Metamorphic conditions across the Seve-Köli Nappe boundary, southeastern Trondheim region, Norwegian Caledonides: Comparison of garnet-biotite thermometry and amphibole chemistry

Elizabeth A. McClellan

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Rocks of the lower Seve Nappe and overlying Köli Nappes are exposed by a domal structure in the Einunnfjellet-Savalen area, southeastern Trondheim region, south-central Norway. In a traverse across the dome, garnet-biotite thermometry from pelitic schists and amphibole compositions in coexisting mafic rocks indicate a normal metamorphic gradient from structurally higher to structurally lower rocks that proceeds uninterrupted across nappe boundaries, including the Seve-Köli nappe boundary. Above the garnet isograd, biotite-bearing pelites indicate lower greenschist facies. Below the garnet isograd, temperatures recorded in garnet-bearing pelites range from 480-515°C east of the dome, to a maximum 550-580°C in the core of the dome. The transition zone between the greenschist and lower amphibolite facies is indicated by mineral compositions of amphibole, plagioclase, and muscovite. The data presented are consistent with nappe stacking prior to or coincident with peak metamorphism, which is interpreted to have occurred during an interval broadly correlating with Scandian orogenesis (c. 435-400 Ma). If this interpretation is correct, there is little evidence from this particular area for the presence of an earlier Paleozoic, widespread regional metamorphic event. Comparison with well-studied Tethyan-type convergent margins may provide a realistic model for pre-Scandian metamorphism in this region.

E. A. McClellan, Department of Geology, University of Kansas, 120 Lindley Hall,1475 Jayhawk Blvd., Lawrence, Kansas, 66045, USA.

Introduction

The central Scandinavian Caledonides are a historic area for the study of thrust and nappe tectonics and metamorphic petrology. One of the classic works of the early days of metamorphic petrology, Goldschmidt's metamorphic map of the central Caledonides (1915, in Miyashiro 1973), portrays the remarkable symmetry of the Trondheim region in which a relatively high-grade thermal axis parallel to the trend of the orogen gives way to belts of decreasing metamorphic grade on either side. Despite the comparatively uncomplicated appearance of the orogen on this scale, detailed work over the past few decades has uncovered evidence for multiple phases of deformation involving varied strain histories, multiple metamorphic events, and inverted metamorphic gradients both between nappes and within single nappes (e.g., Roberts 1971, 1979; Andréasson & Gorbatchev 1980; Andréasson & Lagerblad 1980; Boyle 1987; Rice et al. 1989). Traditionally it has been accepted that final nappe stacking post-dated peak metamorphism and development of the regional foliation (Andréasson & Gorbatschev 1980; Andréasson & Lagerblad 1980; Roberts & Wolff 1981). This relationship is not resolved in every case, however, and while some nappe boundaries consistently juxtapose rocks of contrasting metamorphic grade, others have uncertain or ambiguous histories. As noted by Roberts & Sturt (1980), there is a growing body of evidence for episodic thrusting throughout development of the orogen.

Paleozoic orogenesis comprises several major events that affected the central Caledonides (e.g., Roberts 2003). During the Siluro-Devonian Scandian orogeny (Gee 1975), closing of the Iapetus ocean basin and collision between the Baltic and Laurentian continents produced intense deformation, metamorphism, and major nappe transport. Although overprinted by the pervasive Scandian deformation, earlier tectonothermal events in the central and southwestern Caledonides have long been recognized (Holtedahl 1920; Vogt 1945; Kvale 1960; Sturt & Thon 1976; Naterstad 1976; Guezou 1978), and may have involved emplacement of fragments of the Iapetus ocean basin (Furnes et al. 1980) during Early Ordovician (e.g., the Trondheim disturbance of Holtedahl 1920) and Middle Ordovician time (Sturt & Ramsay 1994). In recent years, debate has centered around the paleogeography of formation of the oceanic fragments, and the timing of their accretion to Baltica (see discussions in Stephens et al. 1993 and Grenne et al. 1999). The traditional interpretation (Gale & Roberts 1974; Sturt 1984; Roberts et al. 1985;

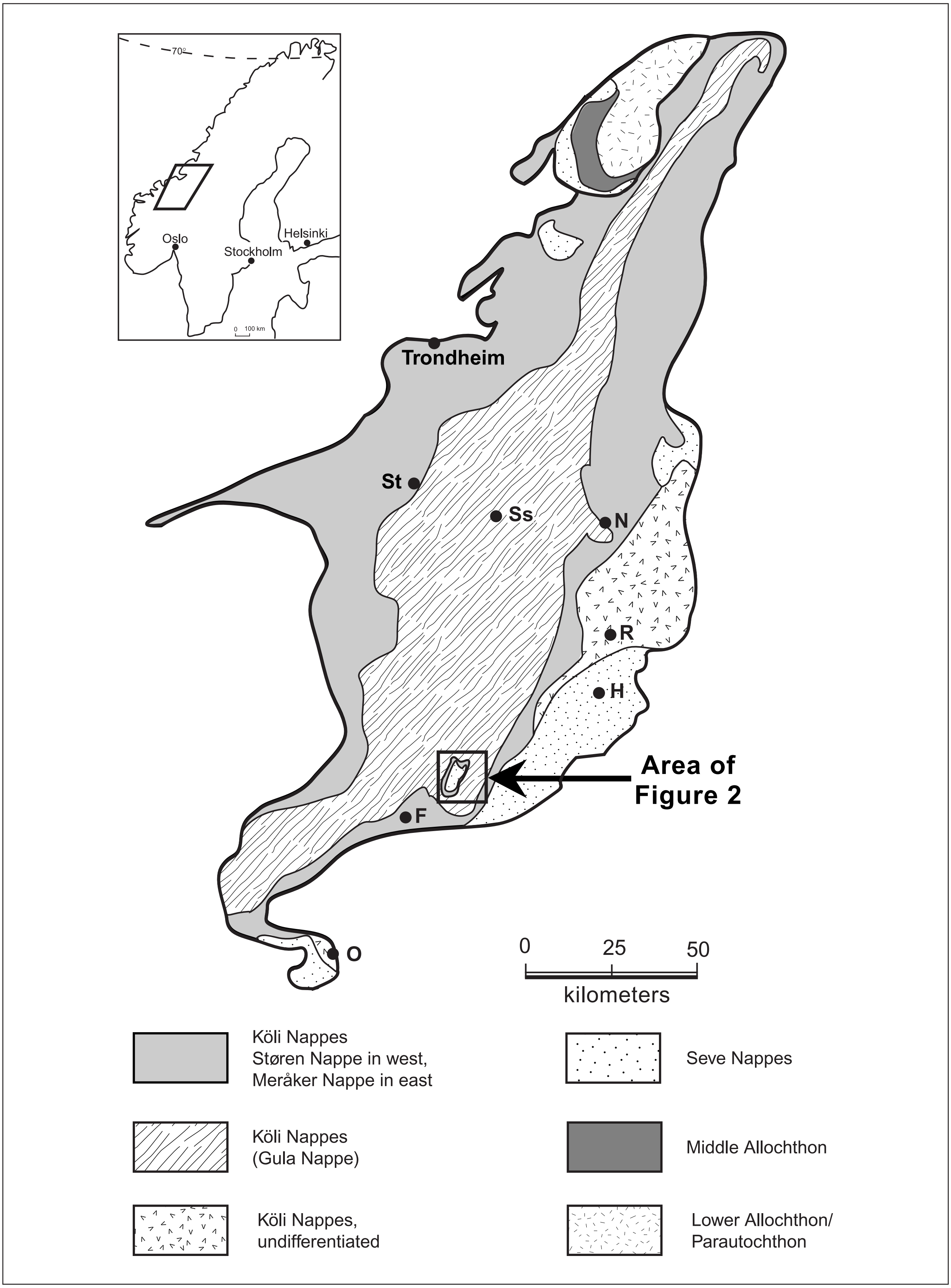

Fig. 1: Simplified tectonic map of the Upper Allochthon in the central and southern Norwegian Caledonides. Location of the study area (Figure 2) shown. Abbreviations refer to place names mentioned in text: F -Folldal; H -Hummelfjell; N – Nordaunevoll; O – Otta; R – Røros; Ss – Singsås; St – Støren.

Sturt & Roberts 1991) holds that all the oceanic terranes formed outboard of Baltica and were accreted to the Baltoscandian margin during the Ordovician. Major justification of this model involves the presence of a "belt-length, terrane-linking unconformity" that overstepped both Baltoscandian margin strata and obducted oceanic fragments, prior to Late Ordovician time (Sturt 1984; Sturt & Ramsay 1999). Concomitant with this interpretation, terrane accretion was accompanied by widespread deformation and metamorphism, and the unconformity therefore marks a significant break between rocks that experienced a complex tectonic history and those affected only by the Scandian event (Sturt & Ramsay 1999). An alternative view, based primarily on the presence of North American fauna in strata associated with some of the oceanic sequences (Neuman & Bruton 1974; Bruton & Bockelie 1980; Bruton & Harper 1981), holds that all (Pedersen et al. 1992), or many (Stephens 1988; Stephens & Gee 1989; Stephens et al. 1993) outboard terranes formed on the opposite side of Iapetus, near the Laurentian(?) margin, and were not accreted to Baltica until the time of the Siluro-Devonian Scandian orogeny. For example, isotopic and structural work from the Uppermost Allochthon, north-central and northern Norway, now clearly demonstrate that these rocks were deformed and metamorphosed during Taconian orogenesis near the Laurentian margin, prior to amalgamation with the Scandinavian Caledonides in the Scandian orogeny (Roberts et al. 2002a; Melezhik et al. 2002; Yoshinobu et al. 2002).

Both end-member interpretations described above are complicated by recent paleogeographic reconstructions showing Baltica in an inverted position during the Early Ordovician, facing Siberia across the Eastern Iapetus Ocean or Ægir Sea, and subsequently rotating counterclockwise rapidly to face Laurentia by Late Ordovician time (Torsvik 1998; Torsvik & Rehnström 2001). However, it is clear that pre-Scandian tectonothermal events in some terranes occurred as they were far removed from the Baltoscandian margin, and thus cannot be correlated with those affecting the margin or even, necessarily, with each other (Stephens et al. 1993).

In south-central Norway (Fig. 1), a domal structure exposes the contact between rocks representing Baltoscandian margin strata and the exotic terranes. This contribution examines metamorphic conditions, based on garnet-biotite thermometry in pelitic rocks and amphibole chemistry in coexisting mafic rocks, recorded in rocks from across the study area. Implications of these data are then explored for 1) the relative timing of metamorphism and nappe stacking; 2) the absolute timing of metamorphism; and 3) evidence for the existence or regional significance of pre-Scandian metamorphic events in this particular area.

Regional tectonostratigraphy of the Upper Allochthon

A majority of the exotic and suspect terranes in the Scandinavian Caledonides occur in the Upper Allochthon, a composite thrust sheet composed of heterogeneous lithologies with wide variations in strain and metamorphic grade. The Upper Allochthon is divided into two major complexes (Roberts & Gee 1985), the structurally lower Seve Nappes and the overlying Köli Nappes (Fig. 1). The Seve Nappes comprise several thrust-bounded units of schist, gneiss, and amphibolite, metamorphosed from epidote-amphibolite to granulite facies (Andréasson & Gorbatschev 1980), or locally eclogite facies (van Roermund 1985). The lower units are, in part, correlative with Late Neoproterozoic dolerite-intruded sandstone of the underlying Särv Nappe (Middle Allochthon); sheeted dike complexes in the Seve Nappe are similar to the ca. 650 Ma Ottfjäll Dolerite of the Särv Nappe (Claesson & Roddick 1983), interpreted to represent the rifted Baltic margin (Gee 1975; Solyom et al. 1979; Gee et al. 1985; Andréasson 1994). Precise U-Pb zircon dates on similar sheeted dikes intruding part of the Seve complex in northern Norway indicate ca. 610 Ma for rifting in that area (Svenningsen 2001). Thus, the Seve Nappes represent thrust sheets derived from the outermost Baltoscandian margin of Baltica, and contain the continent-ocean transition between Baltica and either Iapetus (Stephens 1988; Andréasson 1994) or the Ægir Sea. Conversely, the Köli Nappes contain extensive metavolcanic units representing a variety of oceanic tectonic environments, interpreted to have developed outboard of the Baltoscandian platform and miogeocline (Stephens & Gee 1985). Metamorphism in the Köli Nappes is commonly in greenschist facies.

The Seve Nappes were metamorphosed prior to emplacement onto the underlying rocks of the Middle Allochthon, and the contact consistently juxtaposes rocks of contrasting metamorphic grade (Andréasson & Gorbatschev 1980). Internal thrusts within the Seve Nappes are also interpreted to have occurred after peak metamorphism, based on discontinuities in metamorphic grade and retrogression along contacts (Stephens et al. 1985, and references therein). The Seve-Köli contact, on the other hand, is much more enigmatic. Although traditionally interpreted as a thrust (Williams & Zwart 1977; Gee et al. 1985), and in some areas marked by retrogressed mylonites (Sjöström 1983), in other areas no tectonic break is apparent (Stephens et al. 1985) and the contact appears concordant and transitional (Gee 1978). Lisle (1984) suggested that the contact is an unconformity, upon which only minor, local movement had occurred. Clearly, the nature of the Seve-Köli contact is of fundamental significance for interpretations of Caledonian tectonostratigraphy.

In the eastern Trondheim region, including the immediate study area, the Upper Allochthon includes metasedimentary and metavolcanic rocks of the Essandsjø, Meråker, and Gula Nappes (Nilsen & Wolff 1989), overlying and in fault contact with quartzite and schist of the Hummelfjell Nappe. The Hummelfjell Nappe represents Late Neoproterozoic sandstone and quartzite of the rifted outer Baltoscandian margin (Gee et al. 1985), and in the type area is characterized by abundant mafic dikes within quartzite (Rui 1972). Nilsen & Wolff (1989) assigned the nappe to the Middle Allochthon, whereas Gee et al. (1985) equated it with the Seve, considering it a higher grade equivalent of the Särv Nappe. The lithologic character and tectonic position is consistent with placement within the basal Seve Nappes, that represent "a deformed and metamorphosed variety of the next underlying Särv Nappe" (Andréasson 1996, p. A29).

The Essandsjø and Meråker Nappes are generally interpreted to include oceanic sequences developed outboard of the Baltoscandian margin (e.g., Grenne et al. 1999), and thus correlate with the Köli Nappes. The Gula Nappe is also accepted to be one of the Köli-equivalent nappes (Gee et al. 1985; Stephens & Gee 1985; Roberts & Stephens 2000), although its paleogeographic affinity is still unclear. Whereas the upper, lower grade portion (Åsli Fm. in the eastern Trondheim region) overlies volcanites of interpreted oceanic origin (e.g., Nilsen 1974; McClellan 1995a), the lower, higher grade part (Singsås Fm.) comprises a thick sequence of calcareous psammite and pelitic schist (Nilsen 1978), suggested to be of continental margin or shelf origin (Guezou 1978). Two different interpretations have been suggested for the latter, either as forming part of the Baltoscandian margin, or as a microcontinent off the margin of Baltica (Grenne et al. 1999; Roberts et al. 2002b; Roberts 2003). Several workers have proposed the existence of a major tectonic boundary within the Gula Nappe (Lagerblad 1984; McClellan 1993, 1995a; Bjerkgård & Bjørlykke 1994; Sturt et al. 1997). The exact placement of this boundary is still uncertain, as is its nature, which has been interpreted as either a major unconformity (Sturt et al. 1997) or a fault (e.g., Lagerblad 1984).

Age constraints applicable to the specific tectonostratigraphic units in the present study area are as yet limited. In the central Norwegian Caledonides, U-Pb zircon dating on trondhjemites and plagiogranites intimately associated with ophiolitic units in the Köli Nappes yield ages ranging from 495 ± 3 to 480 ± 4 Ma (see review in Roberts & Tucker 1998). From Folldal, just west of the area of this study, Bjerkgård & Bjørlykke (1994) reported a U-Pb zircon age of 488 ± 2 Ma from a trondhjemite dike intruding the Folldal metabasalts (=Meråker Nappe). To the east, the uppermost sedimentary rocks stratigraphically overlying mafic rocks in the Meråker Nappe contain Silurian (Llandovery) graptolites (Getz 1890). Graphitic phyllite at Nordaunevoll, north of Røros, (Fig. 1) that contains a Tremadoc graptolite (Vogt 1941) is commonly assigned to the Åsli Formation of the Gula Nappe (Rui 1972; Sturt et al. 1997). Minimum ages for rocks of the Gula Nappe are indicated by ca. 435-432 Ma ages of high-Al, continental-margin trondhjemites that cut Gula lithologies, the emplacement of which is interpreted to have have been broadly coeval with the earliest stages of Scandian plate convergence (Dunning & Grenne 2000; Nilsen et al. 2003).

Tectonostratigraphy of the Einunnfjellet-Savalen area

The Einunnfjellet-Savalen area, situated approximately halfway between Otta and Røros (Fig. 1), comprises a domal structure exposing several significant nappe contacts. The structurally lowest strata are exposed on Einunnfjellet mountain, and higher strata dip away from the core of the dome (Figs. 2 and 3). Gently to moderately dipping foliation and overturned folds surrounding the dome pass into a series of steep, NE-SW striking folds in the area of Savalen lake (McClellan, 1994a). The eastern boundary of the study area corresponds with the contact between metasedimentary rocks mapped as Gula Nappe (Åsli Fm.) and metavolcanic rocks of the Meråker Nappe (Nilsen & Wolff 1989). In this region, the contact is a steep shear zone characterized by pervasive greenschist-facies mylonitic fabrics and shear-related structures (McClellan 1993, 1995b). The tectonic significance of this shear zone (i.e. the extent of the zone and magnitude of displacement) has yet to be determined.

Earlier work in and around the area, summarized in McClellan (1994a), includes that of Kleine-Hering (1969), Berthomier et al. (1970), Lieungh (1970), Quenardel (1972), Mosson et al. (1972), Nilsen (1978), and Rui & Nilsen (1988). On the Røros 1:250,000 scale geologic map (Nilsen & Wolff 1989) the Einunnfjellet structure is depicted as a window cored by quartzite of the Hummelfjell Nappe, overlain by and in fault contact with various metasedimentary and meta-igneous rocks of the Essandsjø, Meråker, and Gula Nappes (Table 1). These authors indicate that mainly the upper part of the Gula Nappe (Åsli Fm.) is present in the immediate study area. The Gula-Meråker contact is shown as folded and overturned such that the Gula Nappe locally overlies the Meråker Nappe; however, Nilsen (1988, p. 61) stated that "some uncertainty" still exists as to this interpretation.

McClellan (1994a) interpreted the Einunnfjellet structure as a complex interference dome rather than a

Fig. 2: Geologic map of the Einunnfjellet-Savalen area, based, in part, on Nilsen & Wolff (1989) and, in part, on McClellan (1993, 1994a). Location shown in Fig. 1. Lithologic contacts on the two islands in Lake Savalen after Rui & Nilsen (1988). Cross section A-A' shown in Figure 3.

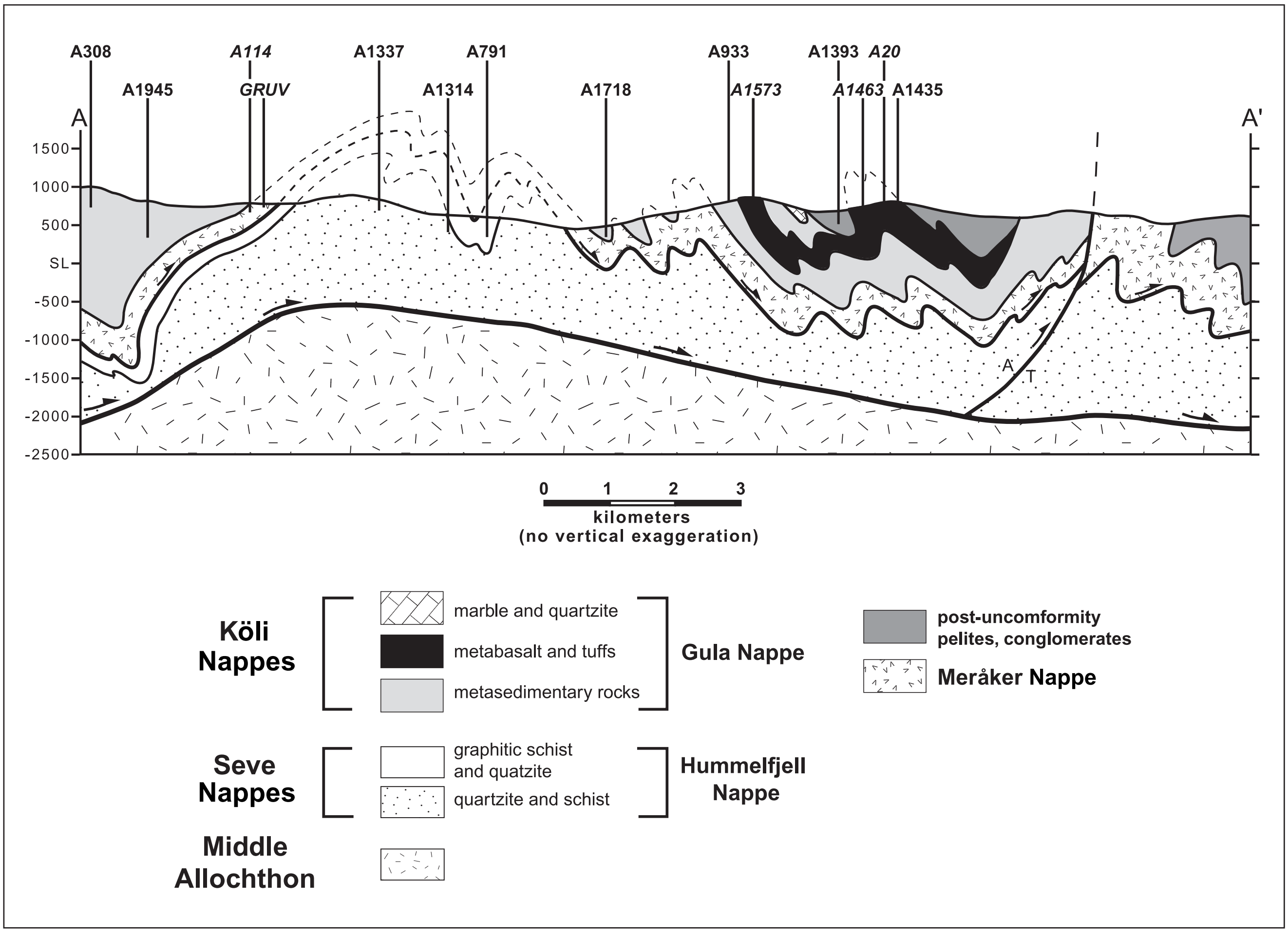

Fig 3: Cross section of the Einunnfjellet-Savalen area, across line A-A' in Figure 2. Locations of samples analyzed for this study are shown; samples off the cross section line are projected to their approximate locations. Sample numbers in plain text are pelitic rocks; those in italics are metabasalts.

simple window, and divided the area into three main tectonostratigraphic units (Fig. 2, Table 1) : 1) a lower continental-margin sequence of quartzite and schist in the core of Einunnfjellet dome, equivalent to the Hummelfjell Nappe of Nilsen & Wolff (1989); 2) a middle thrust sheet, the Savalen thrust complex, containing fragments of oceanic crust, along with pelagic, volcaniclastic and turbiditic sediments, and carbonate rock; and 3) an upper sequence of metaclastic rocks that unconformably overlies the other units. The major differences between this interpretation and the interpretation of Nilsen & Wolff (1989) are in merging of some units, previously assigned to separate nappes (Essandsjø, Meråker, Gula), into the Savalen thrust complex (STC), interpreted as an internally deformed sheet of oceanic crustal rocks and associated sediments; and in assignment of some higher portions of the nappes to an unconformably overlying sequence. Based on geochemical data, McClellan (1995a) interpreted mafic metavolcanic complexes within the STC as a related sequence of boninites and primitive tholeiites of arc/back-arc(?) origin, emplaced onto the continental margin succession (unit 1 above) as a composite thrust nappe. Subsequent geochemical work (McClellan 2002) and unpublished mapping by the author support the general interpretation, although this shows that the unconformable sequence is not as extensive as originally interpreted.

According to this interpretation, the Seve-Köli contact (i.e., the contact between the Hummelfjell Nappe and the Savalen thrust complex) is a tectonic boundary juxtaposing rocks of oceanic crustal origin with a continental-margin clastic sequence. However, in the study area, the contact between the STC and the underlying quartzites and schists is presently concordant and no fault-related fabrics appear to be specifically localized along it. Based on field observations, petrography, and mineral chemistry, McClellan (1993) suggested that there is an essentially uninterrupted increase in metamorphic grade, from lower greenschist to lower amphibolite facies, that crosses the Seve-Köli contact.

In an alternative interpretation Sturt et al. (1995, 1997) suggested major revisions of the tectonostratigraphy of the Upper Allochthon in the Otta-Røros tract, which includes the present study area. In the Otta area, the

Table 1. Regional stratigraphic correlations and nappe terminology in the Einunnfjellet-Savalen area. Regional unit and nappe assignments from Nilsen & Wolff (1989), unless otherwise noted. Köli/Seve nappe equivalents from Roberts & Wolff (1981), Stephens & Gee (1985), Gee et al. (1985).

Informal Unit Name (McClellan 1994; 1995)	Regional Unit Assignment	Regional Nappe Assignment	Köli/Seve Nappe Equivalent
Savalen upper sequence	Åsli Fm., *Gudå Fm. (Dalsbygd Gp. or **Sulåmo Gp. in east)	upper Gula Nappe (Meråker Nappe in east)	Köli
Langåsen quartzite/marble	Åsli Fm. or Gudå Fm.	upper Gula Nappe	Köli
Svartlia schist (in west); Volengsvola phyllite (in east)	Åsli Fm.	upper Gula Nappe (in part Essandsjø Nappe in west)	Köli
Lomsjødalen complex	Gula Greenstone	upper Gula Nappe	Köli
Bangardsvola igneous complex	Fundsjø Gp.	Meråker Nappe	Köli
Einunna mafic complex	________________	Essandsjø Nappe	Köli? Seve?
Moskardsætra graphitic schist	________________	Hummelfjell Nappe (?)	Seve
Strålsjøåsen calcareous schist	________________	Hummelfjell Nappe	Seve
Bjørnkletten quartzite	________________	Hummelfjell Nappe	Seve
Varden quartzite	________________	Hummelfjell Nappe	Seve

*Wolff 1967; **Grenne 1988

Vågåmo Ophiolite (Sturt et al. 1991) overlies a continental-margin sequence (Heidal Group) along a thrust fault. Both the ophiolite and the Heidal Group are overstepped by an unconformable sequence, the Sel Group (Sturt et al. 1991; Bøe et al. 1993). Minimum age is constrained by the Arenig-Llanvirn fauna in serpentinite conglomerate at the base of the Sel Group (Hedström 1930; Bruton & Harper 1981). Sturt et al. (1995, 1997) proposed that these relationships could be traced continuously northward from Otta to the Røros area (Fig. 1), with the unconformity appearing on both limbs of the recumbent Jønndalen syncline (Sturt & Ramsay 1997; 1999). The implications of this proposal are as follows: 1) The lower portion of the Gula (most of the Singsås Fm.) is correlative with the Heidal and Hummelfjell units, interpreted as representing the pre-Caledonian Baltoscandian continental margin. By implication, the Singsås, as well as the Heidal and Hummelfjell Formations, would be equivalent to the Seve Nappes (see contrasting viewpoint in Grenne et al. 1999, p. 434); 2) The upper part of the Gula Group in the Otta-Røros tract (Åsli Fm. and uppermost Singsås Fm.) correlates with the post-unconformity Sel Group, deposited above a 'terrane-linking' unconformity developed after ophiolite obduction onto the Baltoscandian margin; and 3) Interpretation of major structural inversion on the overturned limb of the Jønndal syncline, means that volcanites of the Meråker Nappe (Fundsjø Gp.) occur within the Sel Group, and represent volcanic eruptions into a sedimentary basin developed on continental crust.

The above interpretation invokes widespread regional orogenic deformation and moderate- to high-grade metamorphism as a result of Early Ordovician ophiolite obduction, followed by a weaker greenschist-facies overprint during the Scandian orogeny (Sturt & Ramsay 1999). In this interpretation, the Seve-Köli contact is "a major unconformity, which is also a significant metamorphic hiatus" (Sturt & Ramsay 1999, p. 83-84).

Conditions of metamorphism

Progressive regional metamorphism is most easily characterized by studying pelitic rocks, in which specific discontinuous reactions with rising temperature and pressure produce successive metamorphic zones based on the first appearance of particular minerals (e.g., Barrow 1893). Although mineral equilibria in mafic rocks have been used to characterize the general conditions of metamorphism from some of the earliest studies of metamorphism (Eskola 1920), the continuous variations in mineral composition and abundance with increasing temperature and pressure are still insuffici-

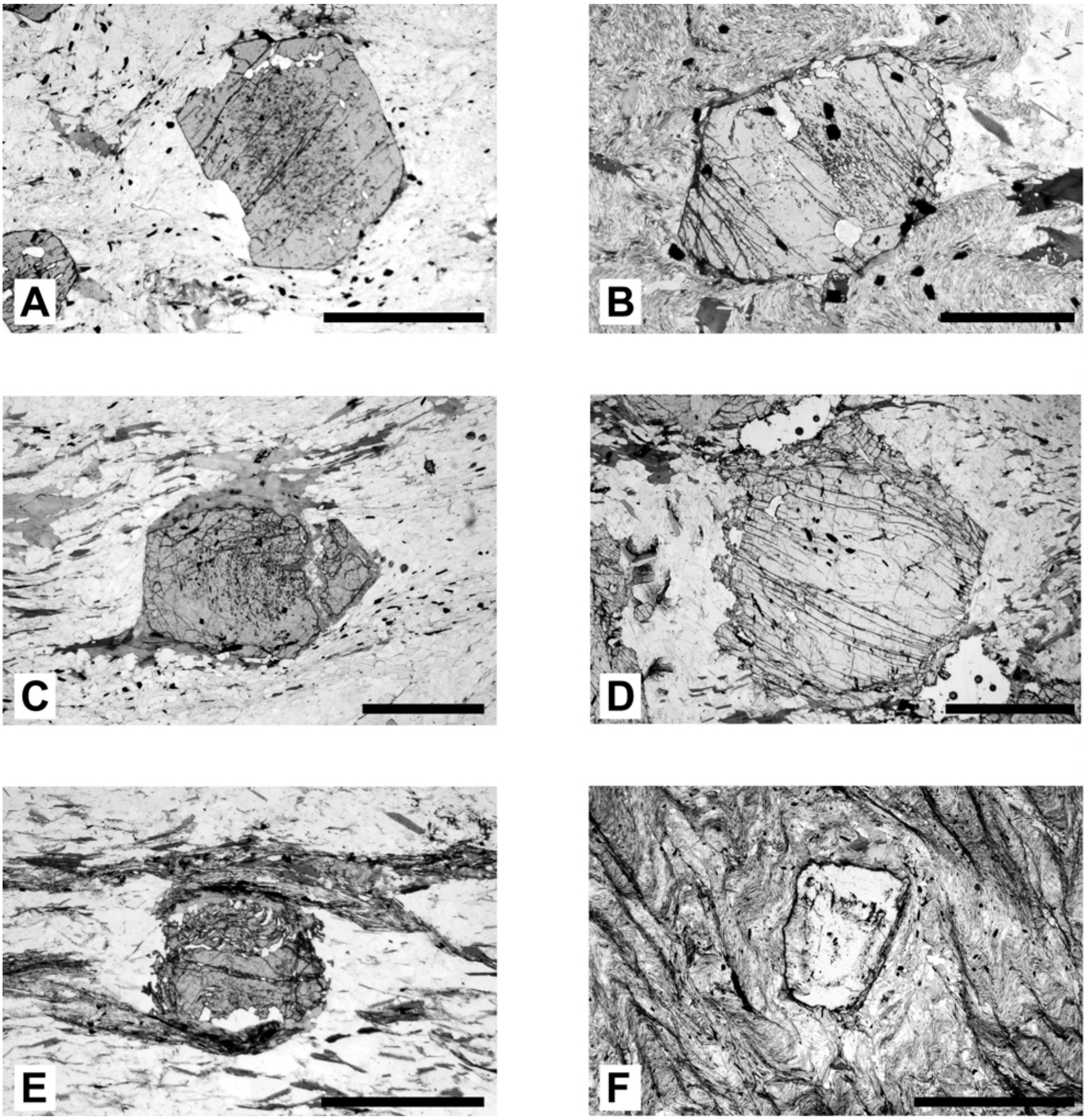

Fig. 4: Photomicrographs of garnet textures and inclusion patterns from the study area. Samples analyzed for this study are identified by sample number; location shown in Figure 3. Inclusions in garnet are dominantly quartz, unless otherwise noted. A: Core of garnet filled with randomly oriented inclusions, surrounded by inclusion-free rim. Sample A1341, Hummelfjell (Seve) Nappe. Scale bar = 2 mm. B: Portion of garnet core filled with random inclusions. Sample A1435, upper Gula (Köli) Nappe. Scale bar = 1 mm. C: Garnet with aligned inclusion pattern, at high angle to external foliation in matrix. Hummelfjell (Seve) Nappe. Scale bar = 1 mm. D: Nearly inclusion-free garnet core, surrounded by thin rim with few coarse-grained inclusions. Sample A1337, Hummelfjell (Seve) Nappe. Scale bar = 2 mm. E: Garnet containing coarse-grained inclusions. Curved inclusion trail is continuous with, although more open than flattened crenulations in matrix. Pressure shadows are developed adjacent to garnet. Upper Gula (Köli) Nappe. Scale bar = 1 mm. F: Garnet in graphitic phyllite. Relatively inclusion-free core is outlined by zone rich with graphite inclusions; outermost rim appears to have overgrown folded graphite-rich foliation. Upper Gula (Köli) Nappe. Scale bar = 2 mm.

ently correlated with specific P-T conditions. Significant strides in this area were made by Laird & Albee (1981), however, in a study of coexisting mafic and pelitic rocks from the Appalachians of Vermont, USA. Subsequently, their technique has been used successfully to investigate metamorphic conditions in mafic rocks from other regions (e.g., Fossen 1988). In this study, metamorphic conditions in the Einunnfjellet-Savalen area are deduced using both garnet-biotite thermometry in pelitic rocks and amphibole compositions in mafic rocks, and the results compared.

Analytical techniques

Analyses of garnet, biotite, amphibole, feldspar, muscovite, chlorite, and epidote were run on the Cameca SX-50 wavelength dispersive electron microprobe at the University of Tennessee, Knoxville, USA. A complete list of analyses is found in McClellan (1993); analyses used in this study are given below. All analyses were run at an accelerating voltage of 15 Kv. A beam current of 30 nA was used for garnet, chlorite, and epidote, and 20 nA for phases containing the volatile elements Na or K (biotite, muscovite, amphibole, feldspar). Beam diameter was dependent on grain size and elements analyzed; generally 5 µm for garnet, and 10 µm for feldspar and micas. Natural mineral standards included the range of compositions of the elements analyzed. Matrix corrections were made by the ZAF procedure.

Pelitic schists used for garnet-biotite thermometry

Nine pelitic samples from an east-to-west traverse across the study area (Fig. 3) were chosen for garnet-biotite thermometry. Typically, the samples are fine- to medium-grained phyllites or schists containing quartz + biotite + garnet + muscovite ± chlorite ± plagioclase ± graphite ± hornblende ± calcite. Epidote, tourmaline, and opaques are common accessory minerals. Staurolite or Al_2SiO_5 polymorphs are lacking in all rocks sampled for this study, although Ramsay & Sturt (1998) reported the occurrence of staurolite-bearing schists in the Hummelfjell Nappe ('Heidal Group' in their terminology) from near the core of Einunnfjellet dome. D. Ramsay (pers. comm.) also reported an occurrence of kyanite near the location of sample A933 described below, although it was not identified in this study despite extensive sampling in the schists of this area.

Mineral zoning and assumptions of equilibrium

A basic assumption of geothermometry is that the mineral pairs used in calculation of temperatures reached an equilibrium state (Spear & Peacock 1989; Bucher & Frey 1994). Petrographic observations of the pelitic schists used in this study indicate that the rocks experienced a complex history of mineral growth relative to deformation; however, evidence for significant chemical disequilibrium, other than the minor retrogression, is absent in most of the samples.

Textural zoning and mineral inclusions in the investigated garnets indicate that crystallization of garnet occurred both synkinematically and interkinematically, reflecting changing stress regimes during mineral growth (McClellan 1993, 1994b). Quartz is the most commonly included phase; opaque minerals (ilmenite, graphite) and epidote also occur in a few samples. Plagioclase and micas occur rarely as inclusions. The cores of some garnets contain fine-grained quartz inclusions that are either random (Fig. 4a, b) or oriented at a high angle to the external foliation (Fig. 4c), although clear, inclusion-free garnet cores (Fig. 4d) are present in some schists from the Hummelfjell (Seve) Nappe. Other garnets contain relatively coarse-grained inclusions, a rotational fabric of inclusions relative to the external foliation, and development of pressure shadows adjacent to garnet grains (Fig. 4e). The rotational fabric typically consists of curved or sigmoidal inclusion trails that may be oblique to or continuous with the crenulated matrix, or more rarely have the classic spiral or snowball texture. Clear, inclusion-free rims are common in garnets from the Einunnfjellet-Savalen area, and occur as either thin, continuous rims around garnet cores that contain fine-grained inclusions, or as discontinuous rims on cores having a rotational fabric. In a few samples, particularly in graphitic schists, garnet rims overgrow the crenulated foliation (Fig. 4f).

The investigated garnets generally display moderate to strong chemical zonation characterized by a systematic decrease in Mn and increase in Fe from cores to rims, coupled with a decrease in Fe/(Fe+Mg) (Fig. 5). Magnesium is low but slightly enriched toward the rims, whereas Ca is variable, although generally decreases toward the rims. Overall compositional variation largely appears to be independent of textural zonation. Only sample A1945 displays a nearly flat zoning profile. Bell-shaped profiles of Mn coupled with a core-to-rim decrease in Fe/(Fe+Mg) are typical of garnet growth with increasing temperature under greenschist- to amphibolite-facies conditions (Spear 1993), and a core-to-rim decrease in Mn and Ca is consistent with garnet growth during a single reaction (Tracy 1982). Diffusion and retrograde exchange of Mg and Fe between garnet and biotite, most common in high-grade rocks (Spear 1993; Bucher & Frey 1994), should not present a significant problem in this study.

Within individual samples, biotite compositions and Fe/Fe+Mg ratios are consistent between core and rim analyses, as well as between matrix grains and those adjacent to garnet, suggesting equilibrium between biotite and garnet (e.g., Whitney & Bozkurt 2002). Retrograde chlorite is most obvious in the graphite-bearing schists A1718 and A933, and some garnets in A933 are partially replaced by chlorite. Care was taken to avoid these areas for thermometry; however, potential deviation from equilibrium in this sample will be considered below.

Table 2. Electron microprobe analyses of garnet-biotite pairs used for thermometry

Sample:	A1435 pair 1	A1435 pair 2	A1435 pair 3	A1393 pair 1	A1393 pair 2	A933 pair 1	A933 pair 2	A1718 pair 1	A791 pair 1	A791 pair 2	A1341 pair 1	A1341 pair 2	A1337 pair 1	A1337 pair 2	A1945 pair 1	A1945 pair 2	A308 pair 1	A308 pair 2
Garnet rim analyses																		
SiO_2	36.43	36.60	36.58	36.89	36.60	36.34	36.71	37.10	36.21	36.29	36.18	36.48	37.19	37.32	37.03	37.17	37.85	37.52
Al_2O_3	20.47	20.68	20.67	20.19	20.68	21.01	21.03	21.25	20.85	20.69	20.87	20.84	21.17	21.21	21.13	21.13	21.19	21.22
MgO	1.16	1.16	1.21	1.41	1.39	1.69	1.90	2.82	2.34	2.69	2.46	2.53	2.55	2.48	3.01	2.89	2.73	2.88
CaO	7.57	7.99	7.52	5.89	6.40	6.87	6.44	5.46	2.29	1.65	3.11	2.95	6.71	7.04	6.44	6.71	6.98	6.67
MnO	1.91	1.76	1.79	2.92	2.91	1.71	0.90	0.63	0.35	0.27	0.01	0.06	1.66	1.78	2.65	2.31	2.39	2.31
FeO	30.90	31.13	31.58	30.97	30.61	30.92	32.37	32.06	36.68	36.87	36.30	36.12	30.64	29.78	28.74	28.57	29.22	29.47
Total	98.57	99.40	99.43	98.36	98.71	98.61	99.43	99.40	98.77	98.48	99.01	99.03	99.98	99.61	99.16	98.89	100.43	100.09
Cations on 12 O basis																		
Si	2.98	2.97	2.98	2.99	2.99	2.96	2.98	2.98	2.98	2.98	2.96	2.98	2.98	2.99	2.98	2.99	3.01	2.99
Al	1.98	1.98	1.98	2.00	1.99	2.02	2.01	2.01	2.02	2.00	2.01	2.00	2.00	2.00	2.00	2.00	1.98	1.99
Mg	0.14	0.14	0.15	0.17	0.17	0.21	0.23	0.34	0.29	0.33	0.30	0.31	0.30	0.30	0.36	0.35	0.32	0.34
Ca	0.67	0.70	0.65	0.51	0.56	0.60	0.56	0.47	0.20	0.15	0.27	0.26	0.58	0.60	0.56	0.58	0.59	0.57
Mn	0.13	0.12	0.12	0.20	0.20	0.12	0.06	0.04	0.02	0.02	<0.01	<0.01	0.11	0.12	0.18	0.16	0.16	0.16
Fe	2.12	2.12	2.15	2.10	2.09	2.11	2.19	2.15	2.52	2.53	2.48	2.47	2.05	1.99	1.93	1.92	1.94	1.96
X_{alm}	0.69	0.69	0.70	0.70	0.69	0.69	0.72	0.72	0.83	0.83	0.81	0.81	0.67	0.66	0.64	0.64	0.64	0.65
X_{pyr}	0.05	0.05	0.05	0.06	0.06	0.07	0.08	0.11	0.10	0.11	0.10	0.10	0.10	0.10	0.12	0.12	0.11	0.11
X_{gro}	0.22	0.23	0.21	0.17	0.19	0.20	0.18	0.16	0.07	0.05	0.09	0.09	0.19	0.20	0.18	0.19	0.20	0.19
X_{sps}	0.04	0.04	0.04	0.07	0.07	0.04	0.02	0.01	0.01	0.01	0.00	0.00	0.04	0.04	0.06	0.05	0.05	0.05
Fe/Fe+Mg	0.94	0.94	0.93	0.92	0.93	0.91	0.90	0.86	0.90	0.88	0.89	0.89	0.87	0.87	0.84	0.85	0.86	0.85
Biotite rim analyses																		
SiO_2	35.41	34.65	35.06	35.33	34.21	35.95	35.5	35.62	35.11	35.04	35.70	35.36	36.70	36.32	36.80	36.38	37.77	37.56
TiO_2	1.70	1.82	2.01	1.67	1.76	1.65	1.74	1.24	1.38	1.65	2.33	1.37	1.42	1.35	1.77	1.75	1.70	1.61
Al_2O_3	17.56	16.99	17.33	18.26	17.22	18.23	17.5	18.19	18.33	18.18	18.03	17.66	17.84	17.57	17.86	18.18	17.74	17.72
FeO	21.03	22.12	21.52	20.58	20.28	19.27	19.45	17.70	20.57	20.41	19.03	20.09	17.46	17.76	15.39	15.02	16.50	16.54
MnO	0.01	0.02	0.04	0.07	0.09	0.10	0.02	0.00	0.03	0.01	0.00	0.00	0.15	0.15	0.02	0.04	0.06	0.08
MgO	8.41	8.42	8.51	8.62	8.11	9.74	9.59	11.37	8.90	8.88	9.65	9.97	11.36	11.59	12.85	12.51	12.27	12.33
Na_2O	0.18	0.11	0.15	0.18	0.18	0.15	0.17	0.16	0.21	0.23	0.25	0.26	0.14	0.15	0.20	0.17	0.16	0.22
K_2O	8.92	8.74	8.90	8.56	8.26	8.86	7.72	8.59	8.64	8.47	9.48	8.66	9.22	8.83	9.10	9.20	8.70	8.59
Total	93.22	92.87	93.52	93.27	90.11	93.95	91.69	92.87	93.17	92.87	94.47	93.37	94.29	93.72	93.99	93.25	94.9	94.65
Cations on 22 O basis																		
Si	5.55	5.49	5.50	5.51	5.53	5.53	5.57	5.49	5.48	5.48	5.48	5.50	5.58	5.56	5.55	5.53	5.64	5.63
Al^{IV}	2.45	2.51	2.50	2.49	2.47	2.47	2.43	2.51	2.52	2.52	2.52	2.50	2.42	2.44	2.45	2.47	2.36	2.37
Al^{VI}	0.79	0.67	0.70	0.86	0.81	0.83	0.81	0.80	0.86	0.83	0.74	0.74	0.78	0.73	0.73	0.78	0.76	0.76
Ti	0.20	0.22	0.24	0.20	0.21	0.19	0.21	0.14	0.16	0.19	0.27	0.16	0.16	0.16	0.20	0.20	0.19	0.18
Fe	2.75	2.93	2.82	2.68	2.74	2.48	2.55	2.28	2.69	2.67	2.44	2.61	2.22	2.27	1.94	1.91	2.06	2.07
Mn	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.00	0.00	0.00	0.02	0.02	0.00	0.00	0.01	0.01
Mg	1.96	1.99	1.99	2.00	1.95	2.23	2.24	2.61	2.07	2.07	2.21	2.31	2.57	2.64	2.89	2.83	2.73	2.75
Na	0.05	0.03	0.05	0.05	0.06	0.04	0.05	0.05	0.06	0.07	0.07	0.08	0.04	0.04	0.06	0.05	0.05	0.06
K	1.78	1.77	1.78	1.70	1.70	1.74	1.54	1.69	1.72	1.69	1.86	1.72	1.79	1.72	1.75	1.78	1.66	1.64
X_{ann}	0.48	0.50	0.49	0.47	0.48	0.43	0.44	0.39	0.46	0.46	0.43	0.45	0.39	0.39	0.34	0.33	0.36	0.36
X_{phl}	0.34	0.34	0.35	0.35	0.34	0.39	0.39	0.45	0.36	0.36	0.39	0.40	0.45	0.45	0.50	0.49	0.47	0.48
Fe/Fe+Mg	0.58	0.60	0.59	0.57	0.58	0.52	0.53	0.47	0.56	0.56	0.53	0.53	0.46	0.46	0.40	0.40	0.43	0.43

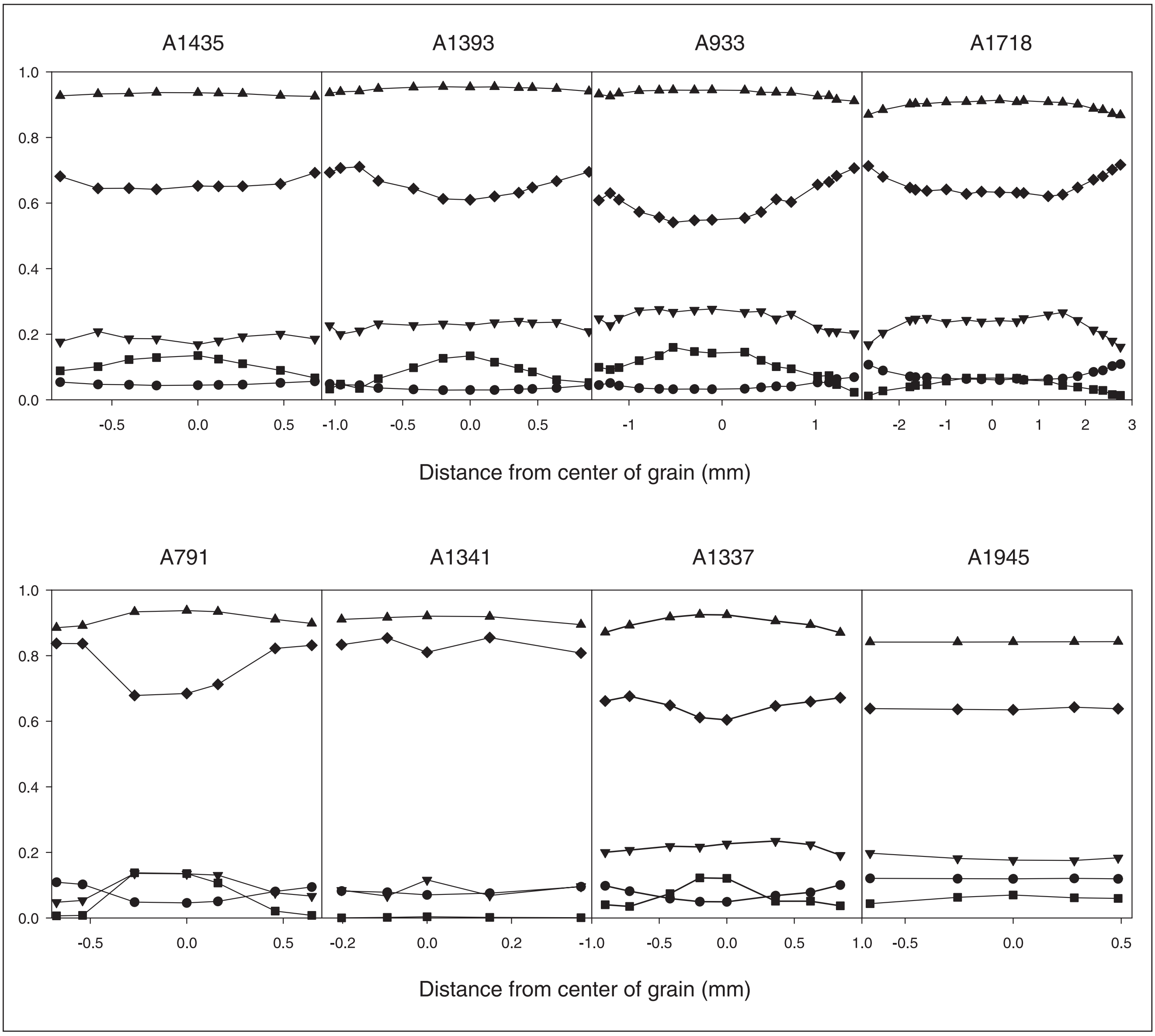

Fig. 5: Cation zoning profiles of garnets used for garnet-biotite thermometry. Components are plotted against distance from center of garnet grain. Ornaments: Almandine (Fe) = diamonds; spessartine (Mn) = squares; grossular (Ca) = inverted triangles; pyrope (Mg) = circles; Fe/Fe+Mg = upright triangles.

Results of garnet-biotite thermometry

The garnet-biotite Fe-Mg exchange thermometer is the most reliable and widely used geothermometer for medium-grade pelitic rocks (Essene 1982). Rim compositions of adjacent garnet and biotite grains (Table 2) were analyzed in the nine pelitic schists discussed above, and temperatures were calculated using the computer program GeoThermobarometry, v. 1.9 (Kohn & Spear 1995). Table 3 shows the results of these calculations, using a model pressure of 6 kbar, based on the calibrations of Ferry & Spear (1978), Perchuk & Lavrent'eva (1983), Hodges & Spear (1982), Ferry & Spear with the Berman (1990) mixing model, and Kleeman & Reinhart (1994).

The experimental studies of Ferry & Spear (1978) and Perchuk & Lavrent'eva (1983) incorporated a model of ideal Fe-Mg mixing in garnet and biotite. Ferry & Spear (1978) indicated that the geothermometer should be used with caution for natural systems in which significant amounts of Ca or Mn are present in garnet, or Ti and Al^{VI} in biotite. Hodges & Spear (1982) refined the Ferry & Spear (1978) calibration, correcting for non-ideal mixing of Ca in garnet. The Ferry & Spear-Berman and Kleeman & Reinhart (1994) calibrations both applied Berman's (1990) mixing model for quaternary Ca-Mg-Fe-Mn garnets. Kleeman & Reinhart (1994) also incorporated expressions for non-ideal Mg,Fe-Al and Mg,Fe-Ti mixing in biotite.

The application of several different calibrations to the samples for this study allows comparison between the

Table 3. Results of thermobarometry.
Temperatures calculated from garnet-biotite pairs, using calibrations discussed in text.

Sample	Unit	T (°C), calculated at 6 kbar P				
		*F & S	*P & L	*H & S	*F & S-B	*K & R
A1435	Gula Nappe					
pair 1		408	480	483	485	516
pair 2		418	487	497	502	523
pair 3		415	485	489	496	519
A1393	Gula Nappe (Gudå Fm.)					
pair 1		443	503	504	513	533
pair 2		453	510	520	527	541
A933	Gula Nappe					
pair 1		441	502	511	519	537
pair 2		464	516	530	541	549
A1718	Gula Nappe					
pair 1		502	540	560	571	570
A791	Hummelfjell Nappe					
pair 1		522	552	548	560	562
pair 2		563	572	582	590	578
A1341	Hummelfjell Nappe					
pair 1		495	536	529	542	547
pair 2		511	545	543	555	560
A1337	Strålsjøåsen					
pair 1		482	528	551	559	562
pair 2		482	528	555	561	563
A1945	Gula Nappe					
pair 1		478	525	545	549	554
pair 2		469	520	539	543	551
A308	Gula Nappe					
pair 1		479	526	550	555	558
pair 2		490	532	557	564	564

**Abbreviations of calibrations: F & S – Ferry & Spear (1978); P & L – Perchuk & Lavrent'eva (1981); H & S – Hodges & Spear (1982); F & S-B – Ferry & Spear (1978) with Berman (1990) mixing model; K & R – Kleeman & Reinhart (1994).*

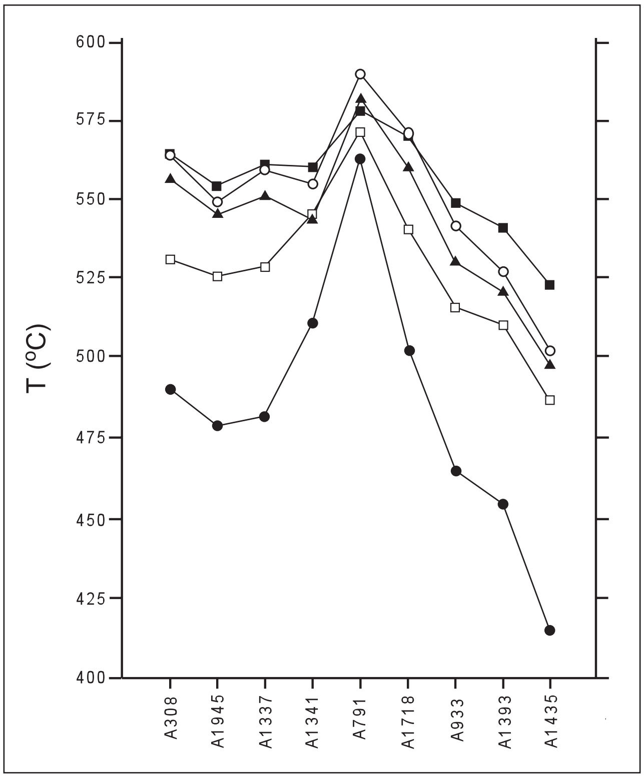

Fig.6: Plot of maximum temperatures calculated for each sample from the garnet-biotite calibrations discussed in text. Samples arranged according to position along the east-west analytical traverse (see Fig.3). Solid circles – Ferry & Spear (1978); open squares – Perchuk & Lavrent'eva (1981); solid triangles – Hodges & Spear (1982); open circles – Ferry & Spear with Berman (1990) mixing model; solid squares – Kleeman & Reinhart (1994).

calibrations, as well as comparison of temperature differences between samples. The five calibrations differ in absolute temperature, by as much as 108°C, between the highest and lowest temperature in A1435 (Table 3, Fig. 6). According to Bucher & Frey (1994), the first garnet + biotite pair may form at temperatures around 470°C where garnet is stabilized by Mn and Ca; therefore, in the lowest-grade samples, the Ferry & Spear (1978) calibration appears to yield unreasonably low temperatures for rocks containing this assemblage. The discrepancy may be due to the calcareous nature of most of the pelitic schists, indicated by the presence of epidote, and in the case of A1945, calcic amphibole. Garnet rim compositions from most of the samples contain (Ca + Mn)/(Ca + Mn + Mg + Fe) > 0.2, excee-

Table 4. *Selected electron microprobe analyses of amphibole

Sample:	A1463	A1463	A20	A20	A1573	A1573	A114	A114	GRUV	GRUV
Total # of analyses:	(10)		(7)		(16)		(15)		(14)	
SiO_2	52.89	45.66	51.92	44.42	48.36	41.37	45.09	43.59	44.58	42.64
TiO_2	0.08	0.30	0.13	0.37	0.17	0.48	0.35	0.25	0.33	0.31
Al_2O_3	3.92	11.94	4.89	11.07	6.90	16.51	13.37	14.28	13.78	14.79
MgO	16.40	11.87	16.38	11.63	13.37	8.25	11.49	12.18	12.25	12.46
CaO	12.61	11.44	11.84	11.11	11.81	11.02	11.17	10.73	11.21	10.47
MnO	0.16	0.26	0.16	0.11	0.21	0.16	0.16	0.21	0.14	0.16
FeO	10.33	13.09	10.68	14.10	12.63	15.70	12.78	12.37	12.39	12.70
Na_2O	0.46	1.51	0.85	1.72	0.71	1.77	1.75	1.57	1.94	1.94
K_2O	0.08	0.33	0.12	0.26	0.18	0.44	0.36	0.35	0.31	0.36
Total	96.93	96.40	96.97	94.79	94.34	95.70	96.52	95.53	96.93	95.83
Cations on 23 O basis										
Si	7.57	6.67	7.40	6.64	7.20	6.20	6.58	6.34	6.45	6.18
Al^{IV}	0.43	1.33	0.60	1.36	0.80	1.80	1.42	1.66	1.55	1.82
Al^{VI}	0.23	0.73	0.22	0.59	0.41	1.12	0.88	0.79	0.80	0.71
Ti	0.01	0.03	0.01	0.04	0.02	0.05	0.04	0.03	0.04	0.03
Fe^{3+}	0.17	0.44	0.49	0.58	0.34	0.44	0.41	0.95	0.60	1.17
Mg	3.50	2.59	3.48	2.59	2.97	1.84	2.50	2.64	2.64	2.69
Fe^{2+}	1.07	1.16	0.79	1.18	1.24	1.53	1.15	0.55	0.90	0.37
Mn	0.02	0.03	0.02	0.01	0.03	0.02	0.02	0.03	0.02	0.02
Ca	1.93	1.79	1.81	1.78	1.89	1.77	1.75	1.67	1.74	1.63
Na^{M4}	0.07	0.21	0.18	0.22	0.10	0.23	0.25	0.33	0.26	0.38
Na^{A}	0.06	0.22	0.06	0.28	0.11	0.28	0.25	0.11	0.28	0.18
K	0.02	0.06	0.02	0.05	0.04	0.09	0.07	0.06	0.06	0.07
Total	15.06	15.27	15.06	15.33	15.13	15.37	15.30	15.16	15.32	15.24

**Amphibole analyses listed contain the maximum and minumum SiO_2 (and minimum and maximum Al_2O_3, respectively) analyzed for that sample. A complete list of analyses is found in McClellan (1993).*

ding the suggested limit for the use of the Ferry & Spear (1978) calibration. The between-calibration differences are reduced to < 40°C if this method is disregarded.

In a between-sample comparison, in which the uncertainties are reduced to those of analytical imprecision (Bucher & Frey 1994), all calibrations yield a similar pattern of temperatures increasing from east to west and peaking near the core of the dome (sample A791), then dropping slightly on the western flank of the dome (Fig. 6).

Metamorphic conditions based on amphibole chemistry

Laird & Albee (1981) compared mineral chemistry in natural mafic assemblages in Vermont to metamorphic grade deduced from interlayered pelitic rocks. Utilizing mafic rocks with the common assemblage amphibole + chlorite + epidote + plagioclase + quartz + Ti-phase ± carbonate ± K-mica ± Fe^{3+}, they characterized the compositional and modal variations in the various phases from biotite to sillimanite grade conditions by a series of formula proportion diagrams. Their results, especially for medium pressure conditions, are supported by experimental studies (Apted & Liou 1983), and clearly demonstrate that the tschermak, (Al^{VI} + Fe^{3+} + Ti + Cr), Al^{IV} ↔ (Fe^{2+} + Mg + Mn), Si, and *glaucophane*, Na^{M4}, (Al^{VI} + Fe^{3+} + Ti + Cr) ↔ Ca, (Fe^{2+} + Mg + Mn), substitutions increase continuously with metamorphic grade, while the *edenite*, (Na^A + K), Al^{IV} ↔ □, Si, substitution increases through garnet grade, then decreases (□ indicates a vacancy in the A-site).

Five amphibole-bearing samples from an east-to-west traverse across the Einunnfjellet-Savalen area (Fig. 3) were chosen for electron microprobe analyses. These specimens contain the common assemblage, and are interpreted to represent basaltic protoliths. Amphibole analyses (Table 4) were normalized following the method of Leake (1978), based on 23 oxygens, using the reasonable crystal-chemical limits set forth by Leake (1978) and Robinson et al. (1982). All mineral grains analyzed are calcic amphiboles, having $(Ca + Na)^{M4} \geq 1.34$, $Na^{M4} < 0.67$, $(Na + K)^A < 0.5$, and Ti < 0.5. The amphiboles analyzed vary from actinolite to tschermakite (Fig. 7). Note that although it was recently recommended that the names actinolitic hornblende and tschermakitic hornblende be abandoned (Leake et al. 1997), these names are retained here to illustrate the range in amphibole composition.

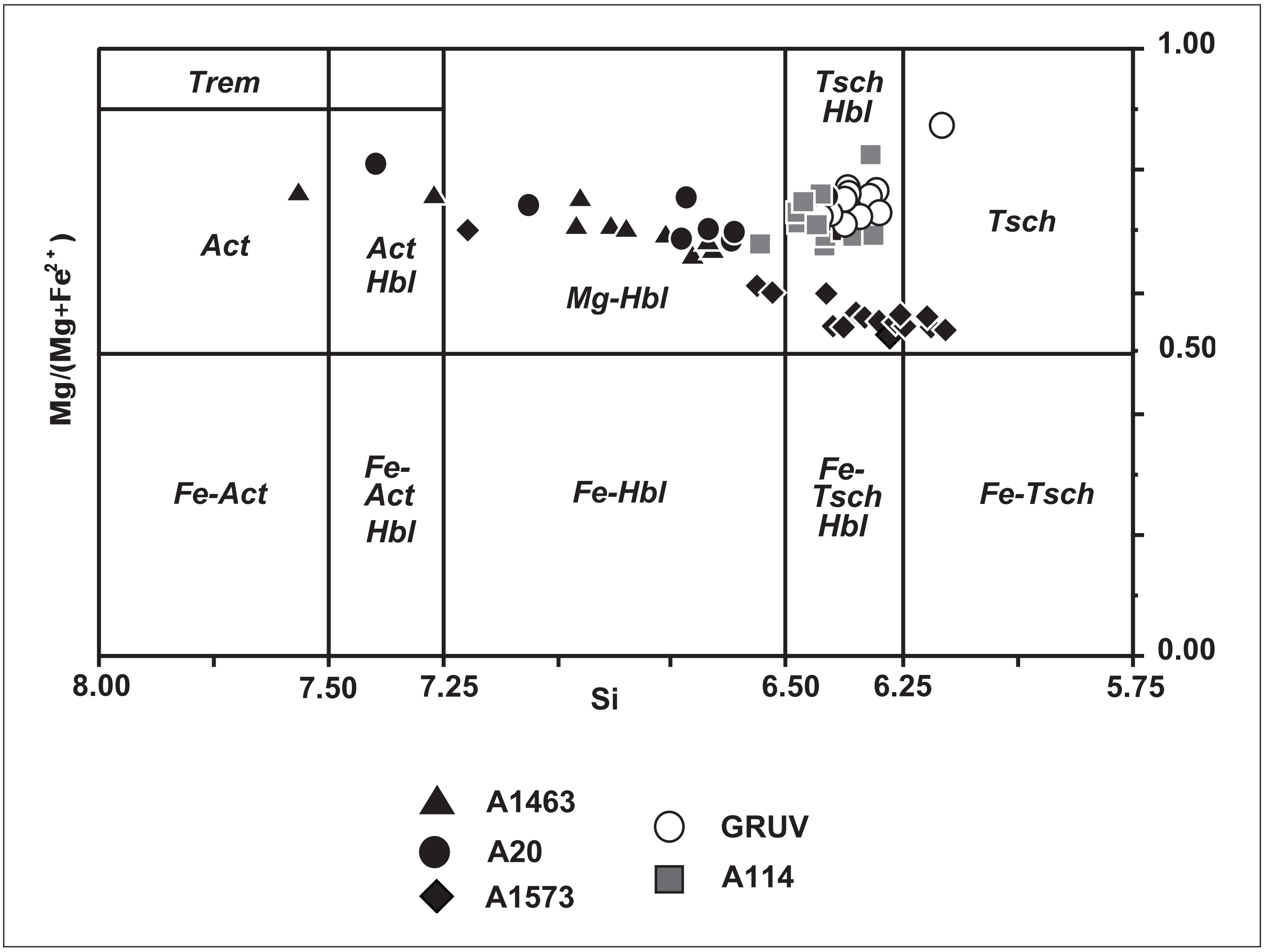

Fig. 7: Classification of calcic amphiboles, based on Leake (1978).

Samples A1463, A20, and A1573 were collected from the Gula Nappe in the eastern part of the study area (Fig. 3). Both A1463 and A20 are fine-grained greenstones containing aggregates of randomly oriented pale green amphibole, along with chlorite and clinozoisite (Fig. 8a). The acicular amphibole grains, generally less than 0.25 mm in length, are dominantly magnesiohornblende, although discrete grains of actinolite to actinolitic hornblende are present. Analyses of cores vs. rims were difficult to obtain because of the fine grain size, although some grains in A20 are optically zoned. In the diagrams of Laird & Albee (1981), in which metamorphic zones were defined based on coexisting pelitic rocks, the amphibole compositions fall into the biotite and lower garnet zone fields (Figs. 9 and 10). Sample A1573 is a slightly coarser-grained greenschist, and contains approximately 2 mm long, gray-green amphiboles in a fine-grained matrix of acicular, randomly oriented grains (Fig. 8b). Amphibole composition is dominantly tschermakitic hornblende (Fig. 7), and the analyses, with the exception of one, fall into the garnet zone field on the Laird & Albee diagrams (Figs. 9 and 10).

In contrast to the samples described above, amphibolites from the Meråker Nappe, closer to the core of Einunnfjellet dome (samples GRUV and A114), are relatively coarse-grained, strongly foliated, and dark-colored. Both samples contain epidote (GRUV) or clinozoisite (A114), although chlorite is of low modal abundance. In thin section, amphiboles are bluish-green, and have a bimodal grain-size distribution (Fig. 8c, d). GRUV contains prismatic grains that range from 0.25 to 2 mm, while A114 contains small acicular grains that commonly appear overprinted by larger anhedral to prismatic grains up to 4 mm long. Despite the overprinting relationships, both the matrix grains and larger amphiboles in both samples form a tight cluster in the tschermakitic hornblende field of the Leake diagram (Fig. 7), suggesting that the amphiboles do not represent growth under significantly different metamorphic conditions. Only two analyses deviate from this composition; one, a rim of a large amphibole from GRUV, is a Mg-rich tschermakite, while the other is a core of magnesiohornblende from a prismatic grain in A114. Amphibole compositions from these two samples cluster fairly tightly in the upper part of the garnet

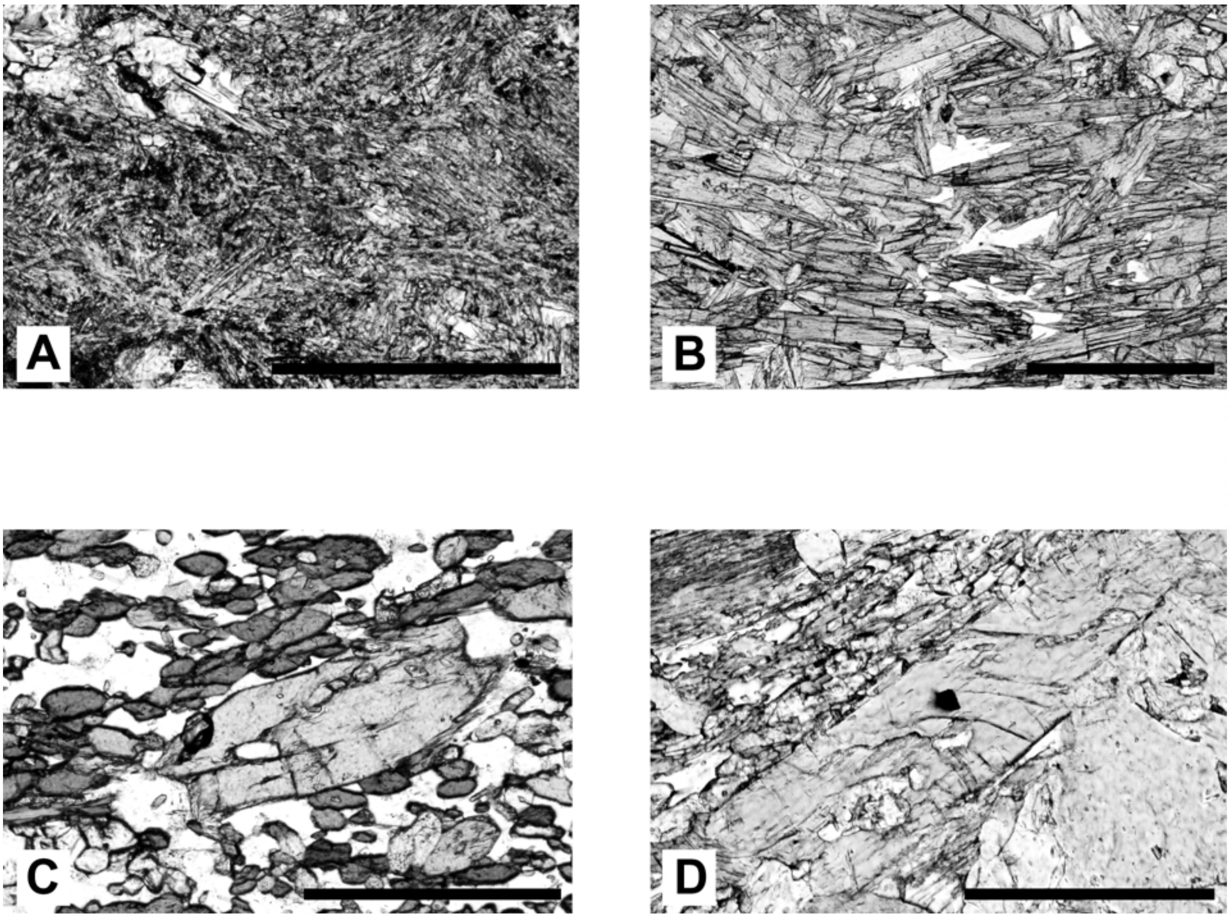

Fig. 8: Photomicrographs of metamorphosed mafic rocks analyzed for this study. Sample locations shown in Figure 3. A: Felted mass of randomly oriented Ca-amphibole, with chlorite, clinozoisite, and calcite. Sample A1463, Gula (Köli) Nappe. Scale bar = 0.5 mm. B: Aligned Ca-amphibole grains, in matrix of clinozoisite, chlorite, quartz, and plagioclase. Sample A1573, Gula (Köli) Nappe. Scale bar = 0.5 mm. C: Blocky Ca-amphiboles with bimodal size distribution, in matrix of quartz, plagioclase, and epidote. Sample GRUV, Meråker (Köli) Nappe. Scale bar = 0.25 mm. Aligned fine-grained Ca-amphibole, cross-cut by larger prismatic or anhedral Ca-amphibole grains. Matrix dominated by quartz and plagioclase, with scattered grains of chlorite and epidote. Sample A114, Meråker (Köli) Nappe. Scale bar = 0.5 mm.

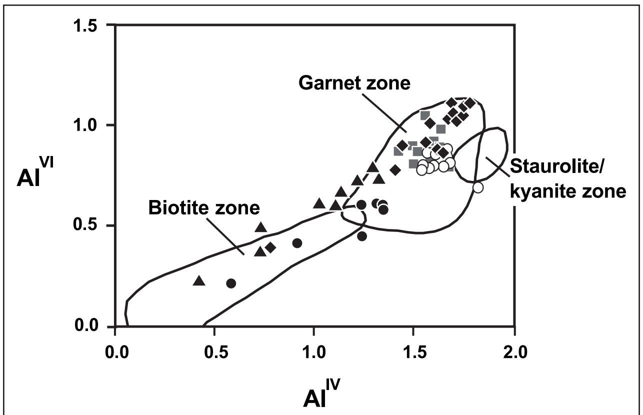

Fig. 9: Al^{VI} vs. Al^{IV} formula proportion diagram for amphiboles. Fields corresponding to the biotite, garnet, and staurolite/kyanite metamorphic zones indicated. After Laird and Albee (1981). S-ymbols as in Figure 7.

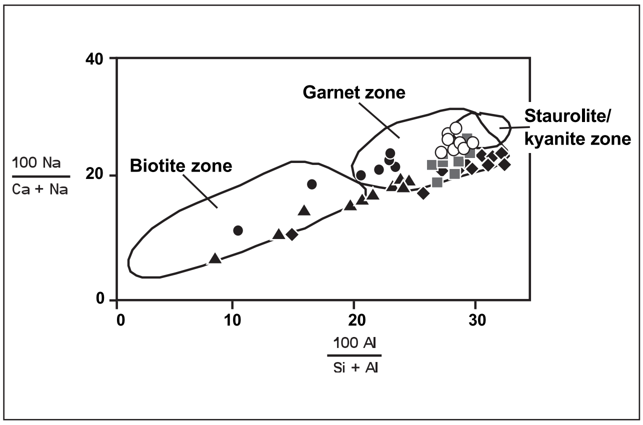

Fig. 10: Na/(Ca + Na) vs. Al/(Si + Al) formula proportion diagram for amphiboles. Fields corresponding to the biotite, garnet, and staurolite/kyanite metamorphic zones indicated. After Laird and Albee (1981). Symbols as in Figure 7.

Table 5. Electron microprobe analyses of plagioclase

Sample:	A1393 matrix	A933 matrix	A1337 core-incl in gar	A1337 rim-incl in gar	A1337 matrix	A791 matrix	A1945 matrix	A20 matrix	A1463 matrix	A1573 matrix	GRUV matrix
SiO_2	68.40	62.12	67.07	66.34	63.53	67.93	62.00	67.46	63.60	62.38	62.01
Al_2O_3	19.71	23.36	20.66	20.64	22.75	19.74	22.96	19.87	22.67	23.41	23.23
CaO	0.18	4.40	1.24	1.38	3.49	0.12	4.29	0.26	3.84	4.77	4.68
Na_2O	11.62	8.99	10.75	10.63	9.41	11.63	9.05	11.49	9.49	8.96	9.07
K_2O	0.06	0.06	0.04	0.21	0.10	0.05	0.08	0.03	0.08	0.06	0.09
Total	99.97	98.93	99.76	99.20	99.28	99.47	98.38	99.11	99.68	99.58	99.08
Cations on 8 O basis											
Si	2.99	2.78	2.94	2.93	2.82	2.98	2.79	2.97	2.81	2.77	2.77
Al	1.01	1.23	1.07	1.07	1.19	1.02	1.22	1.03	1.18	1.23	1.22
Ca	0.01	0.21	0.06	0.07	0.17	0.01	0.21	0.01	0.18	0.23	0.22
Na	0.98	0.78	0.91	0.91	0.81	0.99	0.79	0.98	0.81	0.77	0.79
K	0.003	0.004	0.002	0.012	0.006	0.003	0.004	0.002	0.004	0.004	0.005
X_{ab}	0.99	0.783	0.938	0.922	0.825	0.991	0.790	0.986	0.814	0.770	0.774
X_{an}	0.01	0.212	0.060	0.066	0.169	0.006	0.206	0.012	0.182	0.226	0.221
X_{or}	0.003	0.004	0.002	0.012	0.006	0.003	0.004	0.002	0.004	0.004	0.005

zone field, near or adjacent to the staurolite/kyanite field on the Laird & Albee diagrams (Figs. 9 and 10).

Pressure estimates

Pelitic schists from the Einunnfjellet-Savalen area lack Al_2SiO_5 polymorphs, and consequently the assemblage most commonly used for barometry in metapelites (the garnet-plagioclase- Al_2SiO_5-quartz barometer, or GASP). The assemblage garnet + muscovite + plagioclase + biotite (Ghent & Stout 1981) does occur (although not all are in mutual contact) in some pelitic samples; however, the plagioclase is typically more sodic than An_{20} (Table 5) and therefore inappropriate for use with this barometer (see Ashworth & Evirgen 1985). Although no attempt is made here to quantify the pressures of metamorphism, general pressure conditions can be estimated from the existing mineral compositions and assemblages.

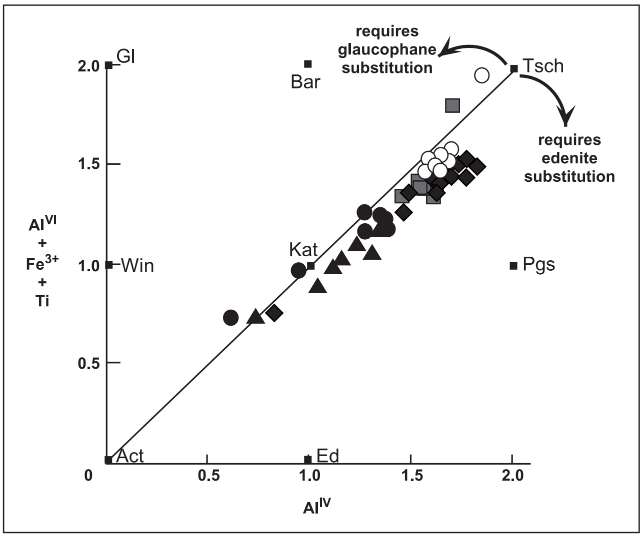

Fig. 11: Al^{IV} vs. (Al^{VI} + Fe^{3+} + Ti) formula proportion diagram for amphiboles. Amphibole compositions from this study are consistent with moderate-P conditions; see text for discussion. End-member amphiboles are: Act – actinolite; Bar – barroisite; Ed – edenite; Gl – glaucophane; Kat – katophorite; Pgs – pargasite; Tsch – tschermakite; Win – winchite. Symbols as in Figure 7.

Based on earlier suggestions that amphibole chemistry is pressure-sensitive (e.g., Leake 1965), Laird & Albee (1981) developed variation diagrams that can be used as a qualitative gauge of pressure. Figure 11 illustrates variation in Al^{IV} vs. (Al^{VI} + Fe^{3}+ + Ti) as a function of common substitutions in amphiboles during metamorphism, where analyses falling above a line from actinolite to tschermakite require glaucophane substitution to maintain charge balance, and those falling below this line require edenite substitution. Laird & Albee (1981) showed that the glaucophane substitution dominates in amphiboles from high-pressure terranes, whereas the tschermak and edenite substitutions dominate at lower pressures. Amphiboles from the study area form a nearly linear trend parallel to and slightly below this line, indicating the dominance of the tshermak and edenite substitutions with increasing metamorphic grade. This trend falls within the limits of amphiboles in rocks from medium-pressure terranes considered by Laird and Albee (1981). As noted by these authors, the parameters used in Fig. 11 are dependent on the normalization method used in converting mineral analyses to structural formulas. Therefore, they also considered cation ratios that are independent of normalization, as illustrated in Fig. 12. Amphibole compositions from this study fall within the medium-pressure (or low- to medium-P overlap) fields on this diagram, in agreement with the conditions indicated above.

Other indications of moderate pressure conditions are (1) the correspondence of the albite-oligoclase isograd with the almandine-garnet isograd (Laird & Albee, 1981) (Fig. 12); and (2) the composition of muscovites from the study area. The phengite component of white mica in pelitic schists appears to be pressure-dependent (e.g., Velde 1967; Powell & Evans 1983; Massonne & Schreyer 1987), and in particular the Si content increases with pressure. White micas in samples from this study are characterized by low phengite components (Si < 6.37 and generally Fe+Mg < 0.55 cations per 22 oxygens; Table 6), consistent with the moderate-pressure conditions indicated by amphibole compositions.

Summary of metamorphic conditions

East of the garnet isograd in the study area (Fig. 13), biotite-grade assemblages occur in calcareous and/or carbonaceous slates and phyllites. West of the garnet isograd, garnet-biotite thermometry and amphibole compositions suggest a progressive metamorphic increase from greenschist to amphibolite facies across the analyzed traverse. A rather abrupt change in the field appearance of both pelitic and mafic rocks corresponds to the transition zone between the two facies, interpreted to occur over an approximately 3 km-wide zone (Fig. 13). In the field, this is indicated by a change from phyllite and greenstone to coarser-grained schist and amphibolite. Within this zone, actinolitic amphibole

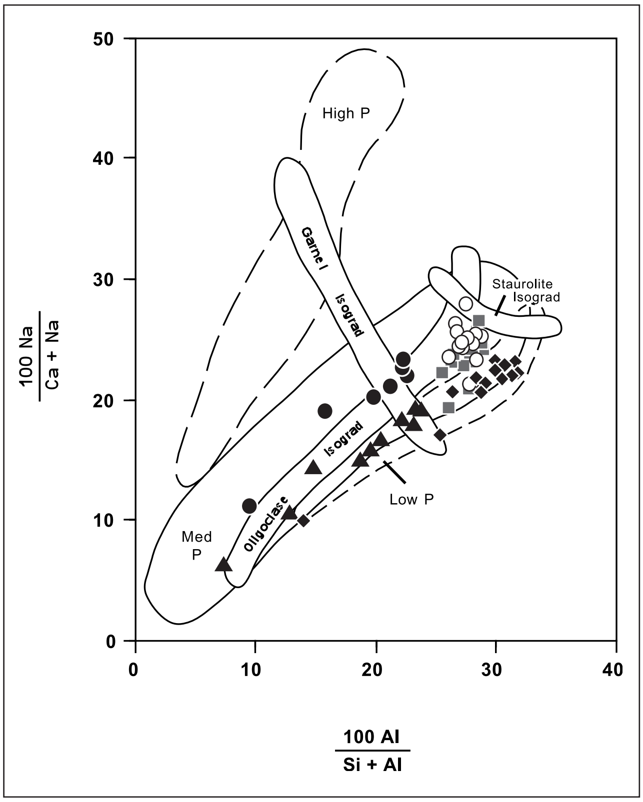

Fig. 12: Na/(Ca + Na) vs. Al/(Si + Al) formula proportion diagram for amphiboles, showing fields corresponding to low-, medium-, and high-pressure metamorphism. Locations of the garnet, oligoclase, and staurolite isograds also shown. After Laird and Albee (1981). Symbols as in Figure 7.

Table 6. Electron microprobe analyses of muscovite

Sample:	A1435	A1435	A1393	A1718	A1341	A1341	A1341	A1341	A1337	A1945	A1945	A1945	A1945	A1945
		matrix		matrix	matrix						matrix			by amph
SiO_2	46.15	45.96	45.98	46.77	45.71	45.98	46.24	46.04	46.02	45.06	45.52	45.88	46.11	46.20
TiO_2	0.32	0.28	0.62	0.46	0.46	0.41	0.25	0.23	0.7	0.49	0.49	0.49	0.43	0.30
Al_2O_3	31.90	33.12	32.92	31.72	35.35	33.83	33.54	34.83	33.99	33.44	33.83	34.07	34.09	33.06
Cr_2O_3	0.01	0.05	0.01	0.02	0.07	0.02	0.03	0.00	0.15	0.08	0.05	0.04	0.00	0.13
FeO	2.85	2.69	1.98	1.67	1.48	1.90	2.09	1.90	1.54	1.19	1.22	1.29	1.29	1.65
MgO	1.39	1.09	1.2	1.84	0.89	1.24	1.24	1.02	1.21	1.37	1.28	1.44	1.29	1.68
Na_2O	0.83	1.07	0.71	0.70	1.77	1.49	1.25	1.27	0.95	1.14	1.34	1.00	1.28	1.18
K_2O	9.58	9.68	10.04	10.12	8.96	9.36	9.31	9.73	9.95	9.26	9.35	9.73	9.51	9.56
Total	93.04	93.94	93.46	93.29	94.69	94.24	93.96	95.02	94.51	92.03	93.08	93.93	93.99	93.76
Cations on 22 O basis														
Si	6.32	6.24	6.26	6.37	6.11	6.20	6.24	6.16	6.19	6.19	6.19	6.19	6.21	6.25
Al^{IV}	1.68	1.76	1.74	1.63	1.89	1.80	1.76	1.84	1.81	1.81	1.81	1.81	1.79	1.75
Al^{VI}	3.47	3.54	3.54	3.46	3.67	3.57	3.58	3.64	3.57	3.61	3.61	3.60	3.61	3.52
Ti	0.03	0.03	0.06	0.05	0.05	0.04	0.03	0.02	0.07	0.05	0.05	0.05	0.04	0.03
Cr	0.00	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.02	0.01	0.01	0.00	0.00	0.01
Fe	0.33	0.31	0.23	0.19	0.17	0.21	0.24	0.21	0.17	0.14	0.14	0.15	0.15	0.19
Mg	0.28	0.22	0.24	0.37	0.18	0.25	0.25	0.20	0.24	0.28	0.26	0.29	0.26	0.34
Na	0.22	0.28	0.19	0.18	0.46	0.39	0.33	0.33	0.25	0.30	0.35	0.26	0.33	0.31
K	1.67	1.68	1.74	1.76	1.53	1.61	1.60	1.66	1.71	1.62	1.62	1.67	1.63	1.65
Fe+Mg	0.61	0.53	0.47	0.56	0.42	0.42	0.46	0.49	0.34	0.41	0.40	0.43	0.40	0.53

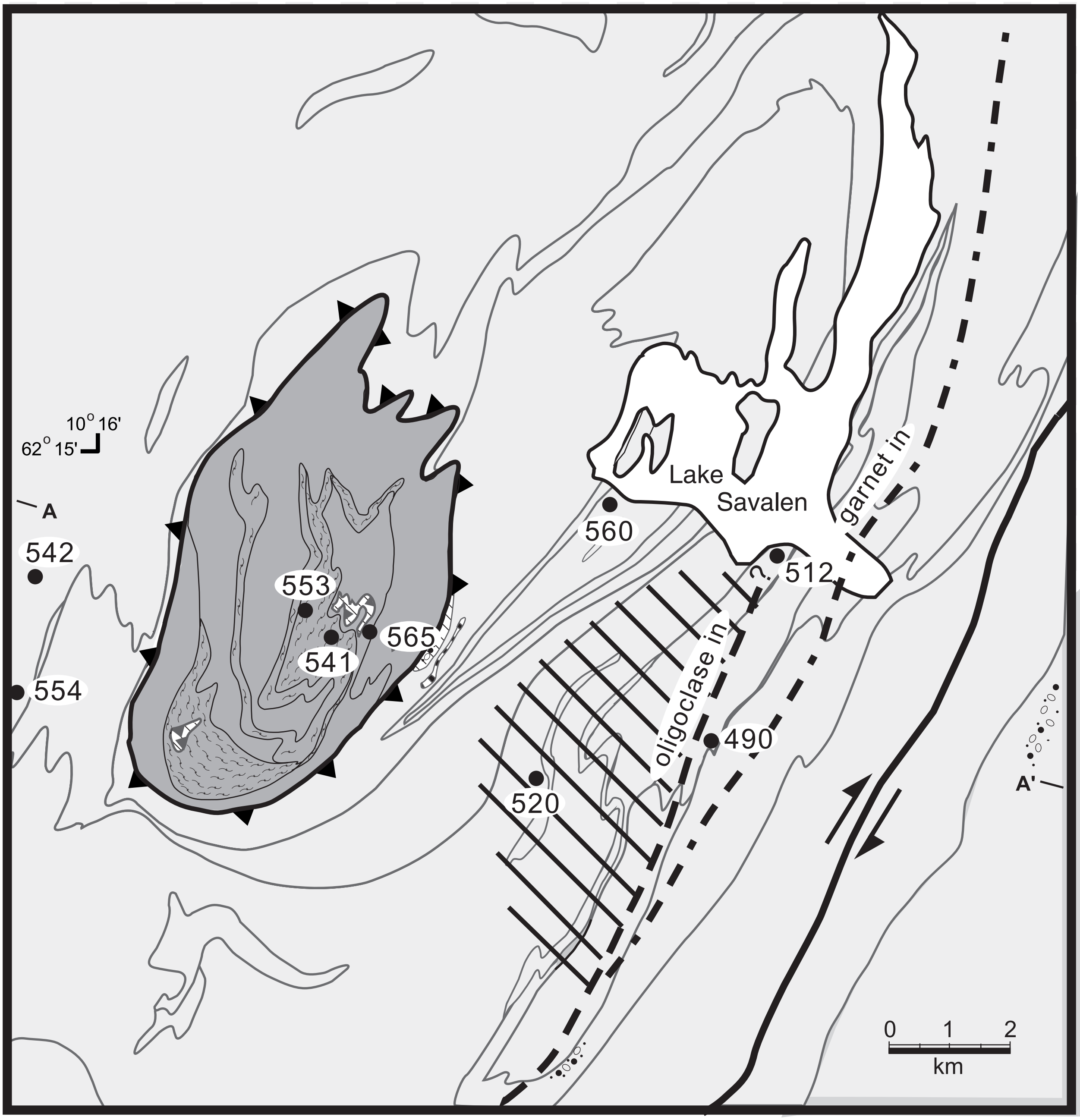

Fig. 13: Metamorphic summary of the Einunnfjellet-Savalen area; Seve Nappe shown in dark gray; Köli Nappe in lighter gray. Temperatures shown are averages for each sample calculated from garnet-biotite pairs using calibration of Hodges and Spear (1982). Dashed lines indicate location of almandine-garnet and oligoclase isograds. Hatched area corresponds to the greenschist-to-amphibolite facies transition zone (i.e. epidote-amphibolite facies) where defined in the region of the analytical traverse.

coexists with magnesiohornblende in mafic rocks, as is commonly observed in rocks across the greenschist – amphibolite facies transition (Spear 1993). Plagioclase composition increases abruptly from almost pure albite in sample A20 to An_{17-22} in sample A1463 (Table 5), and the oligoclase isograd in mafic rocks appears to correspond fairly closely to the almandine-garnet isograd in pelitic rocks (Fig. 13). Mafic samples west of the oligoclase isograd are epidote amphibolites in the sense of Spear (1993); that is, rocks that contain epidote and coexisting plagioclase of approximately An_{25}.

The epidote-amphibolite facies (Fig.14) represents a transition zone, over a fairly narrow range in temperature, between the greenschist and amphibolite facies in medium-pressure metamorphic rocks (Turner 1981). The absolute P-T conditions of this complex, multip-

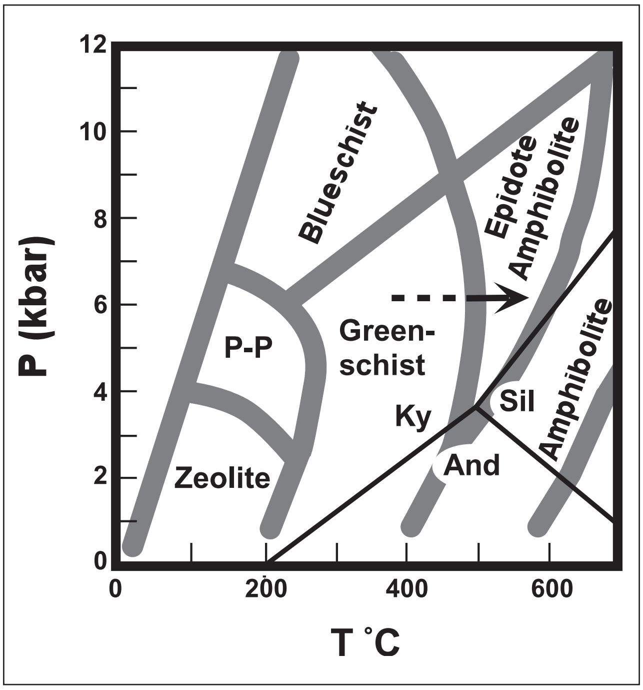

Fig. 14: Metamorphic facies diagram illustrating P-T conditions associated with the epidote-amphibolite facies. Arrow depicts approximate conditions determined from the analyzed traverse in this study, using a model P of 6 kbar; dashed part of line is conjectured from mineral assemblages below garnet isograd. Diagram modified from Spear (1993).

hase transition are difficult to quantify (Apted & Liou, 1983), and depending upon the criteria used to define it, the greenschist-amphibolite facies boundary may vary between 450˚ and 550˚C (Essene 1989). With the exception of those given by the Ferry & Spear (1978) calibration, all temperatures obtained for the pelitic schists occurring in the transition zone (A1435, A1393, and A933) correspond with the interpreted transition (Table 3). The thermometric pattern across the sample traverse assumes that calculations yield maximum calculated temperatures for each sample. As shown earlier, this assumption appears justified except for some evidence for disequilibrium in sample A933. Obviously, sample A933 is critical to the interpreted greenschist-amphibolite facies transition because of its position within the transition zone (Fig. 13). Amphibole chemistry from mafic sample A1573 (Figs. 9 and 10), however, is consistent with the temperatures determined from sample A933. In addition, two pelitic samples collected approximately 75 m west of A933 contain the assemblage quartz + muscovite + chloritoid + chlorite + garnet, with no visible evidence for disequilibrium. This assemblage is stable over the temperature range of 450˚-550˚C (Bucher and Frey, 1994), above which chloritoid disappears in reactions producing almandine + staurolite. Therefore, these samples constrain temperatures in the upper transition zone to < 550˚C, consistent with maximum T calculated for A933.

Westward from the greenschist – amphibolite-facies transition zone, both amphibole chemistry and geothermometry suggest that the metamorphic grade increases continually. The highest temperatures occurred in the structurally lower rocks close to the core of the dome, peaking well above conditions at which staurolite would first appear in rocks of appropriate composition (≥ 500˚C; Bucher & Frey, 1994). Staurolite is generally absent from pelitic schists in the area, most likely due to their calcareous nature (e.g., Boyle & Westhead, 1992); however, the conditions determined in this study are consistent with the previously noted report of staurolite-bearing schists in the Hummelfjell Nappe from near the core of the Einunnfjellet dome (Ramsay & Sturt 1998). The single report of kyanite in the area, as discussed earlier, is also in agreement with these conditions (Fig. 14).

Discussion

In the following section, metamorphic data from the Einunnfjellet-Savalen area are evaluated in the light of regional geologic relationships, in order to examine their implications for tectonic models of the central Norwegian Caledonides. Specifically discussed are 1) the relationship between metamorphism and nappe stacking; 2) the timing of metamorphism; and 3) the existence or regional significance of pre-Scandian metamorphic events.

Relationship of thrusting and metamorphism

In a traverse across the study area from structurally higher to structurally lower units, mineral assemblages, garnet-biotite thermometry, and amphibole chemistry record a continual metamorphic gradient from greenschist through lower amphibolite facies, i.e. a normal metamorphic gradient. Data from this study do not bear evidence for metamorphic inversions across these boundaries, as are commonly associated with major thrust boundaries in the central and northern Scandinavian Caledonides (e.g., Andréasson & Gorbatchev 1980), and which may result from thrusting of a hotter sheet over colder rocks (e.g., Rice et al. 1989), post-metamorphic thrusting of higher-grade rocks over lower-grade rocks (e.g., Barker & Anderson 1989), or structural overturning of isograds subsequent to metamorphism. Conversely, abrupt metamorphic breaks at previously mapped nappe boundaries, a common characteristic of post-metamorphic faults, are not apparent. Therefore, the observed progression is consistent with an interpretation of nappe stacking in the Einunnfjellet-Savalen area prior to or broadly coincident with peak metamorphism.

Thermochronologic constraints on timing of metamorphism in the central Caledonides

Although no thermochronologic data currently exist for the immediate study area, various studies of the Seve and Köli in the Caledonides of central Sweden and Norway are relevant to the present case. Faunal and sedimentological evidence clearly indicates that major nappe translation in the central Scandinavian Caledonides occurred during the Siluro-Devonian Scandian orogeny. However, in contrast to the northern Caledonides where several studies have pointed to Late Cambrian subduction of the outer margin of Baltica accompanied by eclogite-facies metamorphism (e.g., Sturt et al. 1978; Mørk et al. 1988; Dallmeyer et al. 1991; Andréasson 1994; Essex et al. 1997), convincing quantitative evidence for a widespread pre-Scandian metamorphic event affecting the central and southern Caledonides is less clear. Early investigations of Rb-Sr whole-rock isochron dates from the Seve Nappes in the Swedish Caledonides (Claesson 1979, 1980) suggested at least local effects of a pre-Scandian tectonothermal event, whereas U-Pb dating of zircon and monazite from granulite-facies Seve rocks indicated Caledonian metamorphism at 414 – 402 Ma (Claesson 1981). In the latter study, Claesson reported a U-Pb zircon age of 414±27 Ma from a neosome in migmatitic metabasite, from the upper Seve Nappes, and a slightly discordant monazite age of 402 Ma from gneiss in the same unit. More recent dating of metamorphic minerals in the central and northern Swedish Caledonides (Gromet et al. 1996; Essex et al. 1997) also points toward the prevalence of Scandian metamorphic effects. Gromet et al. (1996) argued that U-Pb titanite (sphene) and monazite ages of 440 – 427 Ma from amphibolite- and granulite-facies Seve rocks in Jämtland county, Sweden, signify contemporaneous prograde metamorphism in the Lower and Middle Seve during Scandian orogenesis. Consistent results from both unsheared and sheared rock, and from a range of grain sizes, give no indication of multiple generations or diffusional resetting of earlier formed minerals. Ar^{40}-Ar^{39} studies in the central Scandinavian Caledonides have also yielded conflicting results. Dallmeyer et al. (1985) interpreted Ar^{40}-Ar^{39} hornblende and mica ages from the Seve and Köli Nappes from Jämtland to reflect distinct Early and Middle Ordovician metamorphic events affecting them. These were subsequently overprinted by the Scandian orogenesis. However, in a similar investigation in northwestern Jämtland, Dallmeyer & Gee (1988) suggested that the Köli Nappes in that area had been affected by only the Scandian event (~ 430 Ma), whereas hornblende ages of 471 to 459 Ma from the eastern Seve Nappes indicated cooling through closure temperatures from an earlier event. Recent Sm-Nd mineral isochron dating of eclogites from Jämtland attests to high-pressure metamorphism at ~454 Ma in the Seve Nappes in this area (Brueckner & van Roermund 2004; Brueckner et al. 2004).

Further to the south, and closer to the area of this study, Dallmeyer (1990) described Ar^{40}-Ar^{39} results from the Seve and Köli Nappes (including the Gula and Meråker Nappes) in Sør-Trøndelag county, Norway. With only one exception of a 476 Ma hornblende age from a Seve sample, both units yielded consistent hornblende and mica cooling ages of 435 – 420 Ma. These results are corroborated by muscovite Ar^{40}-Ar^{39} cooling ages between 415 – 402 Ma (Hacker & Gans, in press) from the Seve and Gula nappes along a traverse approximately at the latitude of Singås (Fig. 1). Ion microprobe Th/Pb monazite ages from the Gula and Seve Nappes in the Trondheim region (Hacker & Grove 2002) also suggest that monazite crystallization occurred by 412 Ma.

The results of this study indicate that the Seve-Köli thrust contact in this region, and overlying thrusts within the Köli, were overprinted by greenschist to lower amphibolite-facies metamorphism, suggesting that nappe assembly was earlier than or broadly coeval with the metamorphism. Taking into account the thermochronologic data summarized above, particularly from Köli-equivalent nappes, it is likely that the peak metamorphic assemblages recorded in rocks of this study formed and cooled during an interval between ~435 Ma and 400 Ma, broadly correlative with the Scandian orogenic phase.

Consideration of multiple metamorphic events

Samples from this study, including those from rocks correlative with the Seve Nappe, appear to record effects of a single major metamorphic event of Scandian age, an observation that is inconsistent with the interpretation of an earlier Paleozoic widespread, regional metamorphism affecting the Baltoscandian margin of the central Caledonides. However, numerous accounts from the general region of this study provide evidence for ductile deformation and metamorphism prior to the Scandian orogeny; for example, localized evidence for a polymetamorphic history in rocks of the Gula Nappe (e.g., Guezou 1978; Andréasson & Gorbatchev 1980; Lagerblad 1984; Pannemans & Roberts 2000), the presence of weakly deformed and metamorphosed, Early Silurian trondhjemites cutting rocks that contain a preexisting metamorphic fabric and related folds (e.g., Pannemans & Roberts 2000; Roberts & Sundvoll 2000), and some thermochronologic evidence from the central Caledonides, north of the present study area, discussed above.

In considering the question of multiple metamorphic events, we should ask what type of pre-Scandian metamorphic pattern might we *expect* to see? Our expectations should depend largely on a conceptual model of the Baltoscandian margin prior to Scandian continent-continent collision, and this hinges, in part, on understanding the location and polarity of subduction zones outboard of Baltica throughout the Ordovician, a topic of continued uncertainty (Stephens & Gee 1985; Stephens 1988; Grenne et al. 1999). Sturt & Roberts (1991) proposed westward (oceanward) subduction that generated Early Ordovician suprasubduction zone ophiolites, and eventually led to their obduction onto Baltica, although the authors equated this with Finnmarkian subduction of the Baltoscandian margin in northern Norway. More recently, Roberts et al. (2002) and Roberts (2003) have proposed a model for the central Norwegian Caledonides in which oceanward-directed subduction at ~490-480 Ma resulted in blueschist metamorphism at the base of the Bymarka ophiolite in the Støren Nappe (Eide & Lardeaux 2002), followed by obduction of oceanic crust (e.g., Støren Nappe) onto the Gula microcontinent. Subsequent collision with the outer margin of Baltica emplaced the Gula block and its carapace of ophiolitic crust against the Seve Nappes on the Baltoscandian margin. Recognition of ~454 Ma eclogites in the Seve Nappes in Jämtland provides another important indication of west-dipping, oceanward subduction affecting the central Scandinavian Caledonides, (Brueckner & van Roermund 2004; Brueckner et al. 2004), in this case implying subduction of the Baltoscandian margin beneath an arc/microcontinent composite terrane (distinct from the Gula terrane described above) and subsequent exhumation prior to onset of Scandian orogenesis.

The growing evidence for multiple collision events strongly argues for the existence of complex, 'Pacific-type' oceanic tracts separating Baltica and Siberia during the Early Ordovician, and Baltica and Laurentia by Middle Ordovician (e.g., Eide & Lardeaux 2002; Brueckner & van Roermund 2004). Presently, there is no indication of high-pressure, crustal subduction-type metamorphism in the Gula Nappe (Hacker & Gans, in press), and blueschist-facies metamorphism in the Støren Nappe likely occurred in the oceanic realm, prior to obduction onto the Gula microcontinent (Roberts et al. 2002b; Roberts 2003). Recent thermobarometric data from the Seve and Gula Nappes yield temperatures and pressures for the Seve Nappe ranging from 645°C and 10 kbar, to 724°C and 12 kbar, in contrast to a range of 604-660°C at ~9 kbar for the Gula Nappe, and imply that the Seve and Gula Nappes were structurally separated and buried to quite different depths – 40-50 km and 30-35 km, respectively (Hacker & Gans, in press). These data argue against the Gula-Seve correlation suggested by Sturt & Ramsay (1997, 1999), and consequently against the single widespread obduction event they envisaged. More likely, the Köli-equivalent (Støren, Gula, Meråker) and Seve Nappes in this region were involved in multiple obduction episodes prior to final amalgamation.

Restricting the discussion to the southern Trondheim region, one explanation for the localized nature of pre-Scandian metamorphism could derive from consideration of ophiolites that were emplaced onto continental crust during oceanward-directed subduction (Tethyan-type ophiolites of Wakabayashi & Dilek 2000), but have not been overprinted by orogenesis related to subsequent continent-continent collision. In such situations, high-grade metamorphism is commonly restricted to the metamorphic sole, a zone typically < 500 m thick, of fault-bounded metamorphic rocks structurally beneath the ophiolite (Wakabayashi & Dilek 2000). A steep, inverted, metamorphic gradient characterizes the sole, with the highest grade rocks near the base of the ophiolitic mass and decreasing in grade downwards through the sole. Autochthonous and allochthonous rocks underneath the ophiolite and its sole are commonly deformed but only weakly metamorphic to nonmetamorphic.

A prime example of a Tethyan-type ophiolite is the Semail ophiolite in Oman, which was emplaced, with adjoining oceanic lithosphere, onto the Arabian continent during Late Cretaceous time (Glennie et al. 1973; Searle & Malpas 1980). Beneath the ophiolite is a ≤ 300 meter thick metamorphic sole consisting of upper amphibolite-facies mafic rocks that grade downward into lower amphibolite-facies (Hacker & Mosenfelder 1996; Hacker & Gnos 1997) or greenschist-facies (Ghent & Stout 1981; El-Shazly & Coleman 1990) mafic and quartzose tectonites. The upper, higher-grade portion likely represents oceanic crust, whereas the lower portion includes basalt, shale, and chert protoliths (Searle & Malpas 1980). Directly beneath the sole lie the Hawasina Nappes, a thrust complex of essentially unmetamorphosed Permian-Cretaceous sedimentary and volcanic rocks of the southern Tethyan passive margin (Béchennec et al. 1988). Obduction-related deformation in the Hawasina Nappes is generally characterized by imbricate faulting and open folds with associated fracture cleavage. Only in the lowest structural levels where the nappes experienced very weak (anchizone) metamorphism, are refolded recumbent folds, slaty cleavage, and a weak lineation present (Michard et al. 1984).

A similar analogue is the much older Bay of Islands ophiolite in Newfoundland. The largest Appalachian ophiolite, the ~486 Ma Bay of Islands complex (Dunning & Krogh, 1985) was emplaced onto the Laurentian margin during Ordovician (Taconic) convergence between Laurentia and a volcanic arc complex. A distinct metamorphic sole underlying the ophiolite reaches a maximum thickness of 230 m (Fergusson & Cawood 1995). The most complete section of the sole investigated by Fergusson & Cawood (1995) consists of a few meters of mafic granulite and garnet amphibolite underlain by amphibolite, mafic greenschist, variably deformed pillow lavas, and slaty pelite and psammite at the base. The base of the sole is designated as the contact between slaty rocks and underlying scaly mudstones and/or mélange units.

Taking these observations into account, what predictions can we make regarding the metamorphic character of the Baltoscandian margin in the southern Trondheim region, prior to the Scandian orogeny? Assuming Early Ordovician ophiolite obduction affected portions of the margin, it is plausible that high-grade metamorphism was very localized beneath the ophiolites, whereas the majority of the Baltoscandian margin rocks were only weakly affected. Subsequent tectonic dismemberment of ophiolites and any existing metamorphic soles, along with metamorphic overprinting during Scandian tectonometamorphism, would have led to a fragmented and localized record of the pre-Scandian metamorphic pattern. Therefore, if the interpretation of oceanward-dipping subduction polarity in the Early Ordovician is correct, a model of related ophiolite emplacement does not require that obduction was accompanied by widespread orogenic deformation and high-grade metamorphism.

Conclusions

In south-central Norway, the Einunnfjellet dome exposes the contact between rocks representing strata from the Baltoscandian margin of Baltica (Seve Nappe) and exotic terranes of the Köli Nappes. In a traverse across the study area from structurally higher to structurally lower units, mineral assemblages, garnet-biotite thermometry, and amphibole chemistry record a continual metamorphic gradient from lower greenschist to epidote-amphibolite facies, i.e. a normal metamorphic gradient. Above the garnet isograd, biotite-grade assemblages occur in calcareous and/or carbonaceous slates and phyllites. Below the garnet isograd, garnet-biotite thermometry in pelitic rocks records a continual increase in temperature, from 480°-515°C in structurally higher rocks, to 550°-580°C in structurally lower rocks in the core of the dome. Amphibole compositions in coexisting mafic rocks are consistent with the thermometric results. Although samples from this study lack mineral assemblages appropriate to quantify metamorphic pressures, the amphibole, plagioclase, and muscovite chemistries are indicative of moderate-pressure conditions.

Results of this study indicate that the inferred Seve-Köli thrust contact and overlying thrusts within the Köli were overprinted by greenschist to epidote amphibolite-facies metamorphism, suggesting that nappe assembly was earlier than, or broadly coeval with, the metamorphism. Based upon thermochronologic data available for the central Scandinavian Caledonides, it is suggested that the peak metamorphic assemblages preserved in rocks of this study formed during Scandian tectonothermal activity.

Samples from this study, including those from the Hummelfjell (Seve) Nappe, do not appear to have experienced a complex history involving more than one major metamorphic event. Accordingly, the evidence is inconsistent with the interpretation of a single pre-Scandian widespread, regional metamorphism resulting from Early Ordovician ophiolite obduction, as has been proposed by other workers. However, a model of Early Ordovician, oceanward-directed subduction, leading to obduction of a Tethyan-type ophiolite onto the Baltoscandian margin, does not necessitate that obduction was accompanied by widespread orogenic deformation and high-grade metamorphism, but rather indicates that metamorphic effects could have been restricted to the ophiolite and its underlying metamorphic sole, if present. In such a case, subsequent tectonic dismemberment and metamorphic overprinting during Scandian tectonometamorphism would have led to a fragmented and localized record of any pre-Scandian metamorphic pattern.

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