36	1,8	5063	4,48683	0,01494	0,29982	0,00096	0,922	1775,0	2,4	5,4
F—clfrS	1	461	0,15	3,5	1950	3,33813	0,01280	0,23363	0,00089	0,850	1690,1	3,8	22,1
G—stprclabS	4	491	0,23	2,1	16992	4,28141	0,01224	0,28741	0,00081	0,913	1766,6	2,2	8,8
H—stprclabS	6	2456	0,23	15,6	17079	4,28564	0,01298	0,28791	0,00085	0,943	1765,3	1,8	8,6
I—stprclabS	3	678	0,30	0,7	59527	4,63677	0,02044	0,30776	0,00133	0,984	1787,3	1,4	3,7
J—sprab	1	375	0,29	0,8	8094	4,19008	0,01638	0,28077	0,00110	0,932	1770,0	2,6	11,1

Note: All analyses were corrected for blanks of 2 pg Pb and 0.1 pg U.
*cl—colorless, ab—abraded, fr—fragment(s), l—large, p—pink, s—small, pr—prismatic, S—single, mf—multifaceted, st—stubby.
†Model value calculated using the $^{208}Pb/^{206}Pb$ ratio and the age of sample.
§Total common Pb (including initial common Pb of sample and analytical blank).
#Corrected for spike contribution and fractionation.
**Corrected for spike, fractionation, blank, and initial common lead.
††Degree of discordance.

xenotime has not been reported anywhere else in the Lofoten Complex (Corfu, 2004b; Markl et al., 1998a, 1998b; Rehnström, 2003). It is, therefore, unlikely that the analyzed xenotime is inherited.

The 1800 ± 5 Ma crystallization age for the syenogranitic vein associated with the eclogitized rocks at Myrland is consistent with the 1774 ± 6 Ma (Table 1) age on xenotime, and documents it among the suite of plutons of this age found throughout Lofoten (Griffin et al., 1978; Corfu, 2004b, 2007). Rehnström (2003) observed similar melts associated with gabbros of the Tielma Magmatic Complex, rootless slivers of the Lofoten AMCG complex thrust transported eastwards during the Caledonian orogeny, and interpreted them as palingenetic with intrusion of gabbro into mangerites.

The lower intercept age of 478 ± 41 Ma indicates a strong disturbance of the U-Pb system at this time, and considering isotopic age dates and geological relations from across Lofoten it most likely approximates the time of eclogitization. The granulite-facies event in Lofoten rocks is dated at 1.8 Ga (Corfu, 2004b; Corfu, 2007), but there is no evidence for 1.8–1.7 Ga eclogites anywhere in Scandinavia. Titanite from the Eidsfjord monzonite, which is correlative with the monzonite orthogneiss exposed at Myrland, has a U-Pb age of ≤478 ± 9 Ma (Corfu, 2004b), which overlaps our lower intercept date. Very precise 461–469 Ma U-Pb ages bracket the time of amphibolite-facies metamorphism of the Leknes Group (Figs. 2 and 3; Corfu, 2004a), which resulted in the retrogression of the eclogites (Kassos, 2008), and places a minimum for the age of eclogitization. Mager (2005) reports a precise $^{40}Ar/^{39}Ar$ muscovite plateau date of 457.9 ± 2.3 Ma from migmatite exposed in eastern Lofoten, indicating that some parts of this terrane escaped Scandian reheating above closure temperatures estimated at ~350 °C.

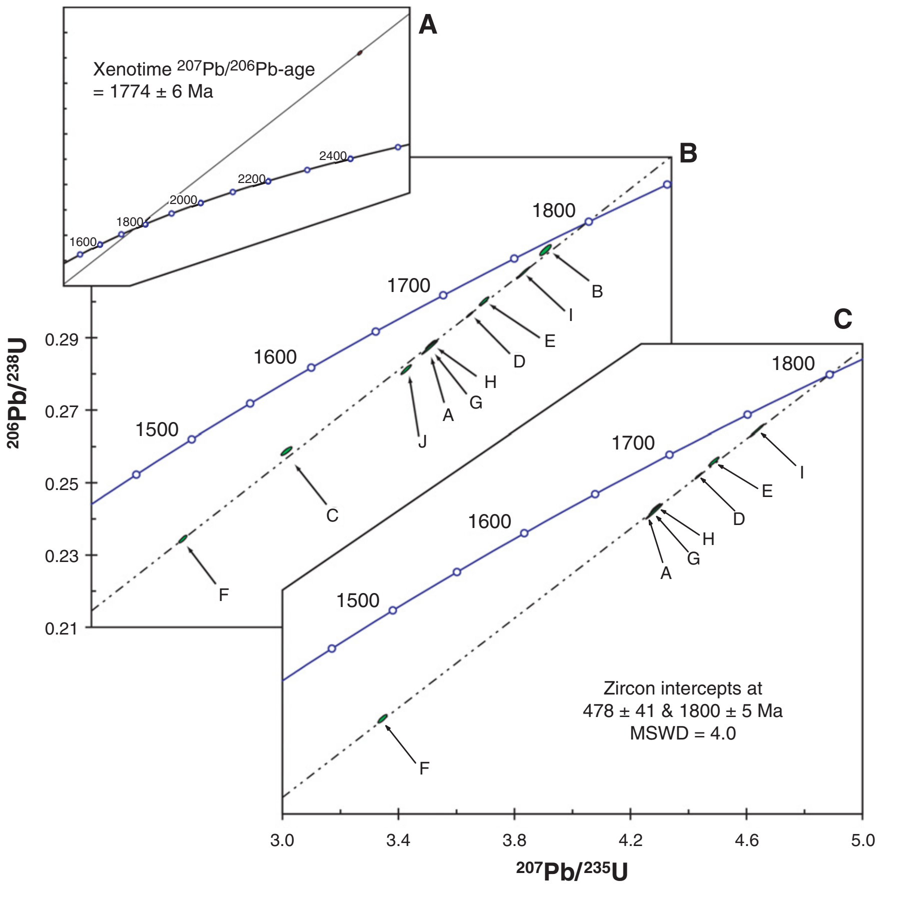

Figure 7. Upper discordia intercepts for xenotime and zircon grains from sample ML-24. (A) Xenotime. (B) All ten zircon analyses. (C) Rejecting zircon analyses B, C, and J; see text.

Steltenpohl et al. (2004) reported an $^{40}Ar/^{39}Ar$ age of 433 ± 9 Ma for hornblende, closure temperature ~500 °C, from the retrogressed eclogite at Nusfjord, which is compatible with cooling from the amphibolite-facies event estimated to have peaked at temperatures of ~780 °C (Mooney, 1997). U-Pb dating of zircons separated directly from one of the Flakstadøy eclogites would improve on narrowing the age range for eclogitization but we have been unsuccessful in recovering any. This might be explained by their general mafic character (gabbros and leuconorites) and/or if the eclogites had crystallized at too low of a temperature as implied by elemental-partitioning geothermometric estimates (lower-end range temperature is 680 °C; Markl and Bucher, 1997). Combined with the lack of accounts of any metamorphic event prior to ~478 Ma (Rehnstrøm and Corfu, 2004) from comparable structural levels or lithologic units in the entire region, inside and outside of Lofoten, it appears that the lower intercept age is the best estimate of the timing of eclogitization.

$^{40}Ar/^{39}Ar$ THERMOCHRONOLOGY

We report $^{40}Ar/^{39}Ar$ analyses for muscovite from four samples collected from each of the Flakstadøy eclogite localities (see Fig. 2 for localities). (Data tables for this study are available in Supplemental Table 1[1]; all age errors are reported as the standard deviation unless otherwise indicated. Methods are described in Appendix 2.) Muscovite grains from each sample were chosen for laser $^{40}Ar/^{39}Ar$ single-crystal total fusion (SCTF) analysis and incremental heating analysis in the Auburn Noble Isotope Mass Analytical Laboratory (ANIMAL) facility (see Supplemental Table 1 [see footnote 1]). Isotopic data were examined through isotope correlation and frequency-distribution diagrams to evaluate total-gas age distributions (Fig. 8). We chose to analyze single crystals of muscovite in view of the potential complexity for samples from this terrane, and also considering that many previous studies have demonstrated the sample preparation procedures for traditional incremental heating analysis of bulk muscovite samples (with many tens of crushed, cleaved crystals) can produce false plateaus that simply reflect homogenization of a sample (e.g., Hodges et al., 1994; Hames et al., 2008). The single-crystal approach of this study is analogous to the many routine $^{40}Ar/^{39}Ar$ single-crystal studies of feldspar phenocrysts from volcanic rocks, or various detrital minerals from sandstones, and our methods and resulting data may be compared directly with previous studies of muscovite in Lofoten (Hames and Andresen, 1996; Steltenpohl et al., 2004). Although dependent on cooling rate and mineral composition, most workers agree that the closure temperature range for muscovite is ~350–400 °C (see McDougall and Harrison, 1999), which is used in the interpretations and discussions below.

Sample BS5–14 is from a felsic gneiss in contact with an eclogite shear zone at the Skagen locality. Muscovite dates from this sample range from 427.0 ± 1.0 Ma to 358.0 ± 1.4 Ma (Fig. 8 and Supplemental Table 1 [see footnote 1]). The oldest muscovite age for this sample is near that determined for the timing of the peak of Scandian metamorphism recorded in the Caledonian nappes on the mainland (i.e., 432 Ma; Coker et al., 1993; Northrup, 1997; Steltenpohl et al., 2003a). One additional muscovite crystal chosen for laser incremental heating (Fig. 9) yielded a plateau age of 367.6 ± 1.8 Ma (2σ). Sample SKFD is from the postkinematic pegmatite dyke that cuts the eclogitic foliation at the Skagen locality (see Fig. 4E). Muscovite SCTF ages from this pegmatite are generally the youngest determined in this study, ranging from 364 to 300 Ma. Sample SR-25 is from a gneiss that encapsulates eclogite lenses at the Storvatnet locality. These analyses are the most tightly clustered with an age range from 355.6 ± 1.0 Ma to 331.3 ± 0.9 Ma and a mean of 343 Ma. Sample ML-24 is from a 1.8 Ga prekinematic felsic vein that is deformed and overprinted by a foliation mapped into the eclogite-facies lithologies at Myrland. The range of SCTF ages for this sample vastly exceeds other samples of this study, with ages as old as ca. 781 Ma and as young as 367 Ma.

In combination, the muscovite ages generally record isotopic closure through an age range between 427 and 300 Ma. The older end of this range is compatible with the 433 Ma hornblende cooling date from the retrogressed eclogite at Nusfjord (Steltenpohl et al., 2003b) and the timing of Scandian metamorphism. The Silurian–Permian ages of this study are consistent with the protracted, diachronous cooling through the muscovite closure temperature interval proposed for Lofoten, with notable timings of muscovite closure of ca. 360 Ma on Vestvågøy to 271 Ma for the outermost islands to the southwest (Hames and Andresen, 1996; Steltenpohl et al., 2004).

The youngest age determined for sample ML-24 is compatible with the general interpretation above. However, the older muscovite ages of this sample are more difficult to interpret. Many studies have documented extraneous argon and anomalously old $^{40}Ar/^{39}Ar$ ages for high-K white micas from ultrahigh-pressure metamorphic settings (e.g., Sherlock and Kelley, 2002). It is also known that muscovite may retain radiogenic argon during polymetamorphism and thus yields relict ages (e.g., Hames and Cheney, 1997). Thus the Neoproterozoic ages from ML-24 could reflect the presence of unresolved nonatmospheric, extraneous argon or they could conceivably be partially reset crystals with inherited ^{40}Ar. Additional isotopic and chemical study of the muscovite would be required to more fully evaluate these alternative explanations.

DISCUSSION

Exhumation of the Flakstadøy Eclogites

Figures 10A and 10B plot temperature versus time (*T-t*) and pressure/depth versus time (*P*/*z-t*) trajectories, respectively, for the Flakstadøy eclogites. Assuming that our ca. 478 Ma lower-intercept date approximates the time of eclogitization, the deep-crustal, anhydrous granulite-facies rocks hosting mafic enclaves and dikes were subducted during an early phase of Caledonian orogeny. Conditions for eclogitization of rocks at Flakstadøy are estimated at 680–780 °C and 11–15 kbar (Markl and Bucher, 1997). In Figure 10A we used the lower-end-range temperature estimate because temperatures this low might explain why zircons apparently did not crystallize during eclogitization. In Figure 10B we arbitrarily picked the higher-end pressure estimate and a density of 2800 kg/m^3 to calculate a depth of 51 km for eclogitization. Upper-amphibolite-facies metamorphism between 469 and 461 Ma (Corfu, 2004a) is estimated at 780 °C and 9.2 kbar (Mooney, 1997). Using our 478 Ma date as the approximate time for eclogitization, this implies that the eclogites were uplifted to depths of ~31 km over a period of ca. 14 m.y. (Fig. 10B). Markl and Bucher (1997) suggested that decompression of the eclogites from 15 to 11 kbar took place under isothermal conditions at ~700 °C. Kassos (2008) interpreted that monzodioritic and anorthositic sweat-out veins associated with eclogites at the Storvatnet locality formed due to this decompression event. Figure 10A illustrates the possibility that decompression was not strictly isothermal but could have been accompanied with rising temperatures, further aiding melting.

$^{40}Ar/^{39}Ar$ dating of hornblende from the retrogressed eclogite at Nusfjord indicates that the eclogites passed through ~500 °C at 433 Ma (Fig. 10; Steltenpohl et al., 2003b). Since plasticity in feldspar sets in at ~500 °C (Gapais, 1989), the 433 Ma date also places a lower constraint on formation of amphibolite-facies shear zones that encapsulate the eclogite lenses. Our estimate of pressure at the time of argon closure in hornblende, like that for muscovite in Figure 10B as well, is largely conjectural. We presumed that the 350 °C isotherm for closure in muscovite lay at the base of the brittle-ductile transition, ~15 km deep (4.4 kbar) for quartz-rich continental basement and a standard geothermal gradient; this may underestimate pressure given that our lithologies are more mafic than average continental crust (Moecher and Steltenpohl, 2009). We then took the mean of 4.4 kbar and 9.2 kbar, the latter being our quantitative estimate for the amphibolite-facies

[1]Supplemental Table 1. PDF file of the argon data table. If you are viewing the PDF of this paper or reading it offline, please visit http://dx.doi.org/10.1130/GES00573.S1 or the full-text article on www.gsapubs.org to view Supplemental Table 1.

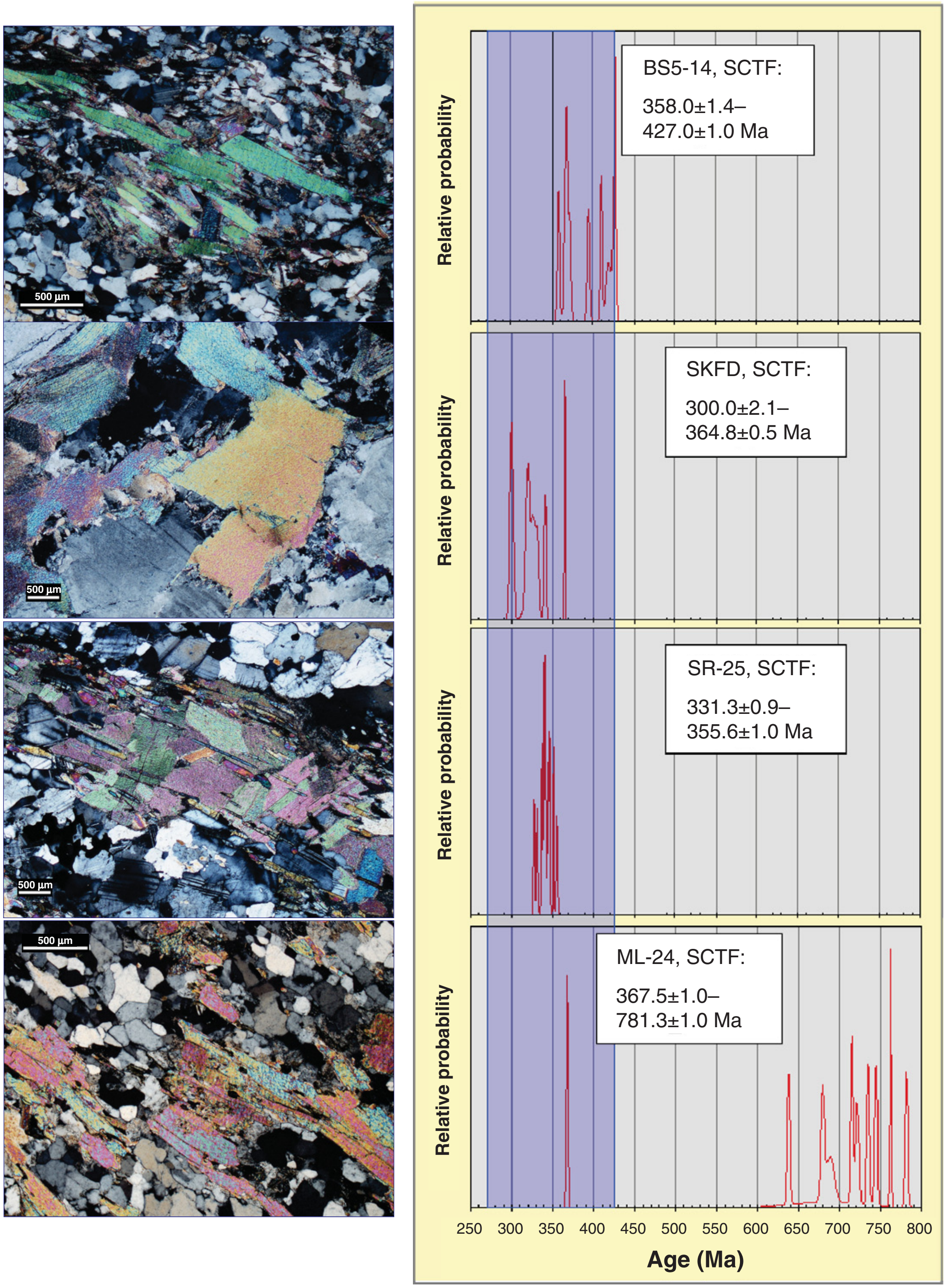

Figure 8. Laser $^{40}Ar/^{39}Ar$ data for muscovite from Flakstadøy eclogite localities. Photomicrographs (left) correspond directly to age probability distributions for four different rock samples with approximately ten single-crystal total fusion (SCTF) ages per sample (right). The maximum and minimum ages obtained for each sample are reported. See text.

peak, to get 6.7 kbar. Although our estimated path following the amphibolite-facies peak in Figure 10B is loosely constrained, it must flatten out with time. Thermal or isothermal decompression from 15 to 11 kbar, corresponding to roughly 11 km of continental crust associated with exhumation, accommodates a temperature drop of ~280 °C in 45 m.y. (mean cooling rate ~6 °C/m.y.). This is supported by the observations of omphacite breaking down via symplectite formation and also by formation of exsolution lamellae, which are reportedly indicative of rapid and slow exhumation, respectively (Markl and Bucher, 1997; Page et al., 2003, 2005; Anderson and Moecher, 2007; Kassos, 2008).

Our estimate of 343 Ma for the time that the Flakstadøy eclogites were elevated above the 350 °C isotherm (i.e., argon closure in muscovite) is based on the muscovite ages for sample SR-25 (356–331 Ma) and their mean (343 Ma), which is corroborated by the three other samples we analyzed and the overall trend of southwestward-decreasing ages throughout Lofoten

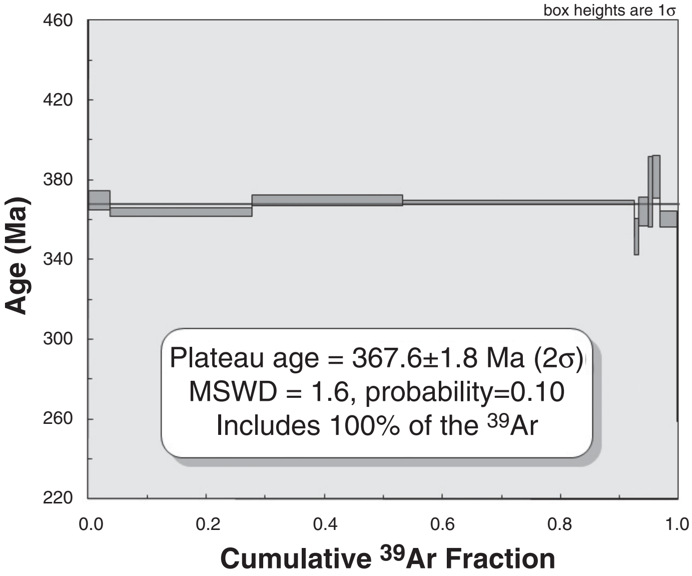

Figure 9. Laser incremental heating analysis of a muscovite grain from BS5–14. See text.

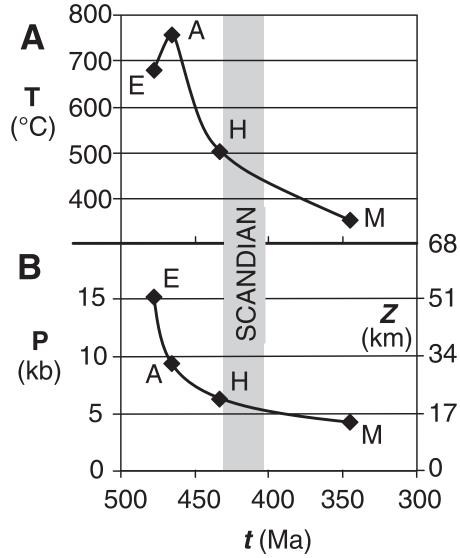

Figure 10. *T* versus *t* (A) and *P*/*z* versus *t* (B) paths estimated for Lofoten eclogites. E—eclogitization, A—amphibolite-facies metamorphism, H—hornblende cooling date, M—muscovite cooling average. See text.

(Hames and Andresen, 1996; Steltenpohl et al., 2004). The Flakstadøy eclogites, therefore, either resided in the middle- to lower-upper crust for ca. 121 m.y. (following the 464 Ma amphibolite-facies metamorphism of the Leknes Group; Corfu, 2004a; Key et al., 2008) or were completely exhumed and then reheated during the Scandian. In either case, Lofoten remained at depths below the 350 °C isotherm far outlasting the Devonian extensional event that exhumed eclogites of the WGR, which also explains the relatively higher degree of retrogression found within the Flakastadøy eclogites.

Scandian versus Mid-Ordovician Effects in Lofoten

The paucity of evidence for Scandian metamorphism and deformation in Lofoten is a long-standing issue (Tull, 1977; Griffin et al., 1978; Bartley, 1982, 1984; Hodges et al., 1982; Hames and Andresen, 1996; Steltenpohl et al., 2004, 2006), and our *T-t* or *P*/*z-t* plots for the Lofoten eclogites (Fig. 10) show no corresponding deflection through Scandian times. We hypothesize that this might be an artifact related to the eclogites having been physically distanced from structures substantial enough to promote fluid infiltration that could cause amphibolite-facies reequilibration during the Scandian. In other words, the eclogites might have been exhumed prior to 433 Ma (i.e., the $^{40}Ar/^{39}Ar$ hornblende date on the Nusfjord eclogite) and then overprinted during a separate Scandian amphibolite-facies peak. Such a scenario would require the Eocaledonian exhumation rate to have been higher than that suggested in our discussion above. Kullerud (1996) reported garnet-biotite thermobarometric estimates of >580 °C for the amphibolite-facies retrogression of the Nusfjord eclogite. This temperature is close enough to the ~500 °C closure temperature for hornblende that it could, hypothetically, relate to cooling from an early stage of the Scandian event (Fig. 10). Muscovite dates for sample BS5–14 reported herein, and ages reported for cores of muscovite porphyroblasts by Hames and Andresen (1996), also are consistent with Scandian metamorphic reequilibration at ca. 427 Ma.

Additional information pertaining to Eocaledonian versus Scandian thermal effects in rocks of Lofoten may be gleaned from U-Pb dates reported by Corfu (2004a, 2004b, 2007). Most of Corfu's U-Pb work was aimed at clarifying Lofoten's Archaean and Proterozoic history, but nearly a dozen dates, mostly lower intercepts, record Paleozoic thermal disturbances. Figure 11 plots these U-Pb dates against distance from

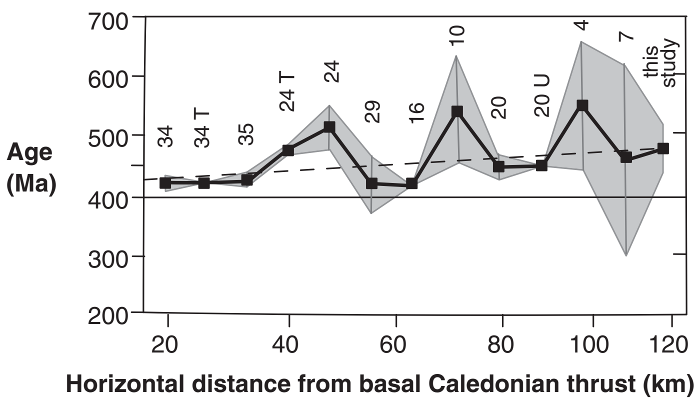

Figure 11. U-Pb age versus distance from the basal Caledonian thrust contact on the mainland east of Lofoten. Sample numbers, dates, and error estimates correspond to those reported in Corfu (2004a, 2004b, 2007) and in this study. All are lower intercept ages except for T (titanite) and U ("uranium-rich mineral"). See text.

the basal Caledonian thrust contact on the mainland. Although there are considerable errors in some of the lower intercept ages, a correlation of ages increasing from the Scandian peak, which is precisely established in the nappe complex to the east at 432 Ma (Coker et al., 1993; Northrup, 1997; Steltenpohl et al., 2003a), westward with horizontal distance from where the basal Caledonian thrust occurs. Note that the lower intercept error ranges appear to generally decrease eastward toward the basal thrust; the smallest errors are those for analyses with the lowest discordance. The dashed line in the plot was drawn from our date for eclogitization from Flakstadøy to the Scandian peak at 432 Ma directly at the basal thrust contact. It is important to note that the basal thrust in this area is syn-peak-metamorphic and is marked in the basement by a structurally downward decreasing mylonitic strain gradient that disappears ~500 m beneath the basal thrust nappe (Tull, 1977; Bartley, 1982; Hodges et al., 1982). We interpret these observations to indicate more complete resetting of the U-Pb isotopic systems of the minerals dated as the basal thrust is approached. Disregarding the lower intercept ages that have the largest errors (e.g., 24, 29, 10, 4, and 7) would tend to lower the slope of the dashed line in Figure 11, pushing our date for eclogitization toward younger ages in our error range; this would require even faster exhumation rates to reach the crustal levels of the amphibolite-facies peak at 464 Ma.

Middle Ordovician Eclogites

Based on our interpretation of the geological and U-Pb results, Middle Ordovician eclogitization in Lofoten occurred substantially earlier (ca. 50 m.y. at 2σ) than eclogitization of the WGR, placing the former within the early temporal group of Caledonian eclogites. If Lofoten is autochthonous Baltic basement, however, our Middle Ordovician date for eclogitization conflicts with the previously established kinematic pattern for Eocaledonian eclogites being restricted to the allochthonous Caledonian terranes. The allochthoneity of the Lofoten terrane has been questioned mainly due to its paucity of Caledonian structures and fabrics (Tull, 1977; Griffin et al., 1978). Hakkinen (1977) suggested that Lofoten might be a microcontinental fragment that was sutured to Baltica during the Caledonian orogeny. Later workers argued, however, that Lofoten represents the western margin of Baltica that was subducted beneath Laurentia during the Caledonian (Bartley, 1982; Hodges et al., 1982; Tull et al., 1985; Steltenpohl et al., 2004). Lofoten is not a simple northward continuation of the WGR, however, as protolith ages for the former are significantly older (2.2–1.8 Ga versus 1.65 Ga). Its occurrence as a northward continuation of the Transscandinavian Igneous Belt (1.8–1.65 Ga; Patchett et al., 1987; Aberg, 1988; Gorbatschev and Bogdanova, 1993), and bounding Archaean rocks to the north, appears to firmly pin Lofoten to the Baltic craton. Finally, no suture has ever been documented in surface exposures to lie between Lofoten and the Baltic basement proper to the east, although one cannot be ruled out hidden beneath thrust nappes in the intervening area. To date, therefore, no evidence is reported to indicate either UHP or Scandian eclogite-facies metamorphism in Baltic basement along coastal Norway north of latitude 64° N. Below we explore potential relationships between eclogitization in Lofoten and the other early group eclogite terranes.

Lying ~300 km northeast of Lofoten at the extreme uppermost preserved levels of the nappe stack are the Middle Ordovician eclogites of the Tromsø Nappe Complex (Fig. 1; Corfu et al., 2003). Mixed miogeoclinal lithologies, such as marbles, mica schists, and paragneisses, that characterize the Tromsø Nappe (Binns, 1978; Andresen et al., 1985; Bergh and Andresen, 1985; Krogh et al., 1990; Coker-Dewey et al., 2000) contrast sharply with the plutonic continental basement and/or rift-cover protoliths of all the other Caledonian eclogite terranes. Steltenpohl et al. (2003b) used similarities in lithologies, tectonic evolution, and timing to hypothesize correlation of the eclogitization event in rocks of the Tromsø Nappe with Middle Ordovician Taconic eclogites exposed in the Appalachian orogen of the southeastern United States (Adams et al., 1996; Stewart and Miller, 2001; Miller et al., 2010). Since workers argue that parts of the Uppermost Allochthon contain Laurentian relics (Yoshinobu et al., 2002; Roberts, 2003; Barnes et al., 2007), Steltenpohl et al. (2003b) further hypothesized that the Tromsø Nappe might be a block of Laurentian "Taconic" crust beached during the Scandian and then left orphaned in Scandinavia when the North Atlantic opened. Correlating the Lofoten eclogites to those of the Uppermost Allochthon, therefore, would require that the former is exotic with respect to Baltica. We are not aware of any provenance studies that would preclude either that the Lofoten basement was sourced from Laurentia, or that Lofoten could have served as basement for deposition of the miogeoclinal lithologies of the Tromsø Nappe. In several areas to the northeast and southeast of Lofoten the Uppermost Allochthon, including the Tromsø Nappe, lies in direct tectonic contact (along extensional faults) with Precambrian plutonic basement of the western internal window (Forslund, 1988; Carter, 2000). The problem, of course, with correlating Lofoten eclogites with those of the Tromsø Nappe Complex is that most workers agree that the former is not an exotic terrane but rather is authochthonous Baltic basement.

Lofoten shares some distinctive characteristics with eclogites preserved in rocks of the Bergen arcs more than 1000 km to the south (Fig. 1). (1) These are the only terranes containing early Caledonian eclogites that formed within parautochthonous Baltic continental basement, and both occur in coarse-grained gabbros and anorthositic rocks (Boundy et al., 1992; Steltenpohl et al., 2003b, 2006). (2) Lofoten and the Bergen arcs are the only places known on Earth where shear zone eclogites and eclogite-facies pseudotachylites occur within continental orthogneisses (Wade, 1985; Austrheim, 1987; Austrheim and Boundy, 1994; Kullerud, 1992; Markl and Bucher, 1997; Steltenpohl et al., 2006; Kassos, 2008). (3) Pressures of eclogitization in these two areas are much less (Lofoten: ~14–15 kbar; Bergen arcs: >17 kbar) than the ≤40 kbar estimated for the minimum pressures of the Western Gneiss region (see Austrheim, 1987; Markl and Bucher, 1997). Whether or not the Bergen eclogites truly belong to the early group of eclogites was recently questioned on the basis of a 423 ± 4 Ma $^{206}Pb/^{238}U$ secondary ion mass spectrometry (SIMS) date for eclogite zircon rims reported by Bingen et al. (2004). The same authors noted, however, that their zircon age is significantly younger than, and consequently incompatible with, muscovite $^{40}Ar/^{39}Ar$ plateau ages of 433 ± 1 to 429 ± 1 Ma from Bergen Arcs eclogite samples reported by Boundy et al. (1996).

The early to middle Paleozoic eclogites of the Seve Nappe Complex lie just 275 km to the east of Lofoten in Sweden (Fig. 1). Neoproterozoic dolerite dikes that had intruded cover sediments deposited upon the extending Baltic margin were eclogitized prior to their Scandian thrust emplacement and translation toward the foreland (Figs. 12A and 12B). In the northern outcrop belt of the Seve Nappe in Norrbotten, eclogitization is generally thought to have predated the Middle Ordovician (i.e., latest Cambrian to Early Ordovician), and was attributed to "Finnmarkian" contraction and exhumation (Mørk et al., 1988; Dallmeyer and Gee, 1986). Seve eclogites to the south in Jämtland, however, are substantially younger, ranging into the Mid-Ordovician (Brueckner and van Roermund, 2001). Rehnström (2003) recently recognized that slivers of the Lofoten AMCG suite are now situated at the top of the Middle Allochthon in the foreland in northern Sweden (i.e., the Tielma Magmatic Complex). These fault slices occur directly beneath the Seve Nappe

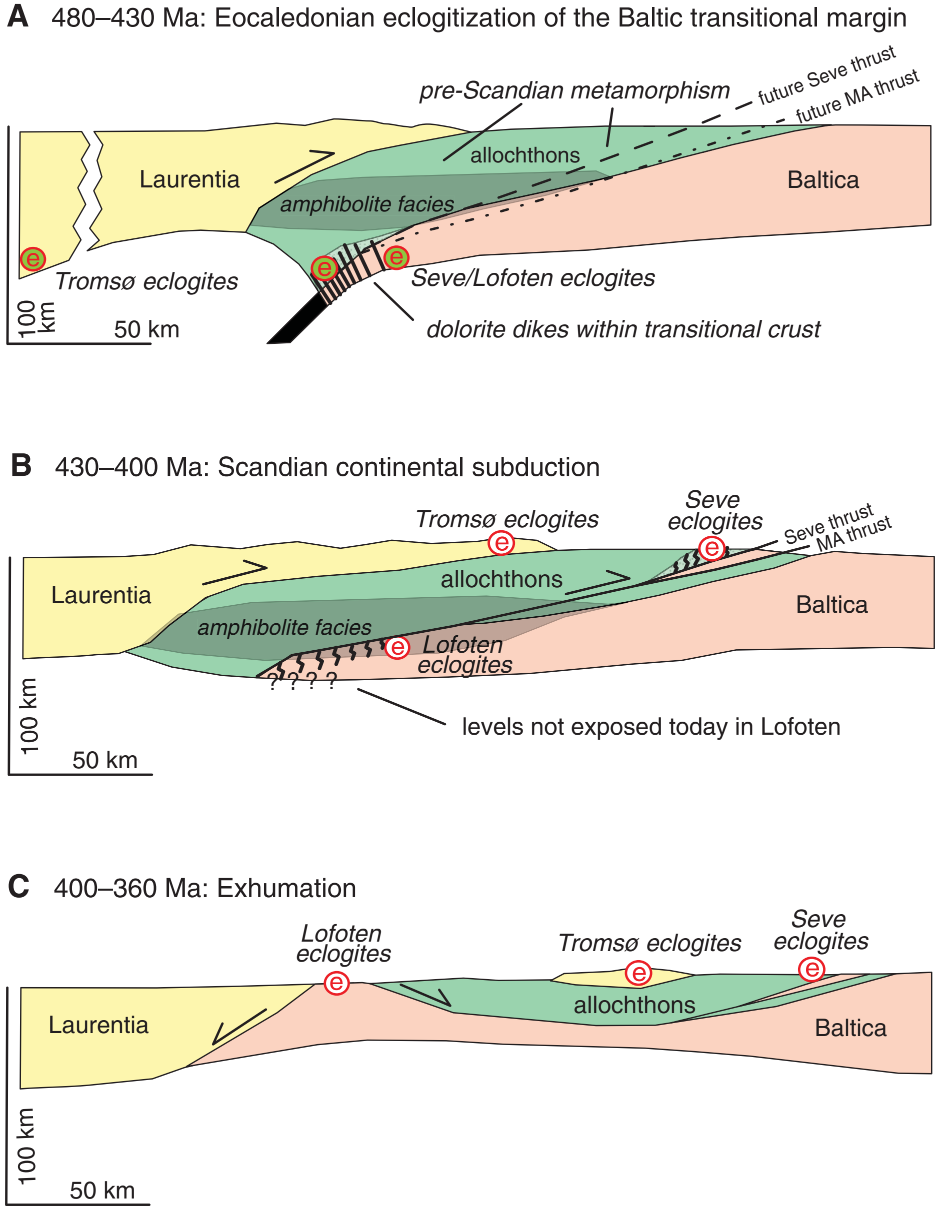

Figure 12. Cartoon cross sections illustrating the authors' interpretation of tectonic evolution of Eocaledonian eclogite terranes in northern Scandinavia; see text. Green eclogite symbols indicate areas of active eclogitization, whereas white ones reflect their passing through lower-entropy retrograde conditions; shaded area at mid-crustal depths is approximate area of amphibolite-facies conditions. "Allochthons" represent juxtaposed arcs and fragments of microcontinents and continental margins. Metasiliciclastics of the Seve Nappe Complex are stippled and lighter green colored. (A) Ordovician–Silurian early collisional phase with initial eclogitization of the Baltic ocean-continent transitional margin. Tromsø eclogites are interpreted as having formed within the Taconic orogen and thus are broken from Caledonian Laurentia. (B) Silurian–Devonian (Scandian) continental subduction. Seve eclogites are plucked from an area outboard of Lofoten and transported in the hanging wall of the Seve thrust. Basal Middle Allochthon (MA) thrust further telescopes Baltic basement and the Seve Nappe while other Scandian thrusts modify the collisional zone, emplacing farthest-traveled nappes with Tromsø eclogites to the highest structural levels preserved in the orogen. Scandian amphibolite-facies retrogression of the Lofoten eclogites commences. (C) Devonian extension, focused along the western part of the transect area, results in exhumation of the Lofoten eclogites above ~350 °C ($^{40}Ar/^{39}Ar$ closure temperature for muscovite) to elevations comparable with those of the Seve and Tromsø eclogites.

Complex sole thrust, and they are interpreted to have been emplaced in Silurian times. The tectonostratigraphic position of these Lofoten-derived basement slivers directly beneath the eclogite-bearing part of the transitional Baltic/Iapetan Seve Nappe Complex, thus, is palinspastically compatible with their having been decapitated from the Lofoten Baltic margin, embedded in the Seve Nappe sole thrust, and transported to their present-day position in the foreland. U-Pb dates on detrital zircons from the Leknes group in Lofoten (see Fig. 2) record a population from 1.6 to 1.0 Ga, which is similar to what has been reported for parts of the Seve nappe (Wade, 1985; Gee et al., 1985; Corfu, 2004a); this Sveconorwegian–Transscandinavian age range also removes Lofoten even further from correlation with a postulated Laurentian-derived Tromsø Nappe Complex. Together with our lower intercept date for eclogitization in Lofoten, the large errors of which overlap the range of dates reported for the Seve eclogites, leads us to hypothesize that both eclogite terranes formed proximal to one another during the same or related event (Fig. 12A). Such a geometric and kinematic plan is also compatible with the stark age difference between Ordovician $^{40}Ar/^{39}Ar$ mineral cooling dates for the Seve eclogites (Dallmeyer and Gee, 1986), which require that they were exhumed (contractionally, in this case) and remained high throughout the Scandian event, while the Silurian through Early Carboniferous cooling dates from Lofoten document its tectonic burial far beyond Scandian times (Fig. 12B). Finally, Devonian through Carboniferous exhumation of the Flakstadøy eclogites, depicted in Figure 12C, is modeled after recent field and $^{40}Ar/^{39}Ar$ studies documenting synchronous gravity-driven, foreland- and hinterland-directed movement away from a thermal dome created beneath over-thickened Caledonian crust in Lofoten, without significant plate divergence (Key, 2010; Anderson et al., 2010; Steltenpohl et al., 2010).

A strong case, therefore, can be made for correlating the formation of the eclogites in Lofoten to either, or both, the Bergen Arcs and the Seve Nappe Complex. The former correlation is more speculative considering the great distance that today separates Lofoten from the Bergen Arcs. It seems counterintuitive that all of the tectonostratigraphic, kinematic, geometric, palinspastic, and, now, age similarities between eclogites in Lofoten and the Seve Nappe could be only coincidental. Paleomagnetic evidence may run contrary to our model because it implies that the continent Baltica was rotating very rapidly in the Late Cambrian–Early Ordovician and this would seem difficult with a well-developed subduction slab. Early eclogitization was probably thus followed by rather fast initial exhumation to facilitate the fast rotation of Baltica (Torsvik and Rehnström, 2001), which is compatible with the exhumation path we suggest for the Flakstadøy eclogites (Fig. 10) and possibly with early stages of strike-slip movement within the Myrland shear zone.

In closing, we suggest a broad twofold subdivision of the early Caledonian eclogite provinces into those that are (1) Baltic derived, and (2) exotic with respect to Baltica. Middle Ordovician eclogites falling under category (1) can further be subdivided into autochthonous (i.e., Lofoten) and, mostly, thrust translated groups.

CONCLUSIONS

(1) U/Pb analyses of zircon and xenotime from a pre–eclogite-facies syenogranitic dyke that cuts the mafic host to a retro-eclogite lens in Lofoten intruded at 1800 ± 5 Ma, and we interpret the strong disturbance to the U-Pb system recorded by the 478 ± 41 Ma lower intercept age as approximating the time of eclogitization during an early stage of the Caledonian orogeny.

(2) Our interpretation for eclogitization at ca. 478 ± 41 Ma is compatible with having preceded regional amphibolite-facies metamorphism at 461–469 Ma, which resulted in the retrogression of the eclogites.

(3) The exhumation path followed by the Lofoten eclogites is estimated as follows: eclogitization at ca. 478 Ma at 680–780 °C and 11–15 kbar corresponding to a maximum depth of 51 km; upper-amphibolite-facies retrometamorphism between 469 and 461 Ma at 780 °C and 9.2 kbar, implying uplift to depths of ~31 km over a period of ca. 14 m.y. under iso- or increasing thermal decompression; $^{40}Ar/^{39}Ar$ dates on hornblende records passing through ~500 °C at 433 Ma; ca. 427–343 Ma $^{40}Ar/^{39}Ar$ dates on muscovite record passage through the 350 °C isotherm at the base of the brittle-ductile transition, ~15 km deep. The Flakstadøy eclogites, therefore, either resided in the middle- to lower-upper crust for ca. 121 m.y. (following the 464 Ma amphibolite-facies metamorphism of the Leknes Group) or were completely exhumed and then reheated during the Scandian.

(4) *T-t* or *P/z-t* data for the Lofoten eclogites do not clearly document a separate Scandian metamorphic/deformational "peak," which we hypothesize to be an artifact of their physical distance away from structures substantial enough to promote fluid infiltration for amphibolite-facies reequilibration.

(5) Middle Ordovician eclogitization in Lofoten is a great deal older than Siluro–Devonian eclogitization of rocks of the WGR (older by roughly 50 m.y. at the 2σ level) but is similar to the timing of the group of Eocaledonian eclogites (ca. 505–450 Ma) found in the Tromsø, Bergen Arcs, and Seve allochthons.

(6) To date, no evidence is reported to indicate either UHP or Scandian eclogite-facies metamorphism in Baltic basement along coastal Norway north of latitude 64° N.

(7) Contrasting lithologies, and inferences based on what is presently known about their provenance, and vastly different tectonostratigraphic postions between the Lofoten and Tromsø Nappe Complex eclogites seem to preclude their correlation, but tend to further indicate that the latter is an exotic Laurentian ("Taconic"?) relic left orphaned on the conjugate side of the orogen.

(8) A strong case can be made for correlating eclogites in Lofoten to eclogites in the allochthonous, oceanic (Iapetan)-continental (Baltic) transitional rocks of the Seve Nappe Complex. Lofoten likely was the source area from which the Seve Nappe eclogites were derived during thrust telescoping of the Baltic margin, providing a link between autochthonous and allochthonous Eocaledonian eclogites.

(9) We suggest a broad twofold subdivision of the Eocaledonian eclogite provinces into those that are (1) Baltic derived, and (2) exotic with respect to Baltica. Middle Ordovician eclogites falling under category (1) can further be subdivided into those occurring in parautochthonous, with respect to Baltica, and thrust translated groups.

(10) Lofoten serves as a spectacular example of how continental crust in a collisional setting can be subducted to mantle depths, and exhumed, without leaving much evidence of it.

APPENDIX 1: U-Pb ISOTOPIC METHODS

Zircon and xenotime were analyzed using isotope dilution thermal ionization mass spectrometry (ID-TIMS) following Krogh (1973) at the Dept. of Geosciences at the University of Oslo. The sample was crushed, ground, and separated on a water-shaking table with a magnetic separator and heavy liquids. Mineral fractions were handpicked and most fractions air-abraded (Krogh, 1982). The grains were washed in dilute nitric acid and rinsed with water and acetone. The fractions were spiked with a mixed $^{205}Pb/^{235}U$ spike and then dissolved in Teflon bombs. Zircon fractions weighing over 3 µg and the xenotime fraction were subjected to chemical separation following the procedures outlined by Corfu and Noble (1992) and Corfu and Stone (1998). U and Pb isotopic ratios were measured on a Finnigan MAT 262 mass spectrometer in static mode using Faraday cups, and for small samples, in dynamic mode with an ion-counting secondary electron multiplier (SEM). Pb and U ratios were corrected for 0.1%/ a.m.u. fractionation with an additional bias correction for the SEM measurements. The correction for initial lead composition was made using compositions modeled by Stacey and Kramers (1975), and the resulting isotopic ratios were analyzed with the Isoplot program of Ludwig (1999).

APPENDIX 2: $^{40}Ar/^{39}Ar$ ANALYTICAL METHODS

Isotopic analyses were performed at the Auburn Noble Isotope Mass Analystical Laboratory (ANIMAL), Auburn, Alabama, under the supervision of Dr. Willis Hames. Samples containing muscovite were crushed, cleaned, and sieved. Standard picking techniques were used under a binocular microscope to pick uncontaminated grains. These grains were then irradiated at the McMaster University Research Reactor in Hamilton, Ontario. The standard used is the Fish Canyon sanidine monitor FC-2 (prepared at New Mexico Tech), with an assigned age of 28.02 Ma (Renne et al., 1998). Ten irradiated grains for each sample were placed in a copper holding disk and analyzed using single-crystal total fusion. The data were reduced using Microsoft Excel and Isoplot 3 (Ludwig, 2003). Mass discrimination and mass spectrometer sensitivity were measured by running air and blank samples every ten and five samples, respectively. Analyses comprise ten cycles of measurement over the range of masses and half-masses from m/e = 40 to m/e = 35.5, and baseline corrected values are extrapolated to the time of inlet, or averaged, depending upon signal evolution.

The ANIMAL facility is equipped with an ultra-high-vacuum, 90° sector, 10 cm radius spectrometer. The spectrometer employs second-order focusing (Cross, 1951), and is fitted with a high-sensitivity electron-impact source and a single ETP electron multiplier (with signal amplification through a standard preamplifier). Analyses are typically made using a filament current of 2.75 A, and potentials for the source and multiplier of 2000 V and –1300 V, respectively. Measurement of atmospheric argon passed through an air pipette monitored sensitivity and mass discrimination. The spectrometer sensitivity was 8.09 × 10–15 moles/V and measured $^{40}Ar/^{36}Ar$ ratios in air were 298 ± 2 during the course of this study.

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