NGU Report 2008.064
A tectonostratigraphic transect across the
central Scandinavian Caledonides,
Storlien --Trondheim – Lepsøy
Part II. Excursion guide in Norway

REPORT

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Summary:
   This guidebook was prepared for the Norwegian part (seven days) of Excursion #34 of the 33rd IGC, Oslo, 2008, to give participants from outside the region as comprehensive a view as possible within 10 days of the stratigraphy, paleontology, tectonostratigraphy, structural geology, igneous and metamorphic petrology, geochronology, tectonics and geophysics, in a traverse across this major thrust-assembled orogen from the foreland in Sweden to places deep in the hinterland in coastal Norway with evidence of ultra-high-pressure metamorphism and melting.
   The first three days in Sweden (Part I of report 2008.064) were led by David Gee[1] and Per-Gunnar Andréasson[2], with co-leaders Erik Sturkell[3] and Anna Ladenburger[1].
   The last seven days (Part II: this report) were led by Peter Robinson[4] and David Roberts[4] with co-leaders Arne Solli[4], Kurt Hollocher[5], Per Terje Osmundsen[4], Tor Grenne[4], Herman Van Roermund[6], Michael Terry[7] and Hans Vrijmoed[8], with contributions by David Gee, Bob Tucker[9], Tom Krogh[10], Alan Krill[11], Emily Walsh[12] and Megan Regel[12]. It was divided by day as follows:
Day 4: *Trondheim Nappe Complex and subjacent units: Storlien to Trondheim.*
Day 5: *Scandian geology of the Outer Trondheimsfjord region.*
Day 6: *Part I: Geology of the Hølonda-Horg-Løkken area.*
Day 6: *Part II: Geology of Trollheimen.*
Day 7: *Garnet peridotites in the northern UHP domain of the Western Gneiss Region, Otrøy.*
Day 7: *Mafic dikes and basement-cover relationships, southern coast of the islands of Midsund.*
Day 8: *Structural and metamorphic relationships between Caledonian nappes and Fennoscandian basement on the mainland near Brattvåg.*
Day 9: *Geology of Haramsøy, Flemsøy and Lepsøy.*
Day 10: *Geology on the route Lepsøya-Brattvåg-Solholmen-Svartberget-Atlantic Highway-Averøy-Halsa-Trondheim.*

   The introduction to this excursion guide, in Norway, includes a dedication to the late Professor Tom Krogh, who died suddenly in the spring of 2008.
   A full list of references pertaining to the geology of the areas covered by the seven days of excursions in Norway is given at the end of the report.

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Metamorphism	Tectonics	Geochronology
Paleontology	Geochemistry	Ophiolites

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Post-meeting Excursion

A tectonostratigraphic transect across the central Scandinavian Caledonides

Preliminary Printing of Itinerary in Norway

Day 4 - Monday August 18, Storlien, Sweden to Trondheim
Day 5 - Tuesday August 19, Trondheim-Ørlandet-Trondheim
Day 6, I and II – Wednesday August 20, Trondheim to Sunndalsøra
Day 7 – Thursday August 21, Sunndalsøra to Lepsøy
Day 8 – Friday August 22, Lepsøy-Brattvåg-Lepsøy
Day 9 – Saturday August 23, Lepsøy-Flemsøy-Lepsøy
Day 10 – Sunday August 24, Lepsøy to Trondheim

Leaders: Peter Robinson and David Roberts,
Geological Survey of Norway, Trondheim, Norway

Co-leaders: Arne Solli[1], Kurt Hollocher[2], Per Terje Osmundsen[1], Tor Grenne[1], Hermann Van Roermund[3], Michael Terry[4], and Hans Vrijmoed[5]

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DEDICATION

This guidebook is humbly dedicated to our friend and distinguished colleague, Tom Krogh, who passed away rather suddenly on April 29, 2008, while this book was in preparation. Tom had ancestors and relatives in Trondheim, so that he enjoyed visiting here with his wife, Kathy, whenever a valid excuse arose.

The world-class geochronologic science that he develop at the Geophysical Laboratory and later at the Royal Ontario Museum had major consequences for the geology of Norway and the Norwegian Caledonides in particular. In 1974, in collaboration with Bjørn Mysen and G. L. Davies, he determined a Paleozoic age for metamorphic zircon in the Ulsteinvik eclogite, at a time when most workers believed the eclogite metamorphism to be Precambrian and continued in this belief for another 5-6 years. Grains from the same zircon separate, using better developed techniques, were redetermined around 1990 by Bob Tucker with the result 402 +/-2 Ma, and more than a decade later these same zircons were found to have inclusions of coesite.

Subsequent to his early endeavor, Tom had an enormous impact on Norwegian geology through his training at the Museum of other workers, who have made, and continue to have, a crucial impact, including Arne Råheim, Bob Tucker, Rolf-Birger Pedersen, Greg Dunning, and Fernando Corfu. In 1997 Tom came to Trondheim for the COPENA Conference (Correlation of Proterozoic Rocks of Europe and North America). I remember, on the last stop of the one-day Conference Field Trip, how Tom stood on the outcrop of pink, strongly lineated, Paleoproterozoic, Ingdal Granite, of the Baltican basement, and enthralled us concerning the remarkable story of metamorphic titanite. At this time I realized that Tom and I had been students together (he at MIT, I at Harvard) when he "took" J. B. Thompson's Phase Equilibrium Course in fall 1960. At that time I was "observing" the course for the second time, having "taken" in fall 1958. Although he could not attend the four-day pre-conference field trip, he took time afterwards to visit several key localities, including the pegmatite cutting eclogite at Averøy, on which he had been working for a decade, based on a sample collected by Arne Råheim about 1983.

Tom and Kathy returned in late summer 1998, and I had the privilege of traveling with him and learning more about his supreme zircon knowledge and feeling for rocks, while more specimens were collected. These became the basis for a series of key age determinations completed in 2003. These results were featured in the post-meeting field trip associated with the 2003 Alice Wain Symposium at Selje, and are also shown in slightly more complete form for this guidebook, particularly on Days 9 and 10. During the Symposium and after, Tom collected an additional group of key samples, and also, incidentally, gained many new, often much younger, contacts and friends among students of Norwegian eclogites. My last meeting with Tom was at the Goldschmidt Conference in Copenhagen, June 2004. At that time, we were struggling to complete revisions after review for a paper (with no less than eight authors) including a discussion of titanite data from Western Gneiss Region basement. Tom took these problems to heart, sequestered himself in his wilderness cabin near Parry Sound for the better part of a week, and came up with the authentic story concerning the data, which to me, was one of the inspiring moments of the whole process.

Upon learning of Tom's passing a younger colleague, who came to know Tom at Selje, wrote: "I will miss his energy, excitement, friendly nature and his ability to do great science!!!! " It could not have been said better.

Peter Robinson, Trondheim, July 29, 2008

Tom Krogh

On the outcrop beside a kyanite-zoisite eclogite with coesite pseudomorphs at Fjørtoft, July 1, 2003, on the Post-meeting Excursion of the Alice Wain Field Symposium. Tom sampled this outcrop in 1998, both the eclogite and the cross-cutting extensional pegmatite. On the latter he obtained a U-Pb zircon age of 394.5 +/-2 Ma (Fig. 9.23, this guidebook), but on the former, despite his best efforts, he was unable to find any zircon. This is now explained by the very low Zr content, because the protolith was a plagioclase-rich cumulate rock (see Stop 9-7 and Fig. 9-11, this guidebook). Tom's zircon knowledge in the lab was profound, but he was also aware of all aspects in the field and with the use of a heavy hammer, so that a collecting tour with him was an experience never to be forgotten. (Photograph courtesy of Helen Lang)
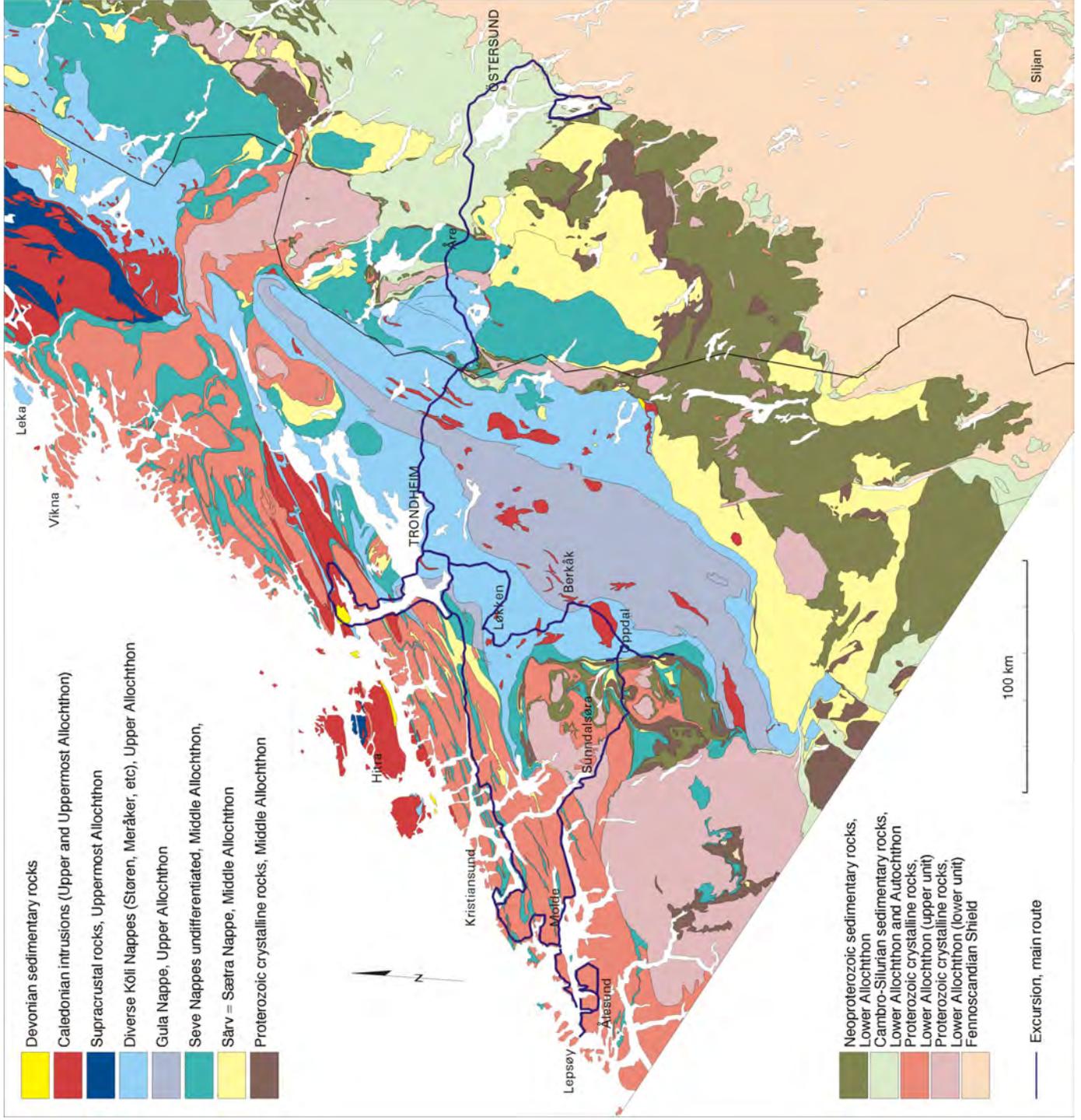
Fig. 0.1
Tectonostratigraphic
map of the central
Scandinavian
Caledonides showing
the general route of 10-
day Post-meeting
Excursion #34.

Devonian sedimentary rocks

Caledonian intrusions (Upper and Uppermost Allochthon)

Supracrustal rocks, Uppermost Allochthon

Diverse Köli Nappes (Støren, Meråker, etc), Upper Allochthon

Gula Nappe, Upper Allochthon

Seve Nappes undifferentiated, Middle Allochthon,

Särv = Sætra Nappe, Middle Allochthon

Proterozoic crystalline rocks, Middle Allochthon

Neoproterozoic sedimentary rocks,
Lower Allochthon

Cambro-Silurian sedimentary rocks,
Lower Allochthon and Autochthon

Proterozoic crystalline rocks,
Lower Allochthon (upper unit)

Proterozoic crystalline rocks,
Lower Allochthon (lower unit)

Fennoscandian Shield

Excursion, main route

100 km

Day 4, Monday, 18 August

Trondheim Nappe Complex and subjacent units: Storlien to Trondheim.

David Roberts
(with contributions by David Gee to Stops 1, 2 and 3, and Peter Robinson to Stop 11)

Summary of route
We leave Åre at 08.00 hours and drive west on the E14 road, arriving at the Customs Station near Storlien just before 09.00 (there is no need to stop here). Proceed to Stop 4.1 (Fig. 4.1). From there, after two more stops just inside Sweden, we will drive west along the wide valley of Teveldalen and, after Stop 4.6, along the narrow, incised valley of Stjørdalen. We continue on the E14 until we reach the E6 road junction on the outskirts of the town of Stjørdal. There, we turn right (north) for stops 4.9 and 4.10 (Fig. 4.1), thereafter returning (south) on the E6 highway, passing Stjørdal, and continuing on to Trondheim. After a short break at the Geological Survey of Norway (NGU) in Lade we visit nearby coastal exposures of higher levels of the Bymarka ophiolite, and then continue to the Trondheim Vandrehjem – arriving at c. 17.45.

Introduction
On this fourth day of the transect we will continue our gradual upward climb in the Caledonide tectonostratigraphy with the focus now shifting to representative rocks of the Köli Nappes as we cross the border from Sweden into Norway. To refresh our memories, however, we begin the day with two stops on the Swedish side, one in the Seve and the other in rocks of the Lower Allochthon at a locality that affords an excellent view to the west, weather permitting, into the Norwegian county of Nord-Trøndelag.

As we approach the border, and especially clear at Stop 2, west of Storlien, we will see that the tectonostratigraphy is deformed across a major, N-S-trending antiform. This structure can be followed over several tens of kilometres, roughly parallel to the Norwegian-Swedish border. Known as the *Skardøra Antiform* (Hurich et al. 1988), it has previously been referred to as the Sylarna or Riksgrense Antiform. This antiformal structure is one of several such open folds that deform the tectonostratigraphy of the Scandinavian Caledonides, and clearly developed during and/or post-dating the principal Scandian thrusting but prior to the extensional deformation that accompanied the late-Scandian collapse of the orogen. Looking southwest from the vantage point of Stop 2, the prominent escarpment of Steinfjellet marks the eastern limit of the Köli rocks of the Trondheim Nappe Complex (TNC), floored by the Steinfjell Thrust which was later reactivated as a normal fault during the late-Scandian extensional deformation (Sjöström & Bergman 1989). This composite, contractional/extensional structure is clearly seen on the seismic reflection profile (Hurich et al. 1988, Palm et al. 1991) (Fig. 4.2), and its mylonitic rocks will be examined briefly at Stop 3. It has been suggested that the late extensional fault at the top of the Steinfjell Thrust may be a continuation of a detachment fault exposed beneath the Devonian rocks at Røragen, 100 km farther south (Norton 1987, Gee et al. 1994). Another contender for the prolongation of the Røragen detachment might be the Kopperå fault, just east of Meråker (Hurich & Roberts 1997, in prep., Mosar 2000), which is also prominent on the seismic reflection imagery (Fig. 4.2).

Thereafter we drive westwards down the valley Teveldalen, initially in low-grade, Lower Ordovician to Llandovery (and possibly Wenlock), turbiditic rocks of the Meråker Nappe, part of the TNC, which we will examine at two stops. At the base of the Meråker Nappe there is a bimodal but largely mafic, amphibolite-facies, magmatic complex (Fundsjø Group) of inferred Cambrian to
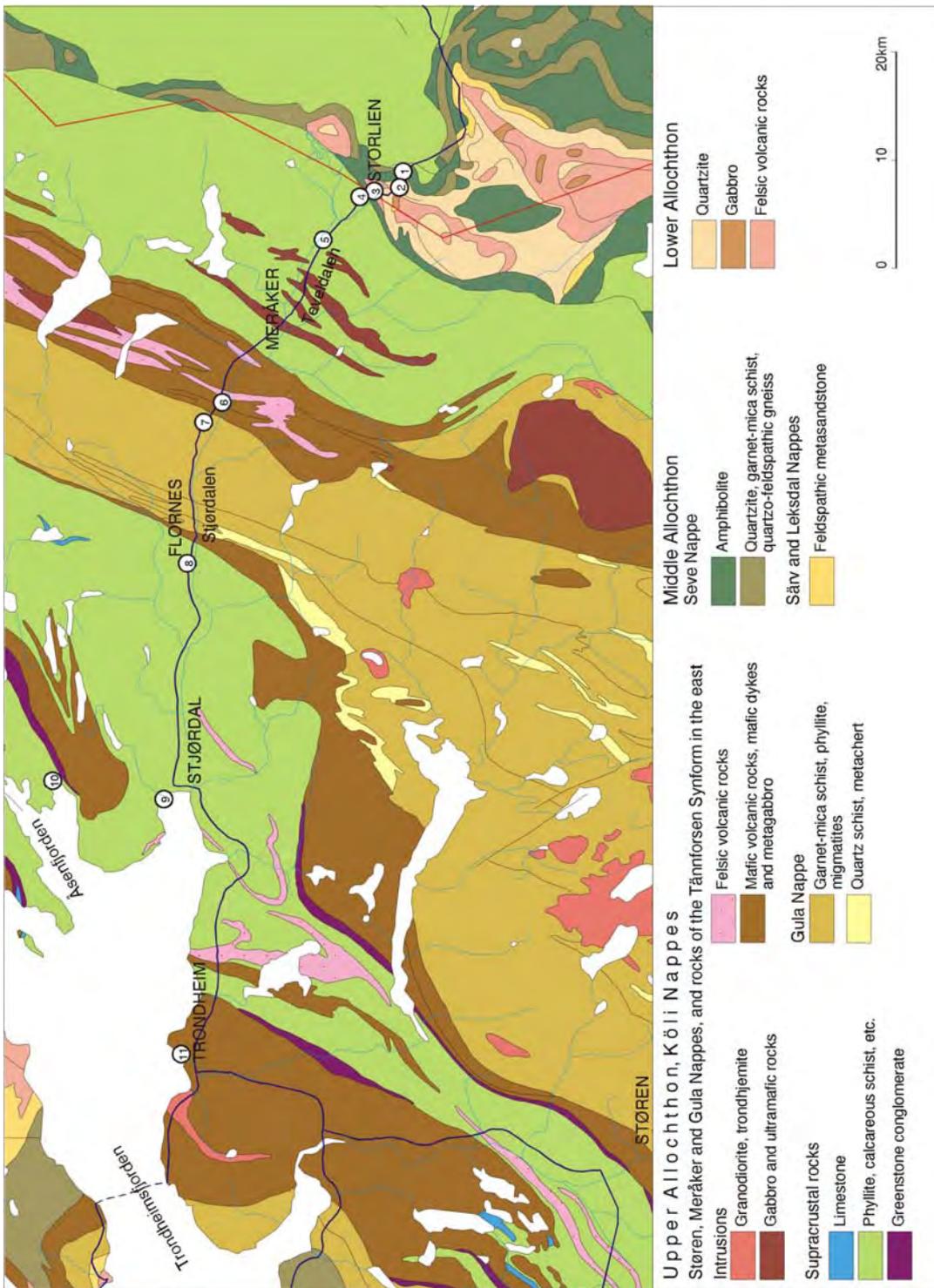
Fig. 4.1. Simplified geological map of the Storlien-Trondheim transect with Day 4 stops marked as encircled numbers.

Tremadoc age. This has provided detritus to a basal conglomerate, above an important unconformity, in the overlying Sulåmo Group (Chaloupsky & Fediuk 1967). Most of the succession in the Meråker Nappe shows westerly dips but the greater part is inverted, except in the east (east of a major syncline exposing the Llandovery-age Slågån Group) where the strata are disposed
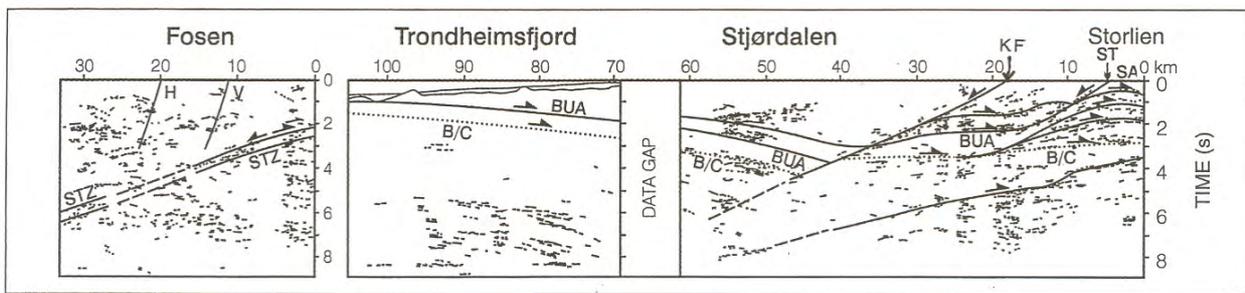
Fig. 4.2. Interpretation of the vibroseis reflection profiles, Storlien (east) to Trondheimsfjord (west), and across Fosen Peninsula (c. NNW-SSE), taken from Hurich & Roberts (1997). B/C – Basement-cover contact, generally tectonised; BUA – Base of Upper Allochthon; H – Hitra-Snåsa Fault; KF – Kopperå Fault; SA – Skardøra Antiform; ST – Steinfjell Thrust; STZ – Stadland shear zone; V – Verran Fault.

normally (Siedlecka 1967). In another interpretation, from an area 30 km farther northeast, inferred younger rocks (?Wenlock, but no fossils) occur in another syncline east of the Slågån Group (Hardenby 1980).

Continuing westwards we pass into the locally higher-grade rocks of the Gula Nappe which are intensely folded in this area and are characterised by a near-vertical foliation. Although the internal structure of the Gula Complex (formerly termed Group) has been controversial, in the far northeast of the Trondheim Region the rocks and the major thrusts define an antiform that is deformed around a younger synformal fold (Roberts 1967). A feature of the Gula Nappe is that it carries evidence of a pervasive, pre-Scandian tectonothermal event. Composite trondhjemite-diorite-gabbro plutons and dykes of Early Silurian age (438-432 Ma) are common and postdate an earlier foliation, inferred to be of Early Ordovician age (Nilsen et al. 2003, 2007). The plutons and dykes were subsequently folded and metamorphosed during the main Scandian event, dated to 430-425 Ma in this particular nappe (e.g., Hacker & Gans 2005).

Farther west along the Stjørdalen valley we encounter low-grade rocks of the Støren Nappe. An important magmatic complex at the base of the nappe is reduced to green phyllonites along this traverse. On the field trip we will see mostly turbiditic metasedimentary rocks of Mid to Late Ordovician age, but at the end of the day we will examine strongly deformed pillow lavas and felsic rocks of the Bymarka ophiolite. Ophiolitic rocks form the basal part of the Støren Nappe over large areas of the western Trondheim Region (Gale & Roberts 1974, Grenne et al. 1980, Heim et al. 1987, Roberts et al. 2002), and plagiogranites and related extrusive rocks in several areas have yielded Late Cambrian (Furongian) to Tremadoc ages. Relics of a blueschist-facies assemblage occur close to the very base of the Bymarka ophiolite (Eide & Lardeaux 2002). The fragmented ophiolites are considered to have been obducted onto rocks now forming the Gula Complex in Early Arenig time, prior to uplift, erosion and deposition of a neo-autochthonous, lower-grade succession (Hovin and Horg groups) of Mid Arenig to Ashgill age. The basal formation is generally a polymict conglomerate (Venna conglomerate), traditionally known as the 'greenstone conglomerate' in older literature, containing many of the foliated rock types reported from the ophiolite. The sedimentary rocks of the Støren Nappe, which are richly fossiliferous in some areas and in specific rock types, and particularly in the area covered by the first part of Day 6, southwest of Trondheim, will be discussed in more detail in the introduction to Day 6.

Stop 4.1: Klevsjön. Seve Nappe. Allow 20 minutes. Leave at 09.25.

Location
Road-cut on north side of E14 road, opposite a large lay-by, c. 800 m west of the Customs (Toll/Tull) Station. Map-sheet 19C Storlien NV, 702885/131315.

Introduction
The Seve Nappe thins rapidly westwards toward the hinge of the Skardøra Antiform, and the upper amphibolite- to granulite-facies units encountered yesterday are completely excised. The retrogressed rocks at this Klevsjön locality become mylonitised farther west (see next stop description) and the Seve Nappe is locally missing in parts of central Norway.

Description
The outcrop provides excellent exposure of variably retrogressed (chloritised and epidotised) amphibolites, schists, calcareous schists and thin calcite marble ribs, all part of the Seve Nappe. The lithologies show a very gentle northeasterly sheet dip at this locality, just east of the crest of the Skardøra Antiform. Small, syn-schistosity, isoclinal folds are present here, with indications of east-southeast vergence. In the western parts of the outcrop there are several low-angle shear surfaces, and amphibolite layers are dissected in lensoid fashion. Quartz veins and segregations are fairly common. Several minor structures show a top-east shear sense, relating to the principal stage of thrusting (here toward ESE, marked by a pervasive lineation) during the Scandian orogeny. However, there are also a few, east-verging, small-scale, asymmetric folds that deform the schistosity, as well as low-angle, west-dipping and high-angle, east-dipping normal faults, all of which are indicative of overall post-thrusting extensional deformation (see also Sjöström & Bergman 1989).

Along the road some 300 m west Klevsjön, north of a small lake, there are psammites and greenschists which are thought to represent retrogressed rocks of the Särv Nappe. These are underlain by phyllitic schists and quartzites of the Lower Allochthon. Särv Nappe metasandstones cut by dolerite dykes reappear in a tectonic lens in the western limb of the major Skardöra Antiform, some 16 km south-southwest of Storlien. As we will see later, comparable psammites with metadolerite or amphibolite dykes occur in lenses at the same structural level in different parts of central Norway.

Stop 4.2: Brudslöjan. Lower Allochthon of the Skardøra Antiform (Sylarna Window). Allow 25 minutes. Leave at 10.00.

Location
Bend on road E14. Road-cut on north side. Small hill on south side c. 50 m from the lay-by; can be reached by a small path. Map-sheet 19C Storlien NV 70298/131240.

Introduction
In this general area, in the core of this antiformal structure, the Lower Allochthon is composed of porphyritic rhyolites, metagranites, quartzites and quartzo-feldspathic schists with thin phyllites. Thin flagstone units have also been reported, akin to lithologies in the Offerdal Nappe. The rhyolites and metagranites are inferred to be part of the 1.85-1.63 Ma Transscandinavian Igneous Belt.

Description
At this locality, quartzo-feldspathic rocks show a penetrative foliation dipping at c. 30° north. Rare isoclinal folds are present and there is a marked stretching lineation trending WNW-ESE. Looking southwest, we have an excellent view of the combined thrust front and extensional detachment fault of the Meråker Nappe of the Trondheim Nappe Complex, on the east-facing escarpment below Steinfjellet (Fig. 4.3). A thin, attenuated part of the Seve Nappe is sandwiched between the Meråker Nappe and the Lower Allochthon, but is mostly covered by scree.

Fig. 4.3. The eastern thrust contact of the Meråker Nappe, Trondheim Nappe Complex, in the scarp below Steinfjellet (909 m a.s.l.), c. 1 km west of the Swedish border, looking southwest. A thin, attenuated slice of Seve Nappe rocks, mostly hidden by scree, overlies felsic volcanites and quartzites of the Lower Allochthon. The Steinfjell thrust was reactivated during late-Scandian, top-west, extensional deformation, and possibly coincides with a northward prolongation, or splay, of the Røragen detachment. It is also inferred to have been active again during the Late Palaeozoic crustal extension that created rift basins in the Norwegian Sea.

Stop 4.3: Stor-Helvetet. Complex mylonite zone, Lower to Upper Allochthon. Allow 25 minutes. Leave at 10.30.

Location
Long, deep road-cut on E14 road, 200-800 m east of the international border. Difficult parking here for a fleet of cars, so we will park just beyond the downhill (northern) end of the road-cut and walk back to examine just the lowest section. Note – keep a close eye on the traffic here, especially in the fast downhill lane.
Map-sheet 19C Storlien, 703030/131250.

Introduction
This 600 m-long road-cut exposes a variety of protomylonites and mylonites, ranging from foliated rocks of the Lower Allochthon at the southern end, through strongly retrogressed representatives of the Seve Nappe (local name, Essandsjø Nappe) and then into pervasively mylonitised rocks, possibly Köli protoliths, at the northern end. This marks the base of the Trondheim Nappe Complex (Meråker Nappe), as on Steinfjellet. This major Scandian thrust (Törnebohm 1896, Roberts 1967, Wolff 1967, Gee 1975) was later reactivated during late-Scandian (Devonian), or younger, extensional deformation (Sjöström & Bergman 1989, Hurich & Roberts 1997), and is clearly imaged on seismic reflection profiles (Fig. 4.2) (Hurich et al. 1988, Palm et al. 1991).

Description
The northernmost parts of the road-cut, on both sides of the road, expose pervasively mylonitised, dark grey rocks of uncertain parentage, but scattered lensoid relics of 'garbenschiefer' are suggestive of Köli affiliation. Pods and lenticular segregations of quartz are common. Farther south (uphill), porphyroclastic schists, mylo-amphibolites and some thin, marble ribs are reminiscent of lithologies of the Seve Nappe, and rare, detached isoclinal folds are present. The WNW-ESE lineation seen at Stops 1 and 2 is also visible. Higher up (but down tectonostratigraphically) we encounter white, blue-grey and rusty quartzites and phyllonites of the Lower Allochthon, below which there are foliated felsic volcanites similar to those at Stop 2.

        We are now located on the western limb of the Skardøra Antiform and it is possible to see, in these road-cuts, that 'late' normal faults and other small-scale structures are pointing to top-west/northwest extensional movement, indicating a reversal of shear sense along the original thrust fault. Such structures, including shear bands (termed *strain-slip cleavages* by Roberts 1967) that deform the earlier, ductile, mylonitic foliation, have been described by Sjöström & Bergman (1989). By comparison with 'late', brittle, normal faults that have reactivated detachment faults or shear zones elsewhere in Norway (e.g., Braathen et al. 2002, Osmundsen et al. 2003), the steeper dipping normal faults at this Stor-Helvetet locality could conceivably be of Carboniferous (Eide et al. 2002), Permo-Triassic or younger age (Fossen 2000, Mosar 2000).

        Should we have time to view the deforested, unguarded, international border separating Sweden and Norway, a small road-cut there (now above the mylonite zone, in the Meråker Nappe) exposes brittly deformed, low-grade metagraywackes and pelites of the Kjølhaugen Group (Siedlecka 1967). Do not spend much time here -- there are far better exposures to be seen at the next stop.

Stop 4.4: E14 road-cut, 200 m west of the border. Turbidites, Kjølhaugen Group, Meråker Nappe. Allow 20 minutes. Leave at 10.50.

Location
Road-cuts along north side of the E14 road, close to a 'Radio 90.5' signpost. This is roughly 200 m west of the Norwegian-Swedish border. With several vehicles involved, it may be best to park on the south side of the road, near a '16 km' signpost. One or two cars can just about squeeze off the road to the right (north).
Map-sheet Meråker 1721-1, UTM coordinates 33VUL525/266.

Introduction
Here, and over a distance of 4-5 kilometres to the west, there are excellent road-cuts exposing low-grade turbiditic greywackes and phyllitic claystones (Siedlecka 1967). The strata dip consistently at 30-50° to the northwest, with graded bedding and a variety of sole structures showing the succession to be the right way up, at least as far west as the hinge zone of the Teveldal syncline. The turbidites in this area form part of the Kjølhaugen Group (Siedlecka 1967, Wolff 1973). Based on mapping in another area, some 30 km to the northeast, Hardenby (1980) reinterpreted these particular turbidites (his Liafjellet Group) as possibly being the youngest (?Wenlock) rocks in the Meråker Nappe.

Description
The northwest-dipping, medium-bedded strata here consist essentially of greenschist-facies, brownish-grey, calcareous greywackes and thinner interbeds of cleaved phyllites. Graded bedding shows that the rocks are younging normally, i.e., they are the right way up. The pelitic interbeds in this road-cut (and elsewhere within the Meråker Nappe) show a pronounced, fairly flat-lying cleavage, of crenulation or spaced type, which gradually refracts into a SE-dipping, steeper attitude in the greywacke beds (Fig. 4.4).

Fig. 4.4. Turbiditic, calcareous greywackes and pelites of the Kjølhaugen Group with a prominent, flat-lying, S3 crenulation cleavage in the pelitic layers, refracting into a steeper attitude in the graded-bedded greywackes. Looking c. northeast.

In some of the pelitic units, thin sandy ribs are offset by a centimetre or more in a reverse sense (top-NW) along some of the S3 cleavage surfaces. A feature of the phyllite beds in many parts of the Meråker Nappe is the presence of tiny (2-4 mm) porphyroblasts of biotite which postdate the main S1 schistosity but predate the S3 cleavage (Roberts 1967). Porphyoblasts of pyrite are also present locally. Towards the base of the nappe, the biotites are chloritised and the pyrites partly converted to hematite and limonite.

On first sight, the bedding/cleavage relationship in these outcrops might seem to contradict the sedimentological evidence and point to inversion. However, the cleavage is a comparatively young structure (S3 of Roberts 1967) and is axial planar to many mesoscopic and small-scale F3 folds (not seen here, at Stop 4) which consistently verge down-dip across the entire Trondheim Region, irrespective of pre-F3 stratal dip. These features, and especially the ubiquitous, fairly flat-lying cleavage, led Roberts (1967, 1969a, 1971, 1979) into interpreting this late deformation phase as relating to a gravity-induced collapse and spreading of the stacked allochthons. The true significance of this collapse came later, with the discovery of major extensional detachment faults in western Norway (Hossack 1984, Norton 1986, 1987, Séranne & Séguret 1987) and subsequently in other areas (Fossen & Rykkelid 1992, Séranne 1992, Gee et al. 1994, Braathen et al. 2000, 2002, Nordgulen et al. 2002, Osmundsen et al. 2003, 2006).

A feature of the Kjølhaugen Group turbidites in this general area – but not along the E14 road traverse – is the presence of polymict conglomerates. Both clast-supported and matrix-supported types are present. Siedlecka (1967) interpreted the conglomerates as submarine-fan accumulations, with paleocurrent data suggesting a source area to the *east*; an interpretation also favoured by Hardenby (1980) for his Liafjellet Group turbidites.

Stop 4.5: Fjelldalsbekken, Teveldalen. Kjølhaugen and Slågan groups. Metagabbro in the Kjølhaugen Group. Allow 20 minutes. Leave at 11.20.

Location
Road exposures east and west of a gravel lay-by on the north side of the E14 road, next to the river Tevla. Map-sheet Meråker 1721-1, UTM coordinates 32VPR475/300.

Introduction
We have now passed across a major, tight syncline (Teveldal syncline) containing the Slågån Group in its core. Based on an early find of graptolites some 20 km farther northeast (Getz 1890), these mainly pelitic Slågån rocks (exposed here at Stop 5) were accorded an Early Silurian (Llandovery) age. Kjølhaugen Group rocks, in a road-cut on the south side of the main road, reappear here in the western limb of the syncline in an inverted position. They are cut by a body of metagabbro, one of many lensoid gabbro sheets that intrude the succession in this part of the Meråker Nappe.

Description
On the south side of the road, Kjølhaugen Group greywackes and phyllites are intruded by a c. 20 m-thick body of saussuritised metagabbro. Sedimentary structures in the west-dipping strata indicate that the succession is inverted, younging east toward the Slågån Group rocks. Although igneous textures in the gabbro are fairly well preserved, the mineralogy is dominated by hornblende and strongly saussuritised plagioclase. In some of these gabbros, chalcopyrite, pyrite and pyrrhotite are common, as are epidote segregations. The gabbro here shows chilled margins against the sediments. Hornfelses have been recorded adjacent to many of the thicker bodies of gabbro. The many dozen bodies of metagabbro in this area, reaching several kilometres in length and up to 500 m in thickness, have been interpreted as concordant or transgressive sills (Chaloupsky & Fediuk 1967, Siedlecka 1967). Although none of these bodies has yet been dated, since they intrude the Kjølhaugen Group but, apparently, not the Slågån, they are presumably of latest Ordovician to Llandovery age and thus comparable, agewise, to gabbros elsewhere in the Meråker and Gula nappes. The major, mafic, Fongen-Hyllingen Complex c. 30 km SW of here, for example, has an emplacement age (U-Pb zircon) of 438 ± 2 Ma (Nilsen et al. 2007).

   On the river bank, there are thin-bedded, dark grey to black phyllites with carbonate concretions, typical of the Slågån Group. Such graphitic and grey phyllites are also exposed a hundred metres or so farther east in road-cuts and along the river banks. There, the early S1 schistosity and, locally, a steep S2 cleavage can be seen to be transected by the flat-lying, late crenulation cleavage, S3.

   There is a c. 20 minute drive to the next locality at Gudå bridge, c. 7 km west of Meråker. On the way we will pass across a metabasalt unit, the Turifoss greenstone, situated at the top of the Sulåmo Group, at the bridge crossing the Tevla river just before the minor road signposted to Kopperå.

Stop 4.6: Gudå Bridge. Magmatic rocks of the Fundsjø Group, Meråker Nappe. Allow 15 minutes. Leave at 11.50.

Location
High road-cut on north side of the E14 road, immediately after (west of) the bridge crossing the railway line, near the small settlement of Gudå; c. 7 km west of Meråker.
Map-sheet Flornes 1721-4, UTM coordinates 32VPR3085/378.

Introduction
The varied amphibolite-facies magmatic rocks of the Fundsjø Group are best exposed above the treeline, notably at and above the top of the skilift (not operational during the summer) on Kirkebyfjellet some 3 km east of here. There, schistose basaltic lavas, and mafic and felsic tuffs, are cut by plagiogranites and younger dolerite dykes. The road-cut near Guda, while not ideal, affords the best outcrop of this complex along the E14 road traverse.

Description
The rocks exposed here are mainly amphibolitised greenstones (i.e., metabasaltic lavas), with some thin units of felsic volcanites, known in Norway as quartz keratophyres. There are also concordant sheets of porphyritic metadolerite. Elsewhere in the Fundsjø Group, such dolerite dykes and sills are seen to cut a pre-existing foliation and folds in the bimodal volcanite assemblage, and there are also some foliated plagiogranites present (Lagerblad 1983, Grenne & Lagerblad 1985). The dolerites are themselves sheared and boudinaged. Such porphyritic dolerite dykes also occur in the overlying Sulåmo Group. In the 1970s, when this road-cut was new and fresh, isoclinal folds could be detected here in the steeply dipping foliation and layering, but they are virtually invisible today. A steep stretching lineation is also present.

The earlier structures in these Fundsjø rocks, including isoclinal folds, are similar to those in the nearby Gula Complex (Roberts 1967, Lagerblad 1983). A study of the porphyritic dolerites by Grenne & Lagerblad (1985) showed that they had a common geochemical signature in both the Fundsjø and the Sulåmo groups. A major conclusion by these authors was that the rocks of the Fundsjø Group were juxtaposed against the Gula Complex, and deformed and metamorphosed, in Early Ordovician time. The only reliable isotopic age from the Fundsjø comes from the southern Trondheim Region, a U-Pb zircon age of 488 ± 2 Ma from a foliated plagiogranite (Bjerkgård & Bjørlykke 1994).

Stop 4.7: Reinåbølet, c. 5 km WNW of Gudå. Schists and pegmatite, Gula Complex, part of the Gula Nappe. Allow 20 minutes. Leave at 12.15.

Location
Road-cut on north side of the E14 road. There is a large lay-by and picnic spot (+toilet) about 100 m east of here between the road and the river, which provides suitable parking for a fleet of cars. Map-sheet Flornes 1721-4, UTM coordinates 32VPR2665/395.

Introduction
Across the Trondheim Region, the rocks of the Gula Nappe vary from low grade to high grade. Low-grade rocks include a mélange with dark-grey phyllitic matrix (Horne 1979) and black slates containing the Tremadoc, Baltican graptolite *Rhabdinopora flabelliforme* (Vogt 1889, Vogt 1941), the latter showing an unusual trace-element chemistry akin to the that of the Alum Shales of the Oslo Region and the Baltic states (Gee 1981). Along our traverse, garnetiferous schists prevail, but south of the river Stjørdalselva, and farther south towards Tydalen, there are kyanite- and sillimanite-bearing rocks (Roberts 1968a, Olesen et al. 1973), partly with migmatites.
The Gula Complex along the valley Stjørdalen is characterised by a steep to vertical foliation and transposed layering, with a steep (in places vertical) stretching lineation. Interpretations of the major structure in this central area of the Gula Nappe have varied greatly over the last 130 years, although most geologists have preferred variations on the antiform theme (reviewed in Roberts 1967, fig. 43). This also fits best with what we know of the map picture in the far northeast of the Trondheim Region, south of the Grong-Olden Culmination.

Description
Garnetiferous schists and fine-grained gneisses are exposed here on near-vertical, glacially smoothed surfaces. Early (F1) isoclinal folds were clearly visible in and above a 2 m-wide and 2 m-deep Pleistocene pothole some 30 years ago, but the pothole has since been filled in by a landslide. A near-vertical stretching lineation can be detected on schistosity surfaces. It is noteworthy that F3 folds and associated S3 cleavage are absent throughout this belt of near-vertical foliation. The reasons for this, and the pattern of development of the F3 collapse folds across the region, are discussed in Roberts (1971).
Veins and lensoid bodies of trondhjemite-pegmatite are present – now partly covered and overgrown – with the following minerals: quartz, plagioclase, muscovite, garnet, tourmaline and apatite. Some 150-200 m east of here there is an exposure of pale grey migmatitic gneiss with amphibolite layers.
A K/Ar analysis of muscovite from the main pegmatite at this locality yielded an 'age' of 412 Ma (Wilson et al. 1973). More reliable ^{40}Ar/^{39}Ar mineral analyses from rocks of the Gula Complex have shown that Scandian metamorphism reached its peak in this nappe during the period 430-425 Ma (Dallmeyer 1990, Hacker & Gans 2005).

Stop 4.8: Flornes bridge. Low-grade metasedimentary rocks of the Lower Hovin Group, Støren Nappe; refolded folds. Allow 15 minutes. Leave at 12.40.

Location
Long road-cut on north side of the E14 road, close to Flornes bridge. Map-sheet Flornes 1721-4, UTM coordinates 32VPR1725/393.

Introduction
We have driven through the steep-sided valley represented by the steeply foliated Gula Complex and across an attenuated, phyllonitic greenstone, all that remains here of a fragmented ophiolite at the base of the Støren Nappe. We will see the ophiolitic rocks later today and/or on Day 6. Note that the dip of bedding and earliest schistosity, S1, from here on towards the coast near the town of Stjørdal is almost consistently eastward – opposite to that in the Meråker Nappe.

Description
The long road-cut close to the bridge to the small settlement Flornes exposes low-grade, thin-bedded, grey, silty sandstones and phyllites of the Lower Hovin Group, part of the Støren Nappe. Although no fossils have been found here, correlation along strike to the southwest indicates that these rocks are of Late Arenig to Llanvirn age. The strata dip steeply to the ESE but are younging to the WNW, as seen from sporadic graded bedding. Stratigraphically older sedimentary rocks just east of here are more green-grey in colour, reflecting a detrital input from the eroding ophiolitic rocks. The basal 'greenstone conglomerate' has not been recorded along the Stjørdalen valley, but reappears a few kilometres to the southwest.

   The prominent feature at this stop is that of many chevron or 'zigzag' F3 folds, with a variably developed, flat-lying, crenulation cleavage, S3. Close inspection will reveal several, tight to isoclinal, syn-schistosity F1 folds with moderately steep NE axial plunges (also seen as a bedding/schistosity intersection lineation). Detailed mapping and structural work has shown that mesoscopic F1 folds point to the presence of an overturned syncline to the WNW.

   Between here and the coast, stratal dip is less steep and mesoscopic, asymmetric, F3 folds are consistently verging down-dip. In some areas the flat-lying S3 cleavage is particularly penetrative. We will be moving up in the lithostratigraphy into an area of complicated refolded folds.

   There is a c. 25 to 30-minute drive to the next stop, depending on the traffic situation around Stjørdal (new road constructions). Where the E14 meets the E6, turn right (north), then after c. 400 m turn left at a temporary roundabout, signposted Havna (harbour); continue for 1 km to a bus-shelter (named 'Storvika') and then follow the signposts along a bumpy dirt road to Storvika.

Stop 4.9: Storvika-Havna road, west of Stjørdal. Upper Hovin Group turbidites, Støren Nappe; primary and tectonic structures. Lunch break c. 30 minutes. Allow 40 minutes on the outcrops. Leave at 14.15.

Location
Start at Storvika -- a local bathing beach and picnic area, and eminently suitable for our lunch stop -- and walk eastwards for c. 500 metres on or above wave-washed exposures to an old quarried rock-face close to the road on the breakwater leading to the harbour. Hopefully, the vehicles can meet us at the quarry car park. Some of the best outcrops can be seen only at low tide. Map-sheet Stjørdal 1621-1, UTM coordinates 32VNR937/3945 to 9415/3935.

Introduction
At this coastal locality we have passed up into the Upper Hovin Group, in an area dominated by a thick turbiditic succession of discontinuous fan-conglomerates and banded greywacke-phyllite units. Structurally, this shore section is located in the lower limb of an early (F1) recumbent syncline (Skatval syncline; Roberts 1968b) which is deformed by two, upright, open to tight, F2

antiforms and a complementary synform (Fig. 4.5), the latter passing through the greenstones that crop out, in inverted position, on Forbordfjellet (590 m a.s.l.) 7 km north of here.
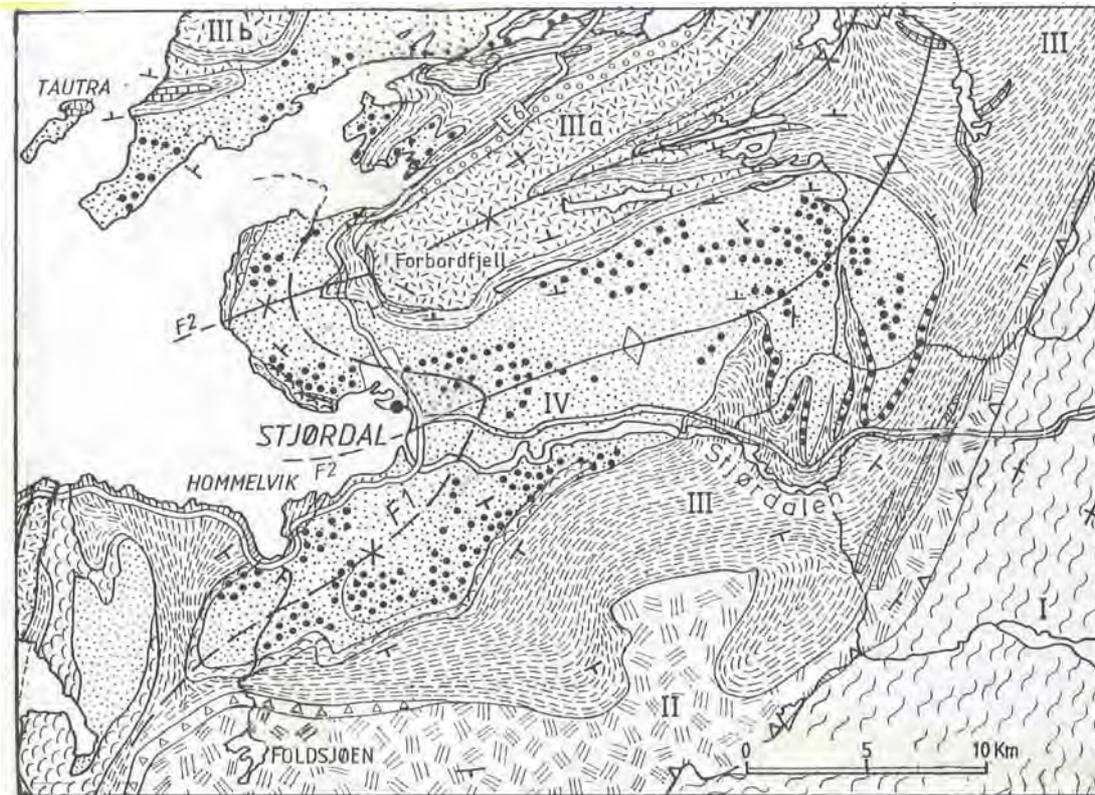

Fig. 4.5. Geological map of the Stjørdal-Forbordfjell district showing the axial traces of principal F1 and F2 folds; modified from Roberts (1968a). I – Gula Complex (Nappe); II – Støren ophiolite; III – Lower Hovin Group (triangles, Venna conglomerate; rings, Stokkvola conglomerate; brick ornament, limestones); IIIa – Forbordfjell greenstone; IIIb – Frosta greenstone; IV – Upper Hovin Group (large dots, main fan-conglomerate units).

Description

    At Storvika and along the shore exposures to the east, thin- to medium-bedded, calcareous turbidites with diverse penecontemporaneous deformation structures may be observed, interbedded with erosively based, massive greywacke-sandstones, lenticular clast-supported conglomerates and chaotic, pebbly, silty greywackes. Rapid facies changes are common, and submarine erosional channels can be seen in a few places. Partly detached slump folds of several metres amplitude are also present (Roberts 1972) (Fig. 4.6), and there is one clastic megadyke comprising a chaotic mélange of deformed strata and small blocks (Fig. 4.7). This association of facies is suggestive of accumulation on a large submarine fan, the coarse clastics probably being representative of channel-fill in inner-fan and/or suprafan lobes on a mid fan (A.Siedlecka, pers. comm. 1979; also Siedlecka 1967). Similar but more regularly banded turbidites a few kilometres north and northeast of here have yielded a deep-water, distal fan, *Nereites* trace fossil association (Roberts 1969b, 1984, Uchman et al. 2005).

        Along this shoreline traverse, F3 folds are common and there are a few F1 folds, the former with their characteristic, flat-lying, spaced S3 cleavage and attendant quartz-calcite veining. The vein history is complex (and not yet studied in detail), also with syn-F1 and post-F3 quartz-calcite veins.

Fig. 4.6. Slump
folds and disrupted
bedding in
calcareous pebbly
turbidites, Upper
Hovin Group,
shoreline east of
Storvika; looking
c. north.

Fig. 4.7. Clastic megadyke
transecting well bedded, calcareous
turbidites, Upper Hovin Group,
shoreline east of Storvika; looking
north-northeast. The dyke is
inferred to have intruded as a
'porridge' of water-charged, poorly
consolidated (thixotropic) sediment
that penetrated a syn-sedimentary
fault on the seabed. A seismic event
may have triggered the process of
dyke intrusion. Smaller examples of
such clastic dykes have been
observed elsewhere in this region.

Stop 4.10: Blekkpynten, Fættenfjorden. Turbidites of the highest part of the Lower Hovin Group, and the Volla conglomerate (Upper Hovin or Ekne Group). Allow about 30 minutes. Leave at 15.30.

Location
At and close to the end of the promontory Blekkpynten, Fættenfjorden. This is about 20 km northeast from Storvika (Stop 4.9). Drive north along the E6 to the farm Fætta, also called Paradisbukta on some maps (3 km NE of Langstein railway station). At Fætta (a camping site), turn left, and sharp left again, driving gradually uphill on a narrow dirt road (marked 'Private') for c. 1.5 km, to a new metal gate. Park there and walk c. 600 m downhill to the sheds and boat-mooring pontoon – or drive if the gate has been left open. Map-sheet Frosta 1622-2, UTM coordinates 32VNR9480/4935.
Note – this can be regarded as an extra stop, as it will take c. 2 hours, driving there and back from Storvika, walking to the outcrops, and studying the sedimentary and tectonic structures.

Introduction
At this locality we see some of the features already seen at Stop 4.9, including spectacular, smaller-scale examples of syn-sedimentary slump folds in banded
turbidites, and there is the added attraction of a superbly exposed conglomerate at the tip of the peninsula (Fig. 4.8). This is the Hopla conglomerate, the base of the Upper Hovin Group (also called the Ekne Group in this part of the Trondheim Region).

Fig. 4.8. Contact between calcareous turbidites of the Lower Hovin Group (to the right) and the polymict Hopla conglomerate (base of the Upper Hovin or Horg Group), Bekkpynten, Fættenfjord in Åsenfjorden. Looking northeast.

From here, we can see Forbordfjellet, underlain by 'greenstones', to the south-southwest. Along the E6, directly south of here, we also have exposures of the Stokkvola conglomerate (Roberts 1975) (shown in Fig. 4.5), another polymict conglomerate of uncertain age but possibly Late Llanvirn judging by an occurrence of Caradoc fossils at nearby Åsen. It was along the track close to Blekkpynten that some of the first trace fossils were discovered in rocks of the Trondheim Region (Roberts 1969b), but sadly the best exposures were blasted away a few years ago during the construction of a driveway to a new cottage.

Description
Sedimentary features in the mainly thin-bedded, calcareous greywackes and intercalated cleaved shales, or phyllites, are best seen along the rocky shore exposures between the boat-mooring facility and the end of the peninsula at Blekkpynten. For trace fossil enthusiasts, one remaining example of a freely meandering trail, in concave hyporelief, is exposed on a bedding surface next to the track behind the new shed. Details of the ichnocoenoses and facies, as well as new finds of trace fossils in the region, the reader is referred to the review by Uchman et al. (2005).

Penecontemporaneous slump folds within particular beds, both underlain and overlain by undisturbed banded turbidites, are common in many parts of the succession in this area, but some good examples are located just above high-tide level below the outdoor toilet at the end of the track, and below the walled 'garden' of the adjacent cottage (Fig. 4.9).

Fig. 4.9. Penecontemporaneous slump fold in an otherwise undisturbed, bedded turbidite succession, Blekkpynten; looking c. southeast.

The turbiditic strata are younging to the northwest, and are overlain by the polymict Hopla conglomerate (Fig. 4.8). At Blekkpynten, this is a mainly matrix-supported conglomerate which here and there contains large fragments of bedded turbidites torn from the substrate or sidewalls of the submarine canyon. Details of the sedimentology, petrography and sequence-stratigraphical analysis of the succession in this Åsenfjord-Frosta district have been presented by Pedersen (1981). Recordings of paleocurrent flow over a wide region have indicated transport from both NW and SE into a back-arc basin system with axial flow from NE to SW (Pedersen 1981) (Fig. 4.10). Moreover, clast material derived from the northwest was predominantly volcanogenic with scattered blocks of island-fringing limestones, whereas material entering the basin from the southeast was largely of continental character (Pedersen 1981, Roberts et al. 1984).

The area and the outcrops on this peninsula are also interesting in showing a prominent steeply dipping, early (slaty) cleavage which is axial planar to tight to near-isoclinal F1 folds (Roberts 1985), which relate to the Skatval syncline. A small example of such a fold can be seen in a small recess in the wall of the track leading to the cottage on the point. Small-scale F3 folds are also present with the associated flat-lying, S3, crenulation cleavage. In some places where bedding surfaces are exposed, these structures appear almost like ripple marks, but on close inspection it is clear that they are tectonic.
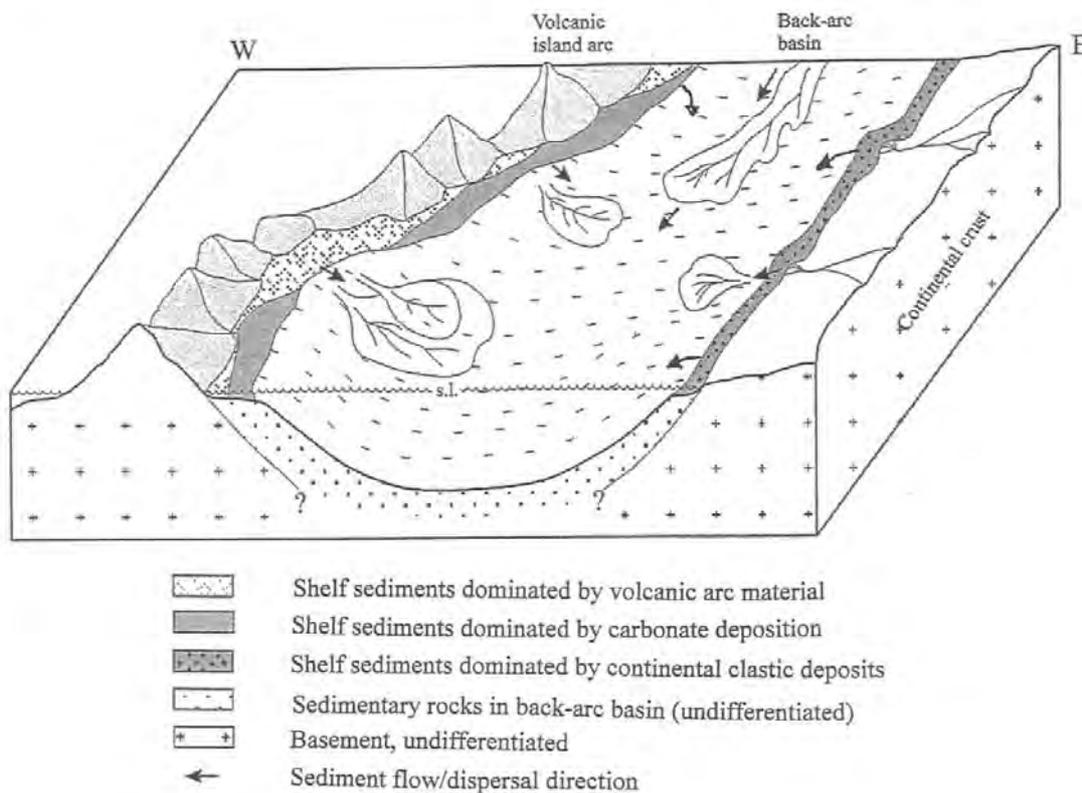
Fig. 4.10. Palaeogeographical model for the depositional basin in the Åsenfjord-Frosta region in Mid to Late Ordovician time. This depicts an island arc source, with fringing, shallow-marine, carbonate rocks, for debris shedding into the basin from the 'northwest' via channels and submarine fans, whilst material sourced to the 'southeast' is largely of continental (non-volcanic) character. The basin became narrower and shallower to the 'northeast'. A subduction zone is inferred to have lain to the northwest of the arc at this time. Modified from Pedersen (1981) and Uchman et al. (2005).

From this stop, we will return to the E6 road and drive to Trondheim, a journey of about 50 minutes. Follow directions to the suburb of Lade, aiming for NGU – the Geological Survey of Norway. There we can use the facilities and perhaps visit the Goldschmidt exhibit (15 minutes). Leave NGU at about 16.40. There will be one more stop on the shore near NGU (4.11, below), before we proceed to the Trondheim Vandrehjem, where participants will stay for two nights.

Stop 4.11: Korsvika, Lade peninsula, Trondheim. Bimodal magmatism, Bymarka ophiolite, Støren Nappe. Allow 40 minutes. Leave at c. 17.30.

Location
Coastal exposures at Korsvika, c. 800 m southwest from NGU (along Leiv Eirikssons vei). There is a small car-park just above Korsvika. Three brief stops, A, B and C, over a stretch of c. 500m. We may omit one if time is short. Map-sheet Trondheim 1621-4, UTM coordinates 32VNR716/367 to 713/364.

Introduction
The Bymarka ophiolite underlies the city of Trondheim, beneath Quaternary deposits, extending from its tectonic base on the Byneset Peninsula in the west to a few kilometres to the east of the Lade peninsula (Fig. 4.1; also Fig. 6.1). Gabbroic rocks predominate in the lowest parts of the pseudostratigraphy in the west, giving way eastwards to a mafic dyke complex and then to pillowed

metabasalts with sporadic metarhyodacites, hyaloclastites, jasper beds and rare coticules (Slagstad 1998, 2003). The highest unit is dominated by mafic and felsic agglomerates with thin magnetite-rich cherts (known locally as 'blue quartz'), jaspers, felsic volcanites and tuffites. and there are rare intercalations of pelitic schists and marble. U-Pb zircon dating of a low-K rhyodacite and an arc-related trondhjemite from the coastal area west of Trondheim have given ages of 482 ± 5 and c. 481 Ma, respectively (Roberts et al. 2002), interpreted as crystallisation ages. As noted earlier, in the main introduction to this day, relics of a blueschist-facies paragenesis have been reported from the basal part of the ophiolite (Eide & Lardeaux 2002). In general, the rocks of the ophiolite complex dip at moderate angles to the east or southeast and are overlain unconformably by a low-grade volcanosedimentary succession equivalent to the Lower Hovin Group. Just south of Trondheim, the Bymarka ophiolite passes directly into the Vassfjellet ophiolite (Grenne et al. 1980), a plagiogranite dyke from which has yielded a U-Pb zircon age of 480 ± 4 Ma (Dunning 1987).

An historical note
The Lade peninsula is steeped in history and was the seat of the powerful 'Ladejarlene' (Earls of Lade) from late in the 9th century. At that time, Korsvika (the bay of the cross) was an important harbour, and was the site of launching of what was supposedly the largest Viking boat ever built (named the *Ormen Lange*, the 'Long Serpent'), by King Olav Tryggvason, who founded the city of Trondheim (then called Nidaros, named after the mouth of the river Nid), in 997 A.D. From that time up to 1217, Nidaros was the capital of Norway and the throne of many of the early kings. In the year 1000, a visitor from Greenland, Leiv Eiriksson (Leif Ericson), was urged by Tryggvason to 'Christianise' his country, but on sailing back to Greenland Eiriksson's boat went off course and he thus accidently discovered Vinland and the new world. In the Middle Ages, the Nidaros Cathedral was the northernmost of four major goals of Christian pilgrimage, with many thousands of pilgrims visiting the tomb of St. Olav each year (named after the religious crusader, Olav Haraldsson, who became king in 1015). The pilgrimage route is still in use today, but only a few hundred pilgrims make the 640 km-long journey from Oslo each year.

Description

Stop A: From the car-park, walk northeast across grassy fields, and the mini football pitch (known as 'Wembley' to the local kids), to the old quarry beneath the hill Kjerringberget. The rock here was quarried by the Germans during World War II with the aim of constructing a breakwater across Korsvika, but the project was abandoned at an early stage.

At the old quarry we see a foliated, medium-grained, metamorphosed trondhjemite with a NE-plunging lineation. The sheared contact with strongly deformed metabasalts and tuffs can be seen up to the right. In this general area the trondhjemite is rich in chlorite, and epidote veining is common. Inclusions of mafic and felsic volcanites are also present. This metatrondhjemite, which also extends along the shore towards the small sandy beach of Korsvika, may be a northeastward extension of the Fagervika intrusion (Slagstad 1998, 2003). The geochemistry of this and other felsic intrusions and extrusions in the Bymarka ophiolite are shown in Fig. 5.6, taken from the publication of Slagstad (2003).

Stop B: Walk to southwest of the sandy beach, around the small hill (if the tide is low; otherwise walk over the top) to the shore area near an old boathouse.

On the cliff below the hill, and on the small, low peninsula, there are diverse, strongly deformed, sheared and schistose, mafic and some felsic extrusive rocks, hyaloclastite layers and some hornblende gabbro and fine-grained trondhjemite sills.

Fig. 4.11. Strongly sheared metabasalt with lenses of what are likely to have been pillows, prior to deformation, Korsvika, Lade peninsula; looking northeast, down the plunge of the prominent lineation. All transitions from well preserved pillow lavas through to isolated, sheared, 'pillow lenses' can be seen in this part of the Bymarka ophiolite.

Fig. 4.12. Tight to isoclinal, syn-foliation folds in banded felsic tuffs and greenschist, small shoreline cliff, just southwest of Korsvika; looking approximately east. It is difficult to detect any consistent fold vergence in this area.

It is difficult to detect true pillow forms in the metabasalts just here because of the intense deformation, with anastomosing shear surfaces defining boudins and other lensoid forms, enclosing likely remains of pillows (Fig. 4.11). There is a prominent NNE-SSW to NE-SW-trending lineation, and syn-foliation sub-isoclinal folds can be seen in some layers Fig. 4.12).

In some horizons, on steep surfaces parallel to the lineation, one can detect late, small-scale structures indicating top-W to -SW extensional shear. Even younger structures include NNW-SSE

to N-S-trending normal faults, a common feature in the Trondheim area and central Norway as a whole, and probably of Late Palaeozoic or Early Mesozoic age.

In the vicinity of the boathouse and southwest from there, we see more felsic bodies, including trondhjemite sills and sporadic felsic extrusive rocks.

Stop C: Return to the main coastal trail and walk southwest past a tunnel entrance and climb down onto a concrete platform (this was originally the base of a tall, lifeboat-testing facility). Exposures are both above and beneath a concrete overhang.

Here, there are diverse mafic and some felsic volcanic rocks, including small examples of coticule beds. The finely laminated coticules occur higher up in three, competent lozenges canted toward the northeast and separated by intense syn-schistosity shear deformation and later, top-SW (to -W), extensional shear with a prominent NNE-SSW-trending lineation. In the vicinity, quartz veins that formed coevally with the main metamorphism are highly disrupted with lensoid form (and rare isoclinal folds are seen elsewhere along this coastline), and subsequently affected by the top-SW shear bands. Beneath the concrete overhang, there are good examples of spaced, top-SW, extensional shear bands. To the far left there are some WNW-verging, asymmetric folds, nicely picked out by felsic layers, that deform the main schistosity.

For those of you who are unfamiliar with the name *coticule*, it is an ancient name for whetstone and used to describe fine-grained, Mn-garnet rocks quarried in the greenstone-facies rocks of the Ardennes in Belgium. The name was imported to the northeastern United States by an American geologist who studied in Heidelberg in the 1880s, and is used routinely in rock descriptions in the NE Appalachians where such rocks are well known in Ordovician, Silurian and Devonian strata up to granulite facies.

Finally, as a curiosity, we can mention that at the end of the last Ice Age, c. 10,000 years ago, sea level in the vicinity of Trondheim lay at circa 180 m above the present-day sea level. The land has been rising since then, and the current rate is c. 2-3 mm per year. This compares with a maximum of c. 8 mm/year in the Gulf of Bothnia, and 0 mm in the outer coastal regions of western Norway.

Day 5: SCANDIAN GEOLOGY OF THE OUTER TRONDHEIMSFORD REGION

Tuesday August 19, 2008

by Peter Robinson, Arne Solli, Kurt Hollocher, Per Terje Osmundsen, and David Roberts
with contributions by Bob Tucker

Summary of Route: The route for the day involves travel southwest from Trondheim to Orkanger, partly through a new highway-tunnel project, then northwest along the southwest shore of outer Trondheimsfjord to Agdenes and the Valset Ferry. We take the ferry at lunch time (25 minutes) to Brekstad on Ørlandet, the southwesternmost part of Fosen Peninsula. We follow a sinuous route on Fosen along the northeast margin of outer Trondheimsfjord to a dinner location at Hysnes Fort, Hasselvika, then continue to Rørvik Ferry for the crossing (25 minutes) to Byneset Peninsula at Flakk, thence to Trondheim in mid-evening. En route to Rørvik we stop briefly to view the results of the devastating 1978 Rissa marine quick-clay slide.

Practical Note to Participants: Departure will be from the Trondheim Vandrerhjem at 8:00, and return will be to the same location about 21:00. The evening meal will be 17:28-18:08 at Hysnes Fort, Hasselvika, on Fosen Peninsula. Most stops are near the road, and walks will be short though locally steep. Expected weather can be anything from hot sunshine to strong wind with horizontal rain. Waterproof boots, rain suit, and pack to carry them are highly recommended. In addition to the rocks, this trip will cover some pleasant ocean, forest, and hill country scenery. The trip was designed to give participants the maximum amount of insight into the geology, within a single day. Emphasis is on "seeing the rocks", admittedly the most easily accessible rocks, and not on extended descriptions and discussions, which may nevertheless take place in unscheduled moments. The distance travelled is about 200 km, mostly on paved roads and some of the stops are quite short, so participants will have to acquire a certain degree of self-discipline to gain full benefit from the experience. A further degree of discipline is imposed by the schedule of two ferries, which must be conformed to. Therefore, when leaders indicate that it is time to leave a stop, the intention is serious and your welfare is involved. In return the ferries do supply an opportunity for informal refreshments and abundant public toilets, which should not be missed. The ferry fares of all participants will be paid by the leaders. During ferry trips it is always possible to climb to a sun deck for further enjoyment of scenery and geology. This is the first and simplest of ferry days, which become increasingly complex on days 7, 8, 9, 10, but by then all will be well trained!!!

Introduction to Geology: Geologically, as we travel westward we first examine units underlying the Støren and Gula Nappes of the Trondheim Nappe Complex, in descending order Seve=Blåhø=Skjøtingen, Särv =Sætra=Songa, and Risberget (see Figs. 5.1, 5.2). We then examine rocks of the Lensvik synform including Sætra Nappe against basement, Blåhø Nappe, and tonalitic and gabbroic intrusive rocks now assigned to the Støren Nappe. Farther northwest we examine infolded schist and marble probably of the Blåhø Nappe in contact with complexly migmatised Baltican basement, and before boarding the ferry at Valset, an extensive exposure of the gneisses of the Ordovician magmatic arc complex constituting part of the Støren Nappe just above the Scandian Agdenes detachment.

When crossing from Valset to Brekstad on Ørlandet we go from the Ordovician gneisses across the trace of the Hitra-Snåsa Fault (part of the Møre-Trøndelag Fault Complex) and onto rocks structurally above the Høybakken detachment, including the high hill, Lerberen, composed of undeformed 442 Ma granite, and Devonian sandstone and conglomerate with poor plant fossils at Døsvika. We then study platy mylonitised Ordovician gneisses just below the Høybakken Detachment and a cliff exposure of cataclasite of the detachment itself. This will be accompanied by a description of complex Mesozoic extensional faulting shown by recent offshore seismic surveys. Further travel is in the highly deformed Ordovician gneisses below the Høybakken detachment, and southeast across the on-land Hitra-Snåsa Fault to an actual exposure of the Agdenes detachment with high-grade Blåhø rocks below and Ordovician gneisses above. We then travel back toward Trondheim in a region dominated by basement with local infolded nappe units, first at high grade, then with decreasing grade, including a spectacular, recently rediscovered, exposure of Sætra Quartzite with dikes and the Mesoproterozoic Ingdal Granite in the core of a late-Scandian antiform.

During the day there will be emphasis on U-Pb geochronology of Paleoproterozoic to Mesoproterozoic, Ordovician and Siluro-Devonian intrusive igneous rocks, and of rocks metamorphosed in the Silurian and in the Devonian with a special emphasis on the U-Pb geochronology of titanite as a key to exhumation processes. There will also be emphasis on major- and trace-element geochemistry of a suite of Late Neoproterozoic mafic dikes, on Ordovician mafic volcanic rocks and gabbros, and on an extensive terrane of tonalitic intrusions apparently formed in the core of an Ordovician magmatic arc.

GEOLOGIC FRAMEWORK OF THE TRONDHEIMSFJORD REGION

This introduction for Day 5 is a modification and upgrading of material presented by Tucker et al. 2004. The area under consideration is shown in Figure 5.1, which also shows the locations of Phanerozoic U-Pb age determinations given in that paper, but not of age determinations on Baltican basement. The Caledonide geology is composed of Baltican continental basement and a series of far-traveled thrust sheets all encountered in some detail during Days 1-4, but commonly given different names as indicated in Fig. 5.2. In addition, there are Devonian clastic sedimentary rocks
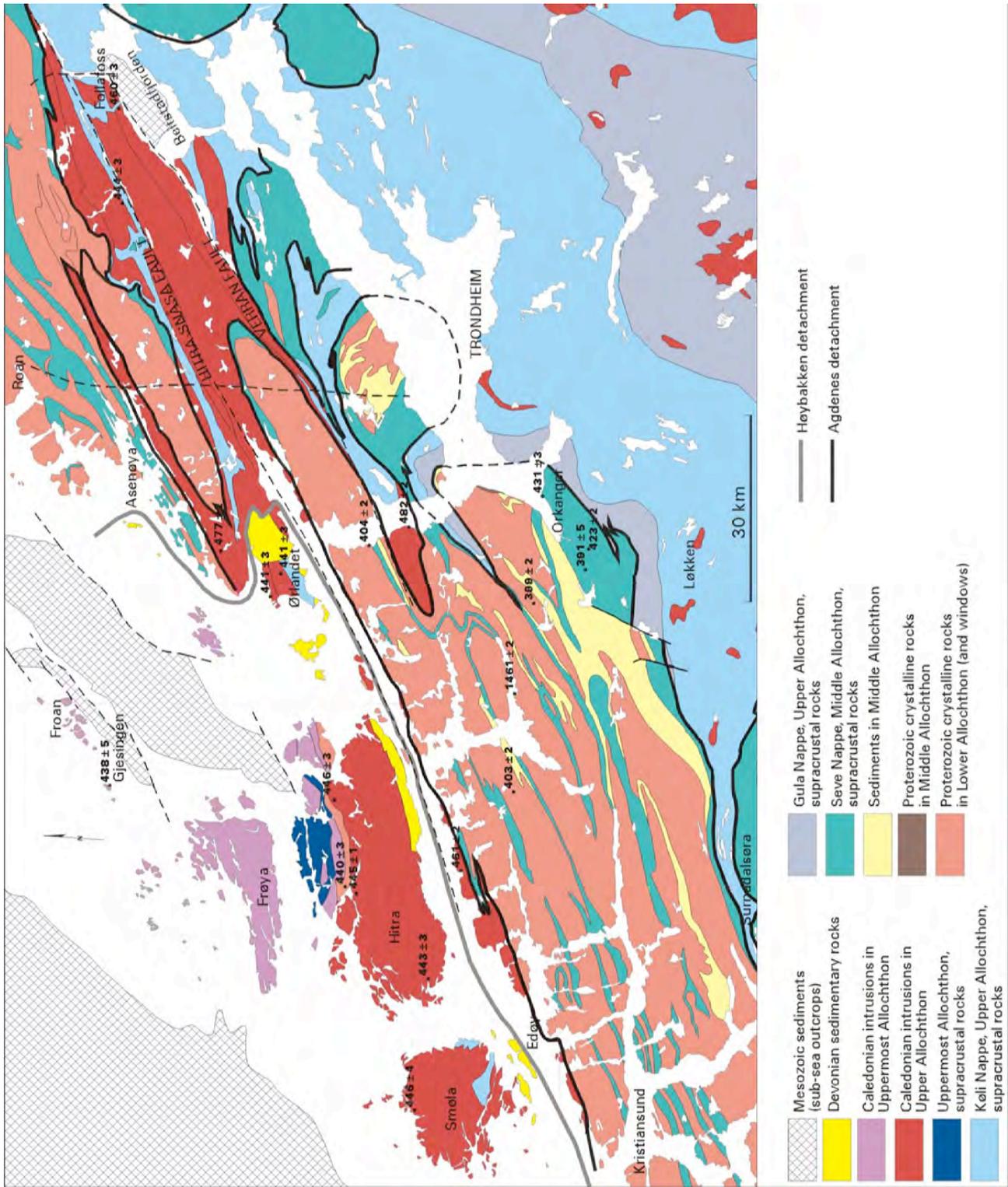
Fig. 5.1 Generalized tectonostratigraphic map of the Trondheimsfjord region, including near-offshore Mesozoic basins, also showing locations of all U-Pb ages reported by Tucker et al. 2004. Of particular interest for Day 5 are the locations of highly metamorphosed basement gneiss and adjacent cover rocks (orange), the locations of only moderately metamoprhosed Ordovician to earliest Silurian intrusions in the Støren Nappe of the Upper Allochthon (red), the locations of Devonian sedimentary rocks (bright yellow), and the locations or implied locations of the Agdenes and Høybakken extensional detachments.

deposited during late-Scandian extension, and offshore sedimentary basins formed during Mesozoic extension, eventually leading to the Cenozoic opening of the modern North Atlantic.

The southeastern part of Figure 5.1 is occupied by a part of the Trondheim Nappe Complex containing large areas of exposure of the Upper Allochthon (Støren, Gula, and Meråker Nappes) of amphibolite-facies to greenschist-facies rocks. Just off the map southeast of Surnadalsøra lies the NE-plunging antiformal hinge of Trollheimen in an eastern promontory of the Western Gneiss Region (WGR) of exposed Proterozoic basement. The NE-plunging Surnadal synform separates Trollheimen from the more highly deformed and metamorphosed region centered on Kristiansund, where mafic rocks in the Proterozoic basement are consistently in the eclogite or high-amphibolite facies. This region, containing rocks at most tectonostratigraphic levels, extends continuously in a NE-SW-trending band from Kristiansund to the coast northwest of Follafoss. The map pattern is dominated by prominent antiformal and synformal hinges, mostly plunging NE, that are merely the latest in a sequence of fold deformations that deformed the pre-existing thrust nappe stack (see more detailed discussion below). Eclogite-facies metamorphism dies out along a vaguely defined line about 30 km west of Orkanger, but relict eclogites are also reported in the Roan Window on the coast northwest of Follafoss (Möller 1988).

The northwestern part of Figure 5.1 is occupied by rocks of the Støren Nappe of the Upper Allochthon, unconformably overlain by Devonian clastic sedimentary rocks, and the Helgeland Nappe Complex of the Uppermost Allochthon. Together, these form part of an extensional allochthon transported west-southwestward many tens of kilometers during late-Scandian sinistral transtension (Krabbendam & Dewey 1997) along a major ductile-to-brittle detachment fault (Séranne 1992, Osmundsen et al. 2006), the Høybakken detachment, highlighted in the figure. Movement on the fault was essentially parallel to regional strike. Where the fault is curved near Ørlandet it outlines a paired SW-plunging syncline and a SW-plunging anticline. The Devonian strata within the syncline dip east, contrary to the dip of the detachment, and form an E-plunging fold. In the ductilely deformed terrane east and southeast of the Høybakken detachment, evidence given below indicates the presence of another major detachment, named the Agdenes detachment (Robinson et al. 2004). Other late- or post-metamorphic faults in Figure 5.1, include the Hitra-Snåså Fault, and the Verran Fault. These and other faults in the region are part of the Møre-Trøndelag Fault Complex (Gabrielsen & Ramberg 1979, Grønlie & Roberts 1989). The youngest bedrock geologic features are fault-bounded submarine basins of Jurassic sedimentary rocks, outlined in offshore regions (Bøe & Bjerkli 1988, Bøe 1991, Sommaruga & Bøe 2002).

HISTORICAL NOTE

It is interesting to look back into earlier interpretations of Fosen geology. Over a century ago, Kjerulf (1871) regarded the 'basement gneisses' of what he called 'Vestranden' as consisting mainly of Precambrian 'granites'. Holtedahl (1944), on the other hand, believed the foliated granitoid rocks to be products of Caledonian granitisation and migmatisation of older supracrustal rocks. Birkeland (1958) totally rejected the Caledonisation idea, noting that "there is no doubt that the rocks of the basement complex existed in their present state long before the beginning of the Cambrian". He also considered the ubiquitous foliation in the granitoids to be a primary igneous structure. Moreover, he favoured the notion that all the Lower Palaeozoic volcanosedimentary rocks in the 'nappes' were lying unconformably upon the Precambrian basement complex. This primary contact hypothesis was also supported by Oftedahl (1964).

Detailed mapping and structural studies in Trøndelag during the 1970s and 80s supported the idea that the intensity of Scandian deformation increases markedly from east to west at any one given tectonic level, not least in the Proterozoic rocks of the Lower Allochthon. Suprajacent nappes, readily identified in eastern districts and in Sweden, were also strongly affected, in some cases attenuated or even excised. The extent of (Scandian) Caledonisation in the Fosen district is such that protolith, granitic to tonalitic, migmatitic orthogneisses and basic rocks of Palaeoproterozoic age have generally been thoroughly reworked and transposed into strongly layered, gneissic L-S tectonites (Johansson 1986a, Roberts 1986, 1998, Möller 1988, Solli et al. 1997b). Möller (1988), in fact, referred to the entire assemblage as the *'Banded Gneiss Complex'* (BGC). Here and there, however, there are relict lensoid bodies and structures present on scales ranging from a few metres to lenses up to 5-6 km in length. These commonly consist of coarsely foliated and migmatitic orthogneiss and basic rocks, some coronitic textured, which along their margins show rapid transpositions through high-strain zones into the ubiquitous, heterogeneous layered gneisses (Johansson 1986a, b, Roberts 1986). Fieldwork subsequent to an air-photo study also showed that Scandian migmatisation is represented locally on the Fosen Peninsula (Roberts 1986), following peak-PT, granulite-facies conditions, a feature confirmed by Möller (1988).

TECTONOSTRATIGRAPHY

Introduction. Understanding of the tectonostratigraphy is based on 1) identification and mapping of a complex sequence of units and 2) recognition that these represent rocks generated in a wide variety of settings that were later assembled in continent-arc accretionary events, and then in a major, terminal, continental-continent collision. Knowledge of these units as studied in the less deformed and metamorphosed regions observed in Days 1-4 is essential. The array of units and unit names is challenging (Gee & Sturt 1985), and discussion of correlation, even for west-central Trøndelag (Fig. 5.2), is beyond the scope of this guidebook.

Baltican basement. The lowest tectonostratigraphic unit exposed in the area of Figure 5.1 is Baltican crystalline basement, the Baltoscandian margin of the former craton of Baltica or the Fennoscandian Shield, dominated by Late

Tectonic Units	Northwestern Areas		Southwestern Areas			Central Areas		Eastern Areas
	Days 5-10	Hemne *Tucker (1986)*	Molde *Robinson (1995)*	Trollheimen *Gee (1980)*	Oppdal *Krill (1980,85)*	West-central *Roberts and Wolff (1980)*	East	Central and southern Jämtland *Gee (1975)*
Late orogenic sedimentary rocks	Devonian					Late Silurian/ Mid-Devonian	Early Devonian	
Upper Allochthon	Støren Nappe		Støren Nappe		Tronget Unit	Støren Nappe / Gula Nappe	Meråker Nappe / Gula Nappe	Köli Nappes
Middle	Seve = Blåhø Nappe	Gagnåsvatn Nappe Unit	Blåhø-Surna Nappe	Blåhø Unit	Surna Unit Blåhø Unit	Levanger/ Skjøtingen Nappes	Øyfjell/ Essandsjø Nappes	Seve Nappes
	Särv = Sætra Nappe	Songa Nappe Unit	Sætra Nappe	Sætra Nappe	Sætra Unit	Leksdal Nappe	Remsklepp Nappe	Särv Nappe
Allochthon	Risberget Nappe	Rønningen Augen Gneiss	Risberget Nappe	Augen Gneiss	Risberget Unit	Hærvola Nappe		Tännäs Augen Gneiss Nappe
Lower Allochthon and Parautochthon	Absent or too thin to show	Øyangen Formation	Åmotsdal Quartzite	Gjevilvatnet Group	Åmotsdal Unit	Quartzite, rhyolite, granitic gneiss Bjørndalen Formation *Gee (1977)*	Quartzites, black phyllites	Jämtland Nappes
Autochthon	Baltican Basement Gneiss	Våvatn Migmatite	Baltica Basement		Lønset Unit	Precambrian crystalline basement		Precambrian granite and gneisses

Fig. 5.2 Correlation of main tectonic units in the central southern part of the Scandinavian Caledonides

Paleoproterozoic granitoid intrusive rocks, in the age range 1686-1650 Ma (Tucker et al. 1990) that were locally intruded by rapakivi granites dated in one location near Molde at 1508 (Tucker et al. 1990, Austreim et al. 2003) and by gabbros dated at 1657 +5-3 Ma and 1462 ± 2 Ma (Tucker et al. 1990). The oldest dated rock in the region 125 km NE from Roan in Figure 5.1 is a 1.84 Ga tonalitic gneiss south of Leka (Schouenborg et al. 1991). The Paleoproterozoic granitoids are similar in age to part of the Transcandinavian Igneous Belt, but are geochemically unlike it (Roland Gorbatschev, pers. comm. 1997). Approximately 1650 Ma U-Pb ages on zircon from partial melt leucosomes in some of the gneisses (Tucker et al. 1990, 2004) suggest there may have been a widespread high-grade regional metamorphism at this time. Generally there is evidence of an early-Scandian overprint ranging in intensity from low-amphibolite facies to eclogite facies, followed by a general late-Scandian amphibolite-facies overprint (Terry & Robinson 2003, Tucker et al. 2004).

The best recent understanding is that the Baltican basement in the area, although providing the basement for emplacement of the sequence of far-traveled thrust nappes, is allochthonous with respect to the Fennoscandian Shield exposed in front of the orogen in Sweden and southern Norway (Gee et al. 1985). Evidence for this comes in two main forms. In the part of the WGR exposed in Trollheimen (see Day 6), maps delineate two levels of basement, each capped by a quasi-continuous layer of variably deformed Neoproterozoic quartzite and pebble conglomerate. The upper level of basement forms a vast and highly folded thrust sheet emplaced above the lower level of basement and its autochthonous cover (Nilsen & Wolff 1989). Robinson (1997) has deduced a similar situation within the anticlinal culmination centered on Rekdalshesten (see Fig. 8.4). Farther north, in the Grong-Olden Culminations, there is a similar map pattern of two levels of allochthonous basement (Roberts 1989, 1997), with an intervening authochtonous cover (Gee 1980),, and the higher thrust sheet (Formofoss Nappe Complex) can be floowed directly into the northern part of Fosen Peninsula. The second form of evidence is in the deep-seismic profile from Trøndelag across the orogen to the autochthonous basement in Sweden (Hurich et al. 1988, 1989; Palm et al. 1991, Hurich & Roberts 1997, Juhojunttii et al. 2001). This shows that the Baltican basement exposed in various tectonic windows near the Norwegian-Swedish border lies above a major shallowly dipping reflector that appears to correspond to the top of the autochthon exposed at the front of the orogen.

Autochthonous cover and Lower Allochthon. Above the Paleo- to Mesoproterozoic Baltican basement in various parts of the map area are vestiges of a very thin, autochthonous sedimentary cover sequence (Gee 1980), usually consisting of probable Neoproterozoic quartzite and conglomerate, overlying pelite representing metamorphosed Cambrian alum shale, and discontinuous limestone. Within the area of Figure 5.1, these units of the Autochthon and Lower Allochthon are too thin to show. In many parts of the map area, especially to the NW, they are absent or recognized with difficulty, and usually not distinguishable from each other. They are, however, reported locally, for example in the previously mentioned quartzites of Trollheimen, and in the Øyangen Formation (Tucker, 1986) of quartzite and overlying mica schist in an area west of Orkanger. This contrasts with the unmetamorphosed lowest thrust allochthon of Days 1 and 2, where the thin autochthonous sequence is overlain by a thicker allochthonous sequence, including Baltican basement, Neoproterozoic sandstones, a Cambrian- Ordovician limestone-shale sequence, and a thin Silurian sandstone and limestone succession grading upward into turbidites at least as young as Wenlock (Gee 1975).

Middle Allochthon. Above autochthonous cover and Lower Allochthon, commonly in direct contact with Baltican basement, are the rocks of the Middle Allochthon. In Figure 5.1, the lowest recognized unit, too thin to show, is the Risberget Augen Gneiss Nappe correlated with the Tännäs Augen Gneiss Nappe in Sweden. This is dominated by variously deformed Mesoproterozoic rapakivi granites with associated other granitoid rocks, gabbros, anorthosites, and mafic dikes. Earlier assigned a Rb-Sr age of 1500 Ma (Krill, 1983), new concordant zircon ages (Handke et al. 1995) indicate that these distinctive rapakivi granites fall into two age groups, 1659-1642 Ma and 1190-1180 Ma, the latter much younger than any known age group in the adjacent less allochthonous basement.

The next key part of the Middle Allochthon, finally extensive enough to show on Figure 5.1 , consists of feldspathic quartzites derived from Neoproterozoic feldspathic sandstones, interlayered with amphibolites derived from Late Neoproterozoic diabase dikes. This is the Särv Nappe in Sweden, variously known in Norway as the Sætra, Songa, and Leksdal Nappes. Furthermore, a few areas of very feldspathic rocks, intricately interlayered with amphibolites, appear to be Baltican basement to the Särv Nappe, heavily intruded by the same Neoproterozoic dike swarm. As seen on Days 2 and 3 in Sweden, the Särv Nappe is as much as 2 km thick. By contrast, in parts of the WGR, mapping has shown that the equivalent quartzite-amphibolite sequence ranges from 10 m (Robinson 1995) to as thin as 1 m (Terry & Robinson 1996, Robinson et al. 2003). For purposes of Day 5 near Trondheimsfjord, we use the name Sætra, to emphasize correlations with regions to the south and west, though Songa and Leksdal have been applied on local maps.

The mafic dikes of the Särv Nappe and potential correlatives in Norway have been a target for major-and trace-element geochemistry, because they represent a potentially powerful tool for correlation, and because they appear to represent MORB-type magmas produced during the initial opening of Iapetus. Most recently, Hollocher et al. (2007a) carried these concepts from regions of well defined dikes near Oppdal into regions a greater deformation and metamorphism toward the northwest, including locations where the Sætra Nappe is only 1 m thick, and even to areas where the dikes were converted to eclogite. A surprising and gratifying result of the geochemical study (see summary below) was that dikes with a geochemistry identical to those known in the Sætra and Särv Nappes, also intrude the Risberget Nappe and even a few sections of highly deformed basement gneiss. The implication of these results is that

the quartzites of the Sætra Nappe, the Risberget Nappe and some basement segments all belong to the Middle Allochthon, and are far-traveled with respect to the adjacent allochthonous Baltican basement.

New assignment to the Middle Allochthon. In Sweden, the Upper Allochthon has traditionally been mapped as two major units, the lower Seve Nappes, containing medium- to high-grade metamorphic rocks, and the upper Köli Nappes containing medium- to low-grade metamorphic rocks, locally with Ordovician fossils. In Norway, in the area of Figure 5.1, traditional equivalent names for the Seve are the Skjøtingen and Blåhø Nappes, whilst the Köli Nappes have a variety of names. In the Trondheim Region (SE part of Fig. 5.1), the Köli Nappes of the Upper Allochthon are represented by the Trondheim Nappe Complex, comprising the Støren, Meråker and Gula Nappes (e.g., Roberts & Stephens 2000). Geotectonically, the rocks of Seve have been considered as the extreme outboard assemblage of the Baltoscandian continental margin, subjected to high-grade, even locally high-P metamorphism in pre-Scandian and/or early Scandian time. In agreement with this are several belts within the Arctic Seve composed of sandstones identical to those of the Särv, but different only in metamorphic grade. The Støren and Meråker Nappes are considered to have a largely exotic origin from the Iapetus Ocean as Late Cambrian - earliest Ordovician, deformed ophiolitic and volcanic-arc sequences overlain unconformably by Mid Arenig to possibly early Silurian volcanosedimentary basinal successions. Early Ordovician fossil affinities in parts of the Støren Nappe at Hølonda and on Smøla suggest deposition of the rocks close to Laurentia, and possibly even that the strata were thrust or obducted onto Laurentia in the Ordovician Taconian orogeny, and then later transferred onto the Baltica margin in the Scandian collision (see below). Because of the contrasting geotectonic settings of the Seve and Köli Nappes and equivalents indicated above, Andréasson & Gee (2008), have made the proposal that the Seve be reassigned from the lower part of the Upper Allochthon to the upper part of the Middle Allochthon. That major change has been agreed to by all the compilers of this guidebook. It is logical from the broadest geotectonic perspective, because the boundary between the Middle and Upper Allochthons now represents, in most respects, the fossil trace of the pre-Scandian Iapetus, whereas previously the boundary was based on a contrast of metamorphic grade between the Seve and underlying Särv Nappe. Moreover, assigning the Seve to the Middle Allochthon, falls in line with the situation in northern Norway, where the Seve correlative – the Kalak Nappe Complex – has traditionally been regarded as part of the Middle Allochthon.

In detailed subdivision of likely Seve equivalents in the Trollheimen region of Norway, Krill (1987) identified an upper Surna Nappe characterized by higher grade metamorphism and abundant intrusions of trondhjemite, and a lower Blåhø Nappe, usually slightly less metamorphosed and lacking such intrusions, and corresponding to the type locality on Blåhø mountain. Robinson (1995) did not attempt to use this distinction in his correlation within the Moldefjord region and adopted the composite term Blåhø-Surna. This was applied to medium- to high-grade mica schists, commonly with garnet, kyanite, sillimanite, and with variably distributed granitoid intrusions and pegmatites, by abundant coarse amphibolites, commonly with garnet and pyroxenes, and by fairly common layers of coarse-grained marble. For the Ålesund and Ulsteinvik 1:250,000 map sheets Tveten et al. (1998) initially simplified Blåhø-Surna to Blåhø, but later changed this to Surna.

Upper Allochthon. Within the area of Figure 5.1, excluding the Gula to the southeast, the Upper Allochthon consists entirely of rocks assigned to the Støren Nappe. This is dominated by low- to medium-grade, metamorphosed volcanic and related intrusive rocks including ophiolite fragments, by metamorphosed volcanosedimentary sequences, and by metamorphosed black shale. A common distinction across the boundary between the Seve=Blåhø Nappe and the Støren Nappe is that the Seve contains medium- to high-grade rocks with coarse garnet, whereas Støren contains low- to medium-grade rocks with no more than 2-3 mm garnets if any at all. An additional component of the Støren, recognized earlier (Kollung 1964, Tucker 1988, Gautneb & Roberts 1989) but receiving special emphasis by Tucker et al. (2004), is a variety of Early Ordovician to Early Silurian medium- to coarse-grained, calc-alkaline, intrusive igneous rocks that were broadly contemporaneous with the extrusion of the arc-volcanic rocks. Above the Høybakken detachment these rocks are generally little deformed, but between the Høybakken and Agdenes detachments, where they are in the age range 482 to 444 Ma, they are strongly gneissic, but minimally recrystallized, and containing very few pegmatites.

Hollocher et al. (2007c and in preparation) have made a major- and trace-element study of the metamorphosed mafic volcanic rocks and gabbros of the Støren Nappe from the Trondheim Region southwestward to the Moldefjord region, and also of the calc-alkaline intrusive suite below the Høybakken detachment, where extensive collections were made near all five of the U-Pb zircon age locations shown in Figure 5.1. This supplements the extensive data base of analyses of mafic rocks in the Støren Nappe by many workers (Gale & Roberts 1974, Loeschke 1976, Grenne & Roberts 1980, Grenne et al. 1980, Roberts 1980, 1982a, 1982b, Heim et al. 1987, Grenne 1989, Slagstad 2003), and the extensive data base of calc-alkaline intrusive rocks from above the Høybakken detachment by Gautneb & Roberts (1989) and Lindstrøm (1995). Preliminary results of these geochemical studies are summarized in separate sections below.

Uppermost Allochthon. In Nordland and Troms, North Norway, the Köli Nappes are overlain by the Rödingsfjället and Helgeland Nappe Complexes and other nappes of the Uppermost Allochthon (for review see Roberts et al. 2007). These consist of Neoproterozoic to Early Paleozoic, metamorphosed sedimentary rocks including migmatitic paragneisses and thick marble units, as well as ophiolitic rocks with overlying cover sequences of Late Ordovician – Early Silurian age. There is evidence in southern Nordland for crust-derived Early to Late Arenig (477 – 466 Ma) peraluminous intrusions (Yoshinobu et al. 2002), and emplacement of Ashgill to Llandovery (448 – 430 Ma), calc-alkaline intrusive rocks (Nordgulen et al. 1993, 2002, Eide et al. 2002, Nissen et al. 2006, Barnes et al. 2007). Recent

structural, geochronological and stable isotope data from different parts of the Uppermost Allochthon support a Laurentian, or possibly a Laurentia-related microcontinental origin for the rocks and have shown that west-directed contractional deformation and granitoid magmatism occurred during an Ordovician Taconic episode (Roberts et al. 2001, 2002a, 2007 Melezhik et al. 2002, Yoshinobu et al. 2002). Reconnaissance (Solli and Robinson, unpubl. data 1997) suggests strongly that the rocks exposed in the northern part of Hitra and on Frøya in Figure 5.1 should be assigned to the Helgeland Nappe Complex. Immediately south of this there is a narrow belt of strongly layered, pink, granitoid gneisses and amphibolites, closely resembling the Baltican basement of the WGR. However, both this basement and the adjacent, inferred, Helgeland Nappe rocks are dominated by steep linear fabrics and lack the dominant sub-horizontal linear fabrics characteristic of adjacent parts of the WGR. Thus, we suspect that all these rocks may be above the Høybakken extensional detachment associated with the Devonian basins, and have been extensionally transported from an original location far to the northeast.

Late- to post-orogenic clastic sedimentary rocks. At the top of the tectonostratigraphic section, but not really part of it, are the moderately deformed and slightly metamorphosed conglomerates, sandstones, shales and rare limestones of the Devonian 'Old Red Sandstone' basins (Steel et al. 1985). The rocks unconformably beneath the basins have escaped much of the deformation and metamorphism characteristic of the WGR and appear to belong to parts of the Caledonides normally exposed far to the east and notheast and also, in some of the West Norwegian examples, rather low in the regional nappe tectonostratigraphy. The `Old Red´ basins are exposed on Smøla, Edøy, Hitra, and Ørlandet and a large number of nearby small islands, at Røragen close to Røros inland, and in four major basins in West Norway. The latter are from north to south the Hornelen, Håsteinen, Kvamshesten, and Solund basins. According to a structural study by Seranne (1992), and in agreement with studies of other Devonian basins in West Norway (Andersen & Jamtveit 1990, Andersen 1993), all of these strata lie on the upper plates of major top-west to -southwest extensional detachment faults that carried both the Devonian strata and their immediately underlying igneous-metamorphic substrate for tens of kilometers southwestward or westward from their original sites of deposition.

The majority of sedimentary rocks in the Old Red basins are conglomerates and sandstones that were deposited on alluvial fans and in braided river environments. In the best preserved and exposed basins, conglomerates typically fringe the basin margins whereas sandstones occupy the central basin areas. In addition, silty redbeds are quite abundant in some of the basins, giving evidence for intermittent floodplain and lacustrine deposition. The Devonian Limestones described by Bryhni (1974) from the Hustadvika area were also interpreted as lacustrine. The occurrence of landslides in the basins in western Norway attests further to a rough, mountainous topography in the surrounding areas, lending support to models of fault-controlled deposition. During deposition of much of the basin stratigraphy, fault-controlled topography was developed mainly inside the hanging walls of the great detachments, although during deposition of the relatively young succession at Asenøy, the basin appears to have sampled the adjacent footwall rocks (Eide et al. 2005, Osmundsen et al. 2006).

The rocks on Hitra contain a phyllocarid and euryptyrid fauna that may be as old as Mid Silurian (Størmer 1935, Bassett 1985). The best palynological data from Tristein islet, north of Ørlandet (Fig. 5.1) suggests a Late Emsian age of deposition (Allen 1976), about 403-394 Ma based on the Tucker et al. (1997b, 1998) Devonian time scale (but see Gradstein et al. 2004 for slighlty different interpretation). An ^{40}Ar-^{39}Ar study of detrital K-feldspar and white mica from Asenøya, northeast from Tristeinen, suggests a Givetian (387-382 Ma) or even younger age of deposition, for the middle part of the `Old Red´ section there (Eide et al. 2005), such that the youngest strata may be Early Carboniferous.

The Devonian strata on Hitra rest with demonstrated unconformity on both the volcanic and the intrusive rocks of the Støren Nappe on the southeastern side of the island, but their relationship with the basement rocks and rocks of the Helgeland Nappe Complex of northern Hitra, also proposed to lie above the extensional detachment, are unknown. In the case of the basins near Trondheimsfjord, detailed studies (Séranne, 1992; Watts 2001; Osmundsen et al. 2006) suggest that movement on the detachment fault was west-southwestward, in essentially the same direction as the slightly earlier and more ductile folds and dominant lineations that pervade the region. Similar, top-SW shear bands have been reported from other parts of the Fosen Peninsula in the footwall of the detachments (Piasecki and Cliff 1988). However, the basins are also affected by contractional deformation in the form of folds and reverse faults that are parallel to the maximum elongation trend. The extensional detachment fault exposed at Høybakken is banked against the WSW-ENE oriented Hitra-Snåsa Fault of the Møre-Trøndelag Fault Complex (MTFC), and several WSW-ESE-trending, smaller fault strands affect the Old Red basin. The faults show sliplines that range from strike-parallel groves and mullions to oblique and dip-slip lineations, with sinistral strike- to oblique slip as the most dominant phase(s) in the outer Trondheimsfjorden area.

Late Paleozoic, Mesozoic and Cenozoic strata. The areas offshore from Møre and Trøndelag are characterised by a series of basins that evolved in Late Palaeozoic, Mesozoic and Cenozoic time. Late Permian into Early Mesozoic rocks are preserved in half-graben basins beneath the Trøndelag Platform (e.g., Müller et al. 2005), which was left in a structurally high position after the main Late Jurassic-Early Cretaceous rift phase. Little is known about the detailed stratigraphy of these basins due to limited well control. However, Late Permian carbonates, shallow marine sandstones, anhydrite and turbidites have been described based on drilled sections (Bugge et al. 2002, Müller et al. 2005). The Late Permian anhydrite can be used as a marker horizon in the seismic reflection data under parts of the platform area, depending somewhat on the quality of the seismic reflection data. The Triassic succession contains turbidites and evaporites as well as terrigenous units, such as redbeds and `grey beds´, of Norian to Rhaetian age. Evaporites are

present also in the Middle and Upper Triassic successions. Lower and Middle Triassic strata display syn-rift wedge geometries adjacent to the half-graben bounding faults (Osmundsen et al. 2002, 2005, Müller et al. 2005). The Trøndelag Platform was left tectonically quiet after the Late Triassic and is covered by relatively thin Jurassic, Cretaceous and Tertiary strata.
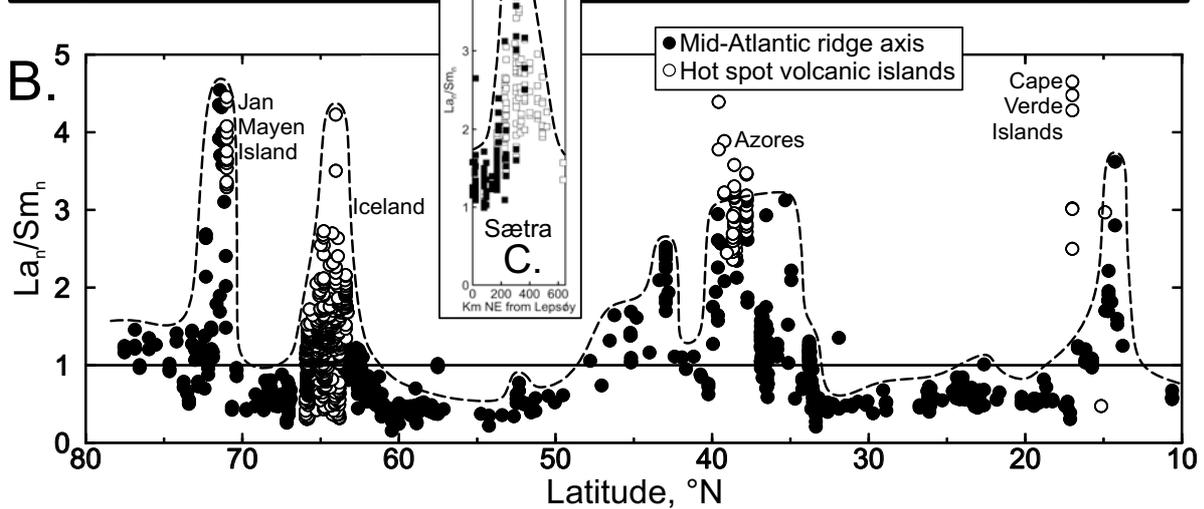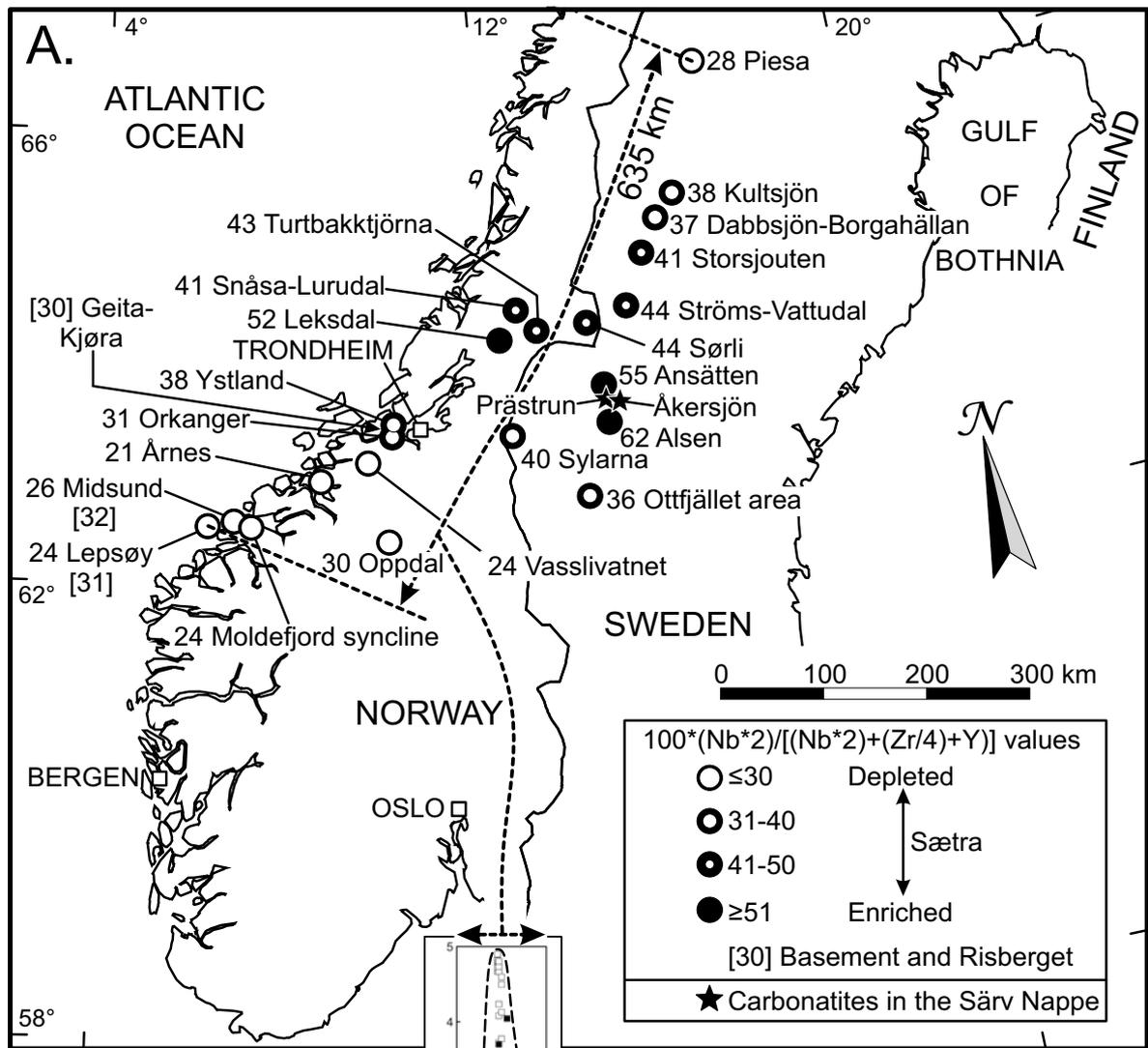
The Halten Terrace (see Fig 5. 11B below) is downfaulted with respect to the platform, is underlain by a thicker succession of Triassic and Jurassic strata and hosts a number of hydrocarbon reservoirs. A number of relationships indicate tectonic activity and rotation of the Halten Terrace in the Jurassic (e.g., Brekke et al. 2000). The Lower Jurassic sedimentary succession on the Halten Terrace contain coal-bearing strata that give rise to well defined reflectors in the seismic reflection data. The Middle Jurassic deposits comprise deltaic and shallow marine sandstones intercalated with shelf mudstone intervals (Dalland et al. 1988). To the southwest of the Halten Terrace is the Slørebotn Subbasin, which contains an up to 1.5 km-thick wedge of Upper Jurassic (Volgian) syn-rift deposits. West and northwest of the Halten Terrace and the Slørebotn Subbasin are the deep Møre and Vøring basins. These are believed to have evolved largely during the Late Jurassic-Early Cretaceous rift phase, which became responsible for very large amounts of crustal extension and for the deposition of relatively thin syn-rift and very thick post-rift deposits. In the deep basins, stratigraphic wedges that overlie deep-seated, rotated fault-blocks are undated and hard to correlate, but are likely to contain Triassic and Jurassic strata (e.g., Osmundsen & Ebbing in press). The rotated fault-blocks are overlain unconformably by Lower Cretaceous units that onlap and drape the fault-blocks, and a strong reflector interpreted to represent the top of the Cenomanian (e.g., Blystad et al. 1995, Brekke 2000) oversteps the southeastern rift flank and thus provides a temporal constraint for the burial of the main rift topography close to the Norwegian mainland. The late post-rift succession is represented by several kilometres of post-Cenomanian, Cretaceous and Tertiary strata. In the more distal parts of the rift, however, in the northwest Vøring Basin, normal faulting continued into, or was resumed in, the Late Cretaceous and Palaeogene. During Campanian and Maastricht times, basin-floor fan deposits were laid down in the outer Vøring Basin (Fjellanger et al. 2005). The basin-floor fan system contains stacked, laterally continuous, sheet-like sandstones of turbiditic origin (op. cit) and reaches thicknesses of up to 1 kilometre as confirmed by an exploration well in the northwestern part of the Vøring Basin (op. cit.).

The Cenozoic stratigraphic record contains Palaeogene lavas in the NW areas of the Møre and Vøring basins, including basinward-prograding units interpreted as lava deltas (Planke et al. 1999). In the southeast, basinward-prograding units include Palaeogene deposits in the Møre area (Fig. 5.11 b), the Upper Miocene/Lower Pliocene Molo formation and the Late Pliocene-Pleistocene prograding wedge (e.g. Henriksen et al. 2005). The Cenozoic stratigraphy is thought to reflect deepening of the post-rift basin and uplift of the Norwegian mainland that followed breakup in the Eocene around 54 Ma (e.g. Faleide et al. 2002). Deposition of the Plio-Pleistocene Neogene wedge is interpreted as controlled largely by glacial activity. Eocene lavas dominate the Cenozoic stratigraphy of the Møre and Vøring marginal highs (Skogseid et al. 1992). In the Eocene, a transgressive event probably led to flooding of large parts of Fennoscandia and to widespread deposition of fine-grained lithologies in the Møre and Vøring basins (Henriksen et al. 2005). In the Oligocene, fine-grained deposition appears to have continued in a deep-marine environment. In the Miocene, a shallowing is interpreted to have taken place as shown by an Early to Middle Miocene hiatus, recognised over the entire continental shelf of the Norwegian Sea (e.g., Eidvin et al. 1993). It has been suggested that the Norwegian mainland as well as the inner shelf areas were uplifted at this time (Jordt et al. 1995, Brekke 2000). Compressive deformation led to the inversion of Palaeogene depocentres in the large arches and domes in the Norwegian Sea; over these, a Mid Miocene unconformity developed (Brekke 2000). Low-angle progradational clinoforms are observed in Miocene strata, downlapping onto the base of the Miocene. In Early Pliocene time, basinward progradation is displayed by the Molo Formation, a very widespread deltaic unit that has been related to epeirogenic uplift of the Norwegian mainland (Henriksen et al. 2005). In the Late Pliocene to Pleistocene, up to 3 kilometres of glacial and glaciomarine sediments were deposited in offshore Norway, indicating high sedimentation rates. Regional uplift of the Norwegian mainland and glacial expansion over parts of the continental shelf led to up to 100 kilometres of westwards progradation of the shelf edge.

GEOCHEMICAL STUDIES OF TECTONOSTRATIGRAPHIC UNITS

Late Neoproterozoic dike swarm of the Middle Allochthon. Figure 5.3 summarizes some of the broad geochemical relationships among basaltic dikes in the Sætra Nappe and related units. Figure 5.3A shows all available data in central Norway and Sweden with respect to the NbZrY value (inset), which is a measure of tholeiitic to alkaline character of basaltic rocks such as these which have no significant Nb anomalies. Sample locations span a geographic distance of 635 km, over which the NbZrY values increase from ~25 in the southwest (Lepsøy, Midsund), up to ~55 at ~64°N

Fig. 5.3 (facing page) Map showing the locations of available Sætra Nappe (and equivalent) dike data in central Norway and west-central Sweden. Symbols and numbers represent the median NbZrY value (see inset) for basaltic dikes at each location. This measure of tholeiitic to alkaline character was used because lanthanide data are lacking for many sites. Dashed lines and arrow represent the 635 km geographic span of data locations. B) Graph showing La_n/Sm_n variations along the Mid-Atlantic Ridge, 10°-80°N (after Schilling et al. 1983), redrafted to include additional data from the PETDB and GEOROC databases (rare rocks with La_n/Sm_n >5 were omitted). Curved dashed line indicates the approximate upper limit of La_n/Sm_n ratios for ocean ridge basalts along the ridge axis. C) All Sætra and equivalent dike data from A, projected onto the 635 km geographic span line and shown at the same vertical and horizontal scales as B.

Filled squares are La_n/Sm_n ratios, open squares are samples without lanthanide data and for which approximately equivalent La_n/Sm_n values were calculated:

$$La_n/Sm_n \approx 0.0015(NbZrY\ value)^2 -0.037(NbZrY\ value) + 1.3;\ R^2 = 0.73.$$

Dashed line shows the approximate upper limit of La_n/Sm_n ratios in Sætra and equivalent rocks as in B for Atlantic ocean ridge basalts. Sætra and equivalent data are from Andréasson et al. (1979), Solyom et al. (1979), Krill (1983), Solyom et al. (1985), Beckholmen & Roberts (1999), and Hollocher et al. (2007a).

(Leksdal, Ansätten, Alsen; plus carbonatites, Jonsson & Stephens 2006), and back down to 28 in the northeast (Piesa). This rise and fall in alkaline character is similar to the influence of hot spots on the composition of mid-ocean ridge basalts. Figure 5.3B shows the change in La_n/Sm_n ratios along the North Atlantic mid-ocean ridge axis. La_n/Sm_n ratios vary along the ridge axis from <1 in areas distant from hot spots to 4 or 5 where associated with hotspot-related islands. The geographic scale of hot spot influence is generally hundreds of kilometers. Figure 5.3C shows the Sætra rocks from A plotted at the same horizontal and vertical scales as B. The width and height of the La_n/Sm_n peak in C is comparable to those on the Mid-Atlantic Ridge. This suggests that the geochemical variation of dikes in the Sætra and equivalent units also records hotspot influence on the Baltica continental margin during rifting that formed the Iapetus Ocean, and the sedimentary basins into which rift-related dikes were emplaced. This indicates that differential thrust transport of these rocks during the Scandian collision was insufficient to erase the original geographic pattern of dike compositional variation.

Figure 5.4A shows dikes to be all mafic and largely tholeiitic, as expected for emplacement in a rifting environment. Samples from Oppdal and the other localities shown with large symbols occupy only the more silica-rich half of the field of all Sætra data. The low silica samples are dominated by more alkaline rocks, the general locations of which are shown in Figure 5.3A.

Figures 5.4B to E distinguish between Sætra and related MORB-like rocks that plot to the right of the vertical line from arc-like mafic rocks that are abundant in other nappes (e.g., Støren and Blåhø Nappes). Oppdal dikes plot in a small box with Nb/La ratios of 0.8-1.5 and La_n/Sm_n ratios of 1.0-1.8. At Orkanger and Vasslivatnet dikes plot on top of those at Oppdal. At Ystland, Lensvik syncline, however, dikes in well constrained Sætra quartzite are more LREE-enriched and set precedent for such compositions elsewhere in the Sætra Nappe. Deformed dikes in basement gneisses and Risberget augen gneiss from the Trondheimsfjord shore at Geita and Kjøra also plot on top of the Oppdal dikes (Fig. 5.4B-E), but extend slightly to higher La_n/Sm_n ratios like the Ystland dikes.

Figures 5.4F to I are REE diagrams for the same locations as Figures 5.4B to E. Samples from Orkanger and Vasslivatnet have slightly flatter and straighter REE patterns than those from Oppdal, whereas Ystland dikes are slightly concave upward with steeper LREE slopes. Dikes in basement and Risberget augen gneiss at Geita and Kjøra span a pattern range between those at Orkanger and those at Ystland. Note that Eu anomalies are small or nonexistent, suggesting little plagioclase accumulation or fractionation.

Figures 5.4J to M show spider diagrams for these same rocks. Though there are small differences in overall pattern shapes derived from the REE patterns, the anomalies are largely similar, and small to nonexistent (i.e., MORB-like). In particular, Nb-Ta anomalies are small or absent, though small negative Zr-Hf anomalies seem to be common. Small uranium anomalies appear to be common, though these may result from minor uranium mobility during alteration or metamorphism.

One sample from Kjøra (Figs. 5.4E, I, M) is of a large, coarse, gabbroic-looking amphibolite boudin that is texturally and chemically distinct from nearby and adjacent Sætra-like crosscutting dikes between Kjøra and Geita. This sample has a low Nb/La ratio (0.29) compared to Sætra (>0.8), a lower and flatter HREE pattern, and large negative Nb-Ta and Zr-Hf anomalies. This rock is clearly unlike the Sætra and related dikes characteristic of the Middle Allochthon. Mafic rocks in the Risberget augen gneiss elsewhere are shown in Figure 5.4E, which indicates a mix of arc-like and MORB- (Sætra-) like compositions, possibly derived from pre- and syn-Iapetan rifting magmatism, respectively.

Metamorphosed volcanic rocks and gabbros of the Støren Nappe. Figure 5.5 shows data for volcanic rocks and some gabbros in the Støren Nappe and related units. In contrast to the chemically rather simple Sætra dikes, this unit is geochemically complex. To help identify the sources of this complexity Figure 5.5A shows an overview of possible tectonic affiliations for mafic rocks in different parts of the Støren volcanics, like Figures 5.4B-E used above for the Sætra Nappe. This shows that most of the mafic volcanics have low Nb/La ratios, and so are probably related to arc volcanism, and many are LREE-depleted, indicating a depleted source. Some alkaline rocks occur at high La_n/Sm_n ratios, mostly from alkali basalts in the Hølonda area (Grenne & Roberts, 1998). Some samples have Nb/La ratios >0.8 and so are more MORB-like. Excluding the alkaline rocks, the data suggest two branches extending from the Kvithyll shipyards samples (Stop 5-16A, large black dots): a branch extending to the upper right (higher Nb/La and La_n/Sm_n ratios), and a branch extending to the lower left.

Figure 5.5B illustrates the two branches more clearly. This diagram has three compositional axes, chosen and oriented to distinguish best between MORB-like and arc-like basalt trends indicated by gray dots. The MORB trend extends between an enriched (or "primitive") MORB source and a depleted MORB source toward the back left corner of the composition space (low Th_n/La_n. Nb_n/La_n, and La_n/Sm_n). The classic arc trend extends between the depleted MORB source and a line of previously depleted sources that have been variably replenished by slab metasomatic fluids. Many mafic volcanics in arcs plot in the MORB trend, representing melts from parts of the mantle wedge not (yet?) affected by fluids from the subducting slab. Most mafic Støren volcanics plot along the MORB trend or slightly away from the axis of this trend toward the arc trend, suggesting that most of these rocks were derived from un- or slightly metasomatized, variably depleted mantle wedge or back arc basin sources. A less abundant but distinct set of Støren volcanics plot along the arc trend and so likely represent melts from previously depleted, metasomatized mantle wedge
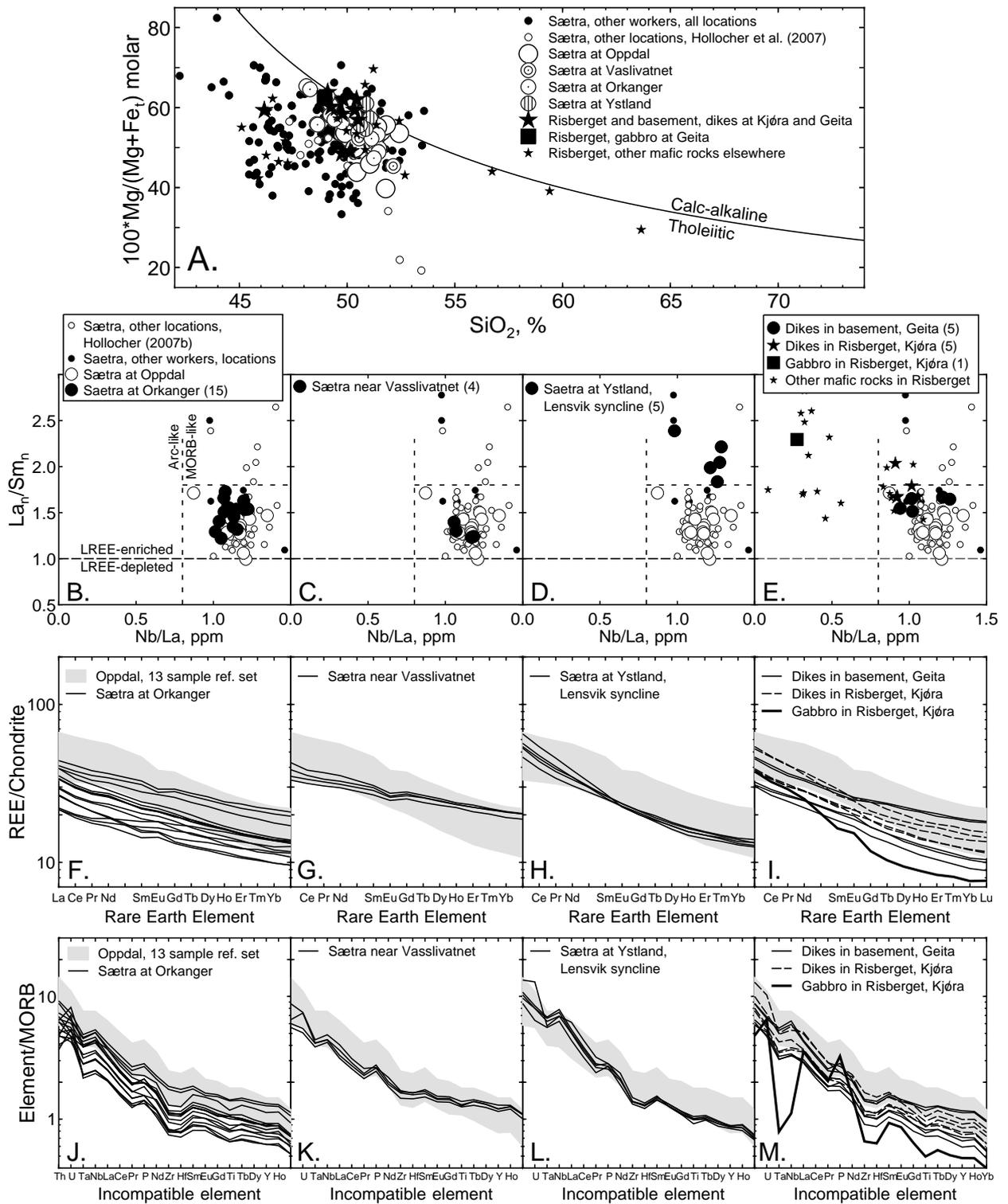
Fig. 5.4 Geochemistry of Sætra dikes at and close to trip stops on Day 5, in comparison to Sætra dikes in the Oppdal quarries (Day 6). A) SiO$_2$ vs. Mg' diagram showing that most Sætra dikes plot in the tholeiitic field. Most samples having SiO$_2$ contents <48% are alkaline. B-E) data are plotted with respect to La$_n$/Sm$_n$ ratio (LREE-enrichment) vs. Nb/La ratio (arc signature; reference lines after Hollocher et al., 2007b). F-I) REE diagrams for the same sample areas. J-M) Spider diagrams for the same sample areas. Data sets are the same as in Figure 5.3. The gray Oppdal reference fields in F-M were defined by the 13 most LREE-rich Oppdal dikes from Hollocher et al. (2007a, two samples with flat REE patterns were omitted). Most of these 13 have slightly Z-shaped patterns, evident in the gray field. REE normalizing factors are from McDonough & Sun (1995), and MORB normalizing factors are from Pearce & Parkinson (1993). The tholeiitic-calc-alkaline discriminant line is from Miyashiro & Shido (1975).

sources. Oddly, the arc- and MORB-like Støren types seem to be geographically and stratigraphically (in the Moldefjord syncline) intermixed, at least using the Hollocher et al. (unpublished) data set.

Figure 5.5C shows important geochemical relationships between the fine-grained volcanics and bodies of coarse-grained gabbroic rocks within the volcanics exposed near Rissa, Litleneset, Storås (near Løkken), and in the Moldefjord syncline. Fine-grained mafic volcanics span a narrow Sc (compatible element) range but over a factor of 100 range in Th (incompatible element) content, representing variable source depletion by melting and enrichment by Th-rich subduction zone metasomatic fluids. The associated coarse-grained gabbroic rocks span a wider range of Sc content, especially at low Th concentrations, that we suspect is caused by variable pyroxene-plagioclase proportions in the protolith gabbroic cumulates. Cumulate layering is found near Brattvåg in the Moldefjord syncline (Stop 8-9). The large gray triangles represent model cumulates derived from an average of mafic Støren volcanic (partition coefficients from Hollocher, 1993), with variable amounts of trapped liquid. The most Th-rich gabbros that have a narrow range of Sc concentrations are interpreted to have magma compositions, whereas the low-Th gabbros are interpreted to be cumulates, ranging from orthocumulates to adcumulates with as little as ~1% calculated trapped liquid.

Figure 5.5D shows all available data for Støren volcanics, with localities seen on Day 5 and felsic rocks from the Bymarka ophiolite, all shown with larger symbols. The data appear to define a bimodal assemblage with few samples having 58-65% SiO_2. Felsic volcanics are dominantly calc-alkaline, whereas the mafic volcanics tend to straddle the line or plot in the tholeiitic field, consistent with the observation that most mafic Støren volcanics plot along the MORB trend (Fig. 5.5B). Most or all of the most Mg-rich mafic rocks are coarse-grained gabbros.

Figures 5.5E and H show REE and spider diagrams for mafic volcanics we will be visiting or near on Day 5. They span a wide range of REE content and pattern shape: LREE-depleted, flat, and LREE-enriched, but all have relatively flat HREE patterns and small or no Eu anomalies. All spider patterns have small to large negative Nb-Ta anomalies indicative of arc-type sources. Th-U anomalies are quite variable, possibly indicating post-emplacement alteration, as suggested above for the Sætra dikes, but the U and Th variability in the Støren is much greater than in the Sætra dikes (Fig. 5.4J-M), and this variability is also characteristic of arc basalts (Fig. 5.5B). We therefore suspect that at least most of the U and Th variation is primary.

The coarse-grained Støren gabbros from around Trondheimsfjord (Fig. 5.5F and I) are all LREE-depleted to variable degrees, and several have positive Eu anomalies suggestive of cumulus plagioclase. The wide range of REE concentrations could indicate highly variable parent magmas (Fig. 5.5E) or variable trapped liquid (Fig. 5.5C). Preliminary modeling suggests that the LREE depletion must be inherited from the parental magma at least for the several most LREE-depleted gabbros. The spider patterns are difficult to interpret directly because of the likely cumulus nature of at least several of the gabbros. The large negative Nb-Ta anomalies suggest derivation from parental magmas with negative Nb-Ta anomalies, and the negative P anomalies are also consistent with some of the gabbros being cumulates.

The felsic rocks of the Støren Nappe are discussed in two sections below, and are compared with three sets of felsic rocks from the Bymarka Ophiolite fragment in Trondheim. These rocks are interpreted by Slagstad (2003) to record felsic magmatism in three successive tectonic environments. The Klemetsaunet rhyodacite is interpreted to be an ocean floor "plagiogranite", derived from fractional crystallization of basalts. The Fagervika Trondhjemite is interpreted to be derived from garnet-free basaltic rocks (melting or crystallization) in an arc environment. The Byneset Trondhjemite is interpreted as being partial melt from the garnet-bearing deep parts of a thrust stack of ocean floor and arc rocks during or after obduction onto the Baltica margin. This diversity of interpreted tectonic environments, large differences in geochemistry, and occurrence in the Støren Nappe make this rock set ideal for comparison to other Støren felsic rocks.

Figures 5.5G and J show dark-colored andesitic and light-colored dacitic rocks (all >55% SiO_2) at localities visited on Day 5, compared to three sets of felsic rocks from the Bymarka ophiolite. The most REE-rich of these rocks are concave upward and saddle-shaped, rather like the Fagervika Trondhjemite in the Bymarka ophiolite. However, the Fagervika Trondhjemites have large negative Eu anomalies that the volcanics do not. The spider patterns show that all of the silica-rich rocks have large negative Nb-Ta anomalies, but other anomalies are quite small, even P and Ti anomalies which tend to be enormous in many felsic rocks because of apatite and Fe-Ti oxide separation (e.g., Fagervika Trondhjemite and Klemetsaunet rhyodacite, Bymarka ophiolite).

Metamorphosed Ordovician-Silurian calc-alkaline intrusive igneous rocks in the Støren Nappe. Felsic (e.g., trondhjemitic) plutons, ranging in size from small dikes to >100 km², cut volcanics in the Trondheim Nappe Compex in many places within and to the south and east of Trondheim (e.g., Støren, Innset, Gauldalen areas; Sigmond et al., 1984; Dunning & Grenne 2000, Nilsen et al. 2003, 2007). Large, relatively undeformed batholithic masses of similar age occur also in the Støren Nappe ~100 km west of Trondheim, such as the Smøla-Hitra Batholith (Gautneb & Roberts 1989, Nordgulen et al., 1995). In addition there are extensive areas of deformed plutonic complexes assigned to the Støren Nappe that occur to the north and southwest of Trondheimsfjord that have little or no associated volcanics except possibly as xenoliths. These rocks include a wide range of deformed rocks from hornblendite to granite, with tonalitic gneisses dominating. Though deformed gneisses, these rocks have extensive evidence of crosscutting relationships that include dikes and intrusion breccias. The geochemistry of these rocks is summarized in Figure 5.6 based on a new data set (Hollocher et al. unpublished). Figure 5.6A shows that these plutons span a continuous composition range from
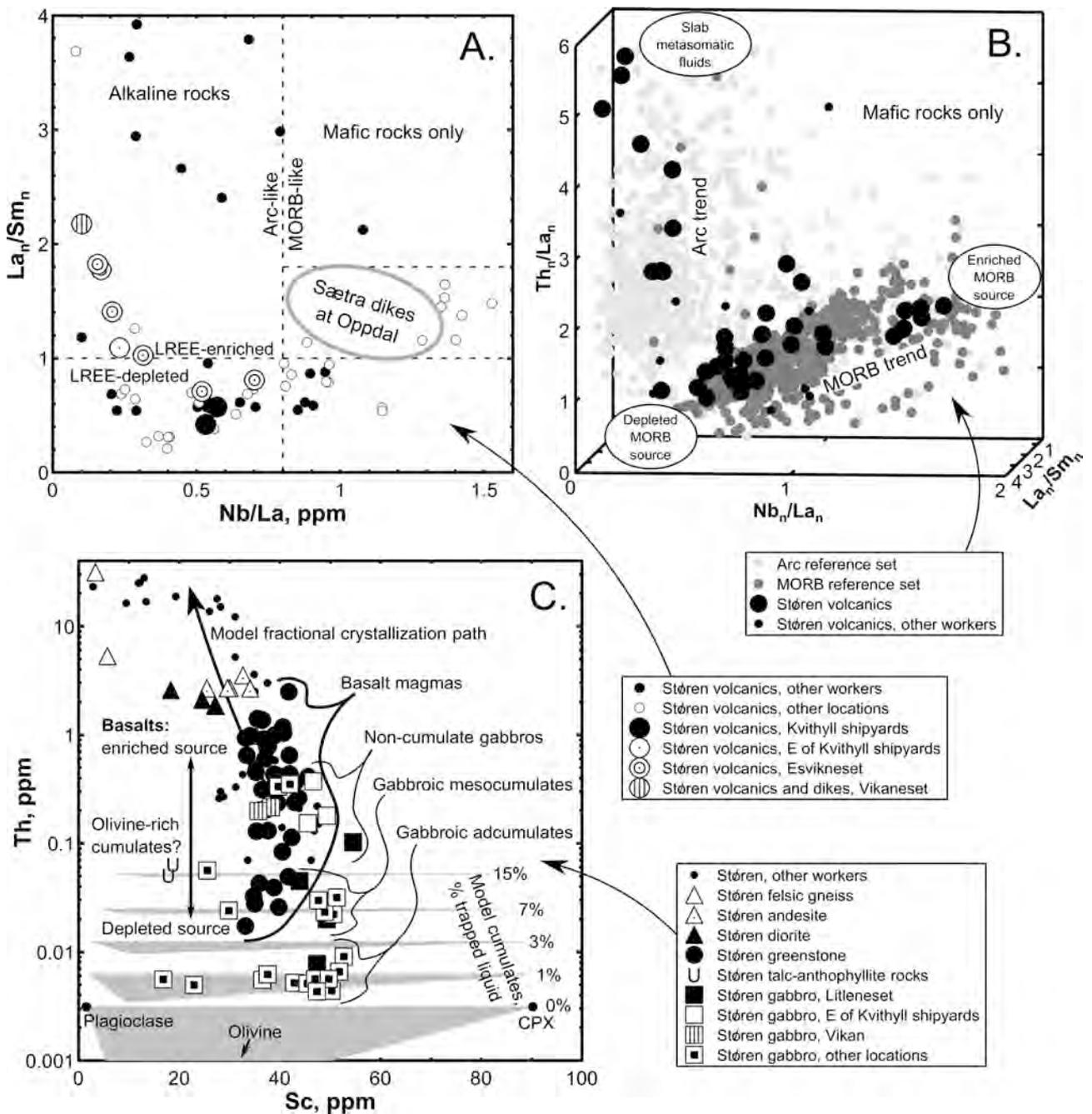
Fig. 5.5A-C Geochemistry of volcanic and plutonic rocks in the Støren Nappe and equivalent units, excluding extensive deformed gneisses to the north, west, and southwest of Trondheimsfjord (Lensvik synform, Valset area, Kjørsvik area, Rissa, Kopparen, Råkvågen area). A) Nb/La vs. La_n/Sm_n diagram comparing mafic Støren volcanics to Sætra Nappe dikes at Oppdal, and illustrating the range of rocks in the Støren Nappe that include MORB-like, arc-like, and alkaline basalts. B) Overview of mafic Støren volcanics in comparison to modern arc and MORB composition fields. The three axes were oriented for best separation of the two fields. C) Sc-Th diagram illustrating relationships between Støren volcanics and coarse gabbroic rocks associated with the volcanics.

45-77% SiO_2, with no bimodal tendency as is seen in the volcanics (Fig. 5.5D). The most mafic of these deformed plutonic rocks are tholeiitic, whereas those with >57% SiO_2 are dominantly calc-alkaline. Though not bimodal, this pattern is the same as the volcanics of Figure 5.5D. As an exception, mafic Lensvik synform rocks (mostly dikes, see Stop 5-6) span the tholeiitic-calc-alkaline line whereas the more silica-rich rocks tend to be tholeiitic.

Figure 5.6B shows in the center a diagram that separates rocks according to LREE-enrichment and saddle-shaped MREE and HREE patterns. First pay attention to the Bymarka ophiolite felsic rocks (gray background, Slagstad, 2003, discussed above), which illustrate the characteristic REE pattern shapes seen in the other samples. The Fagervika Trondhjemite has REE patterns that are typical of many felsic rocks throughout the world: LREE-enriched with flat or slightly saddle-shaped HREE patterns that suggest a source containing considerable residual pyroxene or amphibole but
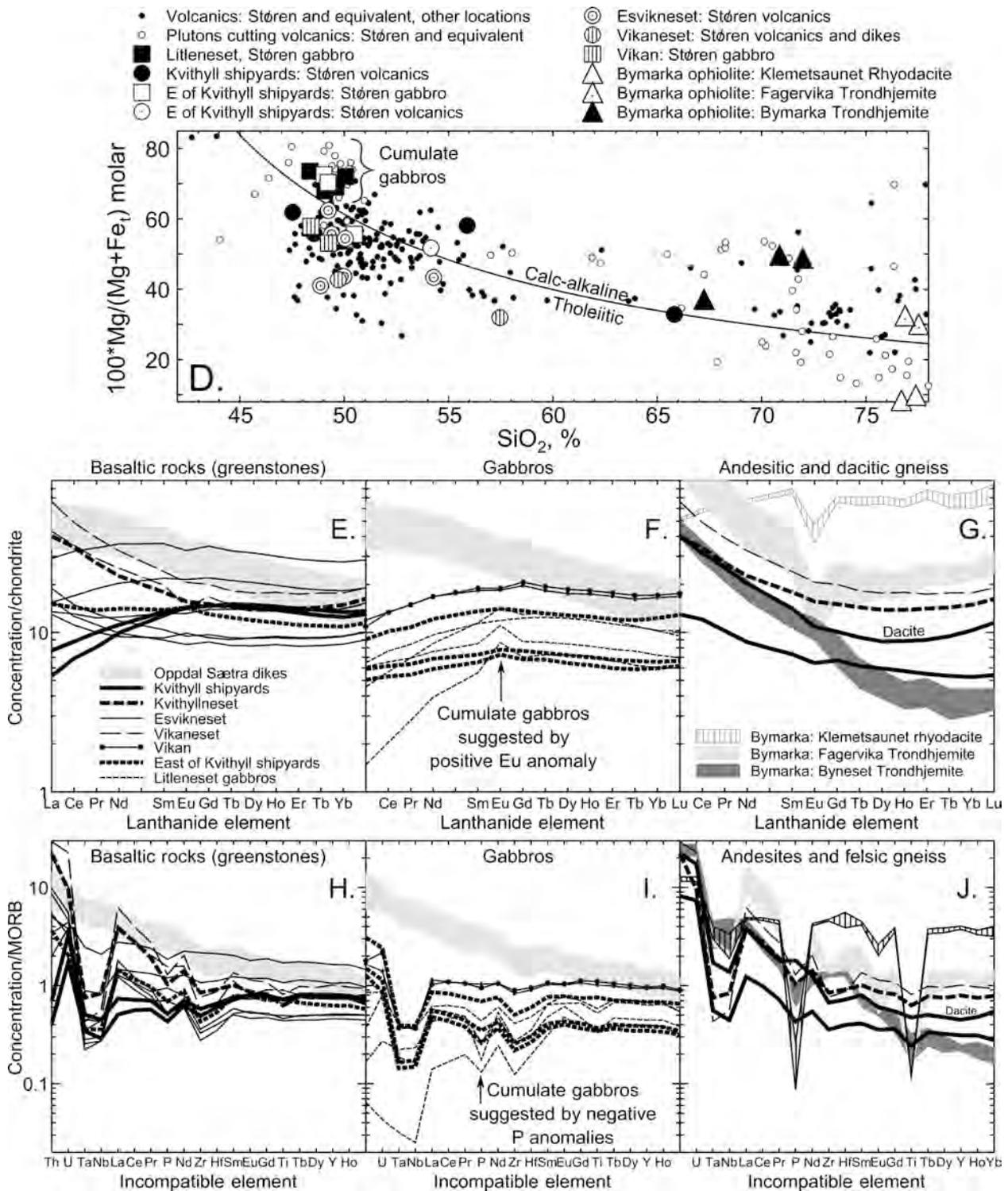
Fig. 5-5D-J D) SiO₂ vs. Mg' diagram showing data from Støren (and equivalent) volcanic rocks and plutons associated with volcanics in central Norway and west-central Sweden. Discriminant line is from Miyashiro & Shido (1975). E and H) REE and spider diagrams for mafic volcanics at and near Støren outcrops visited on Day 5, illustrating the diversity of Støren greenstone compositions in the relatively small area around Trondheimsfjord. F and I) REE and spider diagrams for coarse-grained gabbroic rocks from Trondheimsfjord shores. G and J) REE and spider diagrams for intermediate and felsic rocks, in comparison to felsic rocks in the Bymarka ophiolite fragment, Støren Nappe, Trondheim (Slagstad, 2003; see text). REE normalizing factors are from McDonough & Sun (1995), MORB normalizing factors are from Pearce & Parkinson (1993). Unless noted below, all data are from Hollocher et al. (unpublished). Gray dots in B are nearly 3000 volcanics rocks from the GEOROC and PETDB databases, selected from basalts in modern arc and ocean ridge tectonic environments, including only good analyses of relatively fresh rocks. Støren data of other workers are from Loeschke (1976), Grenne & Roberts (1980), Loeschke & Schock (1980), Roberts (1982), Stephens (1982), Grenne & Lagerblad (1985), van Roermund (1985), Heim et al. (1987), Grenne (1988), Roberts (1988), Grenne (1989), Grenne & Roberts (1998), Dunning & Grenne (2000), Nilsen et al. (2003), Slagstad (2003).
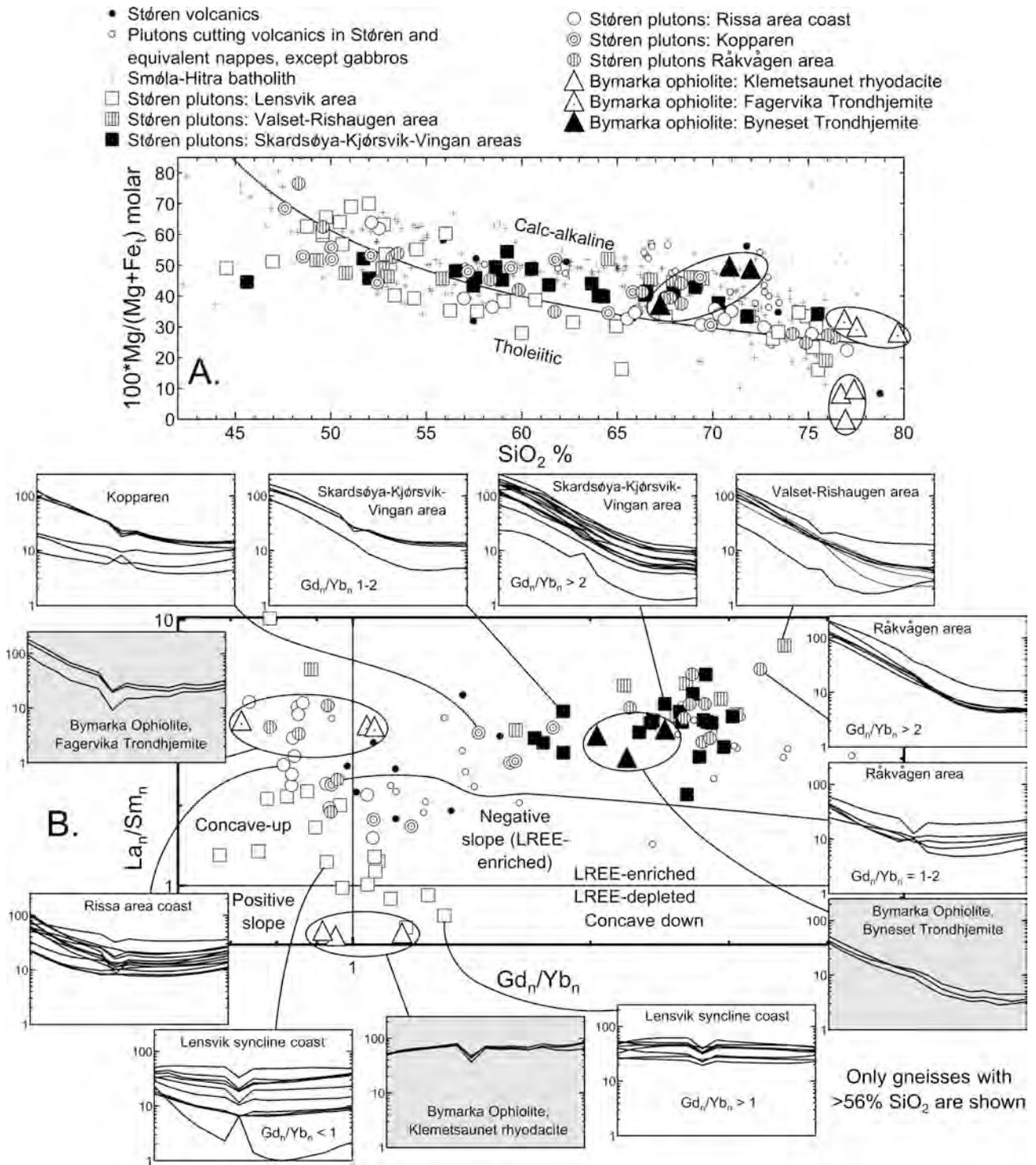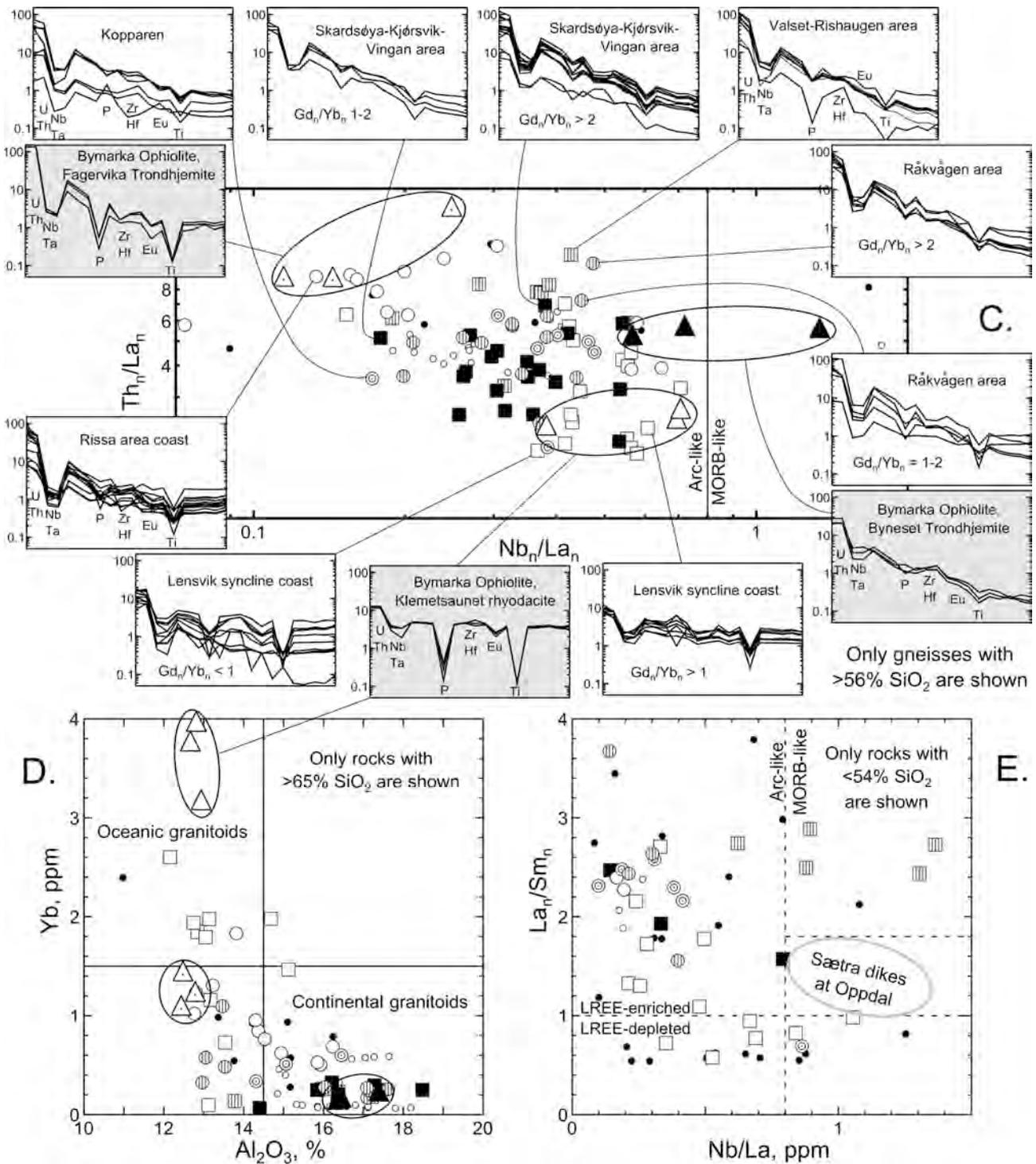
Fig. 5.6. Geochemical summary of deformed plutonic rocks in the Støren Nappe in the Lensvik synform (near Lensvik), Skarsøya-Kjørsvik-Vingan area (coastal region southeast of Hitra and Smøla islands), Valset-Rishaugen area (western Trondheimsfjord), the Rissa area coast, Råkvågen area (northeastern end of Stjørnfjorden, north of Rissa), and Kopparen (northwest of Bjugn). Symbols for the whole figure are defined in A. A) SiO2 vs. Mg' diagram showing plutonic rocks, excluding gabbros associated with volcanics (Fig. 5.5C). The discriminant line is from Miyashiro and Shido (1975). B) Gdn/Ybn vs. Lan/Smn diagram in the center surrounded by REE diagrams (only rocks with >56% SiO2). Bymarka Ophiolite felsic rocks are shown for comparison (Slagstad, 2003, see text). Lines associate diagrams with their respective symbols and Gdn/Ybn ratio ranges. C) Nbn/Lan vs. Thn/Lan diagram in the center, surrounded by spider diagrams showing the same rock sets as in B. The spider diagrams are too small for X-axis labels (see Fig. 5.4B-E), so anomalies are labeled on the less cluttered diagrams. Lines associate spider diagrams with their respective symbols. D) Al2O3 vs. Yb tectonic environment discriminant diagram after Arth (1979),(caption continued next page)

showing only samples with >65% SiO$_2$. E) Nb/La vs. La$_n$/Sm$_n$ diagram showing only mafic rocks, compared to Sætra Nappe dikes at Oppdal (reference lines after Hollocher et al., 2007). REE normalizing factors are from McDonough & Sun (1995), and MORB normalizing factors from Pearce & Parkinson (1993). Data are from Loeschke (1976), Loeschke & Schock (1980), Heim et al. (1987), Grenne (1988), Nordgulen et al. (1995), Grenne & Roberts (1998), Pannemans & Roberts (2000), Roberts & Sundvoll (2000), Nilsen et al. (2003), Slagstad (2003), and Hollocher et al. (unpublished).

little garnet. The Byneset Trondhjemite has similar LREE-enrichment slopes and concentrations, but is depleted in HREE suggestive of considerable residual garnet in the source rocks. In contrast, the Klemetsaunet Rhyodacite has spectacularly high HREE, yet is LREE-depleted. This is suggestive of derivation from a LREE-depleted, garnet-free (noritic?) source, or from crystal fractionation of MORB-type basaltic magma (Slagstad, 2003). The Fagervika and Klemetsaunet rocks also have large negative Eu anomalies suggestive of large amounts of plagioclase separation. Rocks that have REE patterns similar to those of the Fagervika Trondhjemite occur along the Trondheimsfjord coast near Rissa and in the Lensvik syncline (Stop 5-6), and about half of the rocks from the Råkvågen area (Stops 5-14, 5-15) to the northwest of Rissa. Rocks like those in the Byneset Trondhjemite, with HREE-depletion indicative of

residual garnet in the source, include the other half of samples from the Råkvågen area and from the Skardsøya-Kjørsvik-Vingan coastal region southeast of Hitra. Rocks with REE patterns like those of the Klemetsaunet Rhyodacite occur in the Lensvik synform (Stop 5-6) and nowhere else in this data set. Kopparen samples cover the intermediate composition range between the three Bymarka Ophiolite felsic rock types.

Ignoring overall pattern slopes, which are largely defined by the lanthanides that have already been discussed, spider patterns do not vary systematically in any obvious way with region (Fig. 5.6C). All samples have small to large negative Nb-Ta anomalies suggestive of arc-related source rocks. Large negative P and Ti anomalies are indicative of separation of apatite and Fe-Ti oxides, but it is remarkable that many of these intermediate to felsic rocks have only small negative P and Ti anomalies. This indicates little crystal fractionation after extraction of the melts from their sources or, alternatively, extensive mixing of mafic and felsic magmas which would strongly subdue these anomalies. The lack of a bimodal distribution (Fig. 5.6A) may indicate extensive magma mixing processes, a possibility not contradicted by the extensive intrusion breccias visible in outcrop. A few samples have positive P and Ti anomalies, and these may be cumulates, also suggested by the few samples having positive Eu anomalies (Fig. 5.6B), or possibly plagioclase-rich residua from melting (unlikely, considering the general lack of partial melting textures in these rocks).

Figure 5.6D shows a simple discriminant diagram for felsic rocks. Oceanic granitoids that include ocean floor "plagiogranites" tend to have Al_2O_3 contents <14.5% and Y contents >1.5 ppm, like rocks in the Klemetsaunet Rhyodacite and several rocks from the Lensvik synform. Continental granitoids include rocks from post-collision and arc tectonic settings and tend to have >14.5% Al_2O_3 and <1.5 ppm Y, like the Byneset Trondhjemite and many rocks from the Råkvågen (Stop 5-14, 5-15), Skardsøya-Kjørsvik-Vingan, Kopparen, and Valset-Rishaugen (Stop 5-9) areas. Many rocks, however, plot in an unnamed quadrant having <14.5% Al_2O_3 and <1.5 ppm Yb, including the Fagervika Trondhjemite and some rocks from the Råkvågen, Lensvik synform, and Rissa areas. Rocks in this field are not mechanical mixtures of materials (as sources or magmas) in the continental and oceanic granitoids fields, and so must represent a different kind of source rock or processes.

Figure 5.6E is the same base diagram used for Sætra dikes (Fig. 5.4B-E), with only mafic rocks from the deformed Støren plutonic complexes plotted. At first glance there appears to be no particular pattern, with data scattered throughout the composition space. Most samples, however, plot on the arc-like side of the vertical reference line, and a substantial fraction are LREE-depleted as are the Støren volcanics. Four samples from the Valset-Rishaugen area (Stops 5-14, 5-15) are LREE-enriched and MORB-like on this diagram, somewhat like the Ystland-type dikes of the Sætra Nappe (Fig. 5.4D). Figure 5.6E indicates great complexity in source regions and processes in the generation of these rocks.

Overall, the highly deformed plutonic complexes in the Støren Nappe to the north and southwest of Trondheimsfjord tell a story of diverse sources and processes. They seem to have been ultimately derived from mostly arc-like, rarely MORB-like sources, somewhat akin to the volcanics. Some of the felsic plutonic rocks seem to have been derived from garnet-bearing sources, though none of the few available felsic volcanics appear to have been derived from garnet-bearing sources. Some were derived from LREE-enriched sources, and some are clearly from LREE-depleted sources. Some underwent considerable crystal fractionation and others not, and the lack of a bimodal distribution in silica content, and textures in some outcrops, suggest that magma mixing was an important process. Little modeling has yet been done on any of these rocks, particularly those in the Hollocher et al. (unpublished) data set, so much work remains.

STRUCTURAL AND TECTONIC DEVELOPMENT

Introduction. The region covered on Day 5 provides a major change of tectonic style of Scandian deformation. Days 1-4 were concerned with characterization of the strata within the nappes including their ages, differences in metamorphic history and characteristics of fault surfaces between them. By contrast, on Day 5, we enter a region where all of the basement has been involved in one or more stages of ductile deformation, and where the nappes have been deeply, though variably involved with deformation of the basement rocks. In recent years it has become progressively more obvious that the same nappes observed in central Sweden and eastern Norway are present, but extensive work has been involved in identifying them and also in overcoming the contrary opinions of many sceptics. A particular challenge has been to comprehend and trace units that may range in thickness from thousands of meters down to as little as one meter, where units can be found in the field only by determined search, commonly using maps of larger than 1:50,000 scale. Without this tracing of units, it is not possible to create comprehensible cross-sections from which to derive a picture of the structural architecture of the region, much less the kinematic and dynamic evolution, except in the very latest phases of evolution. Figure 5.7 from Tucker (1986) provides some appreciation of multiplicity of deformations of both basement and cover, though in this case on a relatively large scale compared to some areas. In addition, the measurement and interpretation of minor structural features is a continual challenge, particularly due to the fact that the predominant deformation fabric in most outcrops in the region covered by Day 5 was imposed during late ductile phases of extension along axes quasi-parallel to the orogen. This means that observations of earlier fabrics in much of the area of Day 5 have been sparse and uncertain, in contrast to observations made in the region of Day 9. One key feature of the late ductile deformation fabrics is to see that they form a continuum with lower-temperature mylonite fabrics and brittle features associated with extension related to the Devonian sedimentary basins. A further complicating feature as the coast is approached is the appearance of Late Paleozoic, Mesozoic and Cenozoic

extensional faults. The lack of post-Devonian cover in land areas makes their timing and magnitude difficult to assess, but the results of offshore seismic surveys and petroleum exploration provide an exciting picture of a complex history. For all of the above reasons, future detailed tectonic reconstruction of this region will require dedication and hard work on several fronts.
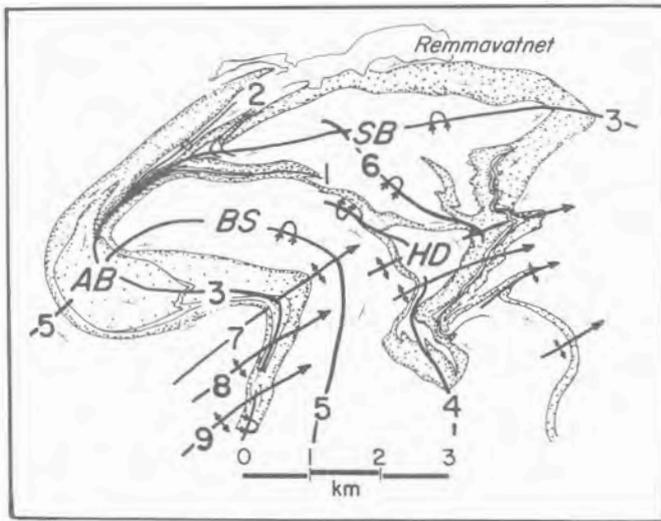

Fig. 5.7 Fold axial surface map of the Remmafjell structure southwest from Lensvik showing major structural features and the traces of various presumed Scandian surfaces (from Tucker 1986). Arrow denote trends and plunges of map-scale folds. Stippled areas indicate cover rocks, unpatterned areas are Baltican basement. AB – Almfjell basin; BS – Bergsvard synform; HD – Heklefjell dome; SB – Skropligdal basin. Numbers refer to axial surfaces of folds described by Tucker, with lowest numbers related to earliest folds, highest numbers to latest.
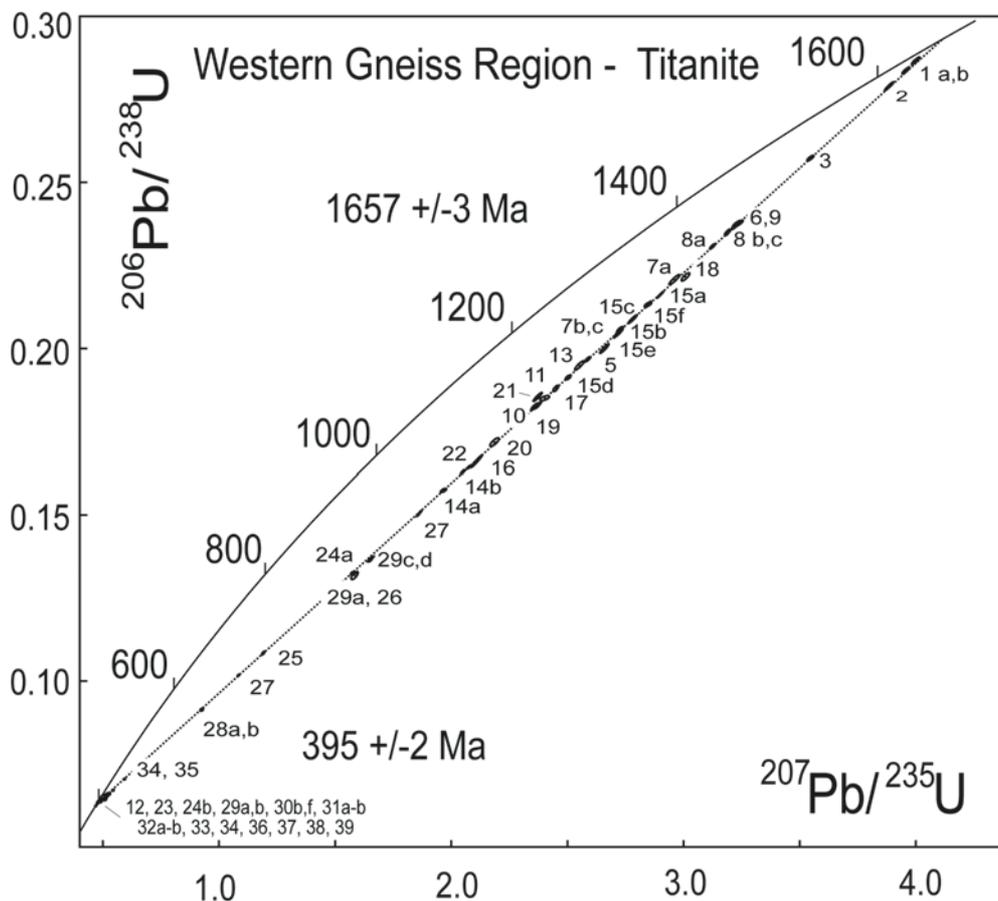

Fig. 5.8 Concordia diagram from Tucker et al. (2004) showing U-Pb ages of variably reset titanites from an extensive area of Paleoproterozoic granitoid gneiss of the Western Gneiss Region. Sample localities are shown on Figure 5.9, which also shows the % resetting of Proterozoic titanite from very low values to 100%. The remarkable fact that all of the titanite analyses fall tightly on a chord between the igneous protolith age of 1657 ± 3 Ma and the end of Scandian Pb loss at 395 ± 2 Ma demonstrates two important features: 1) the lack of significant disturbing events between these times and 2) the brevity and temporal uniformity of the end of Scandian heating over a broad region.

Early Scandian nappe emplacement, deformation and metamorphism. The nappes, as known farther east, were obviously emplaced by long-distance transport from the NW, but unlike locations to the E, e.g. Days 1-4 and also Oppdal Day 6, Stop 6-7, the fabric record of this is scanty, and consists mainly of folds in various locations that are geometrically transverse to the dominant shallow-plunging ENE-WSW trend. One generalized observation is that, as
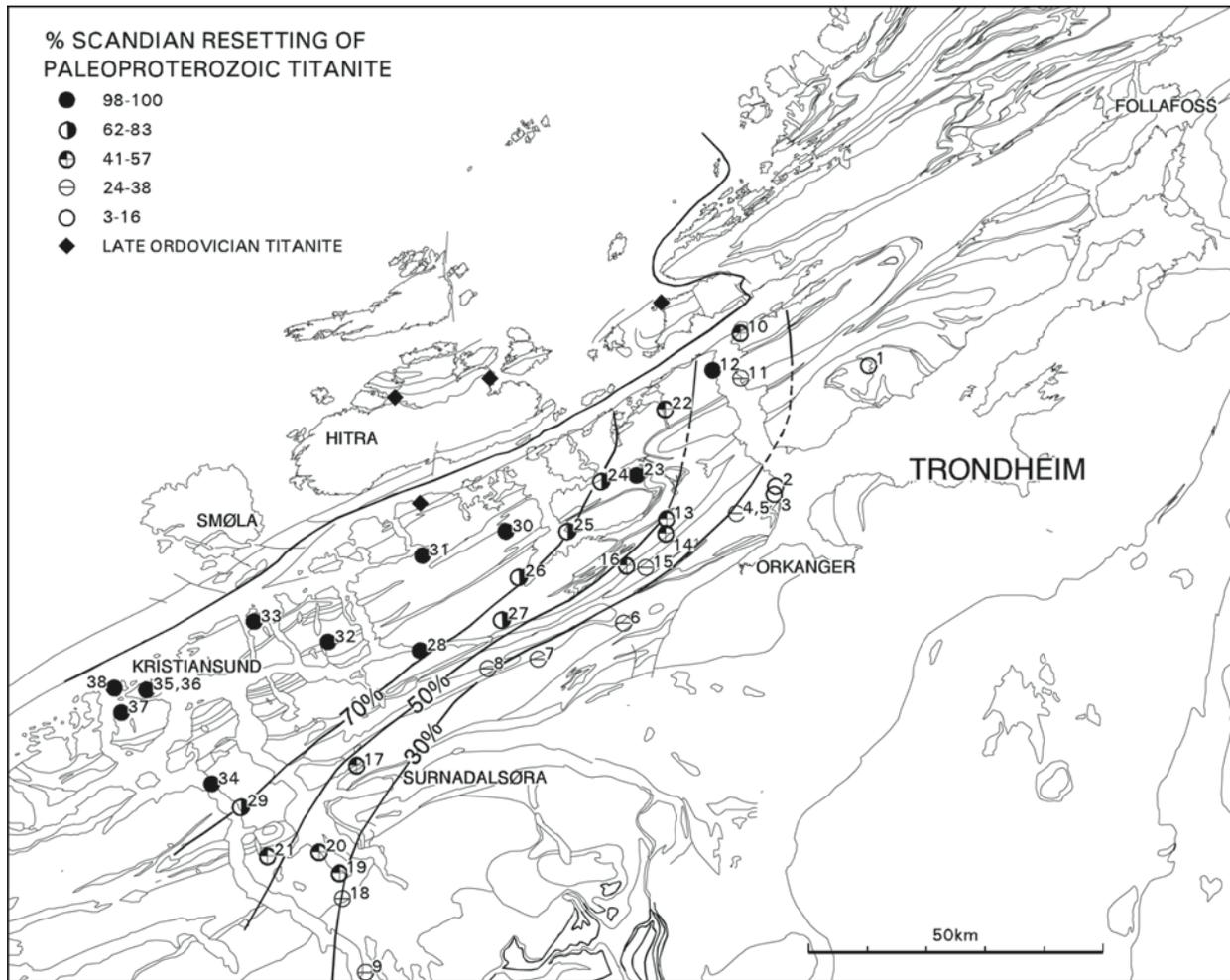
Fig. 5.9 Titanite localities and % resetting contours for basement gneiss samples plotted on the base map of Figure 5.1 and farther to the southwest from Tucker et al. (2004). Data points are divided into five groups according to % resetting. Locality 12, which appears to violate the resetting contours, contains titanite in marble formed only by neocrystallization (Stop 5-8). The intensity of resetting increases in the general direction of the large area of the Støren Nappe on the islands of Smøla and Hitra. Also shown are the four localities of titanite samples from Ordovician intrusive rocks (Fig. 5.10) that show essentially no Scandian resetting.

one proceeds farther and farther NW into the hinterland, evidence for the presence of lower nappes becomes increasingly scarce. Supracrustal rocks that might be assigned to the Lower Allochthon are extremely rare NW of Orkanger. The prominent Risberget augen gneiss nappe appears to die out 10-12 km NW from Orkanger. The Särv = Sætra Nappe can be recognized as far north as Agdenes, though only in layers 1-10 meters thick, whereas the high-amphibolite-facies rocks of the Seve = Blåhø = Skjøtingen Nappe persist still farther west.

Close to Stop 5-1 the Seve rocks do provide key information on an age of deformation and metamorphism to be compared with results from Åreskutan on Day 3. High-amphibolite-facies metamorphic and deformation fabrics are clearly truncated by pegmatites dated at 431 +/-2.9 Ma and 422.4 +/-1.8 Ma (see Figure 5.13A,B), themselves also strongly deformed. So these rocks were strongly influenced by a pre-mid-Silurian high-grade metamorphism, then deformed again. By contrast, the local basement beneath appears to have been overprinted by an eclogite-facies metamorphism only at 401 +/-2 Ma (Early Devonian – Late Emsian) (see Fig.13C) and then returned to lower-amphibolite-facies conditions at 395 Ma (see below).

Early Scandian extension. Evidence for the earliest phase of extension is provided mainly by complex geochronology, in fact a serendipidous result from performing U-Pb measurements on all available titanites from a series of basement gneisses (Tucker et al. 2004), all of which yielded Mesoproterozoic U-Pb zircon ages (Tucker et al. 1990). The titanite results from many samples plot on a chord across Concordia (Fig. 5.8) from an upper intercept age of 1657 +/3 Ma, the age of the Mesoproterozoic basement, to a lower intercept of 395 +/-3 Ma (Early Devonian- Late Emsian). Careful measurements on multiple small samples shows that the titanites are not zoned, but uniformly partially lost Pb by a process poorly understood and found in only a few locations, worldwide.

The positions of the samples on the chord are described in terms of % reset from the primary age, and these percentages are used in Figure 5.9 to create a map of spatial distribution. The least reset samples are from close to Trondheim (1-6) and near Sunndalsøra (9) and percentages increase to a broad region of 98-100% against a sharp line
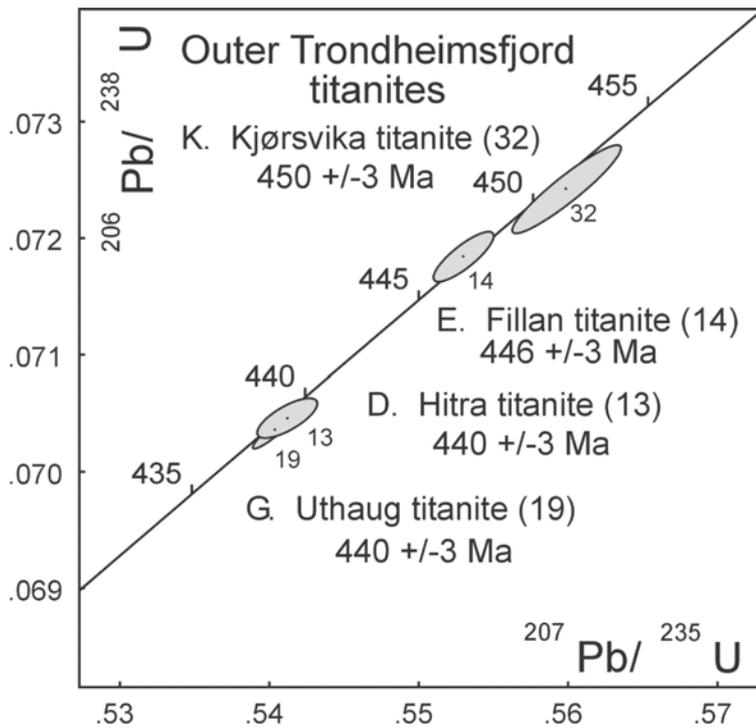
Fig. 5.10 Concordia diagram showing U-Pb ages of unreset or very weakly reset igneous titanite ages from four localities marked on Figure 5.9 of Ordovician intrusive rocks in the Støren Nappe. The oldest age of 455 Ma is from the foliated and lineated diorite at Kjørsvika which lies only 1-2 km across the Agdenes detachment from basement gneiss with 100% resetting of Proterozoic titanite at 395 Ma.
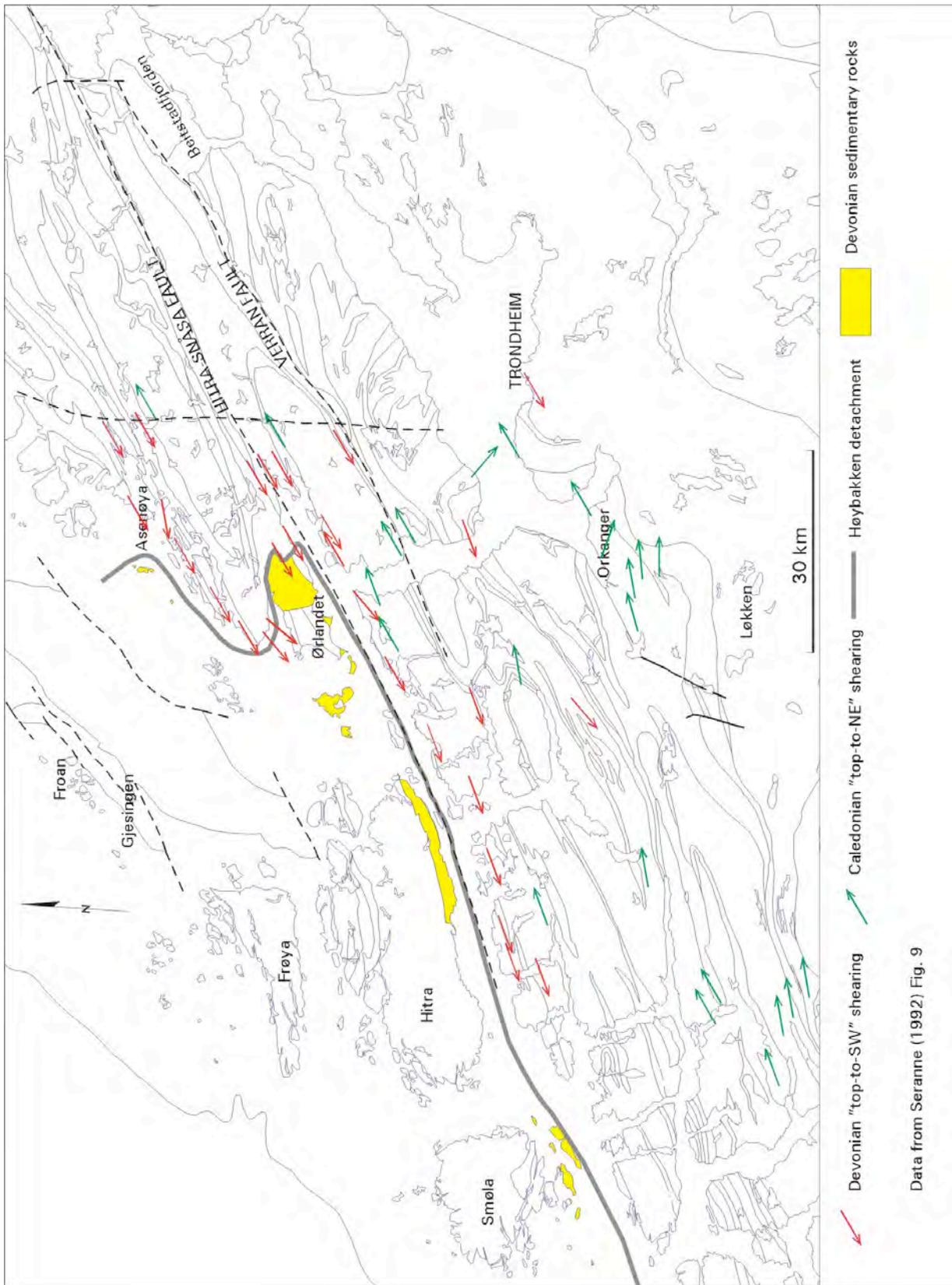
SE of the Hitra-Snåsa Fault. Just NW of this line at Kjørsvika, an Ordovician diorite with a zircon age 460.7 +/- 2.3 Ma (Llanvirn-Caradoc) contains titanite hardly reset (Fig. 5.10) at 455 +/-3 Ma (middle Caradoc). The implication of these results is that the Mesoproterozoic basement was subducted and warmed extensively with Pb loss, until lead loss was stopped by cooling at about 395 Ma. By contrast the Ordovician plutons across the sharp contact were not heated sufficiently during the Scandian to achieve any significant degree of lead loss.

The compelling solution to this problem is to postulate that the contact is an early extensional detachment, the Agdenes detachment, that brought high-level relatively cool Ordovician rocks against high-grade subducted basement. The extension had the double effect of providing a relatively rapid cooling to the uplifted warm basement at about 395 Ma, and providing modest heat to the overlying granitoids. In any case it is evident that after extension both sides of this proposed detachment were ductilely deformed together, producing the common subhorizontal extension fabric that dominates the region. The gravitational potential for the Agdenes detachment might have been supplied slightly earlier by the major mid-Scandian thrust imbrication of the Storli thrust, as described under the Geology of Trollheimen for Day 6.

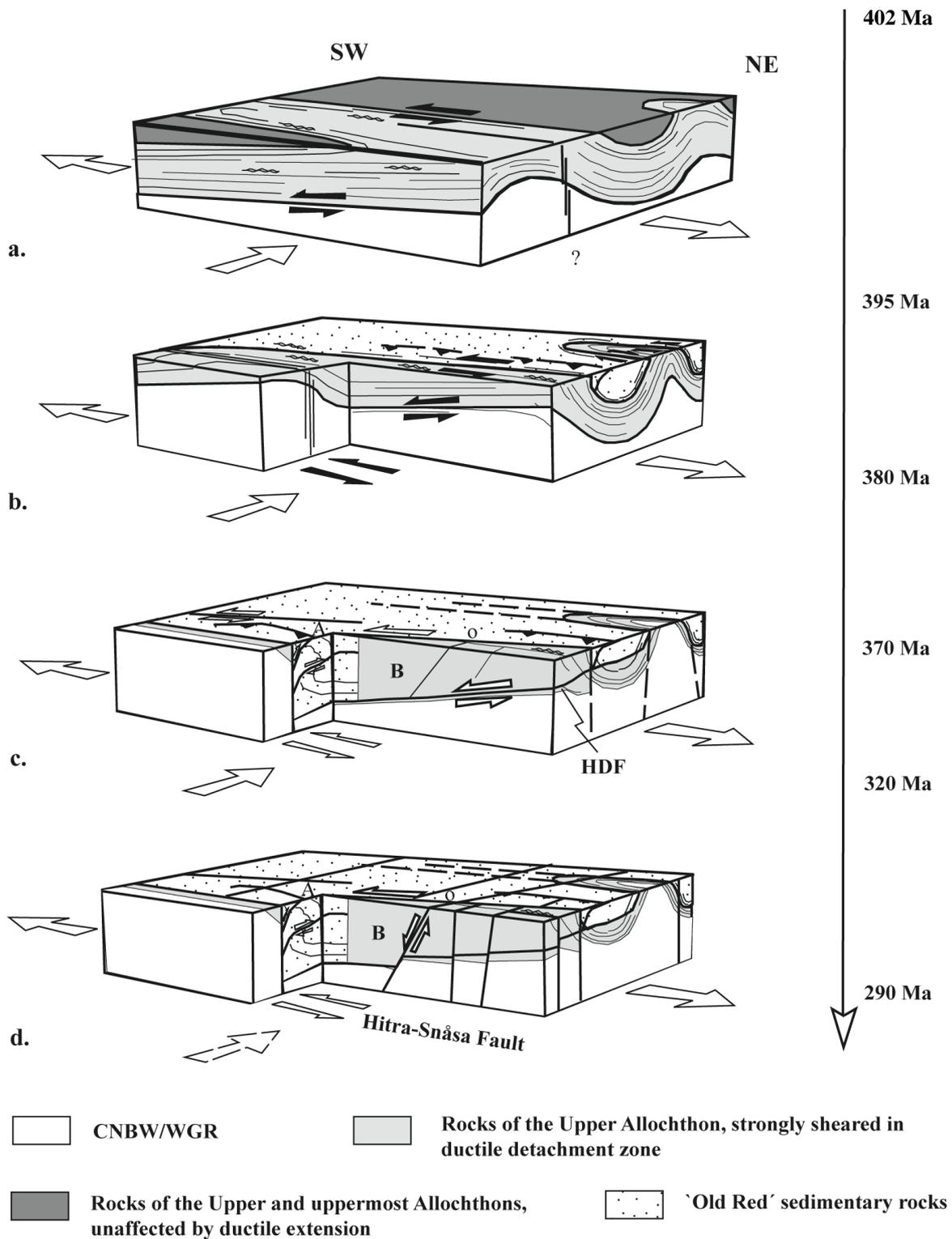**Late Scandian orogen-parallel extension.** The region is pervaded by subhorizontal sinistral or top-W to -SW extensional shear fabrics associated with a powerful subhorizontal lineation. That this fabric occurs both above and below the Agdenes detachment, indicates that it was superimposed near the end of or after movement on the detachment at around 395 Ma. Terry & Robinson (2003) demonstrated that these fabrics are a result of longitudinal constrictional strain, and suggested that the constriction was due to transtensional deformation, where high-level extension was passing into a region of oblique collision while at the same time horizontal oblique convergence at deep levels was providing the crustal imbrication and the gravitational potential for high-level extension. Krabbendam & Dewey (1998) proposed that the identical constrictional fabric was provided by oblique continental separation following the Scandian collision. A characteristic of such constrictional fabrics is that the lineation and fold axis orientation changes very little over long distances, but the orientation of fold axial surfaces can be quite chaotic over short distances, even in a single outcrop. In some locations below the Agdenes detachment, the prominent stretching fabric is assocated with boudinage of more competent layers and infilling of boudin neck lines by pegmatite. In the high-grade regions southwest from Trondheimsfjord visited on Days 9 and 10, these pegmatites have given consistent U-Pb zircon ages of 395-394 Ma.

In studies of the late ductile fabrics, is it multiply evident that the same sinistral strain field accompanied denudation across the brittle-ductile transition, producing progressively more mylonitic and ultramylonitic fabrics in progressively more concentrated zones, and with shear planes gradually oriented progressively from ENE-WSW toward NE-SW. It appears that near the Høybakken detachment zone (Fig. 5.11), the same deformation continued into a late brittle phase, marked by the Høybakken detachment fault, to be discussed at Stop 5-12. The Høybakken detachment and associated Devonian rocks are banked against the Hitra-Snåsa Fault of the Møre-Trøndelag Fault Complex. Based on relative chronology and on a number of ages from U-Pb zircon and ^{40}Ar/^{39}Ar dating, the following chronology appears to apply (Fig. 5. 12): At some time after c. 402 Ma, the rocks of the WGR were cooling in concert with development of a ductile extensional detachment zone. From this time onwards through c. 395-390 Ma, the rocks of the Western Gneiss Region were cooling rapidly. In the mylonites below the Høybakken detachment, exhumation from amphibolite facies at c. 390 Ma into the elastic-frictional regime at c. 370-356 Ma is recorded by ^{40}Ar/^{39}Ar ages from hornblende, micas and feldspar (Kendrick et al. 2004). In the Late Devonian or later, post-dating the Late Devonian cooling ages, the footwall appears to have yielded material to the adjacent Asenøy basin (Eide et al. 2005). As the exposed footwall section entered the brittle regime, the brittle detachment fault sliced through the folded detachment mylonites and captured a slab of mylonites in its hangingwall. This interpretation provides an explanation for the occurrence of sheared and folded rocks in the hanging wall of the Høybakken detachment, that show ductile strain patterns that are similar to those in the footwall (Osmundsen et al. 2006).

Fig. 5.11 Map of same general area as Figure 5.9 showing locations of Devonian sedimentary rocks, and observations of shear sense made by Séranne (1992). The red arrows show his observations of top-west shear sense that he related to ductile extension beneath the Høybakken detachment. The blue arrows indicate his observations of what he believed to be a top-east shear sense older than the extension leading to basin formation. An alternative interpretation that these fabrics are actually part of the same strain field will be offered at Stop 5-1. The one red arrow near Trondheim is an additional arrow added during preparation of this field trip and represents the very clear evidence for top-west shear seen at Stop 4-11 at Korsvika.

Fig. 5.12 Schematic cartoon showing evolution of the Høybakken detachment zone through a series of stages ranging from the early Devonian to the Permian. **a.** After 402 Ma but prior to 395 Ma, top-to-the WSW extensional shearing starts in the ductile Høybakken detachment zone, in a constrictional strain field. Extensional shear zone starts folding. **b.** 395-380 Ma: continued shearing in Høybakken detachment zone. Devonian sedimentary rocks are brought in contact with rocks from the lower plate. Low grade metamorphism in the Devonian rocks. Tightening of extension-parallel folds. **c.** 370-320 Ma: Exposed section enters brittle regime. Brittle Høybakken detachment fault cuts folded footwall mylonites and captures a slice in the hanging wall. Folding and thrusting in the Devonian rocks.**d.** 290 Ma: Brittle activity on Hitra-Snåsa fault (Sherlock et al. 2004). Eventual truncation of the Høybakken detachment by younger faults.
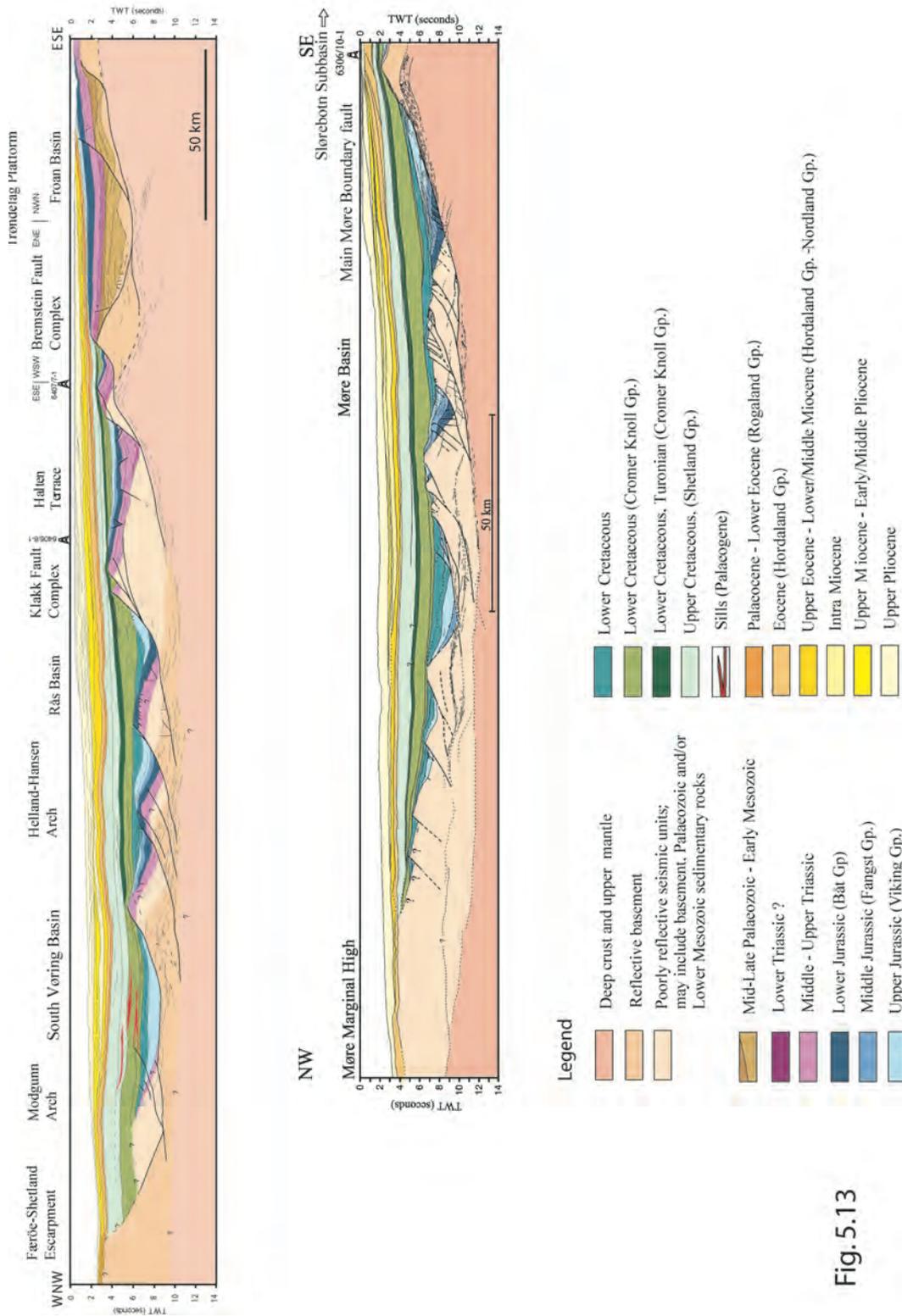
Fig. 5.13 Geoseismic sections from the southern Vøring (upper section) and northern Møre (lower section) basins, showing main elements in the stratigraphy and large-magnitude extensional faults separating principal domains in the rift (slightly modified from Osmundsen et al. 2002). The Bremstein Fault Complex and the main Møre boundary fault are large-magnitude extensional faults, the main Møre boundary fault has a displacement in the order of 30-40 km, the Bremstein fault complex probably less, but the top of basement is not well displayed. Note also extensional (basin-floor) detachments under the central to northwestern basin areas. Sections are in two-way time, but with a relatively modest vertical exaggeration; dips are close to their real values. The top of the mantle is not differentiated on the figures, but the crustal thickness under the most highly extended areas is c. 5 kilometres or less (Osmundsen & Ebbing, in press).

Fig. 5.13

Late Paleozoic, Mesozoic and Cenozoic extension. The Devonian extensional deformation exposed onshore was followed by a series of rift phases from the Permian into the Early Palaeogene that became responsible for the formation of deep basins in offshore Norway and, eventually, for strongly magmatic breakup around 54 Ma. In the area of the Trøndelag Platform, which is located directly offshore from Fosen (Fig. 5.13), Palaeozoic into Early Mesozoic half-graben basins were developed on rotated fault-blocks and are preserved under relatively thin Jurassic and younger successions (Osmundsen et al. 2002, Müller et al. 2005). These were separated by moderately dipping faults with several kilometres of displacement. In some areas, the fault-blocks appear to detach above structural culminations that coincide spatially with strong positive magnetic anomalies (Osmundsen et al. 2002, 2005, Skilbrei & Olesen 2005, Ebbing et al. 2006). A tentative correlation has been made between the highly magnetic rocks offshore and strongly magnetic, granulite-facies rocks in the onshore gneiss culminations. Such a correlation invites to an extrapolation of the onshore detachment zones into the offshore area of the Trøndelag Platform (Skilbrei & Olesen 2005, Ebbing et al. 2006).

The Late Jurassic into Early Cretaceous rift phase did, however, produce a number of extensional detachment faults that accommodated displacements up to 20 kilometres or more, and that appear to have incised and displaced the older Palaeozoic configuration of faults and basins (Fig.5.13) and left structural domains such as the Trøndelag Platform, the Frøya High and the Slørebotn Subbasin in structurally high positions. The Jurassic-Cretaceous detachments define the border between normal to moderately thinned continental crust and very highly thinned crust; northwest of the basin-flank detachments. The crust is less than 10 kilometres thick on the average and in some areas thickness is 5 km or less. These faults are thus important, crustal-scale boundaries. One may speculate whether these faults also reactivated Devonian structural grains, but it must be kept in mind that in mid Norway, the maximum elongation trend changed by almost 90º from the Devonian to the Late Cretaceous (Mosar et al. 2002). Therefore, a one-to-one reactivation of the undulating Devonian detachment faults, for instance, appears as less likely during the main extensional events in the Jurassic and Cretaceous. Still, thick mylonite zones related to Devonian detachment faulting may have guided location of younger detachments.

Strongly magnetic basement culminations appear to be associated also with the Jurassic-Cretaceous large-magnitude extensional faults. This may indicate that warping and truncation of magnetic high-grade rocks in the footwalls of the Jurassic-Early Cretaceous faults produced a new set of basement culminations (Osmundsen et al. 2005, Gernigon et al. 2006) related to the Jurassic-Cretaceous rift phase. Geophysical data indicate the presence of thick, intermediate-density, strongly magnetic rocks under the platform area, but that these rocks become very thin under the deep Møre and Vøring basins (Ebbing et al. 2006 and in press). The areas characterised by the strongest thinning gradients correspond to the locations of the Late Jurassic-Early Cretaceous large-magnitude faults.

The Møre-Trøndelag Fault Complex (MTFC), which is interpreted as a major Devonian sinistral strike-slip zone and which appears to cut the Høybakken detachment, experienced renewed activity in the Permian (Watts 2001) and, probably, in Mesozoic and Cenozoic times. Firstly, the MTFC clearly defines the boundary between crystalline basement (probably rocks of the WGR) and Cretaceous rocks in the area of the Slørebotn Subbasin (Smelror et al. 1994, Grunnaleite & Gabrielsen 1995, Osmundsen & Ebbing in press), secondly, several generations of fault products and mineralizations, including cataclasite, zeolite and carbonate mineralizations as well as unconsolidated breccia and gouge, characterises parts of the MTFC (Grønlie et al. 1991), and are associated with dip-slip lineations (Redfield et al. 2005a, b). Moreover, the southeasternmost strands of the MTFC define a border for the high topography of southern Norway. Jumps in the apatite fission-track age pattern across the MTFC indicates that strands of the MTFC were active in or after the Late Cretaceous, most likely associated with the Cenozoic uplift of southern Norway (Redfield et al. 2005). Thus, as the principal elongation trend evolved from parallel to normal to the MTFC, the kinematics changed from strike-slip to dip-slip. Most likely, the MTFC ended up controlling the present-day topography of southern Norway (op. cit.).

ROAD LOG DAY 5

Time Total time	since prev.	Km Tot. km	since prev.	Notes
8:00				Depart NGU, pick up participants at Vandrerhjem en route.
8:20	20	0.0		Klett traffic circle. Leave E6. Go straight toward Orkanger onto E39. Kilometer log begins here. The entire route for Day 5 including all stops is given in Fig. 5.14.
		2.3	2.3	Bridge over Gaula River. Lowland on the right is the lower floodplain of the river, terminated by Øysand beach, which we pass just before entering the first tunnel. The mountain on the

left and coastal cliffs shoreward of the first tunnel are composed of southeast-dipping greenschists of the **Støren Nappe**. The section on this mountain is a southerly extension of the Bymarka ophiolite. Talc-carbonate rocks in the original basal section here were mined for construction of the oldest parts of Nidaros Cathedral in Trondheim (Tom Heldal, NGU, personal communication, 1997).

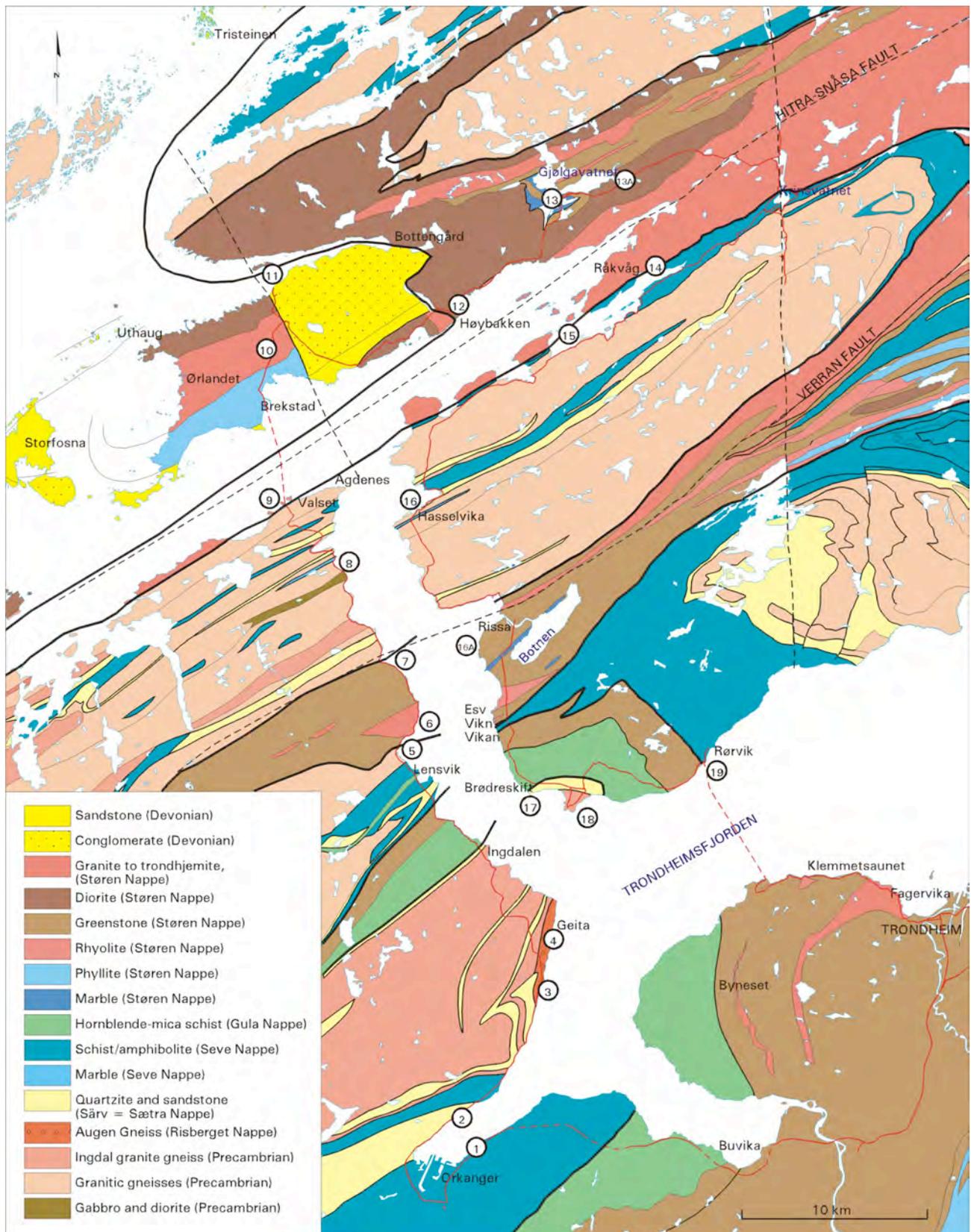|				
8:24	4	4.0	4.0	Bomstasjon (road toll plaza), Øysand (NOK 16). Enter first tunnel of E39 Øysand-Thamshavn Highway Project.

Fig. 5.14 Detailed map of the Trondheimsfjord showing the route of the field trip on Day 5 and all of the stops. Dashed parts of the route indicate ferry crossings or travel in major tunnels. Locations of the Agdenes and Høybakken detachments are highlighted in black and grey.
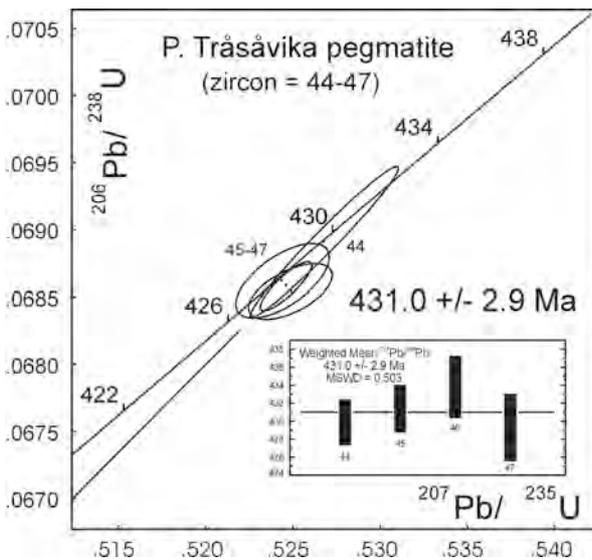
Fig. 5.15A Concordia diagram showing U-Pb ages of deformed early Scandian pegmatites in the Seve Nappe at Tråsåvika (431 Ma on Fig. 5.12).
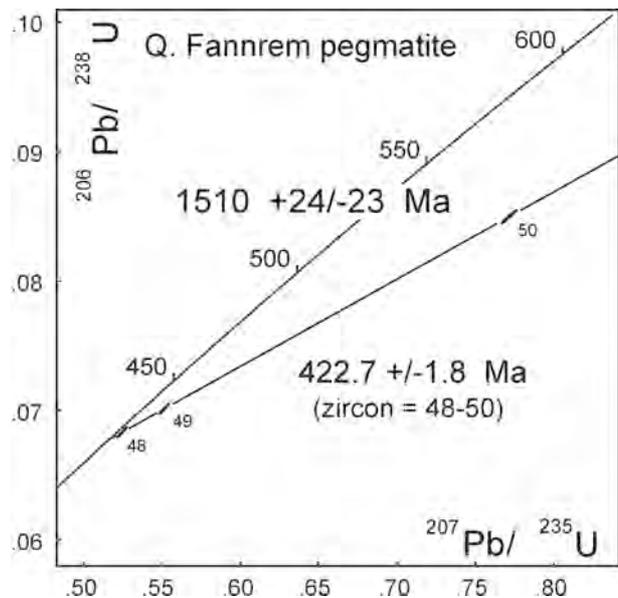
Fig. 5.15B Concordia diagram showing U-Pb ages of deformed early Scandian pegmatites in the Seve Nappe at Fannrem (423 Ma on Fig. 5.12).
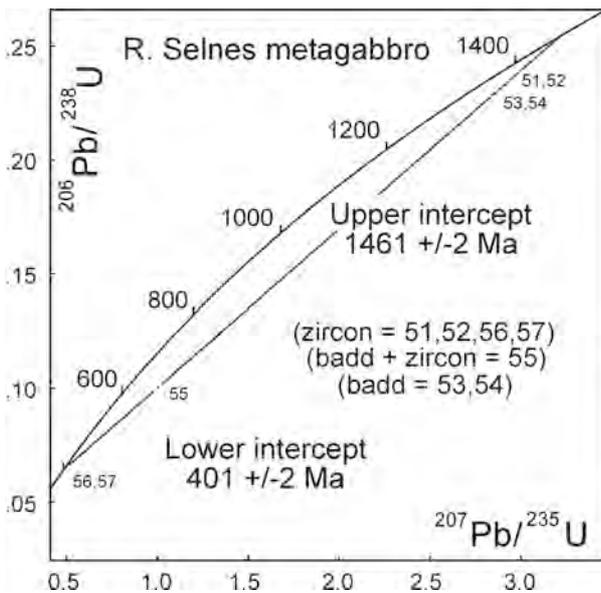
Fig. 5.15C Concordia diagram showing U-Pb ages of baddeleyite and zircon from metamorphosed Mesoproterozoic gabbro at Selnes (1461 Ma on Fig. 5.12). The baddeleyite gives an original igneous age of 1461 ± 2 Ma. Clear metamorphic zircon gives an age of 401 ± 2 Ma, and composite grains lie on a chord between these.

Detailed road log through the project is omitted, but major rock units are described. The entire first tunnel leading to Buvika is composed of amphibole-bearing greenschists of the **Støren Nappe**. Near Buvika are the Stormøllen grain storage towers. Entrance to the tunnel beyond Buvika is still in greenschists of the Støren Nappe, with the base of the Nappe in this tunnel. Through the next tunnels and openings all the way to the next major opening at Børsa and beyond, the rock is a rather homogeneous medium-grained garnet-biotite-hornblende-muscovite schist. On a regional scheme, this can be assigned to the Seve Nappe and is not unlike 'garbenschiefer' seen in the Seve in Sweden. In Norway this is similar to the **Blåhø Nappe** in its type area in Trollheimen 100 km S of here (see Stop 6-9, this guidebook), but in the Trondheim area has been assigned locally to the **Gula Nappe**, though not connected directly with the rocks of the Gula Complex seen on Day 4. In this section there are several highway exposures with important structural relationships. A series of thin, fine-grained, trondhjemitic granitoid dikes, as yet undated, sharply truncate layering and a strong tectonic foliation, but the dikes themselves are also ductilely deformed. Such trondhjemite dikes are common elsewhere in the Gula Nappe, but are not represented in the Seve = Blåhø = Skjøtingen Nappe. Børsa is the business center of Skaun Commune and the turn off for Skaun Village, the fictional site of the farm 'Huseby', made famous in the novel of Sigrid Unset, 'Kristin Lavransdatter'. The base of the 'garbenschiefer' lies within the long tunnel beyond Børsa.

Below lies a major unit composed of high-grade garnet and pyroxene amphibolites, garnet-biotite schists, and pegmatites that is typical of the **Seve = Blåhø = Skjøtingen Nappe** as mapped in much of the WGR. It continues through the next window, the second long tunnel, and the series of huge road exposures beyond, that will be visited briefly at Stop 5-1.

8:39	15	22.0	18.0	Bomstasjon Thamshavn (NOK 16)
		22.6	0.6	Exit right, drive back on old E39.
8:42	3	23.3	0.7	Park by truck turnaround, walk up (south) to Bomstasjon. Cross toll road with caution to N-facing exposures. There will not be time to walk far in either direction.

STOP 5-1: (19 minutes) Huge road cut in Seve =Blåhø=Skjøtingen Nappe with pegmatites.

The south-dipping strata consist of garnet amphibolite, diopside-epidote amphibolite and garnet-biotite schists, locally with sillimanite. A strong early tectonic foliation is cut by pegmatites, which are themselves deformed. These probably belong to the same group of deformed pegmatites which Tucker et al. (2004) dated at 431.0+/-2.9 Ma at Tråsåvika, N of here, and 422.7 +/-1.8 Ma at Fannrem to the S (Fig. 15A,B). These results indicate that the peak high-grade metamorphism in the Seve Nappe took place more than 25-30 m.y. before the peak Scandian metamorphism of the underlying basement at 400-395 Ma (Fig. 5.15C). Here, there is some evidence for late sinistral shear (top-NE), which appears to be typical along what is interpreted as the back-folded northern limb of the Surnadal Synform. This synform is interpreted as an original SE-facing recumbent synform as delineated near Surnadal (Rickard 1985, Krill, 1987). Séranne (1992) interpreted the late structures here as early, top-NE shear features unrelated to the predominantly top-SW and sinistral ductile features associated with the Høybakken detachment (see Fig. 5.10), but we think they are approximately synchronous. Compare these with the top-W to -SW shear features within the Støren Nappe at Korsvika (Stop 4-11).

9:01	**19**			Leave Stop 5-1 toward Orkanger on old E39.
9:07	6	28.0	4.7	Turn right (NW) off E39 toward Orkanger and Agdenes on Route 710.
		28.4	0.4	Traffic circle. Straight on 710. Cross Orkla River.
		29.5	1.1	Road cut. Blåhø schist.
		30.4	0.9	Large quarry on left. Layered amphibolite, +/- garnet, Blåhø Nappe.
9:10	3	31	0.6	Route to Hitra exit left. Stay straight toward Lensvik on Route 710, traveling NE along NW shore of Orkdalsfjorden.
		32.8	1.8	Road cuts in quartzite, amphibolite of Sætra Nappe on coast.
9:15	5	35.5	2.7	Almlia quarry to left.

STOP 5-2: (17 minutes) Sætra Nappe. Folds in metamorphosed diabase dikes cutting feldspathic quartzite.

This description is based on Krill (1981). "The rocks are strongly folded and foliated but cross-bedding and chilled contacts to the dikes are locally preserved. In less deformed parts of these rocks, 15-20 km to the southwest, the dikes clearly cut the sedimentary layering. In that area, cross-bedding and graded beds face southeast, away from the underlying unit of Risberget augen gneiss (Peacey 1963). The rocks in the quarry are folded by near vertical down-plunging open folds and have a marked steep lineation. It is not known whether this structure is of early or late origin. The dikes are 1-1.5m thick, showing a relict plagioclase porphyritic texture. Geochemically (Roberts pers. comm. 1980) and texturally they resemble the Ottfjellet dolerites of the Särv Nappe (Solyom et al. 1979), the dikes in the upper part of Leksdal Nappe (Andréasson et al. 1979) and the metadolerites in the Sætra Nappe at Oppdal (Krill, 1980). Near Oppdal the dikes in the Sætra Nappe are dated to 745+/-37 Ma (Rb-Sr) (Krill 1983b), which is similar to date obtained on the Ottfjellet dikes of the type area (Claesson 1976). The dikes are interpreted to represent magmatism connected to early rifting and opening of the Iapetus Ocean

Hollocher et al. (2007a) report major- and trace-element geochemistry of 15 dike samples from this outcrop, as well as four samples from Vasslivatnet, 50 km SW, giving more detailed confirmation of the geochemical correlations (Fig. 5.4). The ages reported above are questioned because in Sweden the dikes truncate probable Vendian glacial beds indicating the dikes are unlikely to be much older than 600 Ma. In examining the dikes, note that some are dark green porphyry dikes and some cross-cut bedding of feldspathic quartzite. The steep lineation and fold axes are thought to be early (Tucker 1986), and unrelated to the dominant shallow-plunging extensional fabric in the area.

9:32	**17**	35.5		Leave Stop 2. Road cuts in Blåhø schist along coast.
		39.9	4.4	Geitastrand. Baltican basement begins just beyond.
		44.5	4.6	Begin road cuts in augen gneiss of Risberget Nappe.
9:43	11	45.6	1.1	Turn right, side road to Kjøra and descend to village of Rønningen. Use cut-off of old road.
		46.3	0.7	Back to old road. Turn left (N) along coast.
9:45	2	46.6	0.3	Park on right at yellow house. Trail to coast and walk NE on beach.

STOP 5-3: (20 minutes) Risberget Augen Gneiss with Sætra dikes.

Rapakivi granite with a U-Pb zircon age by R. D. Tucker of 1189 +/-2.5 Ma (Fig. 5.16) is cut by mafic dikes. This age is more than 400 m.y. younger than any obtained in the local parauthothonous Baltican basement, supporting the long-held view that the augen gneiss is part of a far-travelled nappe. Both Krill (1981) and Tucker (1986) reported that distinctive foliation and mylonite zones are truncated by the dikes. Tucker (1986) reported that the geochemistry of the dikes is nearly identical that of dikes cutting Neoproterozoic sandstones in the Särv Nappe in Sweden, and this is confirmed in detail by Hollocher et al. (2004) for five samples from these outcrops (Fig. 5.4). A sixth sample is from a mafic boudin in augen gneiss that is cut sharply by a Sætra dike, and has completely different geochemistry, possibly related to mafic magmas associated with the rapakivi granite. If the dikes are Neoproterozoic, as suspected, then the deformation may be Mesoproterozoic, possibly Sveconorwegian. In our examination of the outcrop, we could not find any other clear cross-cutting relations and considered the possibility that one mylonite zone actually might cut through a dike, but in a place where there is no outcrop. The photograph published by Tucker is more convincing than anything we saw in place. It would be ironic as well as informative if evidence of a Sveconorwegian deformation were preserved in the allochthon in this location, and not in the autochthon. That may be the case with the Jotun Nappe and the WGR
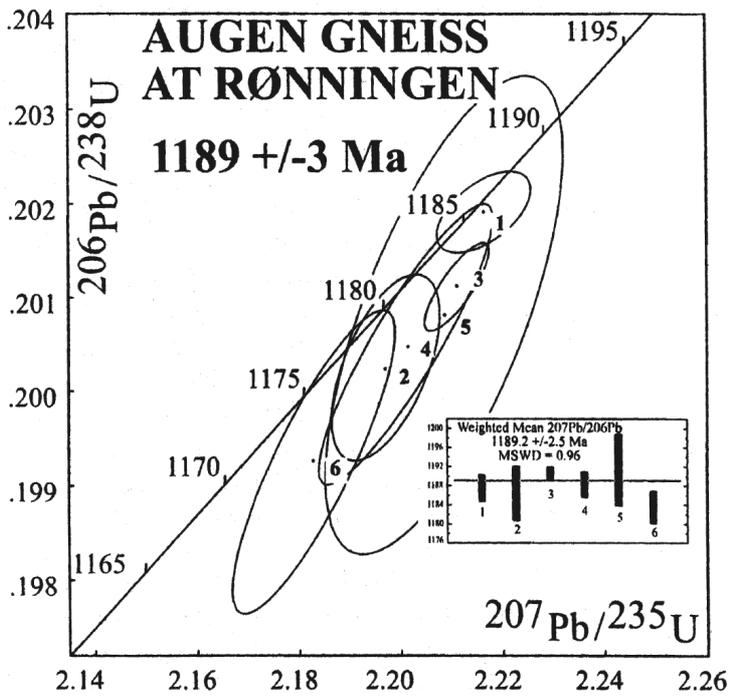
Fig. 5.16 Concordia diagram showing U-Pb ages of zircon in augen gneiss of the Risberget Nappe at Kjøra (Rønningen)

basement beneath it (see Guidebook for Trip A-3, COPENA Meeting 1997).

10:05   **20**   Leave Stop 3.
10:08   3   47.6   1.0   Return to Route 710. Turn right (northeast) toward Lensvik.
10:10   2   48.6   1.0   Turn right off Route 710 on road to Geita.
10:14   4   50.6   2.0   Park at campground office and walk southwest through campground to shore.

STOP 5-4: (20 minutes) Foliated basement gneiss cut by Sætra dike that is itself folded.
The significance of this stop was recognized on a joint trip by Solli, Tucker and Robinson in 2001. Based on five analyzed samples (Hollocher et al. 2007a), the dike has typical Sætra geochemistry (Fig. 5.4), probably identifying it as part of the Middle Allochthon. Details of the outcrop show that although the dike is involved in a medium-sized asymmetric fold, it also cuts a strong tectonic foliation in the enclosing gneiss. The contact between the gneiss and the Risberget augen gneiss to the south is not exposed. The deformation could be

older than the Risberget age of 1189 Ma, or it could just be that the more massive Risberget responded less easily to a younger deformation (Sveconorwegian?). In either case, the planned geochronology of the gneiss from this outcrop of basement of the Middle Allochthon will be interesting.

10:34   20   Leave STOP 5-4.
10:39   5   52.6   2.0   Return to Route 710. Turn right toward Lensvik.
             58.6   6.0   Soon reach shore again at Ingdal in crest of basement antiform exposing pink 1653 Ma Ingdal Granite (see Stop 18).
             58.9   0.3   Road cut in Ingdal Granite. Impossible to park.
             61.8   2.9   Pass large artificial exposures of well foliated rusty-weathering Blåhø mica schist and marble with vertical foliation in deep synformal belt.
10:52   13   67.1   5.3   Center of Lensvik in antiform of Baltica basement, and Lensvik Kirke.
10:55   3   67.7   0.6   Big turnout on right. Park by storage for telephone poles. Cross to road cut on left and walk SW along the exposure.

STOP 5-5: (20 minutes) South limb section of Lensvik synform at Ystland section.
    Walk southwest and down section in rocks dipping steeply north. One first encounters high-grade garnet-biotite schist and amphibolite of the Blåho Nappe. Near the base of the nappe there is quartz-muscovite-calcite schist and marble, most easily recognized where weathered on the coast. South of this is a thickness of about 10 m of feldspathic quartzite with amphibolite layers interpreted as highly deformed diabase dikes in the Sætra Nappe. Major- and trace-element geochemistry of five samples from this exposure (Fig. 5.4) indicate Sætra-like geochemistry (Hollocher et al. 2007a) but with a more alkaline character suggesting approach to a Late Neoproterozoic hot-spot track (Fig. 5.3). The Sætra Nappe here appears to be resting directly on Baltican basement, unless 1-2 m of sheared rock can be assigned to the Risberget Nappe.

11:15   20   Leave Stop 5-5. Continue on 710.
11:19   4   69.7   2.0   Park at small turnout on right surrounded by steel railing. No picnic table. Descend poor narrow trail right to steep shore exposure.

STOP 5-6: (17 minutes) Tonalite gneiss of Ordovician magmatic complex with mafic enclaves and mafic dikes.
    These rocks in the Lensvik synform (Tucker 1986) are easily mistaken for Proterozoic basement of the WGR. These are now assigned to the Støren Nappe of the Upper Allochthon. Tucker (Tucker et al. 2004) dated a sample collected near here that has yielded a zircon U-Pb age of 481.9+/-1.5 Ma (Late Tremadoc, Fig. 5.17). This is similar to the U-Pb zircon ages of 482 +/-5 and ca. 481 Ma reported for felsic rocks in the Bymarka ophiolite (Roberts et al. 2002b). The rocks we see here are not typical for the rocks in the Støren Nappe as they appear in the central and western parts of the Trondheim Region (this Guidebook, Day 4). The eastern equivalent of the Støren ophiolite, the Fundsjø Group, has, however, many intrusions ranging in composition from mafic to felsic (see Day 4), and the same kind of mixing of rocks as we see here is common (eg. Wolff 1973, Grenne et al. 1995). A trondhjemite (plagiogranite) in the Fundsjø
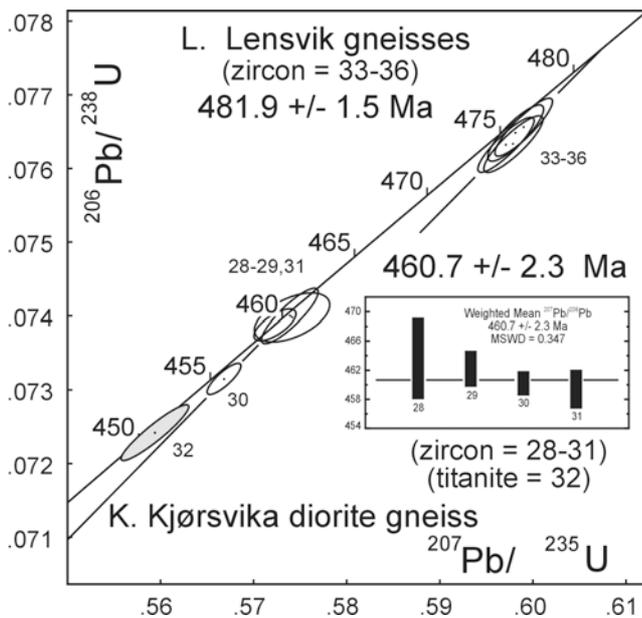
Fig. 5.17 Concordia diagram showing Ordovician U-Pb ages of igneous zircon from granodiorite gneiss near Lensvik (482 Ma in Fig. 5.1) and from diorite gneiss at Kjørsvika (461 Ma in Fig. 5.1).
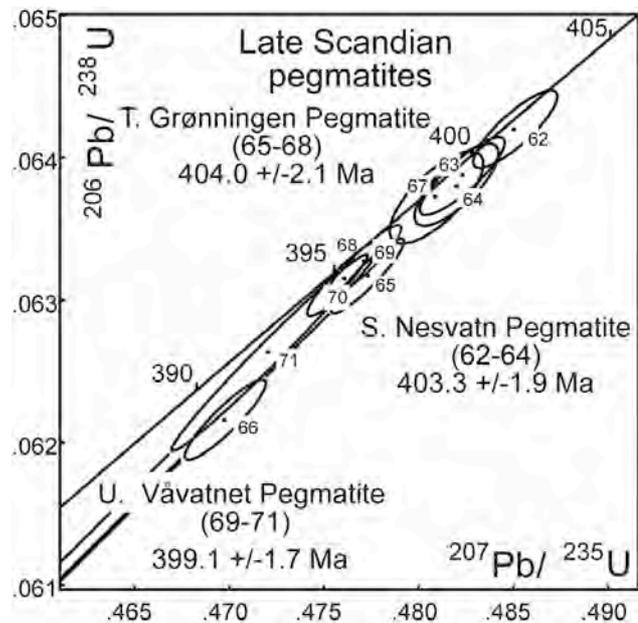
Fig. 5.18 Concordia diagram showing U-Pb ages of discordant pegmatites in Proterozoic gneisses of the Western Gneiss Region at Nesvatn (403 Ma in Fig. 5.1), Grønningen (404 Ma in Fig. 5.1) and Våvatnet (399Ma in Fig. 5.1).

Group has been dated to 488 +/-2 Ma (base of the Tremadoc, Bjerkgard & Bjørlykke 1994), which is slightly older than the intrusions in the Lensvik synform.

Three common features are well displayed on the steep outcrop: a tonalitic host rock with rounded mafic enclaves, and a younger mafic dike. All the rocks are strongly deformed and also contain quartz-rich veins, but lack pegmatites that are prevalent in the adjacent Baltican basement. Trace-element analyses of samples from this and many adjacent outcrops of tonalitic intrusive rocks in the Lensvik synform (Hollocher et al., 2007c and in preparation) have produced a major surprise. Unlike all other samples in this province of intrusive rocks, these yield flat REE patterns (Fig. 5.6a), characteristic of plagiogranites associated with ophiolites, a feature they have in common with some of the felsic intrusions in the Bymarka and Løkken ophiolites near Trondheim. This complicates the proposed interpretation of an Ordovician arc plutonic complex, showing that the complex, spanning an age range 481-438 Ma, based so far on only 9 U-Pb zircon ages, may have begun with an early ophiolite-related phase followed by the construction of a magmatic arc, following a suggestion made by Roberts (1980).

11:36	17		Leave Stop 6. Continue on 710.
		70.3 0.6	Picnic table as advertised. Green Støren volcanic rocks cut by dikes begin just beyond.
		72.9 2.6	Big road cut and shore exposures of mafic Ordovician intrusive and volcanic rocks.
11:43	7	73.5 0.6	Turn into narrow break in steel fence on right, walk ahead to Litleneset shore outcrop.

STOP 5-7: (18 minutes) Coarse metamorphosed gabbro of Støren Nappe in Lensvik synform.
    Strong shear fabrics with subhorizontal lineation. These rocks are nearly identical to some Støren gabbros described by Robinson (1995) near Molde and Brattvåg (Stop 8-9). Recent geochemical studies (Hollocher et al. 2007c and in preparation) support this correlation and indicate that the rocks are cumulates (Fig. 5.5).

12:01	18		Leave Stop 7.
		78.2 4.7	Kjerringneset. Baltican basement at turn out. No stop.
		80.2 2.0	Tunnel. Long cuts in Baltican basement.
12:11	10	82.2 2.0	Pull over on left into road leading to old tunnel mouth at Hambortåa.

STOP 5-8. (15 minutes) Marble and schist, and migmatitic Baltican basement.
    Description based on Krill (1981): "A typical section of the Seve (Blåhø) Nappe consisting of garnet-hornblende schist, amphibolite and white marble is exposed southeast of the Hombortåa tunnel. Small amounts of kyanite can be found in the schists and diopside in the marble. The contact to the basement is at the entrance of the tunnel, but the contact is strongly sheared and not easy to locate. The basement consists of reddish gray migmatitic gneiss of granodioritic composition. Characteristic for the gneisses are granitic neosomes with hornblende selvages of Scandian (??) age."

Marble and schist near the tunnel may belong to the Øyangen Formation of Tucker (1986), below the Middle Allochthon. A U-Pb age of 400 Ma on titanite from the marble (Tucker et al. 2004) is likely a result of Scandian new growth, not resetting of Proterozoic grains. Walk north, down section into big road cut showing a broad antiform in migmatitic Baltican basement. Was partial melting Mesoproterozoic, Scandian, or both? In the general area, U-Pb ages of some pegmatites and melt leucosomes are Devonian (Fig. 5.18).

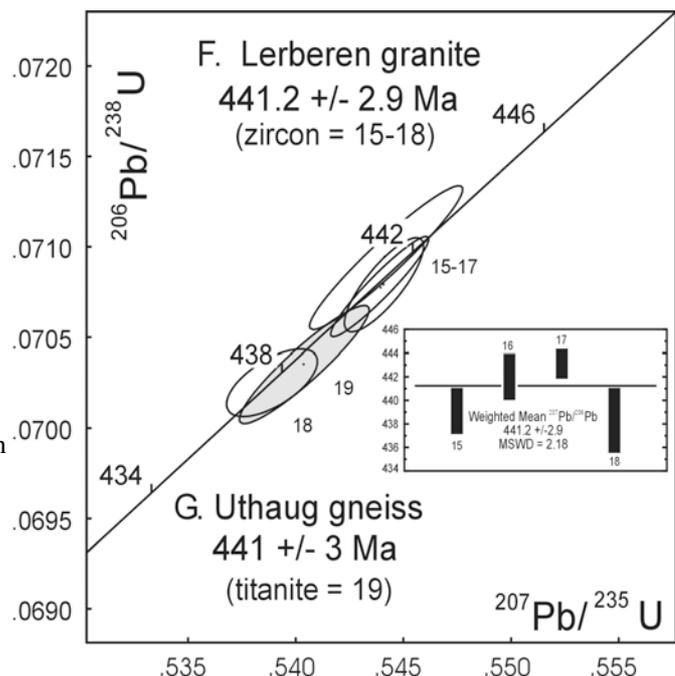|			
12:26	**15**		Leave Stop 8.
	82.8	0.6	Folded interlayered amphibolite and quartz-rich granulite on sharp bend, probably belonging to the Sætra Nappe. The same belt of rock will be viewed in stratigraphic context at Hasselvika (Stop 5-16), and represents the most extreme example of partial melting yet observed in the Sætra Nappe.
	84.1	1.3	Small harbor at Selva. Stay on Route 710 toward Valset Ferry.
	87.4	3.3	Cross pass with huge outcrops and quarry of high-grade pink migmatitic Baltican basement.
12:35	9 88.7	1.3	Descend to sea level and park in waiting line for Valset-Brekstad ferry. The Agdenes detachment lies approximately at the foot of the slope, separating high-grade basement and related cover from relatively low-grade Ordovician intrusive gneisses of the magmatic arc complex. This contact will be seen in outcrop at Stop 5-15. Enter outcrops to left.

STOP 5-9: (25 minutes or until ferry arrives) Strongly deformed mafic and felsic gneisses of Ordovician magmatic arc complex of Støren Nappe north of Agdenes detachment.

The southern part of the outcrop is dominated by simple hornblende amphibolite, probably highly deformed metamorphosed mafic intrusive, rather than metamorphosed volcanic rocks. The northern part is dominated by strongly foliated and lineated diorite to tonalite that probably intrudes the amphibolite. The dioritic gneiss is typical of a very long stretch of coast to NE and SW, including a locality at Kjørsvika from which Tucker (Tucker et al. 2004) has obtained a U-Pb zircon age of 460.7+/-2.3 Ma (Fig. 5.17). An important feature at Kjørsvika is that, although the rock is highly deformed and contains the same late structural fabric as the adjacent basement gneiss, the titanite still retains an Ordovician age of about 455+/-3 Ma, whereas the titanite of the adjacent basement gneiss has been 100% recrystallized during the Scandian at 395 Ma. The implication of this is that the Ordovician intrusions and other rocks of the Støren Nappe have been emplaced against hotter Proterozoic basement during a late-Scandian phase of extensional detachment that took place before the phase of sinistral transtension that produced the dominant fabric (Tucker et al. 1997, 2004, Robinson et al. 1997). This, then, is a key part of the argument for the Agdenes detachment.

13:00	**25**		Drive or walk onto ferry. EAT LUNCH on ferry, use toilets, drinks, coffee. During the ferry crossing, we cross the trace of the Devonian Møre-Trøndelag Fault Complex, which, more toward the northeast, truncates the Høybakken detachment. We thus cross from the relatively low-grade gneissic terrane above the Agdenes detachment to the brittlely deformed terrane with Devonian Old Red Sandstone above the Høybakken detachment.
13:25	25 88.8	0.1	Drive off ferry at Brekstad, proceed east in the direction of Bjugn on Route 710. There are extensive outcrops of Old Red Sandstone on Ørlandet west of Brekstad and on the island, Storfosna. However, we will visit outcrops toward the northeast.
13:30	5 89.7	0.9	Traffic circle go right (north) toward Bjugn.
	91.5	1.8	Right-angle junction. Turn left toward Uthaug and Aune.
	91.8	0.3	Right (northeast) onto gravel road toward Lerberen Granite quarry. Bypass quarry on left and follow up narrow road to parking place near hill top.
13:33	3 92.8	1.0	Road to summit blocked here. Bear left and park by water tank. Climb paved trail with outcrops to summit area.

STOP 5-10. (20 minutes) Lerberen granite above Høybakken detachment. Very fresh exposures of two-mica granite in cuts of paved trail leading to summit. The granite is non-foliated and typical of many of the Late Ordovician to Early Silurian intrusive rocks above the Høybakken detachment associated with the Devonian clastic basins. It has yielded a U-Pb zircon age (Tucker et al. 2004) of 441.2+/-2.9 Ma (Fig. 5.19), placing it in

Fig. 5.19 Concordia diagram showing Early Silurian U-Pb age of igneous zircon from Lerberen Granite (441 Ma in Fig.1) and igneous titanite from Uthaug Gneiss (also 441 Ma in Fig. 1).

the earliest Silurian (Early Llandovery) according to the latest ICS time scale (Gradstein et al. 2004) Geochemical data point to a high-K, calc-alkaline signature (Gautneb & Roberts 1989). The granite is cut by many faults with transitions from coarse-clast red breccias to fine-clast orange breccias and younger green cataclasites with seams of gouge (Roberts 1998). The comprehensive summit view will be explained.

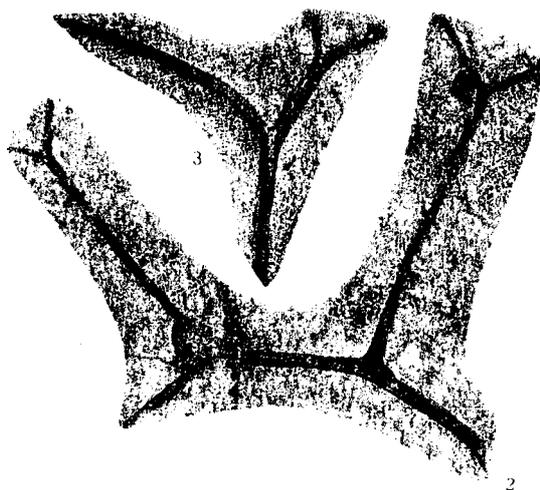|				
13:53	**20**			Leave Stop 10. Return on gravel road to main road.
13:57	6	93.8	1.0	Turn left on main road.
13:58	1	94.1	0.3	Junction. Turn left on 710 toward Opphaug.
14:00	2	96.3	2.2	Center of Opphaug. Junction. Turn left toward Innstrand, Grøtan, Døsvika.
14:01	1	97.1	0.8	Right turn at Grøtan toward Ervik. Follow road straight to coast at Døsvika.
14:03	2	98.4	1.3	Go to shore outcrop at very end of road near shoreline cabins at Døsvika.

Fig. 5.20 Naked-branch systems of *Hostimella* from Døsvika (Stop 5-11), Ørlandet. From Høey (1945).

STOP 5-11: (15 minutes) Devonian sandstone and cobble conglomerate.

The Devonian on Ørlandet belongs to a very long and narrow belt of Devonian rocks. The total length of the belt is 200 km and the width is only 5 to 20 km. Whether the belt comprised one uninterrupted basin or several smaller basins is uncertain (Steel et al. 1985, Osmundsen et al. in prep.). The shape of the basin or basins is thought to be tectonically controlled. The Devonian on the mainland on Ørlandet is subdivided into two formations (Siedlecka1975). The lower Austrått Formation is about 500 m thick consisting of sandstones, pebbly sandstones and conglomerates. It is interpreted to have been deposited in a braided river system. The Bjugn Conglomerate is more than 2000 m thick and consists mainly of coarse-grained, clast-supported, polymict conglomerates. It is interpreted to have been deposited as alluvial fans. Mudstone beds are subordinate. On the islands southwest of Ørlandet there are other types of Devonian sediments not easily correlated to those on the mainland. An interpretation of the sedimentary environments and palaeogeographic reconstructions has been made by Siedlecka (1975), Steel et al. (1985) and Osmundsen et al. (in preparation). The rocks at Døsvika are assigned to the Bjugn Conglomerate (Siedlecka 1975, Wolff et al. 1980). Polymict conglomerates, and fining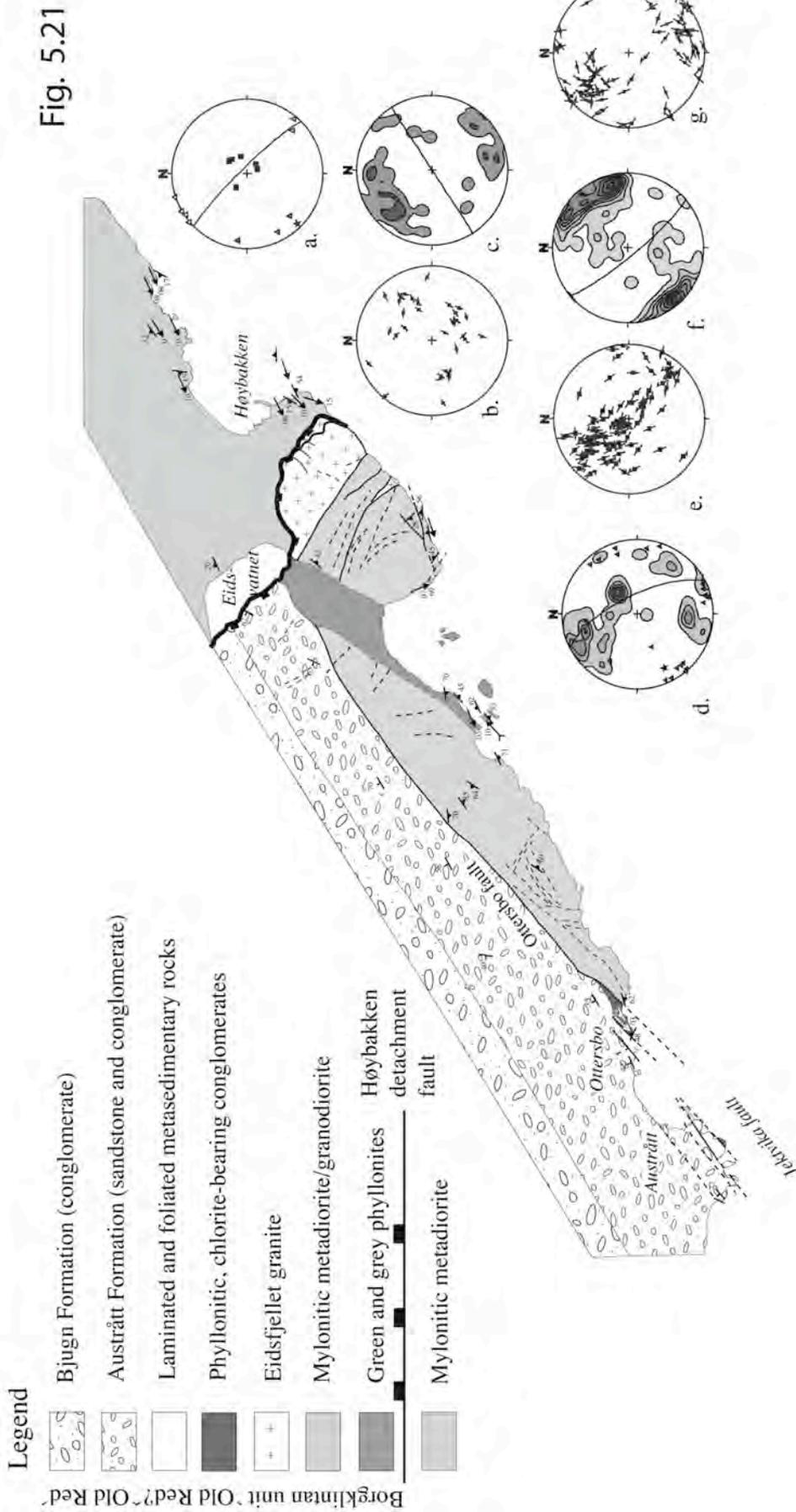-upward cycles in sandstones and minor mudstones are well exposed. The conglomerate is reddish-grey, poorly sorted, and has local cobbles up to 20 cm across. Common pebble types in the conglomerate are granite, granitic gneisses, red jasper, sandstone, quartzite and quartz. The sandstone portions may contain pebbles and exhibit cross bedding. Ripple-cross stratification is present at the transition into the mudstone, which is grey and micaceous.

Mudstones are a subordinate facies of the Bjugn Formation, but important because of the presence of terrestrial plant fossils. At this locality Vogt (1929) and Hoeg (1945, 1966) have described the species *Hostimella and Psilophyton rectissimum*, and Allen (1976) has reported on finds of spores, all of which are indicating an Early Devonian age for the Bjugn Conglomerate. Naked branch-systems of *Hostimella*, a few centimeters in size, may still be found in the mudstones at Døsvika (Fig. 5.20).
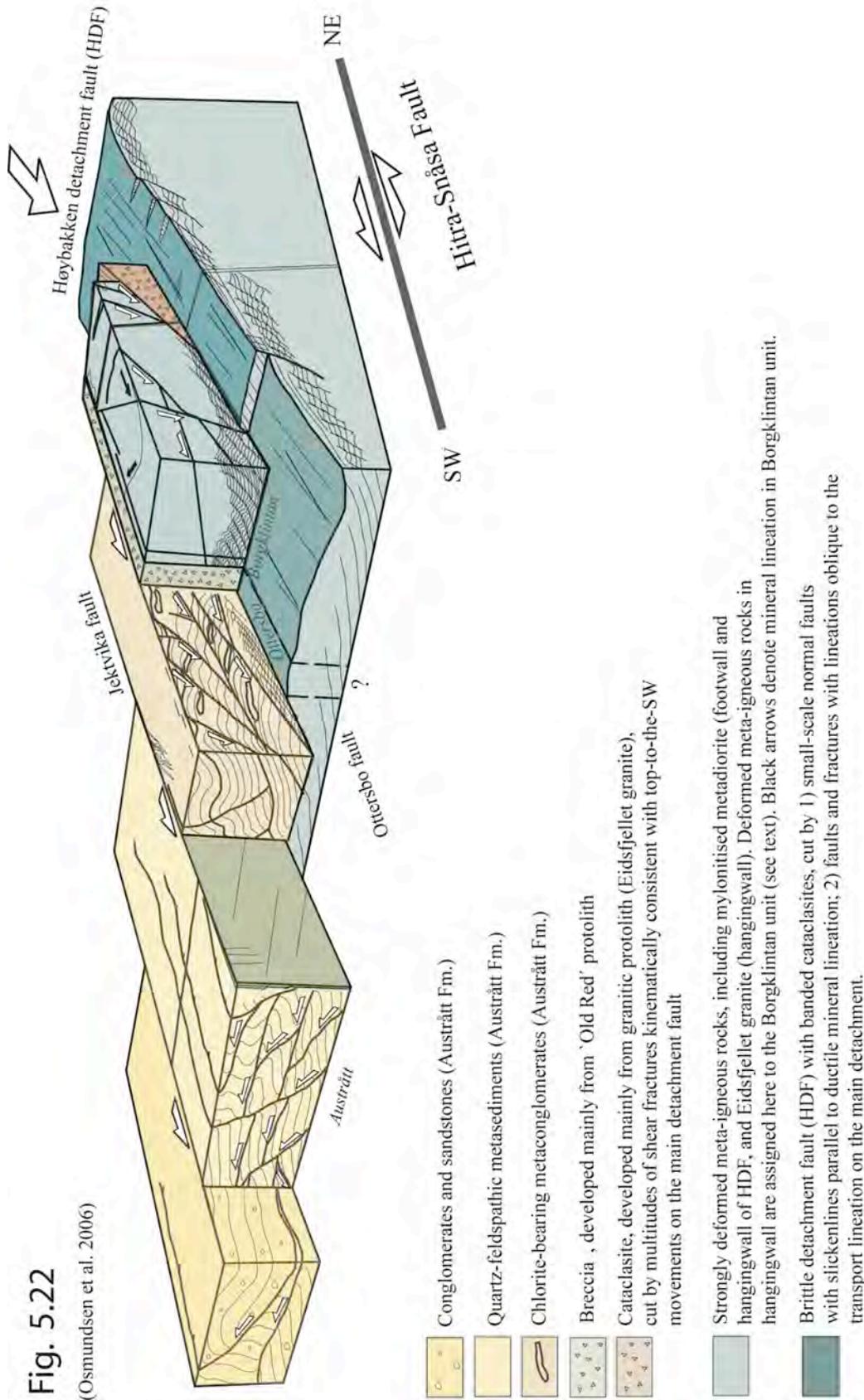
14:18	**15**			Leave Stop 11. Return toward Opphaug.
		99.7	1.3	Turn left at Grøtan.
14:22	4	100.5	0.8	Straight south across main route junction and through center of Opphaug. Road bends much in passing through village.
14:24	2	102.8	2.3	Stop sign. Turn left toward Høybakken. Big Devonian hills on left.
14:31	7	110.6	7.8	Junction. Turn right to Høybakken. The road along Lake Eidsvatnet (see below) is to the left.
14:33	2	111.8	1.2	Sharp right turn toward Høybakken marina.
14:34	1	112.2	0.4	Park, Høybakken near outer breakwater. Hike west on narrow coast trail.

STOP 5-12: (45 minutes) Flaser diorite gneiss and brittle cataclasite of Høybakken detachment.

The outcrops at this stop are situated just below and at the Høybakken detachment (Séranne 1992, Osmundsen et al. 2006, Figs. 5.21, 5.22), one of several extensional detachments that underlie Devonian basin rocks in western Norway. These detachments, or rather, detachment zones, generally consist of a kilometre(s) thick mylonite zone capped by a brittle, low-angle detachment fault, and are thought to have played a major role in the late- to post-orogenic thinning of the Caledonian orogenic crust. The road along the east shore of Lake Eidsvatnet, runs along the top of the footwall of the gently southwest-dipping detachment zone. The lower parts of the footwall are underlain by protomylonites and mylonites of dioritic composition, whereas the upper parts consist of ultramylonite and mylonite. A brittle detachment fault separates the ductilely deformed footwall from the hanging wall. On the western lakeshore, Devonian conglomerates are faulted against the detachment.

Fig. 5.21

Legend

Borgklintan unit 'Old Red?' 'Old Red?'

- Bjugn Formation (conglomerate)
- Auståt Formation (sandstone and conglomerate)
- Laminated and foliated metasedimentary rocks
- Phyllonitic, chlorite-bearing conglomerates
- Eidsfjellet granite
- Mylonitic metadiorite/granodiorite
- Green and grey phyllonites
- Mylonitic metadiorite

Høybakken detachment fault

Geological map of the Høybakken-Ottersbo area showing trace of the Høybakken detachment fault and NE-SW-trending strands of the Møre-Trøndelag Fault Complex. Equal-area stereograms show orientation of important structural elements such as a) undulating detachment fault (squares) with pole to best-fit great circle (star), metre-scale undulations (filled triangles) and younger mineral fibre and slickenline lineation (open triangles). N=16. b) Slip-linear plot of small-scale faults cutting red cataclasite in the hangingwall of the Holybakken detachment. N=35. c) shows contoured poles to m-planes calculated from dataset in b, with mean m-plane (grat circle) indicating SW-NE trending (237 degrees) elongation. d) summarized readings of ductile structures in the so-called Borgklintan Formation, interpreted by Osmundsen et al. (2006) as detachment mylonites captured in the hangingwall of the brittle detachment fault. N=69. The ductile main fioliation has been contoured, best-fit grat circle and pole to best-fit great circle (star) are shown, indicating folding around axis trending 244 degrees, which is roughly parallel to the ductile stretching lineation in the footwall as well as to the generalised brittle elongation direction for the hanging wall in c). triangles represent ductile stretching lineations recorded in the Borgklintan Formation, e) is a slip-linear representation of faults with epidote-bearing mineralizations, largely related to top-to-the-NW thrusting in the Auståt-Ottersbo area (N=99). f) shows calculated poles to m-planes for the data in e., indicating a generalized NW-SE-trending mean m-plane for the faults and thus a generalized movement direction towards 320 degrees. The slip-linear plot in g) summarizes the kinematics of structures recorded from a number of NE-SW-trending fault strands related to the Møre-Trøndelag Fault Complex, including steep strike-slip and dip-slip faults

Fig. 5.22

(Osmundsen et al. 2006)

Høybakken detachment fault (HDF)

Jektvika fault

Ottersbo fault

Austrått

Illvatna

Borgklintan

Hitra-Snåsa Fault

NE

SW

?

Conglomerates and sandstones (Austrått Fm.)

Quartz-feldspathic metasediments (Austrått Fm.)

Chlorite-bearing metaconglomerates (Austrått Fm.)

Breccia , developed mainly from 'Old Red' protolith

Cataclasite, developed mainly from granitic protolith (Eidsfjellet granite), cut by multitudes of shear fractures kinematically consistent with top-to-the-SW movements on the main detachment fault

Strongly deformed meta-igneous rocks, including mylonitised metadiorite (footwall and hangingwall of HDF, and Eidsfjellet granite (hangingwall). Deformed meta-igneous rocks in hangingwall are assigned here to the Borgklintan unit (see text). Black arrows denote mineral lineation in Borgklintan unit.

Brittle detachment fault (HDF) with banded cataclasites, cut by 1) small-scale normal faults with slickenlines parallel to ductile mineral lineation; 2) faults and fractures with lineations oblique to the transport lineation on the main detachment.

Schematic block diagram (not to scale) showing relationships between foot- and hangingwall of the Høybakken detachment and a number of relationships observed in the hanging wall in the Høybakken-Austrått area. Note extensional, top-to the WSW shearing of chlorite-bearing conglomerates adjacent to the contact to the Borgklintan Unit. The Borgklintan Unit also contains abundant, top-WSW and sinistral-oblique ductile to brittle fabrics, and is interpeted as a sliver of mylonites captured by the brittle detachment fault. Most likely, the brittle detachment fault decapitated folded detachment mylonites, explaining the presence of top-to-the SW ductile to brittle fabrics in the sliver of metaigneous rocks. Also, note top-NW thrusting in the Austrått formation. Contractional deformation appears to increase southeastwards toward the Hitra-Snåsa Fault. The relationships between the detachment and steeply dipping, strike-slip fault strands are not always clear, but the detachment is clearly cut by some of them.

At the locality in Høybakken, the detachment separates the so-called Eidsfjellet granite in the hanging wall from mylonitic diorites in the footwall (Séranne 1992 and Osmundsen et al. 2006). The Eidsfjellet granite occurs in a suite of strongly sheared and foliated mafic metamorphosed igneous rocks that strongly resemble the rocks in the footwall. Osmundsen et al. (2006) interpreted these rocks to represent parts of the ductile detachment zone that became incorporated into the hangingwall, due to the decapitation of extension-parallel folds by the brittle detachment fault.

The Marina at Høybakken is situated ca. 50-70 m vertically below the detachment fault in an Ordovician mylonitic diorite. The diorite has a west-plunging lineation and contains strain indicators (in particular ductile shear bands) that all indicate a top-to-SW sense of movement (Séranne 1992, Watts 2001, Osmundsen et al. 2006). Small-scale, lineated semi-ductile faults offset the mylonitic fabric and can be observed close to the waterline. The first large outcrop is a platy flaser diorite gneiss with a prominent gently SW-plunging lineation, part of the Høybakken footwall that was gradually subjected to stronger and stronger grain-size reduction during exhumation. Chlorite largely replaces earlier biotite in the highly sheared dioritic rocks. Outcrops beyond this, and closer to the actual detachment fault, show increasing evidence of brittle fracture as well as thin seams of ultramylonite.

Proceed to the mouth of a small brook just east of a high cliff and find a smaller overhanging cliff in trees which shows the actual Høybakken brittle detachment surface with local breccia and cataclasite. Note in particular 0.5-metre-thick banded cataclasites at the base of the hanging wall succeeded above by pink cataclasite with a granite protolith. The detachment fault itself is gently undulating and largely southeast-dipping at this locality. Southwest-plunging undulations characterise the detachment on the regional as well as the local scale. Note also a vague, SE-plunging lineation on the brittle detachment fault. This likely represents the latest movements on the detachment and could be related either to shortening normal to the extension direction (which affects much of the area, in particular in the vicinity of the Møre-Trøndelag Fault Complex) or to later reactivation in the Mesozoic. Immediately below the brittle detachment fault, cataclastic rocks are developed from the mylonitic diorites in the footwall, and occur intermittently along the fault. Such cataclasites are well developed along many detachment faults elsewhere in Norway.

On the beach, the detachment and the footwall mylonites, are cut by a NW-SE-striking fault that offsets the detachment with by at least several metres and hosts epidote-bearing mineralisations, greenish cataclasites and a WNW-plunging lineation. Similar but smaller faults can be seen cutting the detachment fault at several localities along its strike. NW-SE-trending faults are common on Fosen peninsula and locally host cataclasites. According to Olsen et al. (2007) they are considered to have formed during very latest Scandian extensional unroofing of the region.

In a ^{40}Ar-^{39}Ar study of hornblende, mica and K-feldspar separated from rocks within the Høybakken detachment zone, Kendrick et al. (2004) showed that exhumation and cooling of the footwall is recorded by hornblende ages of c. 400 Ma and mica ages of c. 390 Ma. Mylonites overlying the footwall record Mid Devonian, mica crystallization ages of 384-381 Ma, thus suggesting a prolonged (11-20 m.y.) period of ductile extensional activity. After cessation of ductile extension at 381 Ma (middle Frasnian), brittle extension occurred up to at least Early Carboniferous time, as recorded by K-feldspar in a cataclastic granite from the hanging wall of the detachment (Kendrick et al. 2004, Osmundsen et al. 2006). Outside of the Ørlandet area, an age of 290 ± 10 Ma (Early Permian) has been obtained from a pseudotachylite from the Hitra-Snåsa Fault (Sherlock et al. 2004).

> Return to marina by same trail.

15:19	**45**			Leave Stop 12 and exit from marina.
15:21	2	112.6	0.4	Turn right (NE) on main road. Drive east through Høybakken to Stallvika. All the outcrops are Ordovician diorite with flat foliation below the Høybakken detachment.
15:27	6	118.3	5.7	Junction at Stallvika. Turn sharp left up hill toward Route 710 for Trondheim.
		121.1	2.8	Pass lake.
15:32	5	121.6	0.5	Pass low outcrop on corner to right and park just beyond, near intersection.

STOP 5-13: (10 minutes) White marble and calc-silicate rock of the of Støren Nappe with top-west ductile shear fabric. Shows gentle west-dipping foliation, strong E-W lineation and probable tubular folds parallel to lineation. The shear sense along the lineation is top-west, consistent with observations in the region by Séranne (1992), indicating a ductile shear in the same sense as the more brittle fracture of the Høybakken detachment. About l km north of here the marble has been quarried for industrial purposes.

15:42	**10**			Leave Stop 13. After big view descend toward Route 710.
15:45	3	123.1	1.5	Junction with Route 710 at bottom of hill beside lake. Turn right (east) toward Trondheim.
15:46	1	123.7	0.6	Road cut left on upgrade. Park right and cross highway to middle part of cut.

STOP 5-13A Optional: (10 minutes) Coarse flaser gabbro of Ordovician plutonic complex.
Specimens of this gabbro show spectacular ductile shear fabrics, but the outcrop needs to be examined carefully.

15:46	(10)			Leave Stop 13A. (Including this stop adds 10 minutes to road log.)
15:55	9	131.9	8.2	Large road cut in hydrothermally altered Ordovician intrusive rocks.
		132.6	0.7	Approximate crossing of Hitra-Snasa Fault. Then more intrusive rocks.

15:57	2	134.6	2.0	Junction. Turn right (east) on Route 715 and cross bridge toward Trondheim.
16:00	3	137.7	3.1	Junction. Turn right (southwest) on Route 718 toward Hasselvika.
16:06	6	143.6	5.9	Junction in village. Turn right (north) toward Råkvåg.
		143.9	0.3	Beyond bridge at sea level turn left (west) toward Rakvåg.
16:08	2	145.4	1.5	Pull into huge quarry beneath cliff on right. Avoid steep walls.

STOP 5-14: (10 minutes) Strongly foliated layered mafic and felsic intrusive rocks of Ordovician magmatic complex near Råkvåg. The huge, natural, south-facing cliffs are enhanced by quarrying. The lowest part of the N-dipping section is dominated by mafic rocks which grade upward into more felsic rocks. The ductile deformation is so intense that these rocks could be mistaken for metamorphosed volcanic rocks.

16:18	**10**			Leave Stop 14. Return back toward Route 718.
		146.9	1.5	Junction. Turn right toward Route 718.
16:21	3	147.2	0.3	Turn right (west) on Route 718 and winding shore road with exposures of basement gneiss with subhorizontal lineation. High up in the Fessdal valley to the south there are quarries in quartzite (unlike Sætra) associated with basement.
16:33	12	157.7	10.5	Junction at Selnes. Turn right off Route 718 toward Revsneshagen.
16:34	1	158.5	0.8	Drive to end of road nearest shore and park near wharf. Walk back to low shore outcrops.

STOP 5-15: (15 minutes) Exposure of Agdenes detachment. This exposure shows north-dipping rusty garnet-mica schist, amphibolite, and deformed pegmatite of the Seve=Blåhø Nappe in direct tectonic contact with the overlying Ordovician intrusive complex of the Støren Nappe with felsic and mafic layers including deformed gabbro. This contact is defined regionally as the Agdenes detachment. Both rock types contain a subhorizontal lineation and sinistral (top-W) shear indicators superimposed after detachment faulting.

16:49	**15**			Leave Stop 15. Return toward Route 718.
		160.1	1.6	Junction with Route 718. Turn right toward Hasselvika and winding shore route with exposures of Baltican basement near Hofella inlet. Beyond junction for Fevåg Kai, we climb south through a mountain pass with exposures of Agdenes-type Baltican basement.
17:06	17	176.4	16.3	Roadside park and picnic table at opening of steel fence on right approaching Hasselvika.

STOP 5-16: (20 minutes) S to N coastal section, Baltican basement, Sætra Nappe, Blåhø Nappe.
   Shore exposures show a highly deformed and metamorphosed vertical section. Road cut to south across highway is typical basement gneiss with Scandian hornblende-bearing migmatite zones. The same rock occurs in the innermost part of the shore exposure. This is in contact to the N with very feldspathic, partially melted quartzite with amphibolite and minor calc-silicate layers, that is tentatively correlated with the Sætra Nappe, though the geochemistry of the mafic layers has not yet been tested. It is the most migmatized Sætra we have observed and is exactly along strike from outcrops near Selva, passed just after Stop 5-8. N of the most quartz-rich part of the quartzite, the outcrop continues in garnet-mica schist and amphibolite with very thin minor marble beds, assigned to the Blåhø Nappe. Before leaving, see the World War II Memorial in the roadside park, which is a remarkable extracted 3D fold in quartzite.

17:26	**20**			Leave Stop 16. Continue toward Hasselvika.
17:28	2	177.6	1	At center of Hasselvika, turn south to Hysnes Fort Community Hall where we will have **Dinner (40 minutes).**
18:08	**40**			Leave Hysnes Fort and continue south on coast road.
18:14	6	183.0	5.4	Big coastal road section of pink basement gneiss. No parking
18:15	1	183.3	0.3	Narrow synformal belt of Blåhø Nappe infolded in basement which is abundantly exposed in highway exposures. There is 1 m of Sætra Nappe at the southern contact. Too dangerous to stop with group.
18:27	1	183.6	0.3	Road cut on left with beautiful upright folds in layered basement gneiss.
		184.0	0.4	Road cut in massive basement intrusion, earlier mapped as quartzite or Ordovician intrusion.
		184.6	0.6	Quarry on left, intrusive granitoids of basement.
		187.2	2.6	Bridge at estuary. Basement exposures. Contact with Støren Nappe a short distance south.
18:32	5	188.1	0.9	Hammarberget. Road cut on left and coast exposures of Ordovician diorite gneiss in Lensvik synform. No stop.
18:33	1	189.1	1.0	Junction by gas station. Turn right on Route 717 toward Trondheim.

	1	190.3	1.2	Turn right off main road toward Kvitthyll (white ledge) and large shipyard for Optional Stop 5-16A which consumes an extra 15 minutes, otherwise continue on main road.
			1.7	Djupdelen sign.
	3		0.3	When approaching Kvitthyll Hurtigbåt Terminal, turn right into industrial area and outcrops partly covered by crushed rock.

STOP 5-16A Optional: (10 minutes) Steeply dipping, low-grade epidote amphibolites of the Støren Nappe in Lensvik synform. (Including this stop consumes 15 minutes. Geochemical analytical data are shown in Fig. 5.5)

	(10)			Leave Stop 17. Back toward Hurtigbåt and continue south on coast road.
	1		0.4	Kvitthyll Gabbro. Low exposures, even grained. Flat REE patterns. Follow road away from coast toward Rein Kloster. The name Kvitthyll refers to white rock and is a marble exposed on the next point to the south associated with garnet-bearing amphibolites that may be infolded rocks of the Blåhø Nappe.
18:50	2	193.9	1.2	Sharp right turn back onto Route 717 in front of Reine Kloster, and resume normal road log.

18:35	2	191	1.9	Rein Kloster. Junction. Bear left on Route 717.
18:36	1	192.3	1.3	Turn left off Route 717 toward the site of the Rissa landslide.
18:37	1	193.2	0.9	**Monument of 'Rissaraset 1978' quick-clay slide into the lake. (10 minutes)** Eight farms lost and 1 life. Triggered by garage construction.
18:47	**10**			Leave the 'Rissaraset' site.
18:48	1	194.1	0.9	Return to Route 717 and turn left (south). Climb steep hill toward south underlain by Ordovician intrusive rocks of the Lensvik synform similar to Stop 5-6, then enter a broad region of interfolded Støren low-grade amphibolites with granitoid dikes (in large roadside quarry on left) and Blåhø high-grade schists, amphibolites, and marbles. The latter continue right to a contact with the Sætra Nappe near the bus stop Brødreskift.
18:54	6	200.7	6.6	Park at bus stop for Brødreskift at beginning of farm road. Walk through farm to coast exposure at Storneset.

The name 'Brødreskift' comes from incidents at these farms occurring sometime in the 1200s, and described in a Bishop's Landbook written in 1435. Two brothers were in very strong disagreement over the division of their inheritance of a farm. Specifically the younger brother wanted to have a larger share of the land than their parents had determined in a will. So the local authorities had to take over the case and arrange for a court hearing. On the night before the court appearance, a violent storm broke out with heavy rain. During the storm an enormous rock on the mountain above the farm rolled and slid all the way down to the shore, and in the furrow behind the rock a little brook was running. Both brothers and other people took it to mean that what had happened was God's decision, and they accepted the division of the land that the Almighty had decided. People came from the town and saw, with veneration, what had happened. They called the place Brodreskiftet, and this came to be the name of the two parts of the farm. In 1928 an amateur archeologist investigated the physical basis of this story. The specific stone was not hard to find and was estimated to weigh 9 tonnes (9,000 kg). The valley along the border line was, by this time, 2 m wide and 3 m deep.

STOP 5-17: (40 minutes) Feldspathic quartzite of Sætra Nappe with rare mafic dikes. This spectacular exposure was recorded on the map of Ramberg (1973) but was apparently ignored in later map compilations until 'rediscovered' in 2006, possibly because it lies in an "offshore" corner of the Orkanger 1:50,000 map. It is located on the north limb of an Ingdal basement antiform (see Stop 5-18) and displays two stages of folding, neither clearly related to regional structure at this time . Earlier tight folds are refolded by more open folds with down-to-W vergence. Classic quartz saddle reefs occur in the hinge region of a late fold. There is also graded bedding. In walking southward from first part of outcrop, the quartzite becomes progressively more feldspathic, also less well bedded, yielding the impression that a basement contact was being approached. However, the entire outcrop, for nearly 1 km, along the coast is feldspathic quartzite.

19:34	**40**			Leave Stop 5-17.
19:38	4	202.4	1.7	Turn right off Route 717 toward Rødberg (Raudbergneset) Lighthouse.
19:40	2	204.1	1.7	Turn right at shore.
19:41	1	204.2	0.1	Proceed west to parking area next to old bell house at west end of harbour jetty.

STOP 5-18: (10 minutes) Proterozoic Ingdal Granite Gneiss.
This gneiss with flat foliation and strong lineation is in the core of a NE-plunging antiform (Tucker 1986). The Ingdal granite was first mapped and described by Ramberg (1943). Later Tucker (1986) mapped and classified it into three types, all thought to represent facies of a single intrusive body: fine-grained gneiss, medium- to coarse-grained gneiss, and porphyritic gneiss. The Ingdal granite is a microcline-rich granitic gneiss with a characteristic red colour. In contrast to most of the gneisses in the Baltican basement, which are migmatitic, this is not. The Ingdal granite (or rocks very similar to it) occur over a large area in the basement both north and south of the Trondheimsfjord. Tucker & Krogh (1988) studied the geochemistry and geochronology. They reported a U-Pb age based on zircon and titanite of 1653+/-2 Ma from a rock collected across the fjord, which is interpreted to represent the emplacement of the granite (Fig. 5.23). A lower intercept of 396+/-5 Ma was interpreted as the cooling of zircon and titanite below their blocking temperatures during the Scandian metamorphism. Later work reported by Tucker et al. (1990, 2004) gave a mean protolith age of 1657 +/-3 Ma. Five Ingdal gneiss samples collected in a progression southward from just across the fjord for a distance of about 10 km along strike (see Figs. 5.8, 5.9), showed discordances along a chord of 6.1, 15.5, 34.7, 36.8 and 39.4% pointing toward a lower intercept of 395 +/- Ma, taken to represent the time when late Scandian cooling terminated Pb

loss. These samples are among the least discordant, whereas samples to the southwest in the vicinity of the Agdenes detachment are 98-100% discordant.
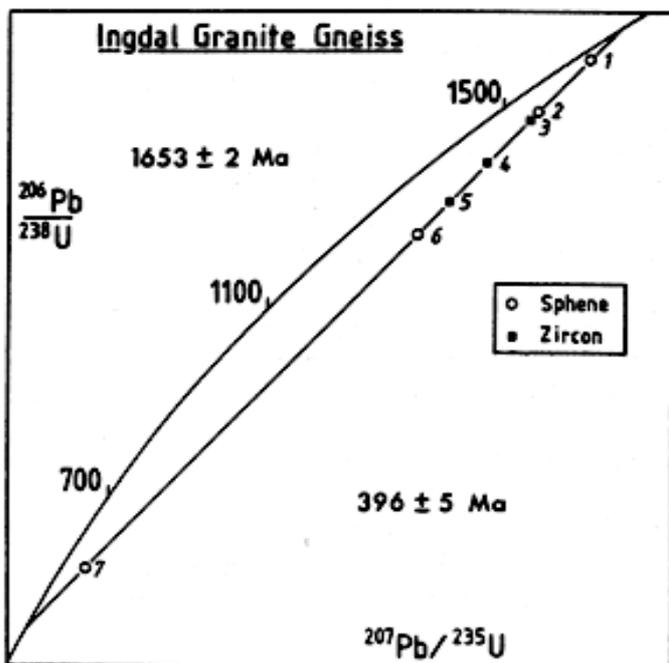

Fig. 5.23 Concordia diagram showing Mesoproterozoic and Scandian U-Pb ages of zircon and sphene for the Ingdal Granite. From Tucker & Krogh (1988).

19:51	**10**			Leave Stop 5-18.
19:54	3	206.0	1.8	Return to Route 717.
19:57	3	208.0	2.0	Sharp right turn uphill, Route 717, center of Statsbygd toward Trondheim and Rørvik ferry.
20:04	7	215.0	7.0	Arrive at Rørvik ferry terminal for ferry to Flakk and Trondheim. It is necessary to pay in full at the small hut before entering the terminal area. (25 minutes crossing time) Large excavation and shore exposures on north side of ferry terminal.

NOTE: Available schedule indicates ferries on Tuesday Rørvik 20:10 -Flakk 20:35 and Rørvik 21:15 - Flakk 21:40. Alternate plans depend on weather and general outlook. 20:10 ferry is most likely.

STOP 5-19: (Time available before driving or walking onto ferry.) Amphibolites in Seve=Blåhø=Skjøtingen Nappe. Predominantly dark-green relatively coarse-grained amphibolites with some felsic layers. Garnets up to 1cm across are common both in the amphibolite and in the felsite. Most of the rocks are interpreted to be of volcanic origin, but concordant porphyritic amphibolites are thought to represent dikes, sills, or microgabbros. The rocks show a prominent SE-plunging lineation considered to be related to thrust emplacement of the nappe. There are also NE-verging folds and many transecting faults trending NE-SW to N-S, some with breccias. Thin pegmatite veins show evidence of strong flattening strains. Note that this outcrop does not agree precisely with contacts shown on earlier geological maps, and was probably not exposed during mapping decades ago. We moved the contact while compiling Figures 5.1 and 5.12.

20:35	**25**			Leave ferry at Flakk.
20:37	2	217.0	1.0	Junction. Left on Route 715. End of road log. Trondheim 11 km, allow 24 minutes.

Day 6, Wednesday, 20 August

Part I -- Morning: Støren Nappe, Upper Allochthon, Trondheim to Løkken.
Part II -- Afternoon: Lower and Middle Allochthons, Oppdal to Sunndalsøra.

David Roberts, Tor Grenne, Peter Robinson & Kurt Hollacher

Summary of route

Departure from NGU at 08.00. We leave Trondheim on the E6 highway, driving south. Initially, we cross the Nidelva river and over the divide to the drainage of the Gaula river, which we follow as far as the exit for Hovin (c. 08.45), where that river is crossed. For the remainder of the morning we travel in uplands between the Gaula and Orkla rivers. Here, in six stops, we examine Ordovician sedimentary, volcanic and intrusive rocks in a classic part of the Støren Nappe, including the distal edge of the Løkken stratabound massive sulphide deposit intercalated with lavas of the Løkken ophiolite. The Løkken Cu-Zn pyrite deposit was mined over a period of 333 years (1654-1987). We have lunch (c. 11.45) at the Astrup Mine Shaft on hornblende gabbro, which stratigraphically underlies but structurally overlies the Løkken ore in the ophiolite sequence.

    After lunch we descend to the Orkla River and drive south up river to Berkåk, where we rejoin route E6 and continue south to Oppdal. All this time we travel in either the Støren or Gula nappes of the Upper Allochthon with related trondhjemite plutons or dykes, or locally in the Seve (Blåhø) Nappe. At Oppdal we join the Driva river and continue south to the famous Engen flagstone quarries in the Sætra Nappe of the Middle Allochthon, showing spectacularly folded Late Neoproterozoic metamorphosed dolerite (diabase) dykes of MORB affinity in feldspathic quartzite. Beyond this we visit road cuts in the Risberget Augen Gneiss Nappe of the Middle Allochthon, and a spectacular mylonite in the same unit. A third stop is made in the Blåhø Nappe at Kongsvoll before we return to Oppdal for a cafeteria dinner.

    After dinner we travel west from Oppdal down the Driva river to Sunndalsøra, en route stopping to see the Neoproterozoic Åmotsdal Quartzite in direct contact with Baltican basement of Trollheimen, and another exposure of Risberget Augen Gneiss with an isoclinal fold. The descent toward sea level is steep down to Gjøra, and then we follow the deep U-shaped valley of Sunndalen to Trædal Vandrehjem just above sea level at Sunndalsøra and near the base of a 1900 m-high valley wall. After settlement in the Vandrehjem, there will be a short drive north to the top of a cliff with a spectacular view over the fjord. This involves an easy ten-minute walk from the parking place.

Part I. GEOLOGY OF THE HØLONDA-HORG-LØKKEN AREA

Tor Grenne & David Roberts

Introduction

The morning of this sixth day will be spent in the Trondheim Nappe Complex, driving south and southwest from Trondheim into the classical Hølonda-Horg district, and westwards towards Løkken, before moving south to Oppdal, via Berkåk (see route, above). Six stops are scheduled for this morning (Fig. 6.1).
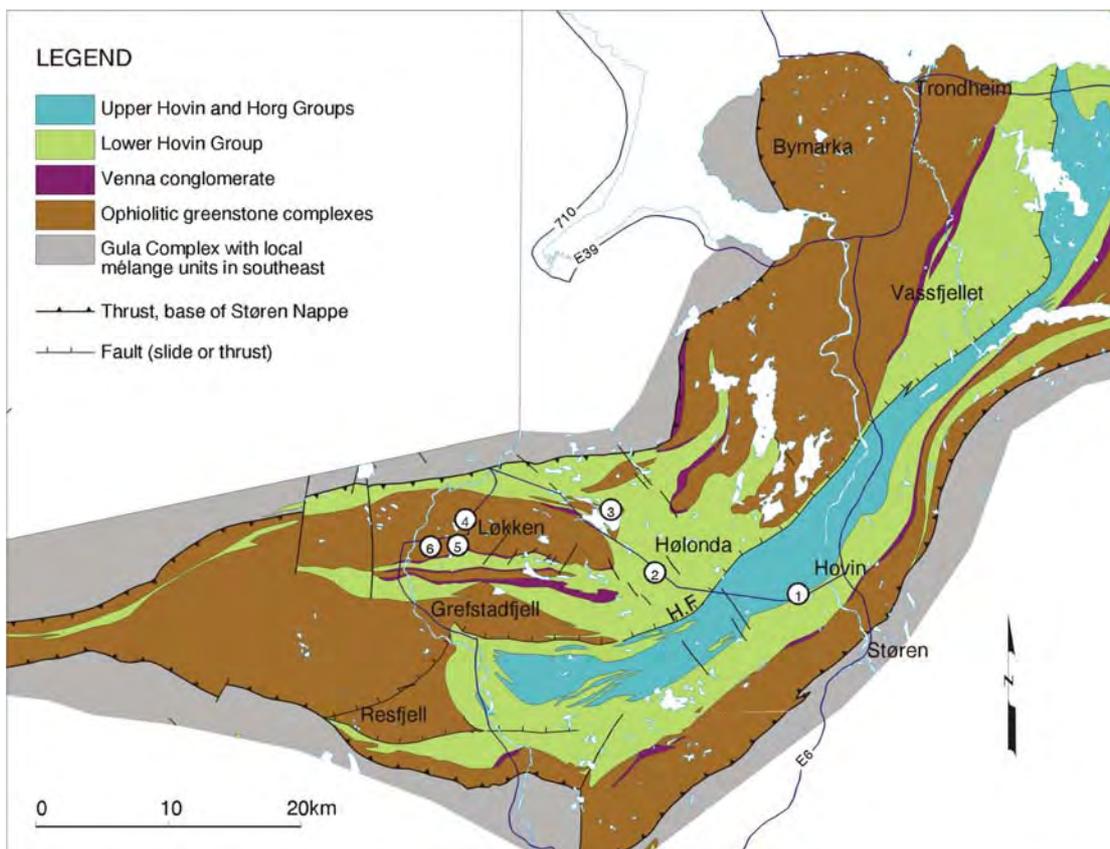
Fig. 6.1. Simplified geological map showing the locations of the principal fragmented ophiolites in the Trondheim-Støren-Løkken district, and the approximate locations of Stops 6.1 to 6.6. The three stops 6.4 to 6.6 are shown again in Figure 6.3. H.F. – Horg Fault.

Driving out of Trondheim on the E6 road we will again see Vassfjellet to our left, essentially an extension of the Bymarka ophiolite and described in detail (as the Vassfjellet ophiolite) by Grenne et al. (1980). As we heard earlier, all of the fragmented ophiolites in this western part of the Trondheim Region have long been regarded as part of one original 'greenstone complex' or 'group' (Carstens 1920), but first recognised as parts of ophiolite assemblages, with metagabbros, sheeted dykes and pillowed metabasalts, by Gale & Roberts (1974). The Hølonda-Horg district, west of the Gaula valley and the E6, has served as a key type-area for the lithostratigraphy of the Støren Nappe, thanks to the work of Vogt (1945) but also by virtue of the fact that large parts of the low-grade succession are richly fossiliferous and have thus served as a mecca for palaeontologists for well over a century (e.g., Brøgger 1877, Getz 1887).

Vogt's (1945) subdivision of the lithostratigraphical succession lying above the 'Støren greenstones' into Lower Hovin, Upper Hovin and Horg series (later changed to groups) still holds in a general way (and will suffice for this field trip), but has been partly revised by Chaloupsky (1970, 1977), Åm et al. (1973), Oftedahl (1979) and Walsh (1986) with the introduction of new (and some old) names. In a nutshell, faunas -- including graptolites, trilobites, brachiopods, gastropods, corals, echinoderms and conodonts -- show that the succession lying unconformably upon the obducted ophiolite(s) ranges in age from Mid Arenig to Caradoc (Kiær 1905, 1932, Strand 1932, Størmer 1932, Bergström 1979, 1997, Bruton & Bockelie 1980, 1982, Ryan et al. 1980, Neuman & Bruton 1989, Neuman et al. 1997). Inferred younger, barren rocks may possibly extend up into Ashgill or, as Vogt suggested, even Llandovery. The faunas are largely of Laurentian affinity, although sporadic Baltican forms occur in the lowest parts of the succession in the southeast (Spjeldnæs 1985, pers. comm. 2002).

The lithostratigraphical succession in this Hølonda-Horg district was originally accorded a synclinal structure (the Horg Syncline), which can in fact be traced northeastwards into the Stjørdal area (Day 4), but a more complicated subdivision into three thrust sheets was advocated by Oftedahl (1979). In particular, an unconformity in the Meldal area, south of Løkken (Ryan et al. 1980) changes northeastwards into a fault (the Horg Fault of Walsh 1986) or slide, cutting out a large part of the northern limb of the Horg Syncline. This fault can be followed, as a thrust, for at least 50 km to the northeast to the coast 10 km east of Trondheim (Solli et al. 2003). Another steeply dipping thrust fault has been identified beneath the inverted Resfjell ophiolite (Fig. 6.1) (Heim et al. 1987) but its precise northeastward extension remains to be traced in heavily forested terrain. Thus, Oftedahl's (1979) proposed subdivision of the rocks of this area into three thrust sheets may well be correct.

The Hovin and Horg group volcanosedimentary successions vary widely across the region but start with a polymict conglomerate dominated by foliated mafic material representing detritus from the deformed, eroding ophiolite. This sedimentation followed directly after the important Early Ordovician *Trondheim event* (Roberts 2003; earlier called the 'Trondheim disturbance', Holtedahl 1920), coeval with ophiolite obduction. Thereafter follow a variety of grey-green slates, sandstones, tuffs and limestones with incursions of pillowed metabasalts and, in one area, a porphyritic 'andesite' of shoshonitic affiliation (Grenne & Roberts 1998) (Fig. 6.2).

Fig. 6.2. Geochemistry of the Hølonda Porphyrites. A) Chondrite-normalised REE patterns; B) chondrite-normalised trace element patterns; C) Th/Yb vs. Ta/Yb discrimination diagram. Modified from Grenne & Roberts (1998). The blue and red colours represent two slightly different magma series.

Higher up there are turbiditic greywackes and shales, sporadic limestones, polymict or monomict conglomerates, and rhyolites or rhyolite tuffs. Accumulation of much of this succession is believed to have occurred in an arc-related marginal basin (Roberts et al. 1984). We will see some of these lithologies at three stops. The very low-grade metamorphism and major structures that affect the succession are part and parcel of the Scandian orogeny. Intra-successional unconformities, and in one case minor folding (Ryan et al. 1980), also termed parorogenic disturbances (*Ekne* and *Horg*, Vogt 1936, 1945), may relate to Taconian orogenesis (Roberts 2003), since the bulk of what is now the Støren Nappe is considered to derive from offshore Laurentia (e.g., Stephens & Gee 1985, 1989, Pedersen et al. 1992) or partly from volcanic islands located within the Iapetus Ocean (Bruton & Bockelie 1980, Neuman 1984, Neuman & Bruton 1989).

The Løkken ophiolite (Fig. 6.3) is located in the inverted limb of a fold nappe (Roberts et al. 1984, Heim et al. 1987) which was refolded to produce a major E-W-trending, open, synformal structure, later dissected by sets of NE-SW and ESE-WNW-trending steep faults and generally west-dipping low-angle faults.
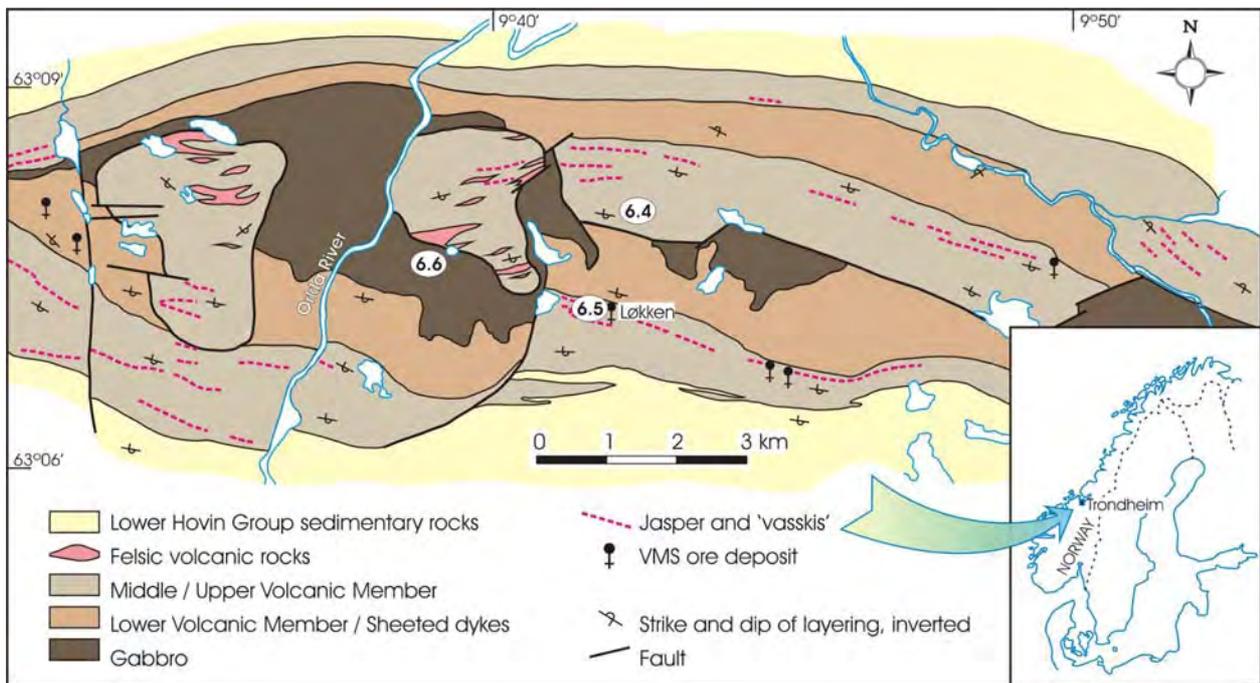
Fig. 6.3. Simplified geological map of the Løkken ophiolite showing the location of excursion stops 6.4 to 6.6. Modified from Grenne & Slack 2003b.

Along with Vassfjellet-Bymarka (Grenne et al. 1980, Slagstad 1998, 2003, Roberts et al. 2002), Grefstadfjell (Roberts et al. 1984) and Resfjell (Heim et al. 1987) it is one of several, supposedly correlatable, fragments of Cambrian-Early Ordovician oceanic crust in the western Trondheim Region (Fig. 6.1). The southern half of the Løkken synform is little deformed, and in some areas the ophiolite pseudostratigraphy is continuous from gabbros in the north to pillow lavas in the south (Fig. 6.4; Grenne, 1989a), the latter in turn being structurally underlain by the younger Lower Hovin Group.

The volcanic sequence is subdivided into three parts, of which the lower member, overlying a sheeted mafic dyke complex, comprises a ~1 km-thick pile of non-vesicular pillow lavas of normal mid-ocean ridge basalt composition (Fig. 6.5). The upper member (0.5–1.0 km thick) is composed of non-vesicular pillowed and massive basalt flows of back-arc basin geochemical affinity, with subordinate rhyolite flows occurring in the western parts of the ophiolite complex; thin units of metapelite are present locally. A variably thick, middle volcanic member is dominated by voluminous, sheet-like, basalt flows but also contains local andesitic to dacitic flows, interpreted to represent rapid tapping of shallow magma chambers that included a significant proportion of intermediate and felsic differentiates ('plagiogranite'). Thick volcanoclastic breccias are abundant, interpreted as fault scarp-related talus deposits. The middle member also contains volcanogenic massive sulphide (VMS) ores, along with beds of jasper (hematitic chert) and a lithological assemblage traditionally referred to as 'vasskis' in Norwegian mining terminology. The latter comprises thin layers of pyrite and/or pyrrhotite alternating with layers composed mainly of iron-rich silicates and magnetite. In many publications the term 'vasskis' is translated as 'iron formation'. A U-Pb age of 487 ± 5 Ma (basal Tremadoc) is indicated for the middle volcanic member based on zircons from a co-magmatic plagiogranite (Dunning & Grenne, unpubl. data), such that older parts of the ophiolite may conceivably be of latest Cambrian age. This falls in line with evidence elsewhere that much of the Iapetus Ocean floor was generated during Cambrian time (Roberts & Gale 1978, Dunning & Pedersen 1988).
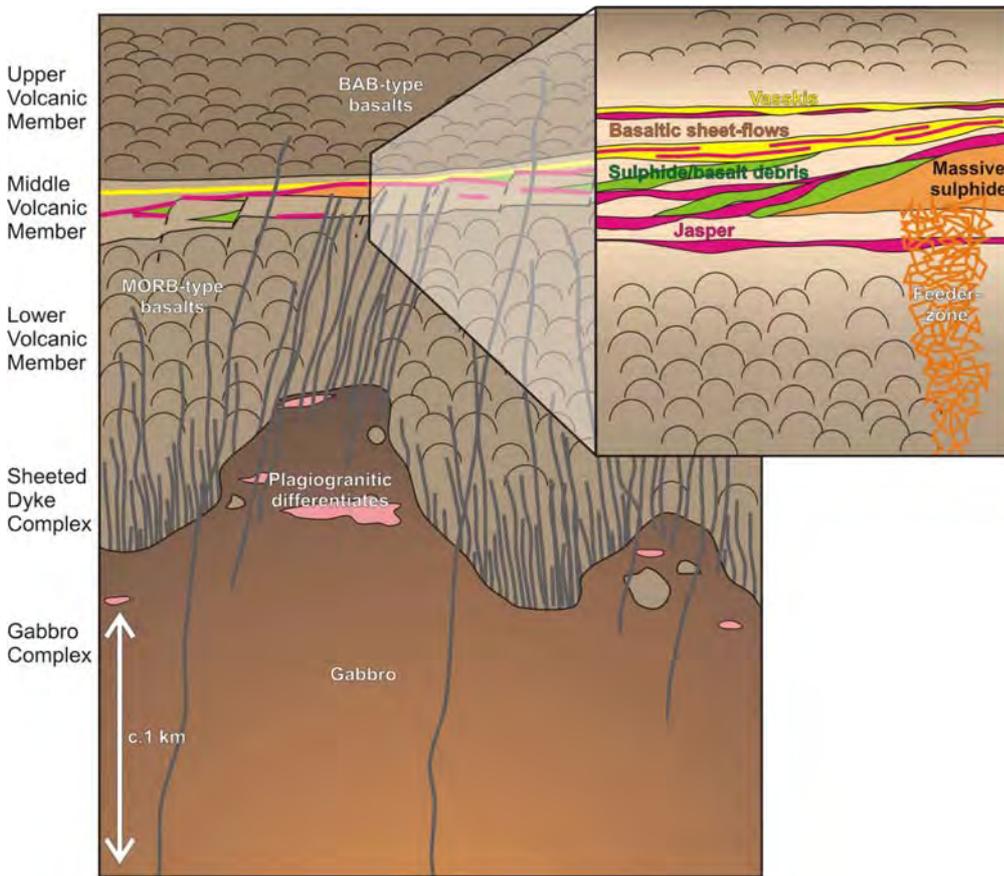
Fig. 6.4. Schematic stratigraphic section of the Løkken ophiolite. Modified from Grenne & Slack 2003b.
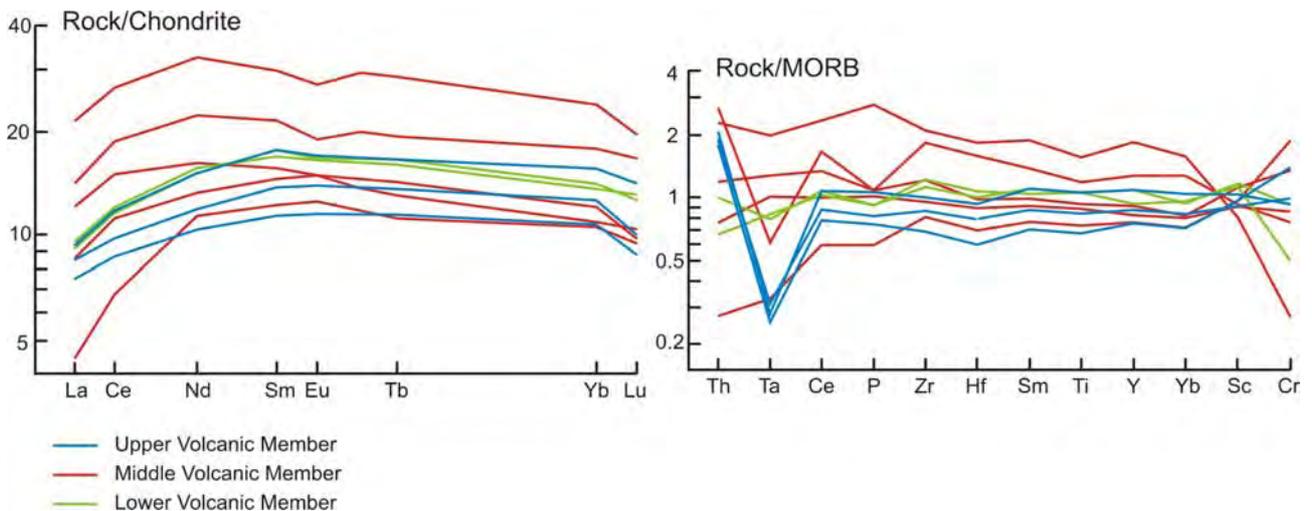

Fig. 6.5. Chondrite-normalised REE patterns and MORB-normalised trace element patterns of basaltic lavas from the Løkken ophiolite. Modified from Grenne (1989a), with supplementary unpublished data for Th and Ta. The diagrams are based on a representative selection from a total of more than 100 analysed samples of mafic magmatic rocks of the Løkken ophiolite.

The Løkken Cu-Zn pyrite ore body is the world's largest ophiolite-hosted VMS deposit (Grenne 1986, 1989b, Grenne & Vokes 1990, Vokes, 1995, Grenne et al. 1999). During the 333 years of its existence, the mine produced 24 million metric tons (Mt) of pyritic ore at an average grade of 2.1% Cu, 1.9% Zn, 0.02% Pb, 19 ppm Ag, and 0.2 ppm Au (Vokes 1995). The main sulphide mass had the form of an elongated body with a total length of about 4 km, an average width of between 150 and 200 m and an average thickness of about 50 m (Fig. 6.6), a morphology which can be ascribed largely to primary features and processes such as subparallel, sea-floor, fault scarps and an extensive hydrothermal vent system on the sea floor (Fig. 6.7).

Fig. 6.6. Vertical longitudinal section and cross-section of the Løkken massive sulphide deposit. Modified from Grenne et al. 1980.

Fig. 6.7. Pre-deformation reconstruction of the Løkken orebody during deposition from hydrothermal vents (including 'black smokers') along fault scarps on the sea floor.

Primary sulphides were precipitated directly above one or more, fissure-related, feeder zones in the hydrothermal vent field, while laminated sediments of sand-sized sulphide debris and coarse sulphide-jasper talus represent reworked material that was deposited 200-300 metres from the main hydrothermal discharge zone as the result of sea-floor faulting and down-slope mass wasting. The feeder zones are represented by sulphide stockworks and penetrative chloritisation of the volcanites, presently found structurally above the massive sulphide ore due to regional inversion of the sequence (Fig. 6.6). Beds of jasper and 'vasskis' can be followed for more than a kilometre from the VMS ores, and represent distal precipitates from the hydrothermal plume (Grenne & Slack 2003a, 2003b, 2005). The size of the ore deposit at Løkken has been ascribed to an unusually large, high-temperature, sea-water convection system in the oceanic crust that resulted from replenishment of hot, primitive magma into exceptionally shallow magma chambers (Grenne 1993).

From Løkken, we continue southwards, meeting the E6 at Berkåk (at c. 12.45) and driving southwest across the Early Silurian, trondhjemite-norite, Innset Massif into the Oppdal area (reach Oppdal at c. 13.20, but continue driving south).

Stop 6.1: Near Sjurdsmoen farm. Diverse lithologies of the Upper Hovin and Horg groups; Støren Nappe. Allow 30 minutes. Leave at 09.20.

Location
Road-cuts in the vicinity of a sharp right-hand (westward) bend in the Hovin to Hølonda road, c. 300 metres east of the farm Sjurdsmoen. This is about 3 km along the road from the bridge over the Gaula river gorge. About 100 m before the big bend, cross a cattle grid (signposted 'Ferist') and park on left close to a side road. Map-sheet Hølonda 1521-2, UTM coordinates 32VNQ6035/9770.

Introduction
The general geology and stratigraphy of this area, between the Gaula valley and Hølonda (c. 10 km west of the Gaula), was first mapped and described in detail by Vogt (1945) and remains today a key and classical area in any discussion of stratigraphic correlation within the Støren Nappe. Although some revisions to the lithostratigraphy have been proposed (e.g. Chaloupsky 1977), and faults are known to disrupt the succession (Oftedahl 1979, Walsh 1986), thus laying doubt on the basic subdivisions (e.g. Neuman et al. 1997), Vogt's main subdivisions still hold in this eastern area. The 1:40,000 bedrock map of Vogt (1945) and the 1:50,000 map of Chaloupsky (1977) will be displayed at this stop.

Description
Almost thirty years ago, Wolff et al. (1980) described this as "a fresh road-cut in the new road Hovin-Hølonda". Now, today, we will hopefully just be able to make out the following lithologies.

At the eastern end of the road-cut there are steeply dipping, thin-bedded, rusty, dark grey, pyritous shales with ribs and thin beds of silty sandstone. This unit is part of the Horg Group of Vogt (1945), of possible Llandovery age. A thin, partly attenuated, quartzite conglomerate (Lyngestein conglomerate) follows (now barely visible here but distinctive along strike), marking the base of the Horg. This is succeeded (to the northwest) by a tuffite or volcanogenic sandstone of rhyolitic composition containing a few flakes or flat clasts of shale. Vogt (1945) described this as his 'Grimsås rhyolite', part of the Upper Hovin Group, and considered that "an effusive origin seems most probable". However, he also realised that whereas some of the 'rhyolite tuffs' elsewhere in the district are massive, others are well bedded with thin laminae of shale, and the latter were duly accorded the name 'tuffitic sandstones' by Chaloupsky (1977).

In the road-cut, the 'tuffite' is succeeded, stratigraphically downward, by a polymict conglomerate (Volla conglomerate) and banded sediments with thin limestone beds containing brachiopods, gastropods, rugose and tabulate corals, *Solenopora* algae and crinoid debris (Bockelie & Bruton, in Gee & Wolff 1981, p. 67), indicative of a Late Caradoc-Ashgill age (D.Bruton, pers.

comm. 2008). A block of this fossiliferous bedded limestone is also found in the Volla conglomerate.

Structurally, this road-cut outcrop is located in the northwestern limb of the Horg Syncline (Vogt 1945), cored by the mainly pelitic rocks of the Horg Group. As noted above, however, parts of the syncline may have been excised along a strike-parallel fault or slide. The spaced cleavage seen here (dipping c. 45º NW) is considered to be a comparatively late structure.

Stop 6.2: Gåsbakken. Hølonda Limestone and Hølonda Porphyrites, Lower Hovin Group; Støren Nappe. Allow 30 minutes. Leave at 10.05.

Location
Road-cut on a sharp bend along the Hølonda-Svorkmo road c. 1 km east of Gåsbakken and c. 250 m west of the farm Øyan. NB! -- This is a dangerous stop, traffic-wise, so take great care. There are also new outcrops on a driveway leading up to a house, about 50 m northwest of the bend. Map-sheet Hølonda 1521-2, UTM coordinates 32VNQ5070/9875.

Introduction
The Hølonda Limestone has featured as a magnet for palaeontologists over the last century, since the initial discovery of fossils by Brøgger (1875). At other localities (but not Stop 6.2) the limestone contains a rich trilobite, brachiopod, conodont and crinoid fauna of Late Arenig to Early Llanvirn age (Neuman & Bruton 1974, 1989, Bergström 1979, Bruton & Bockelie 1980). Occurring in close association with the limestone are 'andesitic' porphyrites (Hølonda Porphyrite), the geochemistry of which is indicative of shoshonitic magmatism (Fig. 6.2) in a mature volcanic arc situated outboard from an active continental margin setting (Grenne & Roberts 1998). Although the faunas in the limestone are largely of Laurentian affinity, some of the brachiopods are completely unknown in well studied North American carbonate rocks, a fact which led Neuman (1984) and Neuman & Bruton (1989) to consider the limestone-fringing island arc as having lain at some unknown distance offshore from the Laurentian continent within the Iapetus Ocean.

Description
The road-cut exposes well bedded Hølonda Limestone and penecontemporaneous porphyrite, with some shales above the limestone. Small blocks of porphyrite occur in the uppermost part of the limestone, and blocks of limestone in the lower part of the porphyrite. The question of the precise nature of the porphyrites, either intrusive or extrusive, has plagued several generations of geologists, and the terminology of the rocks has varied in keeping with the preferred mode of origin (a discussion, with a complete bibliography, is given in Grenne & Roberts, 1998). Remapping of the Hølonda area (Bruton & Bockelie 1980) and subsequent field control and geochemical study (Grenne & Roberts 1998) has convinced us that both extrusive and intrusive porphyrites (lava flows, sills and dykes) are present in this district. Some 10 km farther to the west, porphyrite dykes up to 50-100 m thick cut across strongly sheared, inverted metabasalts of the Løkken ophiolite (Grenne & Roberts 1981).

Although a low greenschist-facies metamorphism has affected these rocks, the massive flows, sills and dykes commonly preserve igneous textures and primary minerals. The dominant phenocryst phase is plagioclase, up to 2 cm across, albeit strongly sericitised, and there are varying proportions of augitic clinopyroxene and orthopyroxene (hypersthene). Secondary minerals are albite, amphibole, epidote and chlorite, with biotite appearing in the northern parts of the Hølonda area. The petrological and geochemical data show that the porphyrites range in composition from absarokitic to latitic and belong to the shoshonite association of mature, subduction-related, arc magma types (Grenne & Roberts 1998), with strong LREE enrichment and La_N/Yb_N ratios between 8 and 14. Shoshonitic lavas have previously been reported from the Støren Nappe in a part of the northern Trondheim Region (Roberts 1982).

Stop 6.3: Sjømoen, Lake Svorksjøen. Venna Conglomerate, basal formation of the Lower Hovin Group, Støren Nappe. Allow 20-25 minutes. Leave at 10.25.

Location
Road-cuts along the northeastern side of Svorksjøen lake, c. 5 km northwest of Gåsbakken. Difficult parking here, so it is best to park outside the gate of the small farm Sjømoen (c. 150 m northeast of the first bend in the road). Because of time, examine just the first 100-200 m of the road-cut. The road is narrow here, so keep an eye on the traffic. Map-sheet Hølonda 1521-2, UTM coordinates 32VNR4675/0125.

Introduction
Traditionally known as the '*greenstone conglomerate*' (e.g. Kjerulf 1875, Törnebohm 1896, Carstens 1920) in central Norway, this basal formation of the Lower Hovin Group is called the Venna Conglomerate (Vogt 1945) in this part of the Støren Nappe. Throughout most of the area (but not along this road section) the conglomerate, or locally a breccia, lies unconformably upon the eroded and deformed Løkken ophiolite. Elsewhere in the nappe, a comparable polymict conglomerate lies directly upon other fragmented ophiolite assemblages (Fig. 6.1; also shown in Fig. 4.5).

    The earlier geologists, from a century or more ago, realised that the clasts and subangular blocks in the conglomerate could not have travelled far, deriving mostly from the subjacent, thick, 'greenstone' complexes. It was also noted (Vogt 1945) that cobbles of a white calcite marble in the Venna Conglomerate in an area north of Hølonda are almost identical to a white marble occurring as thin units in the underlying greenstone complex. Moreover, mafic and felsic clasts are commonly foliated. This and other evidence convinced Vogt (1945) that the "substratum……has been somewhat metamorphosed before the formation of the Venna Conglomerate". As we now know that part of the Løkken ophiolite is of Tremadoc age, and the oldest fossils in the Lower Hovin rocks are Mid Arenig, then the pre-Venna metamorphism would fall in the period ?Late Tremadoc to Early Arenig – the age of the tectonometamorphic *Trondheim event* noted earlier.

Description
The road-cuts expose a NW-dipping succession of clast- to matrix-supported polymict conglomerates and intervening beds of dark greenish-grey, gritty or pebbly sandstones, some with just a few isolated clasts. Much of the clast material in the conglomerates is mafic with a variable foliation, clearly deriving from the ophiolite, and there are also blocks of jasper up to 30 cm across (farther west, 50 cm size boulders of jasper are encountered) as well as diverse clasts of metasandstones, schists and quartz keratophyres (metamorphosed felsic extrusive rocks). Higher up in the succession (westward along the road section) the colour of the rocks is paler greenish-grey and eventually grey, signifying a diminished input of green, chloritic, ophiolitic debris. It may be possible to detect graded bedding in the pebbly sandstones and grits. Higher up, alternations of sandstone and silty shale provide better examples of normal, upward younging.

Stop 6.4: Liabakkan. Pillow lava of the Løkken ophiolite, Støren Nappe. Allow 15 minutes. Leave at 10.55.

Location
Cliffs along west side of main road uphill towards Bjørnli/Storås, c. 150 m after the turn-off from the main road from Svorkmo to Løkken. Park on right shoulder of the road. Note – keep a close eye on the traffic. Map-sheet Løkken 1521-3, UTM coordinates 32VNR 535590-7000720.

Introduction
Pillow basalts of the Upper Volcanic Member here are located in the southern limb of the Løkken synform, where the lavas dip north and are upside down. Similar, but more poorly exposed lavas

may be seen farther uphill (south) along the road. About 450 m to the south, an ESE-WNW-trending fault takes us down to the Lower Volcanic Member with its pillow lavas, hyaloclastite breccias and doleritic dykes.

Description
The pillows seen here have irregular shape and size, ranging from more than 2 m across to less than 10 cm. The small varieties probably represent 'pillow fingers', or protuberances, from the larger pillows. Some pillows show the characteristically 'bun-shaped' primary upper surface providing evidence for inversion of the sequence. The pillows commonly exhibit a concentric structure related to the fast cooling of their exterior and progressively slower cooling toward the interior. The originally glassy rims of the pillows are chlorite-rich and relatively dark. Farther in, there is a variolitic zone which displays <1mm-sized individual varioles that are progressively larger (>1 cm) and more abundant towards the pillow interior where they coalesce and gradually disappear. Under the microscope, the presently epidote-rich varioles show a radiating fibrous texture that originally formed from intimate intergrowths of plagioclase and pyroxene fibres centred on plagioclase microlites, and which formed as a consequence of the rapid chilling of the basaltic lava. Amygdales (filled with calcite that is now weathered out) are scarce and relatively small.

Stop 6.5: Katthullet. Distal part of the Løkken ore body, Middle Volcanic Member, Løkken ophiolite, Støren Nappe. Allow 30 minutes. Leave at 11.25.

Location
Steep cliffs c. 100 m south of the main road from Løkken to Bjørnli/Storås. Park on the left shoulder of the road. Walk southeast on a very old forestry track, then south on a slightly overgrown trail to the outcrops on the right. Map-sheet Løkken 1521-3, UTM coordinates 32VNQ 534930-6999530.

Introduction
This or a near-by locality is thought to be the place of discovery of the Løkken ore in 1654. The present part of the ore represents clastic sulphides derived from the primary sulphide precipitates, re-deposited at a distance of 200 m or more downslope from the hydrothermal vents on the sea floor. The proximal (near-vent) part of the Løkken ore is located down-dip towards the north. Due to inversion of the sequence, a small part of actual seafloor surface on which the sulphides were deposited is well exposed on the underside of an overhang in the cliff.

Description
Stratigraphically below the ore horizon, but structurally above it, a basaltic sheet flow is seen in the northern part of the cliff. The basalt is plagioclase-phyric and slightly vesicular, typical of the Middle Volcanic Member. A more vesicular variety with fairly large vesicles can be seen a few metres to the north. Transitions into irregular, thin-rimmed, pillow-like structures are found locally within the upper few metres of the sheet flow. The flow-top surface displays ropy to lobate pahoehoe structures, now oriented horizontally along strike. Shape and curvature of these structures suggest a flow direction towards the south (in the present inverted position) which is away from, and roughly perpendicular to, the longitudinal axis of the proximal parts of the Løkken ore body, in accordance with interpreted transport direction for the clastic sulphide sediments.

The stratigraphically lowermost part of the c. 90 cm-thick, distal ore horizon, resting directly on (but structurally below) the pahoehoe flow top, comprises 5-10 cm of finely laminated, fine-grained sulphides alternating with greenish bands. These laminated sulphides were overlain by a 40 cm-thick unit of coarse clastic breccia ore composed mainly of fragments of sulphide ore and basalt. The upper 40 cm of the ore zone comprise thin, fine-grained sulphide (pyrite-chalcopyrite-

sphalerite-pyrrhotite-magnetite) layers alternating with fine-grained chlorite-stilpnomelane-magnetite layers.

A 10-15 m-thick mafic sill concordantly overlies the sulphide horizon. The lower and upper contacts are fine-grained, and a poorly defined columnar jointing can be seen in the northern parts. Its plagioclase-phyric texture and chemical characteristics (Grenne 1989a) demonstrate a magmatic affinity to the Middle Volcanic Member, suggesting that the sill intruded at a shallow level below the sea floor.

The clastic ore horizon reappears to the south of the sill, where one can see a c. 70 cm-thick layer of laminated sulphides similar to those on the opposite side. Several of these sulphide layers exhibit a graded bedding defined by the clastic sulphide detritus, in accordance with the inversion of the sequence. A weak tectonic fabric can be seen along both the southern and northern ore horizons, striking c. ENE-WSW with a more gentle NNW dip than the layering. This fabric is axial planar to minor folds visible in the southern horizon. The southern ore horizon is covered by a pillow basalt that is exposed some 40 m to the west along strike.

Stop 6.6: Moshaugen quarry. Gabbro of the Løkken Ophiolite, Støren Nappe. Allow 30 minutes, including <u>lunch</u>. Depart from Astrup Shaft at 12.00.

Location
Smooth and fairly clean rock surfaces just outside a large quarry producing aggregate (crushed rock), east of the 'Astrup Shaft' (former mine shaft of the Løkken ore). Turn left from the main road to Storås west of Bjørnli, along a small gravel road (signposted Moshaugen) towards the Astrup Shaft. The quarry will be seen on the right, after c. 500 m. Note that the actual quarry is not open to the public due to risk of personal injury. For your safety, please stay away from the quarry edges. The entire area may be inaccessible during periods of drilling and/or blasting. Alternative outcrops in gabbro can be found c. 100-150 m to the WSW, just behind (north of) the Astrup Shaft buildings. Map-sheet Løkken 1521-3, UTM coordinates 32VNR 532700-7000050.

Introduction
The gabbros exposed here represent the lowermost preserved parts of the Løkken ophiolite pseudostratigraphy. The outcrops are directly above the Løkken VMS orebody; the nearby Astrup Shaft cuts vertically through more than 600 m of gabbro before it enters metabasalts across a gently west-dipping fault plane. The Astrup Shaft was finished in 1972 and brought the sulphide ore up from depths of 700-1000 m below surface, until the Løkken mine closed down in 1987.

Description
The rocks in the quarry area and near the Astrup Shaft are from relatively deep in the gabbroic complex, typically comprising medium- to coarse-grained hornblende gabbros with varying proportions of clinopyroxene. Doleritic dykes are relatively scarce at this depth in contrast to shallower parts of the gabbroic complex where they become progressively more abundant.

Due to a continuous expansion of the quarry, the outcrops change with time with respect to quality as well as appearance. Features that may be seen here include isotropic gabbro, banded and comb-layered gabbro that was probably formed by crystallisation along the walls and roof of magma chambers, small lenses of pegmatitic hornblende gabbro, and fine-grained doleritic dykes that commonly have somewhat irregular and weakly developed chilled margins suggestive of intrusion into a gabbro that was still fairly hot. Epidotisation, locally associated with sulphide impregnation, is very abundant and affects both the gabbro and the dykes in the form of thin veins, vein networks or pervasive epidote alteration in zones up to a metre in width. In places, the epidotisation zones and networks are cut by dykes that are themselves affected by subsequent epidotisation, providing evidence for contemporaneous hydrothermal activity, dyke injection and gabbro cooling. Similar epidote alteration zones, sometimes referred to as epidosite, are common in ophiolitic gabbros such as at Troodos, Cyprus. They are generally interpreted as a reflection of

hydrothermal circulation deep in the oceanic crust and associated with the formation of seafloor sulphide (VMS) deposits.

On leaving the Astrup Shaft and the Løkken Mine, we have an approximately 1½ hour drive to the Oppdal district and our next stop. The route takes us mostly through lithologies of the Støren Nappe, touching the Gula Nappe in the Berkåk area, and on the way we will catch glimpses, in road-cuts, of many trondhjemite dykes cutting diverse volcanogenic units. South of Ulsberg we pass through the 180 km^2 *Innset massif,* composed of trondhjemite and norite. These bodies have yielded U-Pb zircon ages of 434.8 ± 0.5 and 435.8 ± 0.9 Ma, respectively (Nilsen et al. 2003). Comparable crystallisation ages have been generated for several trondhjemites and diorites in the Gula, Støren and Meråker nappes in these central and southern parts of the Trondheim Region (Dunning & Grenne 2000, Nilsen et al. 2006). These Early Silurian trondhjemites are high-Al/low-Yb plutons and dykes characteristic of continental margin settings – and chemically quite unlike the Cambro-Tremadoc, low-Al/high-Yb trondhjemites (plagiogranites) of oceanic origin. One such Silurian trondhjemite is the Toset pluton near Ulsberg (dated to 432 ± 4 Ma) which has been used as a facing stone in the new Opera House in Oslo. Trondhjemite from the type locality at Follstad near Støren has yielded a U-Pb zircon age of 432 ± 3 Ma (Dunning & Grenne 2000).

Day 6, Part II: GEOLOGY OF TROLLHEIMEN

Wednesday Afternoon August 20

SUMMARY OF GEOLOGY OF TROLLHEIMEN

by **Peter Robinson and Kurt Hollocher with contributions by Bob Tucker and Alan Krill**

Introduction. In the afternoon/evening of Day 6 we will have five stops on the edge of Trollheimen, a roughly anticlinal region at the northeast edge of the Western Gneiss Region (WGR) (Fig. 5.1), where it abuts rocks of the Upper Allochthon exposed in the Trondheim Basin. It is a mountain region with relatively easy access routes and lacks the high relief and very steep slopes of the fjord country to the west. The eastern part of the region was mapped by O. Holtedal (Holtedahl, 1938; Holtedahl & Rosenqvist, 1941) and H. Holtedahl (Holtedahl, 1950), who showed that the deepest layers of gneiss are overlain by a thick quartzite (Åmotsdal Quartzite), in turn overlain by a second layer of gneiss, and then a second layer of quartzite (Fig. 6.8). This layering outlined a series of anticlines and domes that cover most of eastern Trollheimen in the Røros og Sveig 1/250,000 map sheet (Nilsen & Wolff, 1989), and extend westward in the Ålesund 1/250,000 sheet (Tveten et al. 1998) to Sunndalsøra (our overnight location for Day 6).

The structure of an inner part of Trollheimen was studied in detail by E. Hansen (1971), who also mapped stratigraphic units, including strata well above basement, but their significance was not recognized until the studies of Gee (1980), and Krill (1980, 1985, Krill & Röshoff , 1981). This later work established the correlation of tectono-stratigraphic units in and just east of Trollheimen with the more frontal part of the orogen in Sweden. It also allowed correlation around the northeast-plunging anticlinal hinge of Trollheimen to the Surnadal Synform to the northwest (Hernes 1956, Rickard 1985, Krill 1987) and to the regions still further northwest (Gee & Wolff 1981; Tucker 1986; Tucker et al. 2004), and southwest (Robinson 1995) (Day 5 and Day 8).

Tectono-stratigraphic and structural relationships. The deepest levels consist of Mesoproterozoic granitoid gneisses with relatively minor lenses of mafic rock dominated by biotite and hornblende. The overlying quartzite is feldspathic and locally conglomerate with boulders partially of underlying granites. The quartzite is rich in muscovite and hematite, indicating partial derivation by weathering of the granites, and the base is interpreted as a Neoproterozoic unconformity (Gee, 1980). Locally at the top, on Hemre Kam, near Gjevilvatnet, is a thin layer of graphitic sulfidic mica schist with significant uranium, that has been correlated with the Middle Cambrian Alum Shale of the Scandian foreland seen on Day 1 (Gee 1980). The quartzite and schist are overlain by the upper granitoid gneiss layer, slightly more deformed and recrystallized, and with lenses of garnet-corona gabbro and local shear zones with eclogite-facies assemblages as described at Vinddøladalen, a short distance east from Surnadal (Torudbakken 1980, Krill & Röshoff 1981). Fabrics in the contact zone between quartzite and gneiss indicate top-E shear, consistent with SE thrusting of gneiss over quartzite. Maps indicate this zone covers a distance of at least 40 km across strike, providing a minimum value for thrust displacement.

Near Gjevilvasskammene, where Gee (1980) described the thrust, the base of the upper gneiss, resting on mica schist, is overlain by slices of strongly recrystallized quartzite (Hansen 1971) (not shown on any map), that appear to be imbrications of lower quartzite, originally 40 or more km to the NW, interleaved with upper gneiss during early thrusting. They appear restricted to regions near the base of the upper gneiss. Elsewhere, particularly in the Ålesund 1:250,000 sheet, the lower quartzite pinches out, leaving deformed upper gneiss in contact with less deformed lower gneiss, creating problems in recognizing this major thrust through the WGR.

Within the very tight Rekdalshesten antiform, ~100 km west from Sunndalsøra, Robinson (1997) has mapped lower and upper gneisses separated by a highly deformed feldspathic, muscovitic, hematite-bearing quartzite (see Fig. 8.4). The quartzite is 100 m thick on the antiform crest, thinning to 1m on the S limb. Dual gneiss units separated by quartzite are also recognized within the Grong-Olden culmination and in the Tømmerås window (Fig. 0.1) northeast from Trondheim (Gee 1980). Gee (1980), named the thrust for Gjevilvasskamene, also, the type localities for four of Hansen's stratigraphic names, now discarded. Here we prefer Storli thrust (Robinson et al. 2004, Tucker et al. 2004), for a mountain village where it is exposed.

Above the upper gneiss above the Storli thrust, there is a second (upper) Neoproterozoic quartzite unconformable on upper basement (Stop 6-10), followed by a standard sequence of nappes with enormous variations in thickness and locally complete absences. The lowest, above upper Åmotsdal Quartzite, is usually the Risberget Nappe of the Middle Allochthon, of deformed Mesoproterozoic rapakivi granite, locally with metamorphosed gabbro and white anorthosite (see Stops 6-8A,B,C, 6-11). The Autochthon (including the above-mentioned sulfidic schist), and Lower Allochthon are generally absent. The next unit in the Middle Allochthon is the Sætra Nappe (see Stop 6-5). The lower part of the Upper Allochthon consists of high-grade mica schists and amphibolites of the Blåhø Nappe (Stop 6-9), correlative with the Seve Nappe in Sweden. The upper part of the Upper Allochthon consists of much lower grade, locally fossiliferous, Ordovician metamorphosed volcanic and sedimentary rocks of the Støren and related nappes seen on Days 4,5,6. From Oppdal the units trace around the northeastern hinge of Trollheimen, then south into the Surnadal Synform (Hernes 1956, Rickard 1985, Krill 1987) and beyond (Gee & Wolff 1981, Tucker 1986, Robinson, 1995, this Guidebook, Days 5, 8). Commonly the upper quartzite is missing, and the Risberget Nappe rests directly on basement, but other units

MAP OF TROLLHEIMEN SHOWING TRUST REPETITION OF BASEMENT

Based on Røros 1:250 000 Sheet (right) and Ålesund 1:250 000 Sheet (left)

Granitoid intrusions in Blåhø and Støren Nappes

STØREN NAPPE
Mafic to felsic volcanic rocks, graywacke, phyllite, marble, quartzite

BLÅHØ NAPPE
Garnet-feldspar schist and amphibolite
Gabbro/amphibolite, ultramafic rock

SÆTRA NAPPE
Quartzite with amphibolite

RISBERGET NAPPE
Augen gneiss (rapakivi granite)
gabbro/amphibolite, anorthosite

ÅMOTSDAL QUARTZITE OVERLIES UPPER BASEMENT
Conglomerate separately indicated

UPPER BALTICAN BASEMENT
Strongly deformed, recumbently folded, black=eclogite or garnet corona gabbro

ÅMOTSDAL QUARTZITE OVERLIES LOWER BASEMENT
Conglomerate separately indicated

LOWER BALTICAN BASEMENT
Less strongly deformed, less metamorphosed

6-8 Field Trip Stops

Fig. 6.8

including the Støren Nappe are represented. Also from Oppdal (Fig.6.8) units trace W in a syncline across Trollheimen that nearly reaches Eikesdal, 20 km SW of Sunndalsøra. Thus, northern Trollheimen basement areas connect to the rest of the WGR through a corridor barely 10 km wide.

In Trollheimen the nappes and underlying upper basement were subject to folding, completely post-dating thrust assembly of the nappe pile. Prominent are recumbent folds described by Krill (1985, 1986), rooted in central Trollheimen and extending E into the upper Driva Valley (Fig. 6.8), including outcrops of Risberget and Sætra at Stops 6-7,6-8.

Riar Basin, recumbent folds and Storli thrust. A complex recumbent syncline in the central part of northern Trollheimen, was mapped by Hansen (1971). The bottom limb is roughly parallel to the Storli thrust, while the top limb carries basement and its inverted tectonostratigraphic cover at least 8 km SE. The area of this eastward inverted sequence is also the area of the Riaren structural basin and "hand fold" (Fig. 6.8), outlined by the stratigraphic top of the inverted "upper" quartzite, with four fingers to the left and thumb to the right. In the basin center, vertically plunging folds superimposed on all units were Hansen's focus of study. A layer of Risberget augen gneiss, 10 m or less thick, forms a "butter" on the fingers in many places, and separating quartzite from the tectonostratigraphically higher Blåhø Nappe in its type locality, with conspicuous absence of the Sætra Nappe between.

The Riar basin, is a very local structure (Hansen, 1971, Robinson unpublished data 1990). The area of 80° plunges is 2km across, 40° plunges 6 km across, and the linear structure dies out southward ~3 km from the basin center. The earlier recumbent fold is important because its axial surface appears to be truncated by the Storli thrust at Storli. These relationships, suggest the following local sequence of major tectonic-metamorphic events: 1) Early, SE-directed thrusting with establishment of nappe tectono-stratigraphy; 2) SE-overturned recumbent folding of nappe units and underlying basement, including recumbent anticlinal hinge zones extending into cover regions SE of Oppdal and establishment of incipient eclogite-facies metamorphism; and 3) SE-thrusting along the Storli thrust, carrying already recumbently folded/metamorphosed units and upper gneiss over less deformed lower quartzite and gneiss.

Regional implications of local geology. Tucker et al. (2004) suggested that forward thrusting on the Storli thrust could have been related in time around 395 Ma to genesis of the higher-level Agdenes extensional detachment (see Day 5, Stop 5-15), permitting cooling of a large area of the WGR determined by titanite ages (Figures 5.8, 5.9). The Storli thrust would have carried previously deep-seated metamorphic rocks including eclogites, SE over the Baltican margin, providing gravitational potential for the immediately following Agdenes detachment. Along the detachment, basement previously above the blocking temperature of titanite, was brought against much cooler crust of the Upper Allochthon, then both were deformed together in a still later ductile extensional deformation.

Geologic implications of present knowledge of Trollheimen need to be followed up. There is clear evidence that basement gneisses of the WGR need not be, and probably are not, part of a single tectonic slice, not only in early phases of Scandian deformation, but in late ones. The important "lower" quartzite horizon, providing strongest evidence for basement imbrication in Trollheimen, pinches out, leaving the possibility that such boundaries are widespread, but difficult to recognize. Evidence is also provided for major ductile deformation and SE-directed recumbent folding post-dating formation of thrust-related tectono-stratigraphy. This appears to pre-date the major SE-directed Storli thrust, implying that thrusting occurred in two or more phases separated by uncertain intervals. The Storli thrust surface and the slightly later Agdenes detachment appear both to have been involved in later extreme ductile deformation under low-amphibolite facies conditions. This occurred both in large regional strain fields operative near the Surnadal Synform and beyond, and in local strain fields such as that of the Riar Basin.

176.0	13:21	Oppdal. Stay straight on E6.
179.0	13:25	View of Blåhø Mountain and gorge of Driva River.
187.4		Road cuts in Risberget augen gneiss. View of high flagstone quarries.
192.7	13:38	Skiferbrud. Entrance to flagstone quarries on right. Go under underpass, turn right, follow narrow road up hill. Locked gate may require walking here.
193.8	13:41	Quarry building at top of rough road and northwest-facing cliffs.

STOP 6-7: (45 minutes) Eidsvoll Flagstone Quarry in the Sætra Nappe.
Folded late Proterozoic feldspathic sandstone cut by dolerite dikes (Figure 6.9). The folds generally plunge southeast and may be typical of folds oriented quasi-parallel to the nappe transport direction during recumbent folding. The following description was adapted from Krill & Röshoff (1981).

Diabase dikes are abundant in the feldspathic quartzite of the Sætra Nappe, but they are not commonly seen in the quarries. The dikes represented structural heterogeneities in the quartzite and prevented its uniform deformation into well foliated flagstones. Therefore the numerous flagstone quarries here mainly avoid the dikes, which are most common between the quarries. Over a hundred dikes can be found among the quarries. The main diabase at this locality is a very irregularly shaped, stepped intrusion, later deformed. Much of it is weakly foliated, but it is strongly deformed near the quartzite wall-rock, which is tightly folded next to it. South of the three-pronged body of metamorphosed diabase, a well preserved part of the stepped intrusion shows that the dike originally truncated sedimentary layering.
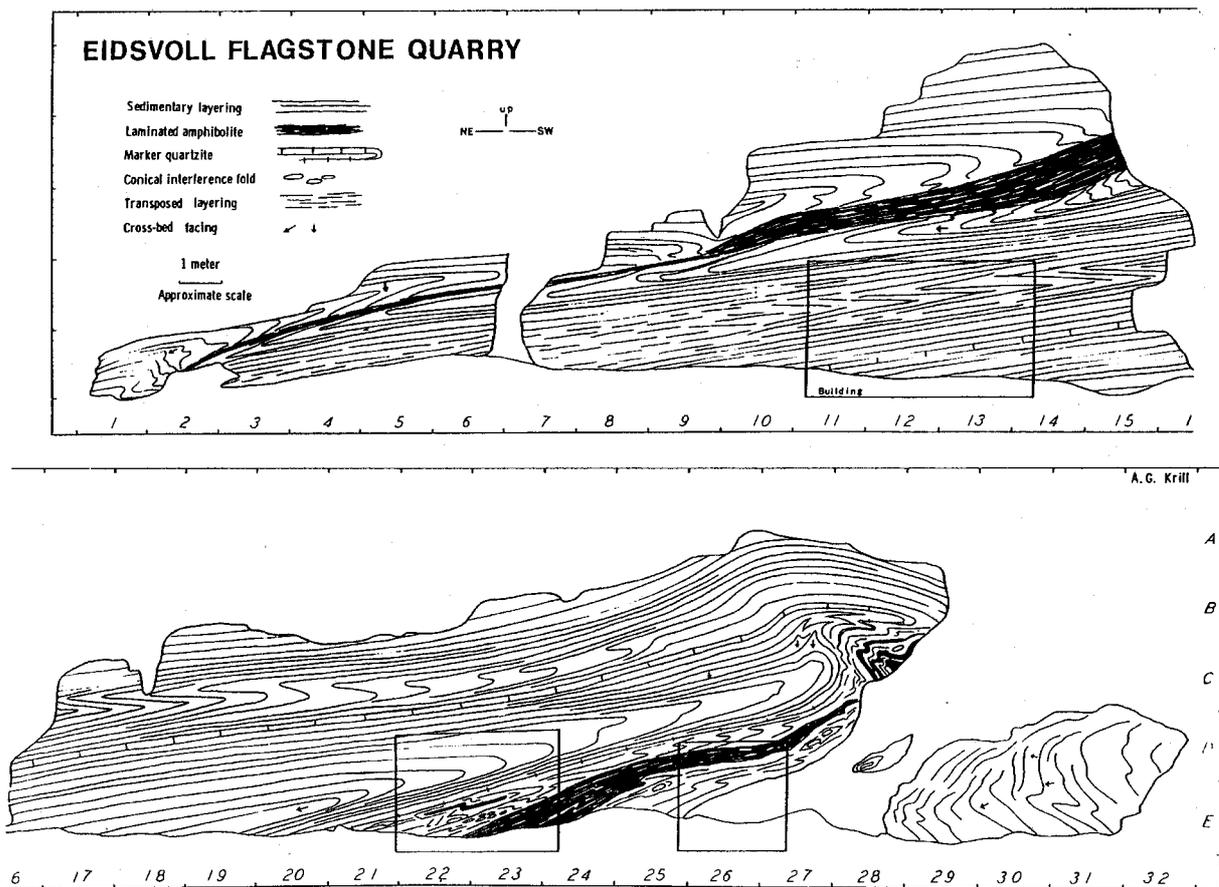
Fig. 6.9 Simplified drawing by A. G. Krill (1986) of west-facing rock wall in the Eidsvoll quartzite quarry at Stop 6-7.

The dike served to shield the quartzite, which is progressively foliated and flattened a few meters away. Farther south, the center of the dike is completely unfoliated. Whole-rock Rb-Sr dating of samples taken from this dike gave an age of 738 ± 65 Ma. On basis of lithology, age and tectonostratigraphic position, the Sætra nappe is considered to be the extension of the Swedish Särv Nappe on the western side of the Trondheim Synclinorium.

Diabase dike samples collected from the vicinity of the Engan quarries served as the key starting point for a recent major- and trace-element investigation of geochemistry (Hollocher et al. 2007a) as a basis for correlation of mafic rocks in the Sætra and Risberget Nappes and adjacent basement through an extensive part of the WGR, as well as correlation with older geochemical data from the Särv Nappe in Sweden and equivalent nappes further north in Norway. The overall results from this study are summarized in the Introduction for Day 5. Localities were visited on Day 5 and additional localities will be visited on Days 7, 8, and 9. See below for results for samples collected near Stop 6-7.

Geochemistry of dikes. Figure 6.10A shows all available data for Sætra Nappe and equivalent dikes in the central Scandinavian Caledonides, with emphasis on samples from near Engan (indicated Oppdal on the figure). All of the dikes are basaltic or nearly so and cluster on the tholeiitic side of the reference line. Dikes at Engan and most other locations are relatively silica-rich (49-53% SiO_2), though generally more alkaline dikes containing 42-49% SiO_2 are found especially near 64°N, northeast of Trondheim. Figure 6.10B compares all Sætra dikes to rocks in different tectonic environments. Most dikes and all at Engan plot in a tight cluster with La_n/Sm_n ratios of 1.0-1.8 and Nb/La ratios of 0.8-1.5. The more alkaline dikes at Ystland and elsewhere plot at higher La_n/Sm_n ratios, up to 2.6, and still more alkaline dikes from elsewhere plot at ratios up to 4.0 (Beckholmen and Roberts, 1999; Solyom et al., 1985). Almost all plot inside the field of East Africa Rift lavas, but are on the low Nb/La ratio side, or outside, of the MORB field which extends to Nb/La ratios >2. Sætra dikes plot far outside the locations of median compositions of arc and flood basalts. Iceland, E-MORB, and ocean island basalts approximate the compositions of different populations of Sætra dikes, though not the most alkaline lavas. Figures 6.10C and E show REE patterns just for the Engan locality. Most of the samples are LREE-enriched, with slightly Z-shaped patterns and small or no Eu anomalies. The sample with the highest LREE concentration and largest negative Eu anomaly also has the lowest Mg' value (40), and so represents a relatively differentiated magma. Two samples have LREE concentrations like most of the rest, but have straighter patterns and HREE are nearly flat, like Sætra dikes at Orkanger and most other localities (Fig. 6.10D). Two samples, unique among all Sætra dikes for which data are available, have low LREE concentrations and flat patterns. All of these samples, however, have similar spider patterns (Fig. 6.x3E) with the exception of positive U anomalies for
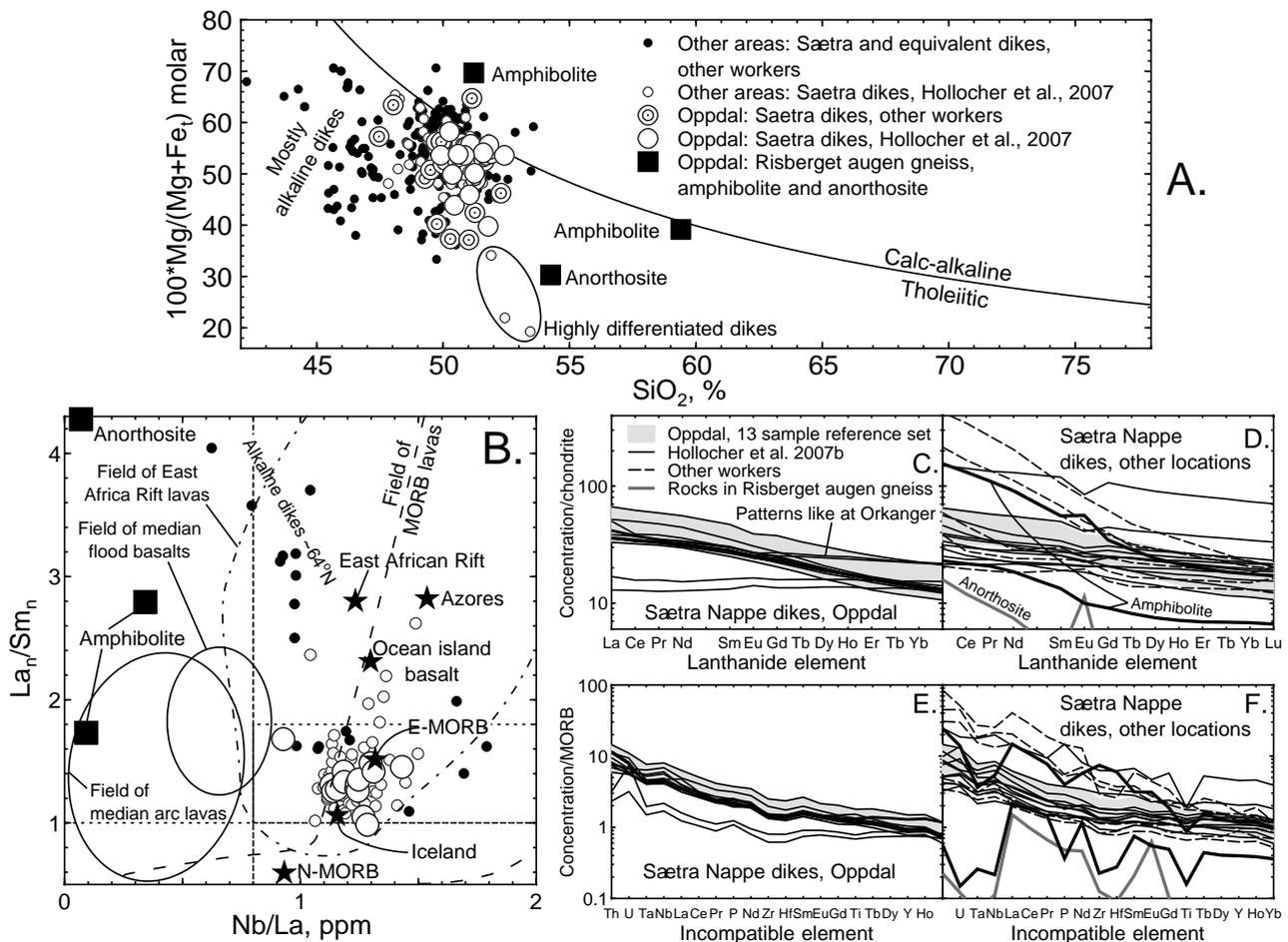
Fig. 6.10 Geochemical relationships of Sætra Nappe dikes at the Engan quarries near Oppdal (Stop 6-7) compared to the same or correlative units elsewhere in the central Scandinavian Caledonides, with two samples from the nearby Risberget Nappe at Oppdal (Stop 6-8A). A) SiO_2 vs. Mg' diagram with a tholeiitic-calc-alkaline reference line (after Miyashiro & Shido, 1975). B) Nb/La vs. La_n/Sm_n diagram comparing Sætra Nappe dike compositions to the fields of mid-ocean ridge basalts and East African rift lavas, the median mafic lava composition from the Azores, East African Rift system, and Iceland, and the fields of the median mafic lava compositions of several arcs, and several flood basalt provinces (see Hollocher et al., 2007a for derivation of these). Also shown are average N-MORB, E-MORB, and ocean island basalt compositions from Sun and McDonnough (1989). See A for key. Sætra dikes from some other locations are shown with 17.9*Ta/La ratios instead of Nb/La ratios (not shown but rationale discussed by Hollocher et al., 2007a). C and E) REE and spider diagrams for Sætra Nappe dikes from Oppdal only. D and F) REE and spider diagrams for representative analyses from other locations, including a representative ~1/5 of samples from Hollocher et al. (2007a) and ~1/3 of samples from other workers. D-F all use the keys shown in D. Plotted data are from Loeschke (1976), Solyom et al. (1979), Loeschke & Schock (1980), Roberts (1982), Stephens (1982), Krill (1983), Roberts et al. (1984), Grenne & Lagerblad (1985), Solyom et al. (1985), Heim et al. (1987), Grenne (1988), Grenne (1989), Grenne & Roberts (1998), Pannemans & Roberts (2000), Roberts & Sundvoll (2000), Nilsen et al. (2003), Slagstad (2003), and Hollocher et al. (2007a and unpublished). REE normalizing factors are from McDonough & Sun (1995), and MORB normalizing factors are from Pearce & Parkinson (1993).

the two lowest LREE samples. Figure 6.10D shows representative REE patterns for Sætra dikes at other localities. Most of these have slightly LREE-enriched, relatively straight patterns like the dikes at Orkanger. The most alkaline Sætra dikes, from the region to the northeast of Trondheim at ~64°N, are strongly LREE-enriched. Interestingly, these most alkaline dikes tend to have more pronounced spider pattern anomalies than do the less alkaline varieties, including small negative Nb-Ta anomalies and, in one case, a large negative Zr-Hf anomaly, characteristics that are not understood at present.

	14:26	Leave flagstone quarry.
194.4	14:29	Pause below quarry. View of cross-cutting dike on right. Turn left through underpass, then turn right.
194.7	14:30	Back on E-6 south. View of Mock Quartzite Factory!!!
197.3	14:33	Turn left on short segment of old E-6. Park and walk back to road cut.

STOP 6-8A: (10 minutes) Anorthosite, hornblende gabbro in Risberget Nappe.
Lenses of metamorphosed anorthosite are associated with various gneisses of the Risberget Nappe. The anorthosite lenses seen here are strongly deformed and fully recrystallized, but a massive anorthositic gabbro in Pineggbekken 4 km further south contains both ortho- and clinopyroxene. Two samples of amphibolite and one anorthosite were collected from this road cut (Hollocher et al. 2007a) , and are shown in Figure 6.10. The amphibolites, one of which has a nominally basaltic composition and the other more andesitic, are clearly different from the Sætra dikes, though in somewhat different ways. The amphibolite with ~51% SiO_2 has lower HREE concentrations than any available Sætra dike analyses, and an enormous negative Nb-Ta anomaly unlike any in even the most alkaline of the Sætra dikes. The other amphibolite has a REE pattern not much different from some of the alkaline dikes, but has a substantial positive Eu anomaly, and large negative Nb-Ta anomaly which, in addition to >59% SiO_2, are unlike any of the Sætra dikes. The one anorthosite from the Risberget Nappe is also very different, mineralogically and chemically, from the Sætra dikes.

North of the metamorphosed anorthosite is the contact (unexposed) with the Sætra nappe. The Sætra nappe lies underneath the Risberget nappe here, as they are inverted on the lower limb of the large-scale Holberget recumbent anticline.

197.5   14:43   Drive or walk to end of segment of old E-6 and examine road cut on left.

STOP 6-8B: (10 minutes) Risberget rapakivi granite.
The rapakivi granite exposed here contains megacrysts of perthitic K-feldspar up to 15 cm long. The rapakivi texture, large K-feldspar megacrysts with thin plagioclase mantles, is considered to be an igneous texture, formed by crystallization from a potassium-rich magma (Tuttle & Bowen 1958, Stewart & Roseboom 1962). This exposure of some of the least deformed parts of the Risberget Nappe is similar to the material from which Handke et al. (1995) obtained a U-Pb zircon age of 1189.0±1.0 Ma (Figure 6.11). This was surprising in view of the Rb-Sr whole rock isochron for this and adjacent outcrops obtained by Krill (1983, see discussion for stop 8C).

14:53   Walk 100m south along new E-6
to road cut on left.

STOP 6-8C: (10 minutes) Mylonite of Risberget Nappe. Krill (Krill & Röshoff, 1981) gave the following description of this outcrop. "The roadcut south of the rapakivi rock contains a common type of augen-gneiss ultramylonite. K-feldspar augen, some with rapakivi mantles, are the only remaining porphyroclasts in the very fine-grained ultramylonitic matrix. Rb-Sr whole-rock analysis of the ultramylonite yielded an isochron: 1533 ± 65 Ma. (Krill, 1983). The ultramylonite may be Precambrian, but its foliation and lineation are parallel to Caledonian fabrics in the nearby Sætra Nappe. Alternatively, the Precambrian date may be the primary age of the rapakivi rock and not the age of the mylonitization."

Fig. 6.11 Concordia diagram showing zircon U-Pb isotopic ratios and igneous crystallization age determined by R. D. Tucker for augen gneiss / rapakivi granite in the Risberget Nappe exposed at Stop 6-8B.

In carrying out his Rb-Sr whole rock study, Krill (1983) noted that the scatter of coarse Rapakivi granite and augen gneiss samples gave a very poor "errorchron" of 1182±115 Ma, which he considered geologically meaningless. In contrast, the ultramylonite samples were much more radiogenic, and when combined with the other samples gave an errorchron indicating an age of 1618±44 Ma. Now that we know, from the zircons. that the age of the rapakivi granite is actually 1189 Ma, it is tempting to speculate that the ultramylonites were produced from another rock in the Risberget Nappe which was not cogenetic with the rapakivi granite and was probably older, possibly the basement into which the rapakivi granite was intruded. This is consistent with the errorchron of 1533±65 Ma obtained from the ultramylonite alone. It seems that this older rock suffered selective transformation to ultramylonite in these outcrops! Krill also obtained a five-point whole rock errorchron of 1163± 80 Ma. for a granitic pegmatite which intrudes the rapakivi granite and is foliated with it as well as sheared along its contacts. Krill's speculation that this pegmatite may reflect a Sveconorwegian thermal event, albeit an early one, seems to be borne out by the new zircon data (see also discussion for Stop 5-3).
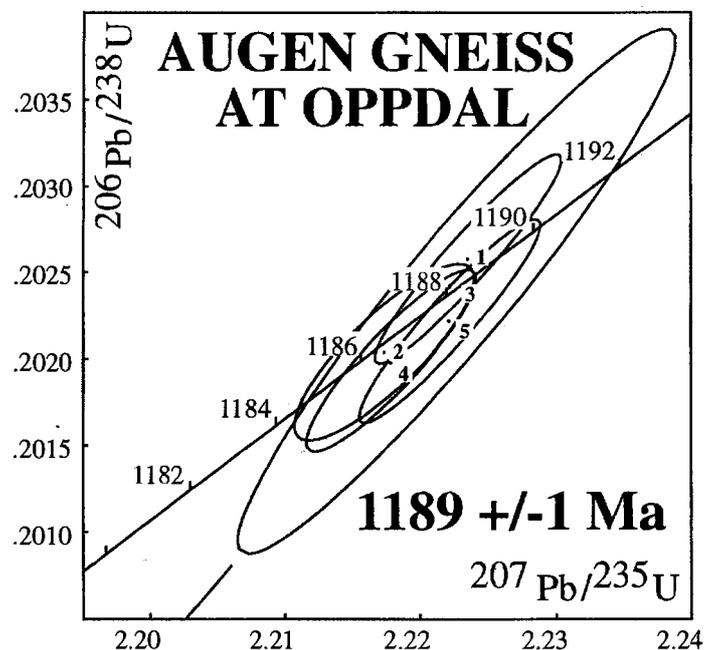
| | 15:03 | Leave Risberget outcrops. Continue south on E-6. |
| 215.0 | 15:18 | Turn right to Kongsvoll Railway Station. Outcrops, loose blocks near west end of bridge over Driva River. |

STOP 6-9: (15 minutes) Biotite-muscovite-hornblende schist and amphibolite of Blåhø Nappe.
    The following note is modified from A. G. Krill, unpublished excursion guide, 1987. The Blåhø Nappe in this area consists of garnet- and hornblende-bearing pelite and amphibolite. In roadcuts about 500m north of here traces of staurolite occur in the schist, and kyanite is found in quartz-veins. The garnets here are retrogressed to chlorite. The basal thrust contact to the Støren Nappe is located on the E6 just south of Kongsvoll Fjellstue, but is not exposed.

	15:38	Leave Kongsvoll. Return north on E6.
252.0	16:13	Back to Oppdal. **Dinner in Cafeteria (45 minutes)**
	16:58	Leave Oppdal toward Sunndalsøra.
273.2	17:19	Junction at Lønset.
278.5	17:26	Quarry on left. Drive carefully into quarry work area and park.

STOP 6-10: (15 minutes) Contact of 'upper' Åmotsdal Quartzite against 'upper' basement.
    A small exposure of steeply dipping basement gneiss occurs at the east end of the quarry work area, close to the main highway. The exact contact with the overlying quartzite is not exposed, but can be located within a few meters. Do not walk onto the highway to examine the small road cuts in quartzite, but instead walk west along the slope into the quarry, where washed natural surfaces as well as blasted surfaces are available for examination. The quartzite contains abundant delicate laminations, all strongly transposed. In past years, when the highway exposures were fresher, delicate cross-lamination could be observed. This is the same 'upper' quartzite horizon that, in inverted position, forms the famous 'hand fold' of Hansen (1971) high on the slope of Blåhø mountain, all tectonically above the Storli thrust. On this excursion we do not see the less deformed 'lower' quartzite and underlying 'lower' basement below the thrust.

| | 17:41 | Leave quarry. |
| 282.9 | 7:49 | Big turnout on left on steep hill, with high road cut on right. Watch for fast traffic. |

STOP 6-11: (10 minutes) Tight isoclinal fold in foliation of Risberget Augen Gneiss.
    This brief stop permits examination of ductilely deformed Risberget Augen Gneiss in a region of complex folds, where the Risberget overlies Åmotsdal Quartzite well above the Storli thrust.

	17:59	Leave Risberget road cut.
287.9	18:04	Pass Gjøra exit and travel down the relatively broad U-shaped Driva Valley. The Storli thrust is locally exposed in the high mountains visible to the right, including a location at elevation ≈1500m where upper gneiss rests above highly strained cobble conglomerate of the Åmotsdal Quartzite.
321.8	18:33	Junction. Sunndalsøra. Turn west and cross Driva River.
323.8	18:36	Junction. Turn left, south toward Litledal. Follow signs for Trædal Camping.
325.3	18:40	Trædal Vandrerhjem at the base of gigantic cliff and overnight.

Scenic STOP 6-12 near Sunndalsøra: After settlement in the Vandrerhjem, there will be a short drive north to the top of a cliff with a spectacular view over the fjord, involving an easy ten-minute walk from the parking place, with outcrops. Sunndalsøra lies on a significant boundary, East of the fjord there are large areas of relatively flat foliation characteristic of the nearby part of Trollheimen, composed mostly of basement rocks. On the west of the fjord, where we are standing, there are steep to shallow SE dips characteristic of the southwestern part of the Surnadal synform (Fig. 6.8) The narrow tail of the Surnadal synform, strongly overturned to the NW or recumbent, is exposed in mountains north from Sunndalsøra, and additional narrow synformal belts of nappe units have been traced partially across the fjords and mountains northwest from Sunndalsøra.

Day 7: GARNET PERIDOTITES IN THE NORTHERN UHP DOMAIN OF THE WGR, OTRØY

Thursday August 21

Summary of Route: From Trædal Vandrerhjem, Sunndalsøra, travel is back 1.5 km to junction with Route 62, then turning left (northwest) into a tunnel to Øksendal , then follow the west coast of Sunndalsfjord to Eidsøra. We follow Route 62 SW across a low pass to Eidsvåg, continuing to Tjelle, then onto a minor road along the entire NW coast of Langfjorden to the ferry terminal Sølsneset. We then turn N on Route 64 toward Grønnes Peninsula, over a long bridge to Bolsøy and into the Fannefjord undersea tunnel. Upon emerging from the tunnel, we turn west on Route E39 to the center of Molde, then slightly by-passing the city center and onto Route 662 so far as Mordalsvågen where we take a 15-minute ferry ride to Solholmen on Otrøy. From Solholmen, it is then a few minutes scenic drive on Route 668 along the north coast to Ugelvik and the garnet-peridotite bodies that are the main focus of the day. On leaving Ugelvik we travel to the center of Midsund and make a slight digression to the southwest corner of Otrøy. From Midsund we travel quickly across Midøy and to the ferry terminal at Dryna for the ferry to Brattvåg (20 minutes). From Brattvåg ferry terminal, we have 50 minutes in which to accomplish 10 minutes of driving to the ferry terminal at Skjeltene for the 15-minute trip to Lepsøy (Lauvsøy in some schedules) where we stay for three nights.

INTRODUCTION TO DAY 7

OCCURRENCE AND INTERPRETATION OF GARNET PERIDOTITES
IN THE NORTHERNMOST UHPM DOMAIN OF THE WGR, OTRØY

by Herman van Roermund

Occurrences of peridotite bodies of meter to kilometer-scale dimensions are a conspicuous feature of the Western Gneiss Region (WGR; Fig. 7.1). Chlorite-amphibole peridotites are the most common and only a few bodies contain the pyropic garnet–olivine assemblage that is diagnostic of the stabilisation of high pressure/ultrahigh pressure (HP/UHP) eclogite-facies assemblages. Nonetheless petrographic features, including the observed retrograde transition from granoblastic garnet-bearing peridotite into foliated chlorite-amphibole peridotite within certain bodies, suggest that at least some of the designated chlorite peridotites in Figure 7.1 may have originally contained pyropic garnets.

The garnetiferous peridotite bodies exposed on the islands of Otrøy, Flemsøy and Fjørtoft (locations indicated in box of Fig. 7.1), two of which will be visited during this field excursion, are especially noteworthy because they contain evidence for the early stability of M_1 megacrystic mineral assemblages that include high P/T enstatites and majoritic garnets (presently exsolved into M_2 and M_3 assemblages). The northernmost garnet peridotite bodies are geographically located within the northernmost UHPM domain of the WGR (Van Roermund, 2008).

The terminology of Scambelluri et al (2008) is used when reference is made to Archean, Proterozoic or Scandian tectonometamorphic events: M1 = Archean; M2 = Proterozoic; M3 = Scandian (Fig. 7.4). In addition M1A refers to a stage of adiabatic decompression at high T (1700-1800°C) from the mantle transition zone that resulted in the extraction of HPT decompression melts and left behind refractory peridotite while M1B refers to the accretion of this refractory peridotite to the continental lithosphere while within the garnet stability field (see Fig 7.4 and further discussion below).

Fig- 7.1 Simplified map showing location of the Western Gneiss Region in the Scandinavian Caledonides (inset), the distribution of large peridotite bodies in the WGR (open circles), peridotite bodies containing garnetiferous assemblages (filled circles) and Fe-Ti peridotites mentioned in this paper (triangles). Modified from Brueckner et al., 2008

Carswell et al (1983) drew attention to the existence of two fundamentally different chemical types of garnet-bearing peridotite bodies in the WGR. The Fe-Ti type garnet peridotites were interpreted to have had a prograde (U)HP metamorphic history that started as the ultramafic portions of layered low pressure crustal intrusive bodies were metamorphosed during Scandian subduction. The Eiksunddal eclogite complex on Hareidland (Fig. 7.1) with its subordinate garnet peridotite layers, well documented by Schmitt (1964; who was the first geologist on the Moon and did his PhD on this complex) and Jamtveit (1987), is a classical example of the Fe-Ti garnet peridotite type. Jamtveit et al (1991) were unable to obtain a meaningful Sm-Nd age for a garnet peridotite sample from the Eiksunddal complex, due to isotopic disequilibrium between the garnet, clinopyroxene and whole rock. However, they did obtain a Scandian metamorphic age of 412+/-12 Ma for coexisting garnet and clinopyroxene from a well-foliated eclogite sample from the same complex, in agreement with other Scandian Sm-Nd mineral ages from the WGR (Griffin & Brueckner, 1980).

Fe-Ti type garnet peridotites also occur in the northernmost UHPM domain of the WGR, examples are the Svartberget- and RaknesTangen bodies (Fig. 7.1). The diamond-bearing UHP mineral assemblage of the Svartberget body was dated by Vrijmoed et al (2006a, b; 2008) as Scandian (390-380 Ma; Sm-Nd grt-cpx). The younger Scandian ages are however interpreted as reset-ages due to a post-UHP HT amphibolite facies overprint

Fig. 7.2 Tectonometamorphic evolution of the mantle-derived, Mg-Cr type, garnet-bearing peridotite bodies in the northernmost UHPM domain, Western Gneiss Region, Norway. A) after Carswell (1986) and Medaris & Carswell (1990), B) after Van Roermund et al. (2000a; 2001), C) modified after Carswell & Van Roermund (2003), Carswell & Cuthbert (2003) and Spengler et al. (2006).

By contrast, the Mg-Cr type garnet peridotites display characteristic upper mantle mineral and whole rock chemistries (Carswell, 1968) and isotopic ratios (Brueckner, 1977) consistent with derivation from very highly depleted sub-continental lithospheric mantle (SCLM). Medaris (1984), Carswell (1986), Medaris & Carswell (1990) and Krogh and Carswell (1995) documented the complex, multi-stage, metamorphic evolution experienced by some of these particular garnet peridotite bodies. They concluded that the peridotites initially contained high-temperature pyroxenes that, on the basis of scarce petrographic evidence, coexisted with spinel rather than with garnet (stage 1 in Fig 7.2a) implying an initial low pressure protolith origin. They suggested the spinel peridotites were converted into garnet peridotites by a subsequent "quartz eclogite facies" tectonometamorphic event of unknown age (stage 2 in Fig 7.2a).

However, Sm-Nd dating of garnet+clinopyroxene+whole rock in samples of Mg-Cr type garnet peridotite and associated olivine-free garnet pyroxenites by Mearns (1986), Jamtveit et al (1991), Brueckner et al., (1996, 2002) and Lapen et al., (2005) has importantly shown that the earliest garnet-bearing assemblages (stage 2 in Fig. 7.2a,b) in these mantle-derived peridotite bodies were of (minimal?) mid-Proterozoic (M2) age and consequently were present as stable assemblages in the sub-continental lithospheric mantle (SCLM) long before the Scandian, subduction-related, (U)HPM event (M3; Fig 7.2C) experienced by the WGR.

A very surprising discovery has been the recognition by Van Roermund et al (1998, 2000a+b, 2001) of pyroxen exsolution lamellae in megacrysts of the earliest (= M1 and M2; Fig 7.2C and 7.3A) garnet generations in samples from the Mg-Cr type peridotite bodies outcropping on Otrøy. This microstructural feature has been taken to indicate that these garnets (M1) originally contained a significant majorite (i.e. high silica) component. Van Roermund et al (2000) deduced that the initial composition of this super-silicic majoritic garnet, had a majorite component of 5-8% indicating minimum pressures of 6-6.5 GPa at high temperature. They interpreted - (see also Drury et al, 2001) the peridotites were originating within a rising mantle diapir or alternatively an uprising part of an asthenospheric convection cell from depths of at least 185 km (Fig 7.4). More recently Spengler et al (2006) deduced initial compositions of this majoritic garnet of more than 20 % implying that these peridotite bodies originated from at least mantle transition zone depths (Fig. 7.4).

"Similar" garnet microstructures (Fig. 7.3B) were later described by Terry et al. (1999) from Fjørtoft and Flemsøy (Fig. 7.1). More importantly, exsolved pyroxene within M1 and M2 garnet (Fig. 7.2C) have now been reported from all major rock types within the garnet peridotite bodies, including garnet websterites and garnet clinopyroxenites (Spengler et al, 2002; 2006). Accordingly, the metamorphic evolution of the garnet peridotites in the northernmost UHPM domain (Fig. 7.1) must be modified to eliminate the proposed (Medaris (1984), Carswell (1986) and Medaris & Carswell (1990)) earliest high-temperature/low pressure assemblage (stage 1; Fig. 7.2A). The existence of aluminous spinel rather than garnet coexisting with the aluminous pyroxenes, which was the key to the earlier interpretation, was based on a single thin section of uncertain provenance (Carswell, 2002, personal communication) and is now believed to be incorrect. This important modification was considered in detail by Van Roermund et al (2000; 2001; Fig. 7.2B) and subsequently further elaborated by Carswell and Van Roermund (2003) and Carswell and Cuthbert (2003).

Fig. 7.3 Optical micrographs illustrating pyroxene exsolution from garnet. A) Garnetite; Otrøy garnet peridotite (M2). B) Garnet websterite; Bardane (M3). Note: scale-difference, both images were taken with the same magnification.

Fig. 7.4 Pressure-temperature (P-T) path of diamond- and majorite-bearing Mg-Cr type garnet peridotites in the northernmost UHPM domain of the WGR. Thick black path: Archean M1(A-B) to Middle Proterozoic M2 events recognized in peridotites. Light gray path—Scandian (M3) evolution. Dashed line—inferred P-T path of associated continental crust. White boxes—diamond crystallization conditions (dia), majorite samples from Sulu (S) and Tibetan Plateau (T). Solid thin black line—hot subduction geotherm (Peacock and Wang, 1999). Thin solid lines labeled 1%, 5%, 20%—experimental isopleths for majorite component in garnet. Dol—dolomite; opx—orthopyroxene; Mag—magnesite; cpx—clinopyroxene; Phl—phlogopite; richt—richterite.

M2 minerals present in megacrystic orthopyroxenite and garnetites (M1 in Fig. 7.2C) were dated by Brueckner et al (2002) and Spengler et al (2006). Brueckner et al (2002) also reported Archean model Re-Os ages as well as an Archean Re-Os "errorchron" age from sulfide grains included within megacrystic (M1b) orthopyroxene. During subsequent cooling this M1b megacrystic orthopyroxenite is interpreted to have transformed into a coarse grained M2 garnet bearing orthopyroxenite caused by exsolution of M2 garnet and M2 clinopyroxene out of Opx (M1b), that apparently grew together with spinel (M2; $Cr^{\#}$=75). Similar, but smaller, high Cr spinels are also included in M1b Opx, giving rise to the M2 assemblage defined by Opx2, Cpx2, Grt2, Ol2 and Sp2 and dated by the Sm-Nd technique as being minimal of mid Proterozoic age (1651 ± 47 Ma, Brueckner et al., 2002; 1405 ± 13 Ma, Spengler et al., 2006). Estimated PT conditions, using recalculated M2 Opx mineral compositions, during M2 are: 1300-1500°C and 3-4.5 GPa (Carswell, 1973; Van Roermund and Drury, 1998; Terry et al. 1999; Van Roermund et al., 2002; Carswell and Van Roermund 2005).

Spengler (2006) and Spengler et al (2006) used LREE partitioning between Grt2 and exsolved Cpx2 in garnetite from the Otrøy garnet peridotites (Fig. 7.3A) to determine that Grt 2, Cpx2, Opx2 and Ol2 were formed at T ≥ 1300°C, implying a pressure of 3.5-4.0 GPa for the formation of the M1b garnet that contained 1% majorite (Fig. 7.4; in which Grt $_{M1B}$ → Grt $_{M2}$ + Cpx $_{M2}$+Opx $_{M2}$). Spengler et al (2006) also used Sm-Nd techniques to date the M2 mineral assemblage in the same garnetite as being (minimal) mid-Proterozoic (1405 ± 13 Ma). Interestingly the garnetites had initial $^{143}Nd/^{144}Nd$ ratios around 0.54756±0.00013, indicating Archean model ages for the M1a garnet (T_{chur}= 2.53 and 2.90 Ga). These model ages clearly indicate that the original, highly depleted, garnet peridotites were formed in the Archaean.

Archean melting of the garnet peridotites is also supported by whole-rock Re–Os model ages from sulfide free rocks from Almklovdalen (2.7 - 3.1 Gyr BP) (Beyer et al., 2004;2006) and Otrøy (Spengler 2006). The Almkovdalen peridotite shows no evidence of garnet formation during the Archean whereas the Otrøy peridotites do. The Otrøy peridotites therefore appear to be the first reported case of orogenic peridotites produced by extensive melt depletion while within the garnet stability field during the Archaean era. The most likely explanation for the origin of the northernmost garnet peridotites at Otrøy, Flemsøy and Fjørtoft is therefore that they were part of a rising mantle diapir and/or an upwards moving asthenospheric convection-cell that crossed the dry peridotite solidus resulting in HPT decompression melting (see also Drury et al 2001). Melting was followed by accretion of the refractory peridotite to the SCLM at depths of the order of 125-130 km where the peridotite cooled down to the local Archean geotherm (Fig. 7.4; M1a → M1b). This interpretation is consistent with a drop in the majorite content in M1 garnet from 20% in M1a garnet to 1% in M1b garnet (Fig 7.4; Spengler et al., 2006; Scambelluri et al., 2008). At some stage during adiabatic decompression, HPT melting and accretion there may have been a stage of refertilisation based on a five point WR Sm-Nd errorchron of 3333±190 Ma ($^{143}Nd/^{144}Nd$ $_{(I)}$ = 0.50851; MSWD=55) for different types of internal garnet pyroxenites from Otrøy and Flemsøy (Spengler, 2006; Spengler et al., 2008). It is possible that the megacrystic orthopyroxenite (M1b) formed during this stage (=M1b) rather than being residue left after melt extraction.

Full details of the Proterozoic history and location of the peridotite bodies within the SCLM remain somewhat conjectural given the scarcity of preservation of M1/M2 assemblages due to widespread overprint by Caledonian (M3) assemblages (see below). It seems likely that M1b garnet-bearing assemblages, containing 1% majorite, are mostly re-crystallised and re-equilibrated into exsolved M2 assemblages, including a normal Cr-pyrope garnet (Fig.7.2C; Fig 7.3A), as these rocks were further cooled (and uplifted?) to ambient Proterozoic geotherms (M2).

A revised, multi-stage metamorphic evolution for the northernmost Mg-Cr type garnet peridotites, that incorporates large scale, Archean, adiabatic decompression, illustrated by the transformation of a 20% (M1a) to 1% (M1b) majoritic garnet, is presented in Figure 7.2C. Archean adiabatic decompression (M1A) is concomittent with HPT melting (decompression melting) when the decompression path of the Otrøy peridotites, at high T, intersects the dry peridotite solidus (Fig. 7.4). Subsequent accretion of the refractory peridotite (M1b) to the SCLM occurs within the garnet stability field such that garnet can still incorporate 1% majorite in its crystal lattice (Fig 7.4; M1b). Needle-shaped pyroxene exsolution from a supersaturated 1% majoritic garnet (M1b; Fig. 7.4) is interpreted to have occurred in mid Proterozoic times (M2) when the accreted and depleted garnet peridotite bodies cooled down from higher Archean (M1b) to lower Proterozoic (M2) geotherms (Fig. 7.2C). A PT path consistent with such an "early" Archean to Proterozoic evolution is illustrated in Fig. 7.4 (after Van Roermund et al., 2000a and b; Spengler et al., 2006 and Scambelluri et al., 2008).
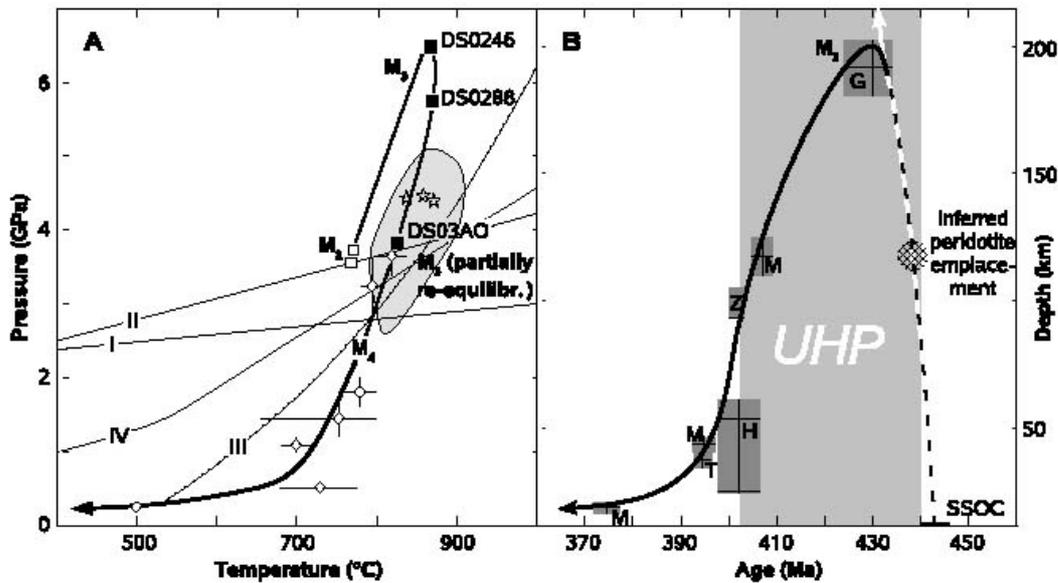
Fig. 7.5. Pressure (P), Temperature (T) and time (t) evolution of garnet-peridotite and other UHP metamorphic rocks in the WGR.

A) P—T diagram showing calculated estimates on Otrøy mantle rocks using one internally consistent set of thermobarometric calibrations: open squares, megacryst mineral assemblages (M2) containing pyroxene lamellae in garnet; filled squares, recrystallized mineral assemblages (M3) lacking pyroxene lamellae in garnet. M4 illustrates the conditions possible for the formation of olivine-ilmenite symplectite after M3 clinohumite (Spengler et al., 2002). Shaded field delineates 'best estimates' for UHP metamorphism of the Otrøy and Fjørtoft gneiss complex determined from eclogites and websterites using combinations of different thermobarometric calibration. Stars, PT conditions of strongly foliated eclogite from the Nordfjord area. The deduced path for continental plate subduction and exhumation (solid thick arrow) shows the formation of the M3 assemblage is prograde. The early retrograde evolution forms a narrow bent hairpin path within the diamond and Ti-clinohumite stability fields and was accompanied with partial mineral re-equilibration of M3 assemblages. M4 mantle minerals and basement rock data (open diamonds) indicate further retrogression within the graphite stability field. Numbers refer to sample numbers measured for P and T. Solid curves indicate the position of phase changes (LP/HP): quartz/coesite (I), graphite/diamond (II), olivine+ilmenite+rutile+H20/Ti-clinohumite(III) and the cratonic geotherm of Baltica (IV).

B) Depth–t diagram showing geochronological data from UHP metamorphic mantle and crustal rocks and the preferred subduction and exhumation paths for a full orogenic cycle. Black arrow and white arrow show the depth–t evolution of the continental lithosphere and oceanic lithosphere, respectively. Light grey shaded field illustrates the time-period, when the western edge of the WGR was buried to UHP metamorphic conditions. G, garnet-pyroxenite from Otrøy and Flemsøy (Sm-Nd on M3); M, Fjørtoft garnet-kyanite-gneiss (U-Th-Pb mean ages on monazite); Z, Hareidlandet coesite-eclogite (U-Pb on zircon) with minimum P estimate; T, WGR orthogneiss (U-Pb on titanite+zircon); H, Sandsøya garnet-kyanite-gneiss (Ar-Ar on hornblende). SSOC, Solund-Stavfjord ophiolite complex (U-Pb on zircon).

New discoveries with important consequences for the P-T-t history experienced by these peridotites were the subsequent discoveries of Scandian-aged microdiamonds and/or M3 majoritic garnets within megacrystic garnet websterite samples (M2) enclosed in the Bardane garnet peridotite on Fjørtoft (Van Roermund et al, 2002; Scambelluri et al., 2008) as well as new microdiamond occurrences within the Fe-Ti garnet peridotite/garnet websterite at Svartberget (Vrijmoed et al., 2006; 2008). Rather surprisingly the microdiamond did not form in association with the M1 / M2 UHP majoritic garnet (Fig 7.2C and 7.4) during the Archean to mid-Proterozoic evolution. Instead combined petrographic, geochemical and isotopic data (Brueckner et al, 2002; Van Roermund et al, 2002; Carswell and Van Roermund, 2005; Scambelluri et al., 2008) provided strong evidence that the microdiamonds formed together with a new generation of M3 majoritic garnet, in response to an influx of subduction-related, supercritical, crustal-derived, high-density, COH-rich fluids during Scandian-aged deformation and recrystallisation of the M1 Archean and M2 Proterozoic megacrystic assemblages. In addition M3 pyroxene lamellae exsolved from M3 garnet have one order of magnitude smaller pyroxene lamellae sizes than M1/M2 (Fig. 7.3A, B).

However, the dissolution of 1 vol % pyroxene in M3 majoritic garnet, increases the Scandian metamorphic conditions from 840-900°C/ 3-4.5 GPa, as reported previously, up to 900-1000°C/5.5-6.5 GPa (Fig. 7.4). Similar PT conditions are derived by Vrijmoed et al., (2006; 2008) and Spengler et al. (2008; Fig. 7.5a), who both used Al isopleths in Opx adjacent to garnet from internal and external garnet websterites in combination with classical geothermometers. This large pressure increase indicates the WGR was subducted during the Scandian much more deeply into the mantle than previously believed (≥180 km; Scambelluri et al. 2008; Van Roermund, 2008).

In conclusion: a surprising new interpretation has emerged for the northernmost, mantle-derived, Mg-Cr type, garnet-bearing peridotites (and the contrasting Fe-Ti type) of the WGR. Evidence from several sources indicate that after a prolonged Archean-Proterozoic history they experienced Scandian, subduction-related, UHP metamorphism (M3a in Fig 7.2c), which was concomittent with the infiltration of a crustal-derived, supercritical, COH-rich, dense subduction zone fluid that also locally impregnated the peridotites with free carbon that resulted in micro-diamond formation. Moreover the garnet peridotites on Fjørtoft (Bardane) preserve evidence of a third (M3) majoritic garnet microstructure (M3a; Fig 7.2C) clearly demonstrating that this part of the WGR has been subducted to depths of the order of 180-200 km (Fig 7.4). Similar Scandian PT conditions, using more classical/conventional geothermobarometric techniques, were derived for the Otrøy Mg-Cr type garnet peridotites (as well as for some surrounding country rock eclogites) and the crustal Fe-Ti type peridotite/websterite at Svartberget (Carswell et al., 2006; Vrijmoed et al., 2006).

The long time-interval between Proterozoic stage 2 and Caledonian stage 3 (Fig 7.2A and B) or M2 and M3a (Fig 7.2C) creates uncertainty over exactly where these mantle-derived peridotite bodies resided. They conceivably may have already been emplaced into the Proterozoic continental crust. However, it is considered more likely that they resided in the uppermost sub-continental lithospheric mantle, with Laurentian affinity, and were not incorporated into the continental crust until deep subduction of the WGR underneath the Laurentian plate during the Caledonian/Scandian plate collision. According to Spengler et al., (2008), the Otrøy, Fjørtoft and Flemsøy peridotites resided in the mantle at a depth of 120-130 km (Fig. 7.5B), until around 440 Ma ago, when subduction of the continental crust started. They were subsequently transfered from the hanging-wall mantle into the subducting continental crust and carried by the subducting crust to depths of 180-200 km and metamorphosed under UHPM conditions around 430 Ma ago, based on the weighted Sm-Nd age of 430 Ma for three samples containing M3 grt-cpx assemblages (Fig 7.5B). The peridotites and their host crustal rocks were back at subcrustal levels around 400-390 Ma, giving a time span for the full Scandian subduction cycle of 40-50 Ma. If true this extended time slot puts severe constraints on the operating geodynamic paradigm that subduction and exhumation of continental crust occurs over relatively short time periods of ca. 20 Ma. Such a new interpretation is also consistent with experimental, theoretical and observational facts that crustal rocks subducted deeply into the mantle at T ≈ 900-1000°C should melt. Within the most northern UHPM domain of the WGR, Norway, evidence for Scandian decompression melting can easily be recognised in the country rock gneisses. This is in strong contrast to observations done in more southerly located UHPM domains of the WGR.

ITINERARY FOR DAY 7

8:00 Leave Trædal Vandrerhjem (with all baggage), Sunndalsøra. From Vandrerhjem, travel is back 1.5 km to junction with Route 62, then turning left (northwest) toward tunnels to Øksendal. Just before the first tunnel there is an opportunity to turn right onto the fjord shore for a short photostop, **8:06-8:11** before entering tunnels.

At Øksendal we are mostly out in the open again and following the west coast of Sunndalsfjord to Eidsøra. Scrutiny of opposite fjord walls beyond Øksendal shows some of the steep narrow belts of cover nappes. As we approach Eidsøra, an enormous cliff on the north side of the fjord exposes a huge anticlinal structure in the foliation. Beyond Eidsøra we follow Route 62 SW across a low pass to Eidsvåg, continuing to Tjelle. Near Eidsvåg there is a view south into Eresfjorden and the childhood home of Bjørnstjerne Bjørnson, 1832-1910, a Norwegian patriot and author (considered one of the four greatest) who composed the words of the Norwegian National Anthem.

8:52-8:57 Just after leaving Route 62 at Tjelle, onto a minor road that goes along the entire NW coast of Lanfjorden, we have a short stop at Tjellefonna landslide/tsunami site, then continue to the ferry terminal at Sølsneset. All the way from Eidsøra to Sølneset we have travelled mostly parallel to a steep foliation in basement gneiss, and outcrops show the characteristic subhorizontal extensional lineation.

At Sølneset we turn N across the strike of basement foliation on Route 64, eventually curving west toward the tip of Grønnes Peninsula.

9:29-9:39 We have a short stop at a public park with toilet at the east end of the long high bridge to Bolsøy, There are outcrops of typical strongly laminated, highly lineated gray basement gneiss nearby, and there are relict eclogites not far away. Basement rocks continue a short distance onto Bolsøy, then we cross diagonally through the south limb of the Moldefjord synform with a northwest-overturned axial surface, successively crossing a 600 m-thick section Risberget, Sætra, Blåhø to Støren fine-grained amphibolites and metamorphosed diorites in the core of the fold. Robinson had the privilege to map these road exposures when construction was in progress in 1990, and Bolsøy could only be reached by ferry.

We are in Støren diorites as we descend into the Fannefjord undersea tunnel, and before reaching the lowest point of the tunnel we cross a much thinner north-dipping north-limb section, Blåhø, Sætra, Risberget and into basement. In the bottom of the tunnel the basement is cut by a late brittle fault. But the same basement rocks continue in the northward ascent to the mouth of the tunnel.

9:51 At junction where road from Fannefjord Tunnel intersects E39 east of Molde, we turn west on Route E39 to the center of Molde, the City of Roses. In spring 1940 King Håkon VII stood under a tree while the city was heavily bombed with the main objective of killing him. The tree became a historic symbol of national resistance. It died a few years ago but has been replaced. We slightly by-pass the city center, coming out on the coast again near the football stadium, and onto Route 662 so far as Mordalsvågen. En route to Mordalsvågen we pass mostly over gneisses, but all of the rocks beside the road leading down to the ferry wharf, including those beside the ferry waiting line, belong to a narrow synformal belt of Blåhø garnet-biotite schists, amphibolites and one thin marble, that continues west through Otrøy.

10:16 Arrival at Mordalsvågen ferry terminal, depending on traffic in Molde. Ferry departs **10:30**. During the crossing there can be spectacular views southward to the Romsdal Mountains, southwest to Reksdalshesten, west to Otrøy, and north up the deep marine channel of Julesundet. Several German fortifications left over from World War II are also prominent in the view.

10:45 We reach Solholmen on Otrøy, where we meet Herman Van Roermund. We then follow Route 668 along the northern side of Otrøy until reaching the village Ugelvik (±15 km), and park vehicles in the church car park on the left side of the road.

INTRODUCTION TO THE GEOLOGY OF OTRØY AND MIDØY

A general introduction on the occurrence and interpretation of garnet peridotites bodies in the northernmost UHPM domain of the WGR has been presented as an introduction to this day. A generalised geological map of the western part of Otrøy and the eastern part of Midøy is illustrated in Fig 7.6A (after Van Roermund 2008). The Proterozoic Baltican basement rocks on northern OtrØy form E-W to NNE-SSW trending belts that consist of interlayered augen orthogneiss (commonly migmatitic) and well layered (migmatitic) dioritic-granodioritic gneiss with abundant eclogites and subordinate garnet peridotites. The dominant amphibolite facies foliation is subvertical, commonly with a well developed subhorizontal E-W amphibolite-facies lineation (Fig 7.6B,C).

In contrast, in the southern part of both islands, Robinson (1995, 1997) recognized some infolded allochthonous supracrustal units now mapped as Blåho and Saetra Nappe equivalents (Fig 7.6A; Robinson, 1995; 1997; Hollocher et al., 2007). More important the allochthonous supracrustal units in the south very rarely contain eclogites in the studied area (Robinson 1995; Van Roermund et al., 2005; Hollocher et al., 2007; Van Roermund 2008). Thus, a fundamental EW-running Scandian tectonic contact divides both islands into northern (U)HPM basement units versus southern basement and allochthonous medium-pressure units (Fig 7.6A).

Three mantle fragments occur on Otrøy within the northern basement unit, the so-called Ugelvik (U), Raudhaugene (R) and Midsundvattnet (M) garnet peridotite bodies (Fig 7.6A), for which detailed geological maps are given in Fig 7.7 and Fig.7.8. Previous studies concluded that incorporation of the garnet peridotites into the Caledonian basement rocks occured during the Scandian Orogeny (430-400 Ma) under lithostatic pressures equivalent to uppermost β-quartz eclogite-facies conditions (Griffin et al., 1985; Carswell et al., 1986; Krogh & Carswell, 1995). More recent studies have indicated crustal emplacement of the garnet peridotite bodies occured under much higher, UHPM, conditions, deep within the diamond stability field (Van Roermund et al., 2005; Spengler 2006; Spengler et al., 2006; Vrijmoed et al., 2006; Scambelluri et al., 2007; Van Roermund 2008).
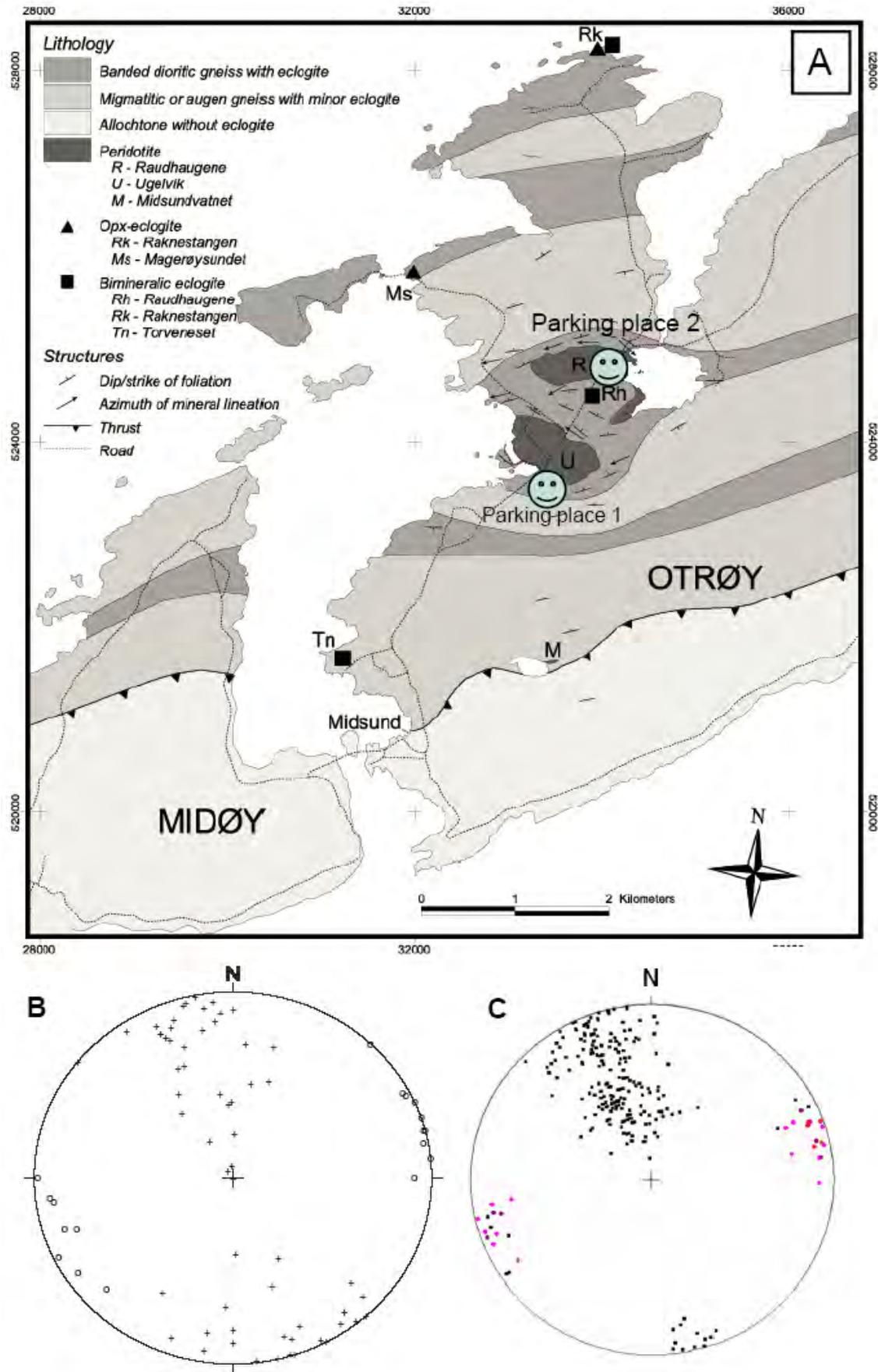
Fig. 7.6 A) Geological map of the western part of Otrøy and eastern part of Midøy (after Van Roermund 2008). B) Structural data (foliation and lineation) of northern domain of Otrøy and Midøy. C) Structural data (foliation and lineation) of southern domain of Otrøy and Midøy.
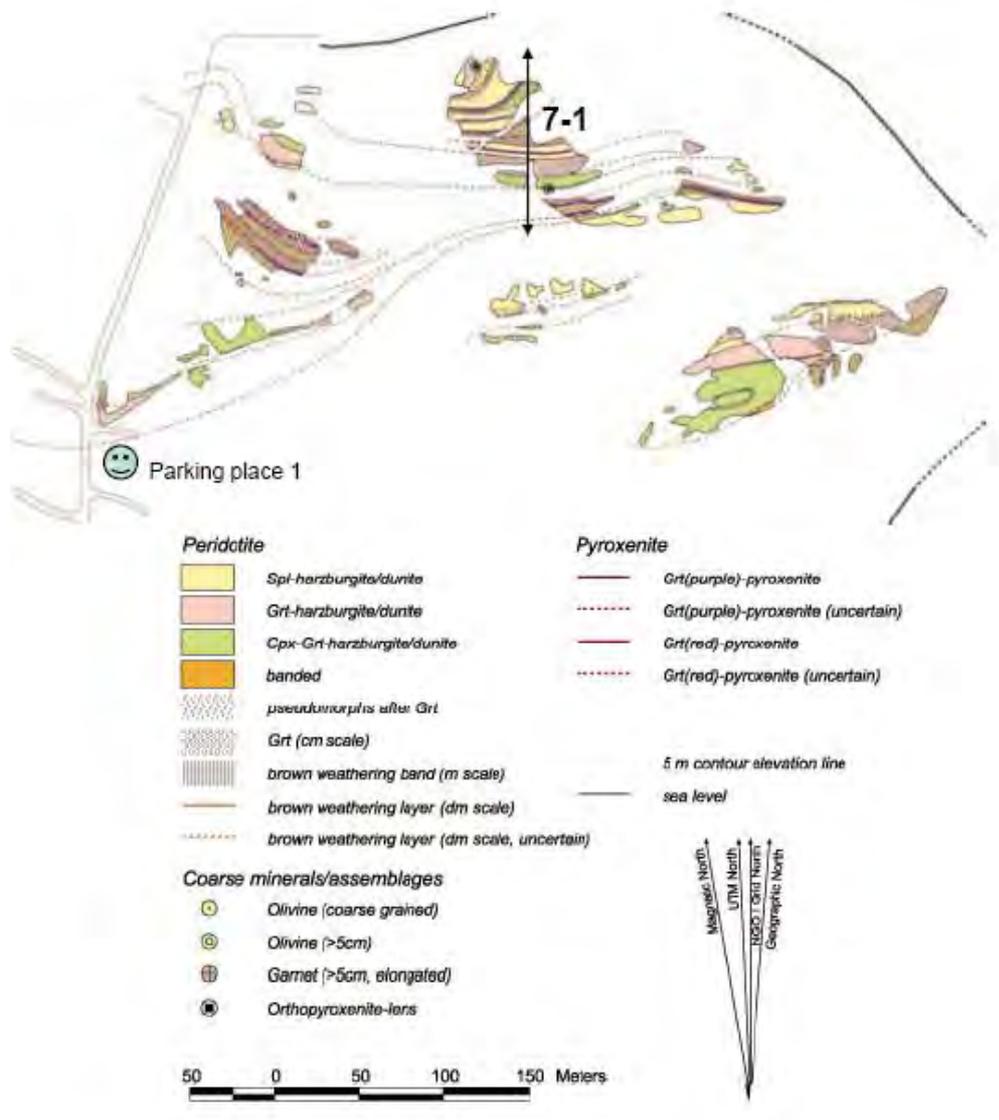
Fig. 7.7 Geological map of the eastern part of the Ugelvik peridotite body.

The three garnet peridotite bodies can be classified as *relict peridotites* according to the nomenclature of Brueckner & Medaris (2000) i.e. are unique in that not only Scandian mineral ages and crustal emplacement conditions can be reconstructed, but, in addition, the Archean to mid/late Proterozoic mantle evolution of the peridotite bodies is still serpentinisation, crustal emplacement at UHPM conditions did not erase (completely) the Archean to mid/late Proterozoic mantle signature of these bodies. Therefore, during this one-day excursion we will concentrate on the following main aspects of the garnet perdotites:

> 1) Archean HPT decompression melting.
> 2) Megacrystic garnet (garnetites), olivine and (garnet) websterite.
> 3) Majoritic garnet(s) and their implications.
> 4) "Cooling" during the Proterozoic.
> 5) Crustal emplacement during "early" Scandian UHPM.

Other aspects that may be studied, though regarded here as being of minor importance during this excursion, are:

> 6) Two garnet pyroxenite types
> 7) Black versus pink garnets
> 8) Unusual fold geometry of garnet pyroxenite embedded within peridotite.

Fig. 7.8 A) Geological map of the western part of the Ugelvik peridotite body (for colour code: see Fig 7.7).
B) Geological map of the central part of the Raudhaugene peridotite body (for colour code: see Fig 7.7)

The terminology of Scambelluri et al (2008) is used here when reference is made to Archean, Proterozoic or Scandian tectonometamorphic events: M1 = Archean; M2 = Proterozoic; M3 = Scandian (Fig.7.4). In addition M1A refers to a

stage of adiabatic decompression at high T (1700-1800°C) starting at mantle transition zone depths that resulted in the extraction of HPT decompression melts and leaving behind refractory peridotite, M1B refers to the accretion of this refractory peridotite to the continental lithosphere at depths still within the garnet peridotite stability field (Fig 7.4).

Fig. 7.9 Generalised phase diagram for fertile peridotitic mantle, based on experimental data for dry peridotite melting (Takahashi 1990; Walter 2003).

LOCALITIES: FOR STOPS 7-1 to 7-13 (see Figs 7.7A, B, C)

Stop 7-1 Ugelvik East. This locality will be visited to demonstrate and discuss the origin of the compositional layering that is so clearly present within the U-, R- and M garnet peridotites (Fig 7.6A). This compositional layering is primarily defined by alternating layers of (spinel) dunites/harzburgites versus garnet dunites/harzburgites (Fig 7.7), locally supported by discrete garnet pyroxenite layers. The clear absence of (more) fertile peridotite is strong evidence that the peridotite has undergone one (or more) period(s) of partial melting. To melt a peridotite T $\geq$ 1120 °C is required, but the melting temperature (T_{melt}) a.o. depends strongly on pressure, degree of melting and availability of H_2O (Fig 7.9). At Otrøy (and surroundings) no evidence was found for the availability of H_2O during peridotite melting, so we are dealing here with the dry peridotite solidus (Fig 7.4 and 7.9). The most important questions concerning peridotite melting are:
  1) What is/are the age(s) of the partial melting event(s)?
  2) What are the physical conditions, in terms of pressure (P) and temperature (T), during (partial) melting? More important did partial melting occur at low ($\leq$ 15 kb) or high pressure ($\geq$30 kb) conditions?

Re-Os bulk rock (WR) dating has demonstrated that peridotite melting took place in the Archean (Spengler 2006; Spengler et al., 2006; Fig 7.10A), in agreement with other Re-Os age dating-work applied to WR or single sulfide grains in peridotites exposed elsewhere in the WGR (Beyer et al., 2004, 2006; Brueckner et al., 2002, 2008).

A generalized phase diagram for fertile peridotite mantle, based on experimental data for dry melting (Takahashi, 1990; Walter, 2003), is illustrated in Fig 7.9. It can be visualized from this diagram that an uprising adiabatic mantle plume (or any other kind of upwelling mantle convection cell) at T = 1600-1750°C will cross the dry peridotite solidus between 4.5-7 GPa. Consequently the refractory peridotite body may contain garnet if it cools down rapidly: before lithostatic pressure (P) becomes as low as 3 GPa. (i.e. the PT path will not cross the red garnet-out phase boundary line in Fig 7.9). Alternatively garnet-free mineral assemblages will be formed in the refractory peridotite if similar decompression paths, as described above, will cross the garnet-out phase boundary line of Fig 7.9. In other words different degrees of HP decompression melting may thus explain the observed compositional layering observed in the Otrøy peridotites (Spengler, 2006; Spengler et al., 2006; Scambelluri et al., 2007). Alternatively a stage of low-pressure

Archean melting will require a UHPM recycling period in Archean and/or mid Proterozoic times in order to explain the mid-Proterozoic Sm-Nd garnet-cpx-opx-wr ages of 1.4 Ga determined for M2 cpx-opx-grt assemblages within garnetite (Spengler et al., 2006, 2008; Fig 7.10B).

In addition, post-Scandian decompression can easily be recognized in the peridotite body using the various mineral assemblages that define (M3b, c and d) replacement-coronas around garnet. Note also that two differently coloured garnets are present within the outcrop: purple-red versus black. In some cases it can be recognized that black garnet rims surround purple-red garnet cores.

Stop 7-2: Ugelvik west Harbour Quarry. This locality (Fig. 7.8A) displays NE-dipping, layered and foliated garnet peridotite. The layering is defined both by the variable garnet content and by discrete garnet pyroxenite layers that are mostly >1cm thick. However, as well seen on the top surfaces of the outcrops on the southern side of this small quarry, some garnet pyroxenite layers are sheared out to only mm thickness, sometimes literally to only single crystal thickness. These surface garnet pyroxenite smears display an obvious mineral orientation lineation that plunges 30-40° to the N. (an orientation clearly different from the late-Scandian amphibolite facies mineral lineation present in the surrounding gneisses (Fig 7.6B,C).
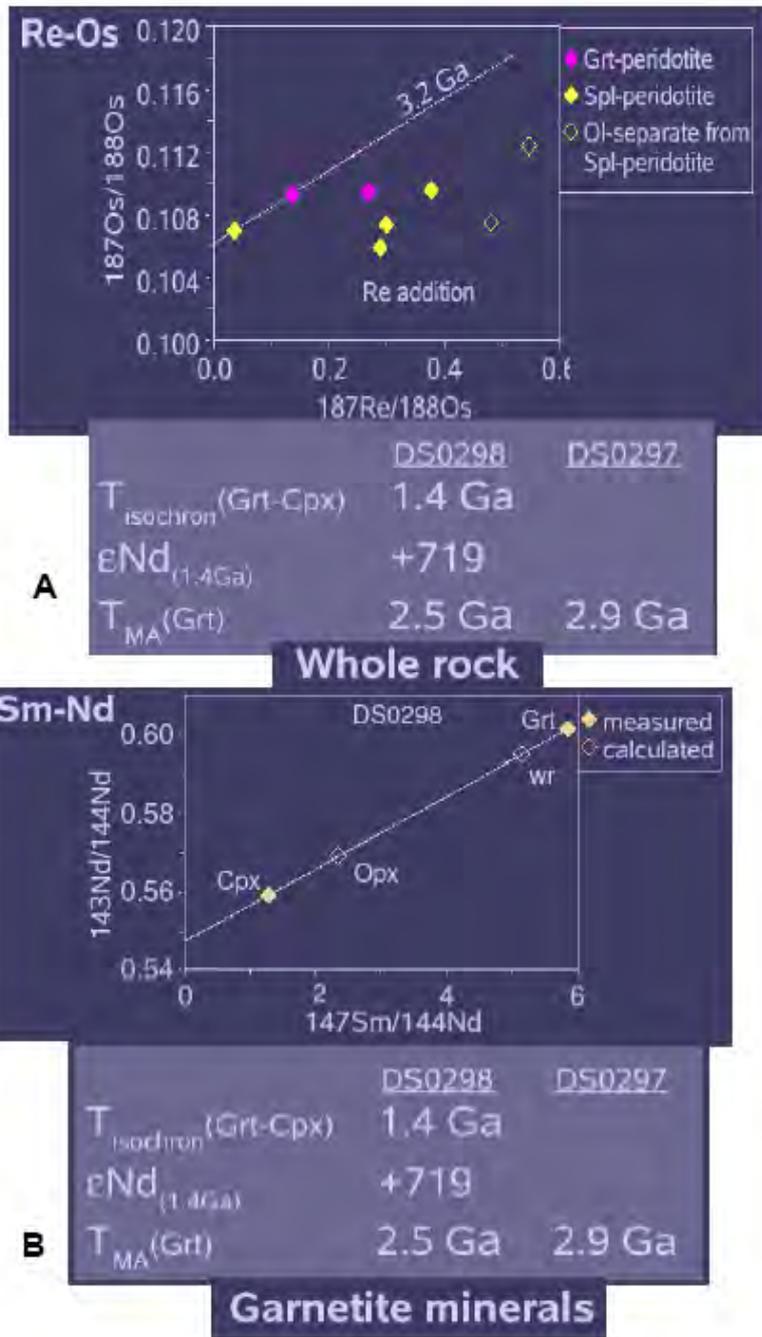

Fig. 7.10 Re-Os and Sm-Nd WR- , isochron and modal ages (after Spengler 2006).

Scattered larger purple red, Cr-pyrope, garnets in the peridotite here are up to 2 cm in size and heavily fractured. They not uncommonly contain inclusions of bright green Cr-diopside, less commonly also of orthopyroxene and olivine. Garnets commonly show the initiation of a replacement corona of kelyphite (a fine-grained intergrowth of (M3b) pyroxenes and (M3b) Cr-spinel). Later-stage (M3c and d) dark green amphibole and pale purple-grey (M3c and d) Cr chlorite are conspicuous on a set of S-dipping fracture surfaces, whereas a later set of near-vertical joint surfaces is lined with (M3e) serpentine.

Garnet peridotite in this quarry passes to the NE into garnet-free dunite, as can be seen in the blocks up above the quarry close to the outcrop of megacrystic garnet websterite: a spectacular exsolved garnet pyroxenite lens within the garnet peridotite that was first reported by Carswell (1973). *(Please note that field excursion participants will be*

*escorted in small groups to see and photograph **but not sample** this lens, during which time other participants should keep well away from the back face of the quarry for fear of falling dislodged rocks).* It is considered that this remarkably fresh megacrystic M1A garnet pyroxenite/websterite forms a deformation-sculptured, lensoid pod of roughly 90 x 60 x 30 cm dimensions with its long axis plunging ca. 30°N and its short axis perpendicular to the NE-dipping foliation in the adjacent peridotite (Fig 7.11C).

Carswell (1973) recognised this garnet pyroxenite pod to comprise mostly deformed M3 sub-grains of original, high-temperature, megacrystic M2 orthopyroxene that has extensively exsolved both M3 clinopyroxene and M3 garnet (Fig 7.11D). Whilst much of the exsolved M3 clinopyroxene forms rational lamellae //(100) in the host orthopyroxene, most of the exsolved M3 garnet occupies a vein network along the M3 orthopyroxene sub-grain boundaries.

Fig. 7.11 Examples of megacrysts in the Otroy garnet peridotites. A) Garnetite megacryst exposed at stop 7-8. B) Olivine megacryst exposed at stop 7-7b. C) Garnet websterite lens exposed at stop 7-2. D) Detail of garnet websterite lens (7-6C) showing garnet and cpx exsolution from kinked Opx megacryst.

Carswell (1973) considered that some scarce larger M1a/M2 garnet grains (up to 5mm size) within the pod might not be of exsolution origin and surmised that the original M1a orthopyroxene (with ca. 4 wt.% Al_2O_3 and 2 wt.% CaO) may have crystallised on the peridotite solidus together with primary garnet and olivine at ca. 1500-1600 °C and 3.5-4.0 GPa. More recent samples extracted from this pod have confirmed that the coarse deformed and exsolved orthopyroxenite part grades within the confines of this lensoid pod into a garnet peridotite with increasing content of olivine that contains M1a/M2 garnet grains of up to 1cm size.

Van Roermund and Drury (1998) and Van Roermund et al (2000, 2001a and b) have recognised that megacrysts of the earliest garnet (M1; Fig.7.11A) in the Ugelvik peridotite body show evidence of pyroxene exsolution and accordingly had an original majoritic composition. They deduced that this majoritic (M1a-M1b; Fig 7.4) garnet (within peridotite) may have contained as much as 3.6-4.0 vol.% pyroxene and thus may have formed at pressures of 6.0-6.5 GPa (M1a; Fig 7.4). However, such pressures and such garnet compositions are hard to reconcile with coexistence with an orthopyroxene (M1b; Fig 7.4) with ca. 4 wt.% Al_2O_3 and 2 wt.% CaO. Given the uncertainties that should be attached to the composition estimates for a primary M1b garnet and M1b orthopyroxene, depending on whether or not all inter-crystalline grains are aggregated together with intra-crystalline exsolution products in determining the original compositions, a possible reconciliation could be the original coexistence of a M1b majoritic garnet with ca.1% pyroxene component with M1b orthopyroxene that had about 3-4 wt.% Al_2O_3 and 2 wt.% CaO at conditions of ca. 1350-1500°C and 3.5-4.5 GPa (Van Roermund et al., 2002; Carswell and Van Roermund, 2005). Cooling from M1b to M2 (Fig. 7.4) would have caused exsolution of intracrystalline pyroxene laths from a 1% majoritic M1b garnet. Majoritic garnets containing more than 1 vol% of pyroxene dissolved in their lattice will accordingly be called M1a.

By analogy with Archean Sm-Nd modal ages obtained for the "earliest" M1 garnet-bearing clast assemblages (Spengler et al., 2006; Fig 7.10B) and Re-Os sulphide ages from single sulphide grains included in M1b Opx megacrysts (Brueckner et al., 2002), it is considered likely that the early megacryst assemblages in the Ugelvik peridotite body formed within an Archean mantle diapir (or any other kind of asthenospheric upwelling; M1a in Fig
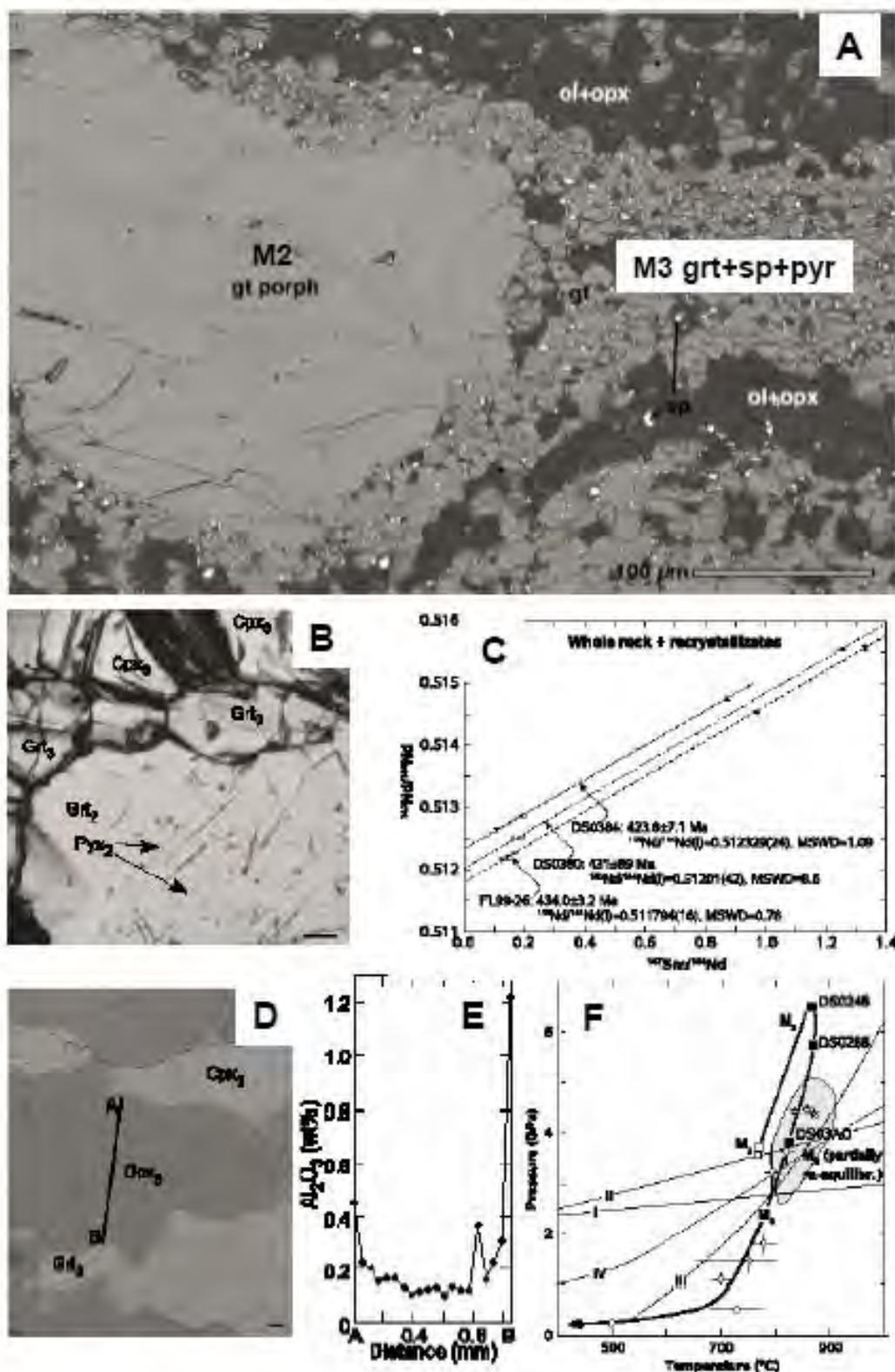
Fig. 7.12 Crustal emplacement history of Otrøy garnet peridotites. A) Strain-induced recrystallisation of M2 grt porphyroblast into an M3 mineral assemblage related to crustal emplacement of the garnet peridotites. B) Detail of recrystallised M3 mineral assemblage (grt3, cpx3 and opx3) surrounding M2 garnet clast containing exsolved pyroxene. C) Sm-Nd age dating results illustrating "isochron" ages of 423, 431 and 434 Ma for recrystallised M3 assemblages (after Spengler 2006). D) Backscattered electron image of M3 mineral assemblage. AB= traject used for microprobe analyses in a M3 opx, grt, cpx assemblage. E) Al_2O_3 wt% analyses of opx3 illustrated in Fig 7.12D (after Spengler 2006 and Spengler et al 2008). F) Inferred (pro- and retrograde) Scandian PT path for Otroy garnet peridotites using all Al_2O_3 wt % analyses in opx in combination with other geothermobarometric techniques (after Spengler 2006 and Spengler et al 2008).

7.8) and/or during accretion of the refractory peridotite to the cratonic lithosphere (M1b in Fig 7.4). Given the Archean ages for HP decompression melting (Fig. 7.10A) and the Archean ages of megacryst formation (Fig. 7.10B) it is thought likely that both are formed during the same HP decompression melting period. In contrast the exsolved and recrystallised mineral assemblages within the M1 pods formed either during continuous cooling till mid Proterozoic times (M2 conditions; Figs. 7.4 and 7.10B) and/or during extensive Caledonian tectonic reworking at PT conditions of ca. 750-850°C and 3.5-4.0 GPa (Jamtveit et al, 1991, Brueckner et al, 2002), recently upgraded to 850-950 °C and 6-7 GPa (Van Roermund et al., 2005; Carswell et al., 2006; Scambelluri et al., 2007; Spengler et al., 2008; Van Roermund, 2008).

Stop 7-3: Ugelvik Harbour, Western Shoreline- opposite pleasure boat jetty. Here at the rope anchorage (Fig. 7.8A), there is a ca. 4.5m length exposure of garnet websterite folded within garnet peridotite. Tight isoclinal folds in the adjacent peridotite plunge ca 10° to 340°. This garnet websterite, in common with that forming rather more conspicuous layers within the Raudhaugene peridotite body, has a distinctly porphyroclastic texture (Fig 7.12A). Garnet clasts of up to 3-5 mm size, may display fine scale pyroxene lamellae in cores, show recrystallisation to smaller M3 neoblasts at margins and are set in a recrystallised granoblastic M3 matrix of fine-grained (0.1-0.3 mm) clino- and orthopyroxene. There is also some conspicuous growth of late amphibole grains of darker green colour than the clinopyroxene.

Stop 7-4: Ugelvik Harbour, Western Shoreline- opposite outer harbour wall. Outcrops here, ca. 130 metres NW of the last locality and close to the end of the rough road to the shore, display peridotite with conspicuously large (1-5 cm) garnets. Most are heavily fractured and hence weathered out as hollows in the rock. They also show breakdown to clots of Cr-chlorite along steep SW-dipping joint surfaces.

Stop 7-5: Ugelvik Western Shoreline Here, about 150 metres further north along this shoreline (Fig. 7.8A), are more outcrops of peridotite containing some conspicuously large garnets. The largest is of ca. 12 x 6 cm dimensions but is highly fractured and only exposed at low tide. Nearby peridotite contains other garnets of 3-6 cm size including one reasonably solid garnet megacryst of 6 x 4 cm dimensions.

Stop 7-6: Ugelvik Western Shoreline- on peridotite point into the bay Here, on the point just a few metres N of the last locality, there is a sheared out distinctly porphyroclastic and sugary textured garnet pyroxenite layer (1-3 cm thick) with a strong mineral orientation lineation that plunges shallowly N. It is enclosed within peridotite with mostly small garnets that passes just to the N into a garnet-free dunite.

Stop 7-7: Ugelvik Western Shoreline- northernmost peridotite / pyroxenite outcrop Here (Fig. 7.8A), there is a conspicuous, sheared out, porphyroclastic textured, garnet websterite layer with some distinctly garnet-rich internal layers. (Note the distinctly purple colour of the garnets in strong contrast to the red garnet colour in Stop 7-3). Such near-monomineralic layering, as also seen in certain other garnet pyroxenite occurrences within the Ugelvik and Raudhaugene peridotite bodies, may signify that such predominantly fine-grained garnet pyroxenites may be the product of shear deformation, tectonic attenuation and re-crystallisation of original megacrystic garnet plus pyroxene concentrations within these peridotites, as seen for example in the coarse garnet-bearing orthopyroxene lens at the Ugelvik harbour quarry locality (Stop 7-2). Interestingly, this particular garnet websterite occurrence is enclosed within peridotite that not only contains scattered large garnets but also some sizeable clasts of early orthopyroxenes. The extent of deformation-induced recrystallisation and of late-stage serpentinisation means that it is hard here to find relicts of early coarse-grained olivines, but such megacrystic olivines can be seen at outcrop 7-7b (Fig 7.11B). Nonetheless it is considered likely that a megacrystic garnet + orthopyroxene + olivine (± clinopyroxene??) assemblage was originally widespead within the Otrøy peridotites. Such a deduction, together with microstructural observations within the dominant peridotites of M1 garnet clasts with recrystallised trails of new finer-grained M2 garnet and of the predominance of layering-parallel mosaic-porphyroclastic and mylonitic M3 garnet pyroxenite microfabrics, signify that the overall very conspicuous mineralogical layering within these peridotites was reworked during two main ductile deformation events (M1/M2 and M3). The first phase (either M1 or M2 in age) was initiated under Pre-Caledonian HT coesite eclogite-facies UHPM conditions, the second phase occured under much lower T (but still 850-950°C) Scandian diamond-facies UHPM conditions (Fig 7.4). In addition to these two deformation events, widely separated in time, the megacrysts- and related HP decompression melting evolution clearly illustrate that the Otrøy garnet peridotites have undergone two UHP metamorphic events also widely separated in time: i.e. an Archean/Proterozoic versus a Scandian event. The Scandian UHPM event was responsible for crustal emplacement of the Otroy garnet peridotites. This topic will be dealt with in stops 7-10 to 7-13.

Stop 7-8: Ugelvik –sand road section Walking from stop 7-7a to stop 7-9 (Rødberg Byggmarket; Fig 7.8A) we walk up hill along a small sandy road. On the left side of this road we pass a peridotite outcrop that contains a beautiful megacrystic M1 garnetite with sizes of 10 to 4 cm. (Fig 7.11A). *(Please note that field excursion participants will be allowed to make photographs **but not sample** this unique M1 garnetite lens).* At Stop 7-9 we will discuss the importance of the garnetite megacrysts and their relationship to other megacrysts like Opx and Ol.

Stop 7-9: Ugelvik – Rødberg Byggmarked The owner of this builder's merchant warehouse (Harry Rødberg) has excavated some remarkable large Cr-pyrope garnet megacrysts from the underlying peridotite during the preparation of foundations for the adjacent new house. With special permission, some specimen samples may be examined and photographed on the garden wall to the house. These solid garnet megacryst fragments, with dimensions up to 9 x 8 x 6 cm, are said to have been extracted from a layer of coarse garnetite about 2 metres long. Importantly some samples demonstrate the coexistence of this early megacrystic garnet with olivine, clinopyroxene and orthopyroxene, the latter pyroxenes sometimes display evidence of internal exsolution of garnet and clinopyroxene, so confirming the important deductions made for the exsolved megacrystal orthopyroxenite pod in the Ugelvik west harbour quarry (Stop 7-2; Fig 7.11C and D).

Similar garnetites have been studied by Van Roermund and Drury (1998), Van Roermund et al (2000, 2001(a and b); Spengler (2006) and Spengler et al., (2006). The most important result was the recognition of three generations of garnet (respectively M1, M1/M2 and M3 in age; Fig 7.11A and Fig 7.12A and F). M1a garnet originally contained a majorite component of 20% (Spengler et al., 2006; Fig 7.4) and transformed during isobaric decompression into M1b garnet that contained a majorite component of 1% associated with interstitial exsolved M1b pyroxene (Cpx and Opx) and M1b olivine. During subsequent cooling till mid-Proterozoic times this M1b supersilicic garnet exsolved M2 pyroxene needles within a recrystallised M2, majorite-free, garnet host (Fig 7.12A). This microstructural evolution is similar to that of the Opx megacrysts seen in Stop 7-2. Exsolution temperatures of ≥1260 °C could be established using LREE geohermobarometric techniques applied to M2 mineral assemblages (Spengler et al., 2006). Interstitial M2 clinopyroxene and M2 garnet define Mid Proterozoic Sm-Nd cpx-grt cooling ages of 1.4 billion years (Fig 7.10B). In addition the initial Nd isotope ratio of the mineral isochron, 0.5476, is extremely radiogenic compared to mantle compositions at 1.4 GyrBP (ϵNd $_{1.4GyrBP}$ = +719). Nd modal ages calculated relative to a chondritic evolution of the mantle are 2.9 Gyr and 2.5 Gyr for olivine bearing and olivine absent garnetite, respectively (Fig 7.10B).

Stop 7-10: Raudhaugene Peridotite Body – folded garnet pyroxenite layers Drive about 1 km NE from Ugelvik and just past the small fire station turn left along a small track and park in the small disused peridotite quarry (Fig. 7.8B). This locality is in outcrops above the quarry that are roughly 50 m to the NW. Here some excellent tightly folded garnet pyroxenite layers are displayed that show tight cuspate closures in the pyroxenite in comparison with more rounded closures in the adjacent peridotite. This rather surprisingly suggests that the more garnet- and pyroxene-rich layers were mechanically weaker than the enclosing peridotite. Is this a grain-size effect, with the finer grain size of the garnet pyroxenite resulting in it being mechanically weaker than the enclosing coarse-grained peridotite? Alternatively, does it signify deformation at very high (≥1260 °C) temperatures when garnet may perhaps become mechanically weaker than olivine? Or does the garnet pyroxenite layer resemble melt during the formation of the fold? Another interesting feature is that these fold structures have much steeper plunges than seen in the Ugelvik peridotite body and also in the dominant folds and lineations seen in the gneisses adjacent to the peridotite bodies (Fig 7.6B).

Stop 7-11: Raudhaugene Peridotite Body – on K2 peridotite "peak" Walk northwestwards from the last locality onto the highest peridotite outcrop (Fig. 7.8B), the so-called K2 peak. Here there are excellent outcrops of peridotite with quite large Cr-pyrope garnets available to sample. There are also further obvious folds and some distinct garnet pyroxenite layers.

Stop 7-12: Raudhaugene Peridotite Body – darker, more Fe-rich, peridotites Just N of K2 there are distinctive outcrops of a darker-weathering, more Fe-rich, peridotite, containing dark green/black garnets, that is interlayered with the more usual paler-weathering peridotite containing the usual purple coloured garnets. Whereas the latter garnets (including the megacrystic majoritic garnets) customarily contain about 4-6 wt.% CaO and 2-4 wt.% Cr_2O_3, as is typical of garnets equilibrated within Ca-undersaturated lherzolite- and harzburgite assemblages (see Fig.11 in Van Roermund et al, 2001), the dark green/black garnets have mostly much higher contents of both CaO (as high as 12 wt%) and Cr_2O_3 (up to almost 7 wt %) probably indicative of the fact that they equilibrated as part of a wehrlite assemblage.

Whilst proto-mylonitic, mosaic-porphyroclastic M3 textures are widespread throughout the U, R and M peridotite bodies (most conspicuously in the garnet pyroxenite layers), in places this darker weathering peridotite displays an extreme ultra-mylonitic fabric in which the larger dark green M2 garnets are smeared out into stringers of recrystallised, lower Cr, M3 garnet coexisting with Cr-spinel (Fig.7.12A).

Stop 7-13: Raudhaugene Peridotite Body – multiple garnet pyroxenite layers The peridotite in the most northwesterly outcrops in this body, about 80 metres to the WNW of the last locality (Fig. 7.8B), displays numerous garnet pyroxenite layers up to 8 cm in width. Some layers are essentially bimineralic (garnet clinopyroxenite) whereas others are of garnet websterite composition, although the orthopyroxene has commonly suffered extensive secondary hydration. The garnet pyroxenite layers typically have markedly porphyroclastic M3 textures (Scandian) and contain garnet clasts (M2) that commonly contain exsolved lamellae of both clinopyroxene and finer-scale rutile (Fig 7.12B).

Three of these garnet pyroxenites/websterites reveal Sm-Nd wr-grt ages of around 430 Ma (Fig 7.12C; Spengler 2006; Spengler et al 2008). This age has been interpreted as the growth of early Scandian UHP minerals within peridotite. More important EMP chemical analyses of recrystallised M3 opx crystals (in contact with garnet; Fig 7-12A and B and D) showed that Al_2O_3 wt% in M3 opx was much lower (0.25-0.15; Fig 7.12E) than those within M2 megacryst assemblages clearly indicating prograde metamorphic conditions during continental subduction (Fig 7.12F; Spengler 2006; Spengler et al 2008). This M3 UHP foliation is thus subduction-related and interpreted to reflect garnet peridotite incorporation into the continental crust during prograde continental subduction. The most likely subduction geometry that allows such a peridotite emplacement model is that the garnet peridotite was originally positioned at sufficient depth within the hanging wall of the continental subduction system in agreement with Brueckner's model (Brueckner 1998,1999) and consistent with the application of this model to the Scandinavian Caledonides (Brueckner and Van Roermund, 2004).

Some of the conspicuous garnet pyroxenite layers within the Raudhaugene body show an Fe-enrichment trend relative to the enclosing peridotites and thus may represent the crystallisation products of more evolved magmatic melts. However, others lack Fe-enrichment trends and may show internal near-monomineralic layering, and thus seem best interpreted as the shear deformation products of original localised concentrations of the primary garnets and pyroxenes present in these peridotite bodies. Such garnet pyroxenite layers are thus most likely to be of tectono-metamorphic origin and to correspond to the extensively recrystallised products of original concentrations of megacrystic garnet and pyroxenes, as suggested in the interpretation of the garnet websterite occurrence at Stop 7-6 in the Ugelvik peridotite body.

The conspicuous garnet pyroxenite layers within the serpentine-veined peridotite at this locality can be followed along strike within further peridotite outcrops to the east. The party should then cut back south across the higher peridotite outcrops to join a track on the other side of the Raudhaugene body and then return to the vehicles parked in the quarry, completing the tour of garnet peridotites.

MAFIC DIKES AND BASEMENT-COVER RELATIONSHIPS, SOUTHERN COAST OF THE ISLANDS OF MIDSUND

by Peter Robinson and Kurt Hollocher with contributions by Emily Walsh and Megan Regel

En route from Ugelvik to Dryna ferry terminal, we will examine two key outcrops showing relationships of basement and cover rocks (Fig. 7.13), where there has been no detectable eclogite-facies overprint. Results of recent geochemical studies indicate the presence of one 1 m-thick remnant of the Sætra Nappe along a contact between the Blåhø Nappe and Baltican basement, and that crosscutting mafic dikes in the adjacent basement have Sætra-like geochemistry. In addition, U-Pb zircon geochronology on a sample of the dike-intruded basement indicates both a protolith age of ≈1600 Ma, like that of other basement in the region, but also evidence of a metamorphic overprint ≈950 Ma (Sveconorwegian) not known nearby. Both these lines of evidence suggest this particular basement should be assigned to the Middle Allochthon.

From Raudhaugene quarry return to Route 668 and head southwest through Ugelvik Village. The route turns gradually left (south) into the outskirts of Midsund. Continue to junction on corner where main road turns steeply downward to right toward center of village. Instead stay straight up short steep section toward Heggdal and Stop 7-14. At big curve in road note cuts in Blåhø mica schist and cross-cutting but strongly deformed Scandian pegmatites. Pass a sharp bend near the southwest corner of Otrøy and proceed a short distance east along the south coast. Pull over at large parking place on right with broad view. Climb down through rocks, including Blåhø mica schist and pegmatite and lots of heather to the coast and walk west a short distance on continuous shore outcrop.

STOP 7-14: (35 minutes) "Sausage rock" basement gneiss intruded by mafic dikes, now amphibolite.

The purpose of this stop is to show some very fine contact relations in a coastal pavement between strongly foliated pink basement gneiss and amphibolite. The outcrops (Localities 193, 194, 195 in Fig. 7.13) give convincing evidence that the amphibolites are dikes cutting a previously highly deformed gneiss. The geochemistry of the dikes (see below) is essentially the same as those intruding the quartzites of the Sætra Nappe, and thus they are probably Late Neoproterozoic. If correct, then the basement is probably that of the Sætra Nappe itself and not directly related to the adjacent basement of the WGR. However, minor differences in geochemistry between the sausage-rock dikes and the dikes in quartzite at Oksvollneset, suggest these two tectonic levels may not have originated in the identical part of the Middle Allochthon. The dikes clearly truncate a strong deformational and metamorphic fabric, the age of which is quite uncertain, but presumably Mid- to Early Proterozoic. The scarcity, but not total absence of garnets in these dikes suggests, but does not prove, that they never went through eclogite facies, unlike the amphibolites of the Sætra Nappe on the south shore of the fjord (see Stops 8-4, 8-5, 8-6, 8-7) as well as those of the basement to the north. In fact, no bonafide eclogites have been identified in any unit on the south coast of Midsund, although very fresh eclogites are well known only 1-2 km to the north.
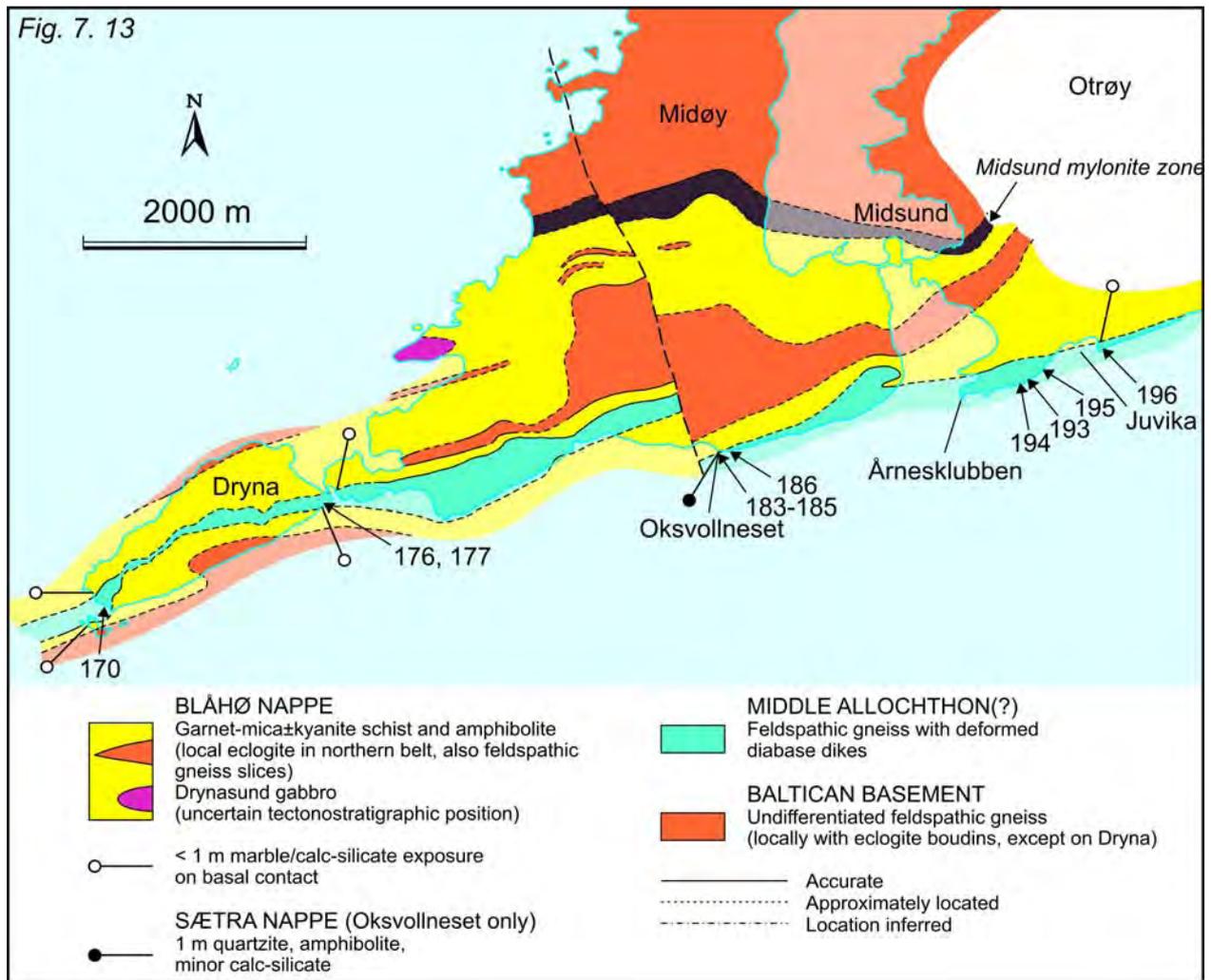
Fig. 7.13 Detailed geologic map of southwestern Otrøy, Midøy and Dryna (after Hollocher et al. 2007a), showing the Midsund mylonite zone separating UHP eclogite-facies rocks from HP eclogite-facies and amphibolite-facies rocks of the southern coast. Key locations include Oksvollneset, where 1 m of quartzite with calc-silicate on a contact between Blåhø rocks and "sausage rock" basement gneiss contains deformed dikes with Sætra geochemistry (Fig. 7.14); and other locations where thin marble/calcsilicate occurs on the same contact. Geochemical sample locations of dikes with Sætra geochemistry in "sausage-rock basement" are shown by number, and results indicate that this segment of basement should be assigned to the Middle Allochthon. Stop 7-14 is at locations 193, 194, 195 near Årnesklubben, and Stop 7-15 is near locations176, 177 at the east end of Dryna, 2 km from Dryna ferry terminal.

In one superb outcrop, the strongly foliated and laminated pink gneiss between two dikes is strongly folded about subhorizontal axes exactly parallel to the sub-horizontal late-Scandian mineral lineation. A first impression that the folds are older than the dikes is not difficult to dispel. In fact it appears that the late Scandian folds probably nucleated on the mechanical inhomogeneity created by the dikes. The older tectonic foliation in the gneiss, is, however, clearly truncated by the dikes.

Geochemistry of mafic dikes. Figure 7.14A shows mostly Sætra quartzite and basement sausage rock dikes from the Midsund area. The Sætra, exposed only at Oksvollneset (Fig. 7.13), is only ~1 m thick, but contains amphibolite boudins that plot with other Sætra dikes, as do the basement sausage rock dikes, which are interpreted to be part of the same dike swarm in a different segment of the Middle Allochthon. Shown for comparison are Sætra dikes from several other locations, basement sausage rock dikes from Lepsøy, and Sætra-like dikes in basement and Risberget augen
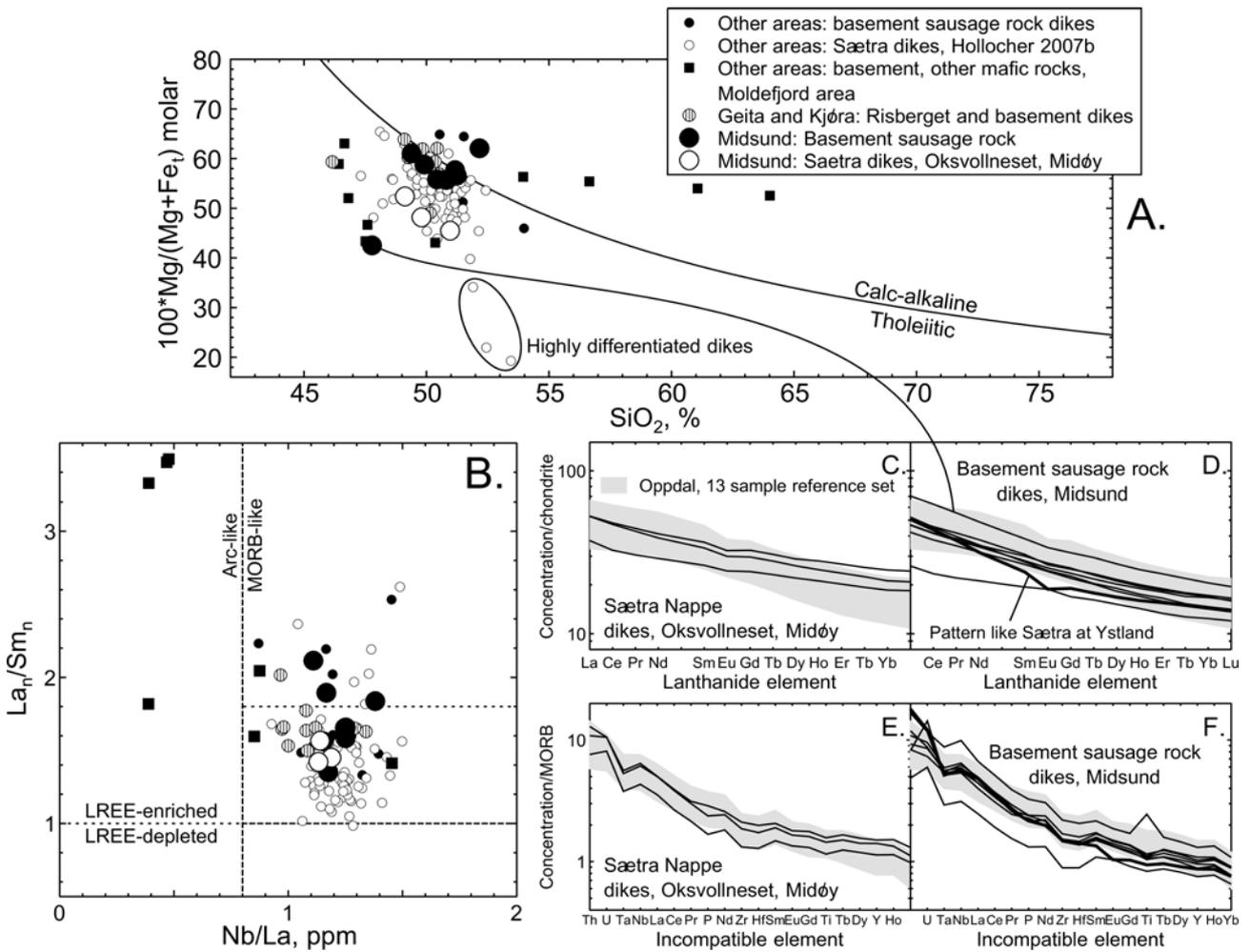
Fig. 7.14. Geochemical data principally showing the similarities of Sætra and basement sausage rock dikes in the Midsund area, and those from elsewhere. Some mafic rocks from non-sausage rock basement in the Moldefjord area are included to illustrate the diversity of compositions. A) SiO_2 vs. Mg' diagram showing rocks in comparison to a tholeiitic-calc-alkaline reference line (Miyashiro & Shido, 1975). B) Nb/La vs. La_n/Sm_n diagram showing the same rock sets and with the same key as A. C and E) REE and spider diagrams, respectively, for amphibolite boudins from a 1 m thick quartzite layer on the south shore of Midøy at Oksvollneset. D and F) REE and spider diagrams, respectively, for basement sausage rock dikes from the south shores of Dryna, Midøy, and Otrøy, Midsund area. All data is from Hollocher et al. (2007a and unpublished) except for the few mafic rocks from the 1500 Ma augen orthogneiss in *basement* on Midøy (Griffin and Carswell, 1985). REE normalizing factors are from McDonough and Sun (1995), and MORB normalizing factors are from Pearce and Parkinson (1993).

gneiss on Trondheimsfjord between Geita and Kjøra. Also shown are several samples of mafic rocks from basement near the Moldefjord syncline and on Lepsøy and Midøy to illustrate their compositional diversity compared to Sætra and basement sausage rock dikes. Figure 7.14B shows that all of the Sætra and basement sausage rocks from the Midsund area plot with other Sætra, Sætra-like, and basement sausage rock dikes from elsewhere, indicating that they are all geochemically related. Figures 7.14C and D show REE patterns for Sætra and basement sausage rocks from the Midsund area only. Most samples resemble Sætra REE patterns at Orkanger and elsewhere, but one of the basement sausage rock dikes has a pattern shape exactly like Sætra dikes at Ystland: slightly concave upward with a steep LREE slope. Figures 7.14E and F show spider diagrams for these same rocks, and the patterns closely resemble Sætra dikes elsewhere. One of the basement sausage rocks, however, has a slightly larger positive U anomaly than most Sætra, and one other sample has a positive Ti anomaly, which is unusual.
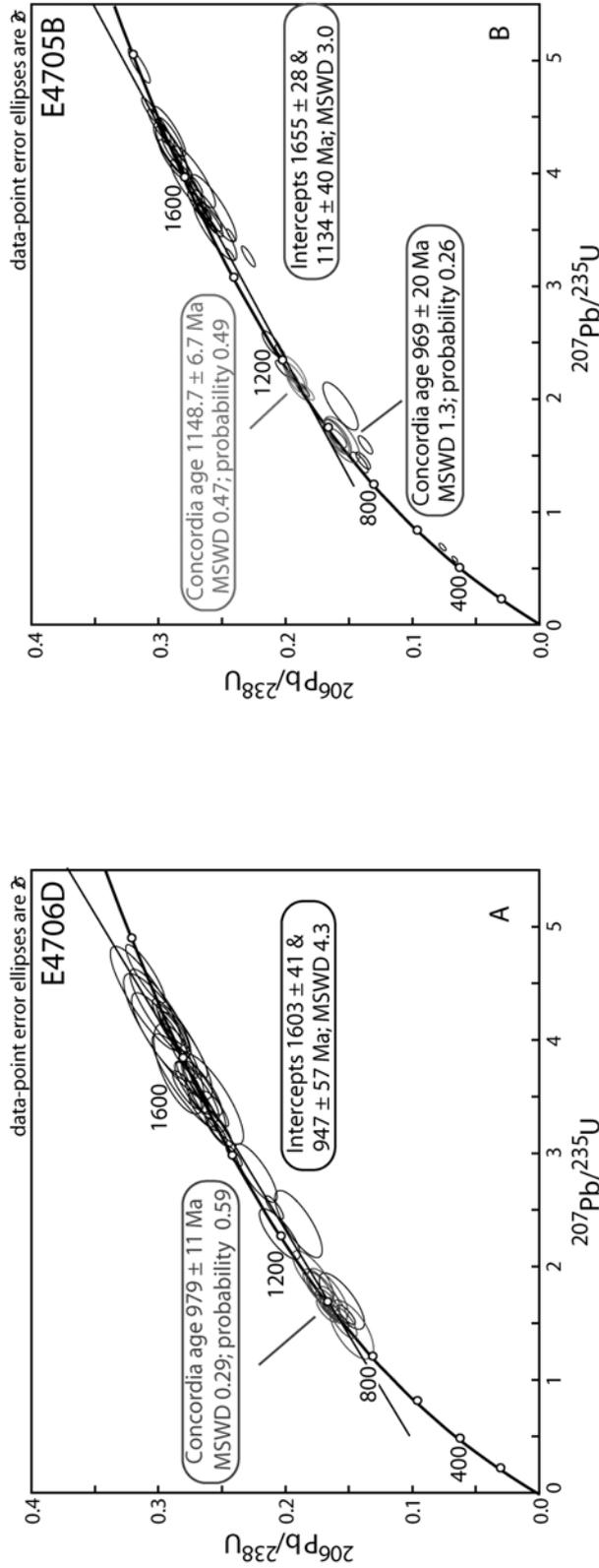
Figure 7.15. Concordia diagrams showing U-Pb ages of zircon from two samples of strongly foliated granitoid gneiss that are the host for metamorphosed Late Neoproterozoic diabase dikes A) on southern Otrøy, (locality 193 of Fig. 7.13), and B) southwestern tip of Dryna, (locality 170 of Fig. 7.13).

U-Pb zircon geochronology of foliated basement gneiss. The strongly foliated pink gneiss was sampled at locality 193 on Figure 7.13 (E4706D) for zircon U-Pb dating. The plentiful zircons are colorless, translucent and subrounded prismatic, up to ~250–300 μm long with length-to-width ratios of 2:1 to 2.5:1. Cathodoluminescence reveals complex zoning, generally including an oscillatory-zoned core, oscillatory-zoned mantle, and a lower U, apparently recrystallized rim. Zones may be divided by resorption embayments or straight-line breaks. Zircons were analyzed using the multicollector, laser-ablation inductively coupled plasma mass spectrometer (LA-ICP-MS) at the University of Arizona. Spot ages (Figure 7.15A) record only Precambrian ages, chiefly Gothian (~1690–1620 Ma; Skår, 2000) and Sveconorwegian ages; the latter is represented by a concordia age of 979 ± 11 Ma (MSWD = 0.29; all error reported at 2σ). Typically, the oscillatory-zoned core yields a Gothian age; Sveconorwegian ages are recorded by oscillatory mantles, or, more rarely, the unzoned, or hazy, rims. This trend is aptly depicted by a discordia that plots through all of the age data, showing a rough upper intercept of 1603 ± 41 Ma and lower intercept of 947 ± 57 Ma (MSWD = 4.3).

The "sausage rock" basement gneiss was sampled again (locality 170 on Fig. 7.13) on the coast west of the Dryna ferry terminal (E4705B; Figure 7.15B), with slightly different results. Petrologically and mineralogically, the two samples are similar; zircons are translucent, subrounded prisms with ~2:1 length-to-width ratios or, in this sample, rounded grains ~100 μm in diameter. Cathodoluminescence reveals equally complex zoning patterns, although many zones that appear to be different record the same age, within error. Oscillatory-zoned cores commonly yield Gothian ages, as with E4706D, and while the Sveconorwegian event is weaker, it is present (concordia age of 969 ± 20 Ma, MSWD = 1.3). E4705B exhibits a stronger ~1150 Ma age signal, as demonstrated by a concordia age of 1148.7 ± 6.7 Ma (MSWD = 0.47), and two Caledonian spot ages. The Caledonian ages were recorded by apparently recrystallized outer rims, one from a Gothian-aged grain and one from a Sveconorwegian-aged grain; evidence of this can be seen in the discordant ellipses of Figure 7.15B. More typical of zircons analyzed from this sample is a Gothian core with mantle ages of ~1150, as is depicted by the discordia with upper intercept 1655 ± 28 and lower intercept 1134 ± 40 Ma (MSWD = 3.0).

U-Pb zircon geochronology of foliated basement gneiss. See Figure 7.15 and complete discussion there.

Leave Stop 7.14. Return toward Midsund. At junction with Route 668, turn sharp left down hill and through village center. Continue west across bridge from Otrøy to Midøy. At T junction turn right (not left toward Ramsvik) staying on route 668 and pass around northern tip of Otrøya, thus passing north across the Midsund mylonite zone and back into eclogite-facies pink augen orthogneisses that dominate the landscape. Ahead there are vast views of Nordøyane (North Islands) including Goaldet (meaning of name uncertain), the peak towering over the Lepsøy Misjonsenter, our home for the next three nights.

We pass by a stone wall on the right with the sign "Riksgrense". This was the E-W international border between Sweden to the north and Denmark/Norway to the south between February 26, 1658 and May 27, 1660. On first date, after a serious military defeat, Denmark ceded Trondelag, Skåne, Bornholm and several other places to Sweden. On the second date, after another war, Trondelag went back to Denmark.

To the southwest on the coast we can glimpse the point, Little Digerneset. Here, Griffin and Carswell (1985) found eclogite-facies assemblages in both mafic rocks and in original igneous "back veins" of melted country rock, making a strong argument that the eclogites are in-situ metamorphic rocks, not tectonic "water melon seeds" from the mantle. Dominant pink rocks are augen orthogneiss of Carswell and Harvey (1985) dated at 1506 ±22 Ma by whole rock Rb-Sr, and more recently at 1508 by Tucker et al. (1990) using U-Pb zircon.

Just before reaching the enormous cliff on the left, we pass over the unexposed Midsund mylonite zone again and enter a thick zone of rocks assigned to the Blåhø Nappe. On the coast west from here, the mica schist contains kyanite, amphibolite contains garnet-clinopyroxene-rutile assemblages with retrograde hornblende-plagioclase pseudomorphs after garnet along walls of pegmatites. There is also minor eclogite.

In the next several km we travel mostly on new roads on the western part of Midøya, passing several contacts between Blåho and basement. There is a view to right of Drynasund lighthouse resting on gabbro. Suspected eclogite-facies shear zones (Griffin and Råheim, 1973) are not supported by recent work (H. Van Roermund, pers comm. 2008). Eventually we swing by a huge coastal exposure of Blåhø mica schists, amphibolites and cross-cutting pegmatites, then cross a very short bridge across the channel from Midøy to Dryna. Just beyond the bridge is a bus stop on the left and just beyond that is a narrow parking place to be used for Stop 7-15 (Localities 176, 177 on Fig. 7.13). Walk forward a short distance to coastal exposure on southeast side of road. Depending on time before the ferry to Brattvåg (18:05) we may omit the last part of this stop or the entire stop.

STOP 7-15: (15 minutes) Contact of Blåhø Nappe and "Sausage Rock" Basement. "Sausage rock" (Norwegian= pølserstein), already seen at Stop 7-14, is the informally named rock unit that dominates the south coasts of Dryna, Midøya and western Otrøy (Fig. 7.13). Fairly similar rock is also known in the central segment of Nordøyane (Stop 9-8). It consists of thinly laminated, highly foliated granitoid gneiss with thick and thin highly boudinaged layers of even-grained, locally porphyritic hornblende amphibolite. Because of the very high ductility contrast between enclosing gneiss and amphibolite, the amphibolite deformed into complex sausage shapes parallel to the late Scandian extensional lineation. When viewed in vertical outcrops cut normal to the lineation, the pattern is commonly chaotic, and contrasts sharply with the "straight" fabric observed in adjacent belts of schist and amphibolite of the Blåhø Nappe. When viewed on horizontal surfaces perpendicular to foliation and parallel to lineation, the boudins commonly show "bookshelf" style boudinage indicating late sinistral shear. There is abundant evidence that the mafic layers were invaded locally by late felsic melts presumably derived from the felsic matrix.

   The purpose of this stop is to examine contact relations on both sides of a narrow belt of sausage rock that extends the entire length of Dryna (Fig. 7.13). The first coastal exposure shows the south contact of this belt with mica schist and amphibolite assigned to the Blåhø Nappe. The contact zone, 1 m thick, is characterized by a thin bed of marble and diopside calc-silicate and also feldspathic schist zones that are unusually quartz-rich. The majority of well exposed contact zones on Midsund contain such marble and calc-silicate layers. The rusty-weathering schist several meters into the Blåhø Nappe is rich in gedrite, suggesting it is likely to be a metamorphosed hydrothermally altered mafic volcanic rock. Gedrite rocks have been found in similar locations on Otrøy as well as on Lepsøya (see Stop 9-8).

   After observing this exposure walk north on the road past vehicles to another pavement of sausage rock which also lies on the coast just before the bus stop. From the pavement walk along the west side of a knob close to the road to a small northwest-facing outcrop at the water line which is beyond the bus stop. Here there is a thin layer of marble and calc-silicate which is on the north contact of the belt of sausage rock. The marble occurs in a steep groove on the facing cliff of Middøya about 50m away, showing that there is no appreciable horizontal fault offset in the narrow gut between the islands.

Return to vehicles after Stop 7-15, and drive 2 km west (about 2 minutes) to Dryna ferry terminal and park in waiting line. The nearest outcrop is Blåhø garnet-biotite schist and amphibolite cut by pegmatite. Drive onto ferry at 18:05. Ferry crossing takes 20 minutes during which various parts of the scenery associated with Day 8 can be pointed out. Upon arrival at Brattvåg ferry terminal, we have 50 minutes in which to accomplish the 10-minute drive to the ferry terminal at Skjeltene no later than 19:15, There are several options, such as examining the outcrops near the Brattvåg terminal, going to one stop related to Day 8, or going to a store in Brattvåg.

Drive onto ferry at Skjeltene for crossing to Lepsøy (Lauvsøy in some schedules). When entering the ferry, each driver must express a preference for destination and will be directed onto the ferry accordingly. The correct answer is Kjerstad (pronounced 'share-sta'). Once ashore in Kjerstad, we drive away from the shore and turn right (north) on the main road, traveling about 5 minutes to Lepsøy Misjonsenter, where we stay for the next three nights. Dinner will be served soon after our arrival, but first it may be a good idea to find your sleeping location, which may be either in the main house, in the small shore house, or in a trailer parked in the yard.

Day 8: STRUCTURAL-METAMORPHIC RELATIONSHIPS BETWEEN CALEDONIDE NAPPES AND FENNOSCANDIAN BASEMENT ON THE MAINLAND NEAR BRATTVÅG

Friday August 22 (potentially August 23 depending on weather)

by Peter Robinson and Kurt Hollocher with contribution by Bob Tucker

   General Route: The day will begin with a morning ferry ride from Lepsøya to Skeltene on the mainland. The trip log begins at the junction for Brattvåg Ferry Terminal. From Brattvåg (Fig. 8.1) we drive south and then east through Vatne and over a low mountain pass to Fikksdal in Vestnes Kommune. From Fikksdal we drive northwest and west along the south shore of Moldefjord back to Vatne, with four stops at Dragneset, Øygaardsneset, and on the east shore of Vatnefjord. From Vatne we return toward Brattvåg, but turn right at the junction to Helland at the bottom of the hill beyond the tunnel, and proceed to the coast east of Brattvåg for five stops. After these stops, the route is retraced to Brattvåg, and the trip log includes three additional stops a short distance west of the Brattvåg Ferry Terminal, if time permits. At the end of the day we will take the ferry from Skjeltene back to Lepsøya.

   Ferry Lepsøya to Skjeltene: Friday (or Saturday) **8:00-8:40** (long way); **9:50-10:10.**
   Ferry Skjeltene to Lepsøya: Friday 18:15-18:30, 19:15-19:30; Saturday 19:15-19:30

INTRODUCTION TO DAY 8

   The day will be devoted to the features of Caledonide nappes on the south limb of the Moldefjord syncline and in the Helleneset syncline, and their structural and metamorphic relations with the underlying Proterozoic basement. The introduction to the nappes in this part of the Scandian hinterland follows up on earlier views of similar stratigraphic units on Days 3-6 and provides a necessary background to understanding the evolution of the more basement-dominated areas further north studied on Days 7, 9 and 10.

Tectono-stratigraphic Units: Moldefjord and Helleneset Synclines near Brattvåg

   A sequence of seven tectono-stratigraphic units was identified that corresponds very closely with units in Trollheimen and elsewhere in the central Caledonides, although some of the units observed elsewhere are absent. This geology is summarized in detail in Robinson (1995c), though tectonic implications of the structural features were poorly understood at the time. Tectono-stratigraphic correlations of rocks assigned to the Middle Allochthon have been tightened considerably through recent major- and trace-element studies of mafic dike rocks by Hollocher et al. 2007a.

Baltican basement. This basement consists of strongly deformed gneisses, gabbros, eclogites, amphibolites, and trace amounts of supracrustal rocks, derived from Proterozoic intrusions with very minor amounts of host rock. Locally gabbro boudins preserve primary igneous textures. The basement is locally separated into **lower basement** and **upper basement** by Åmotsdal Quartzite. By analogy with Trollheimen, the quartzite is probably unconformable on the lower basement and is overthrust by the upper basement. As in Trollheimen, eclogite is found in upper basement (Stop 8-5), but not in lower basement.

Åmotsdal Quartzite. Feldspathic quartzites, locally hematitic and locally with feldspathic pebbles, exposed in an anticline in Vatnefjord and to the east (Stop 8-3). Locally up to 100 m thick, the quartzite has been mapped in areas where it is only 1 m thick (Robinson, 1997).

Risberget Nappe. This thrust nappe consists of augen gneiss and amphibolite that represent deformed middle Proterozoic rapakivi granite with subordinate gabbro, intruded by mafic dikes (Stops 8-2, 8-4, 8-5, 8-6, 8-7, 8-11). Near Brattvåg the Risberget Nappe contains a distinctive mappable unit of amphibolite derived from gabbro (Stops 8-5, 8-6, 8-7, 8-11) in which thin dikes of rapakivi granite have been recognized. In a boudin near Skår the rapakivi granite is preserved in a totally undeformed state with large orthoclase phenocrysts rimmed by plagioclase (Stop 8-6). Recognition of the Risberget Nappe is more difficult in the Moldefjord and Surnadal synclines than in eastern Trollheimen. In eastern Trollheimen the Risberget Nappe is consistently separated from upper basement by Åmotsdal Quartzite and other autochthonous or para-autochthonous supracrustal rocks that are absent in these western areas. Augen gneiss of generally similar character is a major component of the Baltica basement, especially to the north of the Moldefjord (Carswell & Harvey 1985) where it has yielded a U-Pb zircon age of 1508 Ma (Tucker et al. 1990), similar to the Rb-Sr whole rock age obtained by Krill at Oppdal and to the U-Pb zircon age obtained on the porphyritic mangerite at Flatraket (Lappin et al. 1979). In practice the basement augen gneiss is usually fine grained and the microcline phenocrysts are flattened and polycrystalline as a result of relatively homogeneous deformation, whereas the typical gneiss of the Risberget Nappe contains large coarsely crystalline phenocrysts in a micaceous matrix interspersed with zones of obviously mylonitic character. A recent development comes from the U-Pb zircon dating of augen gneisses in the Risberget and Tännäs Augen Gneiss Nappes (Handke et al. 1995) at nine different locations in Norway. Of these 5 yielded ages of 1650-1642 Ma., four yielded surprising ages of 1190-1180 Ma. including the rocks exposed at Skår (Stop 8-7) and Oppdal (Stop 6-8B) and none yielded the "expected" age of 1500 Ma. The best argument that
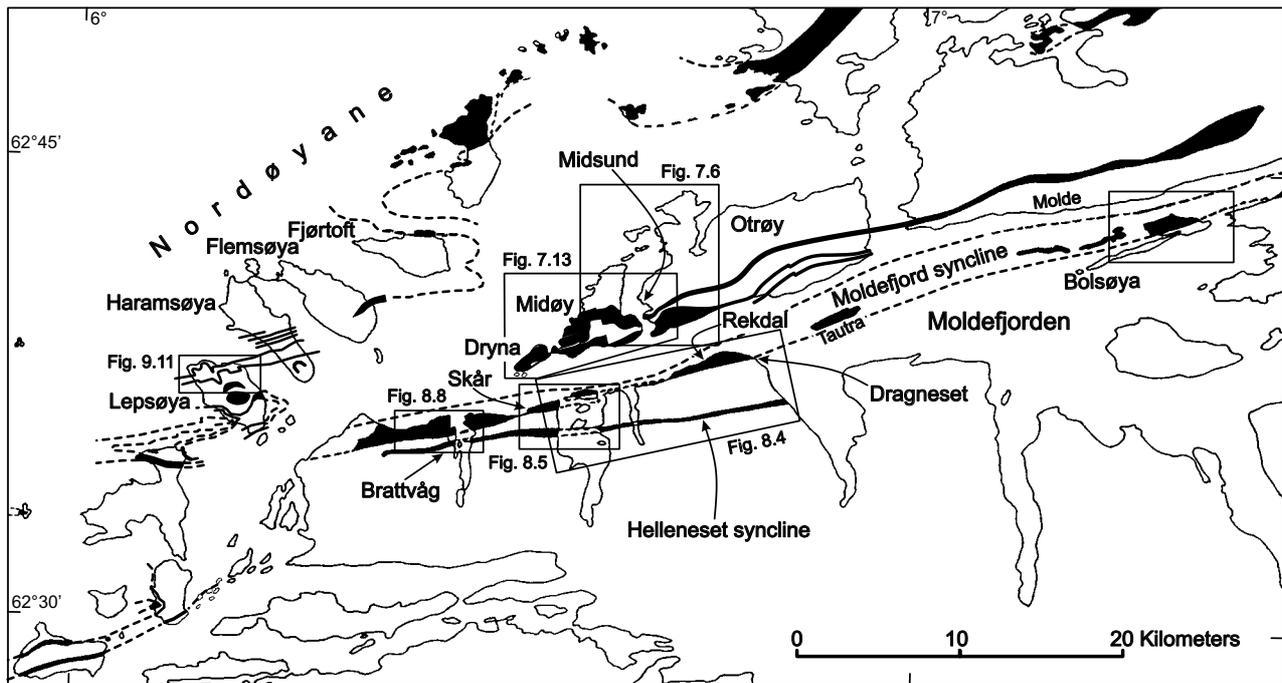
Fig. 8.1 Index map showing the general region to be visited on Days 7, 8, 9, with emphasis on distribution of cover rocks and special emphasis on the Moldefjord and Helleneset synclines to be visited on Day 8. Areas covered by Figures 7.6, 7.13, 8.4, 8.5, 8.8, and 9.11 indicated.

the Risberget unit is a nappe comes from its great lateral continuity and consistent stratigraphic position across the orogen from these areas to the Tännäs Augen Gneiss Nappe in Sweden (Krill & Röshoff 1981). The new ages lend further support to this interpretation. Added support comes from the geochemistry of mafic 'dikes' (see below) showing that they have identical compositions to dikes of the overlying Sætra Nappe (Hollocher et al. 2007a).

Sætra Nappe. This nappe is characterized by strongly laminated pure to feldspathic biotite quartzite and amphibolite, locally garnetiferous, and locally eclogitic (Stops 8-1, 8-4, 8-5, 8-6, 8-7, 8-11, 9-8). These rocks represent metamorphosed sandstone cut by diabase dikes corresponding to the Särv Nappe near the front of the orogen in Sweden. Though locally discontinuous, this key unit occurs throughout the belt, but is commonly no more than 10 meters thick. It is also present on northern Lepsøya (Stop 9-8) and southern Midøy (Fig. 7.13), where it is about 1m thick. Near Brattvåg, the unit, especially its lower part, contains boudins of zoisite±phengite eclogite up to 10 m thick, and prograde growth zoning in garnet (Stops 8-4, 8-5, 8-6, 8-7) . Outcrop relationships demonstrate that the eclogite-facies metamorphism took place before the dike rocks where deformed into very thin amphibolites. In general aspect, these rocks are quite similar to metamorphosed sandstones with diabase dikes metamorphosed to eclogite at Grapesvare, in the Seve Nappe of Arctic Sweden. At that location the eclogite metamorphism has been dated at 505 Ma (Mørk et al. 1988; Essex & Gromet 1996; Essex et al. 1997), opening the question as to whether these eclogites are also pre-Scandian. Unfortunately no attempt has yet been made to do geochronology. Geochemistry of ordinary amphibolites and of eclogites (see below) show they have identical compositions equivalent to mid-ocean-ridge basalts (Hollocher et al. 2007a, 2007b).

Geochemistry of mafic rocks in the Sætra and Risberget Nappes. Figure 8.2 shows dikes from the Sætra Nappe and a variety of mafic rocks from the Risberget Nappe in the Moldefjord syncline. Figure 8.2A shows that Sætra dikes in the Moldefjord syncline have similar compositions despite considerable geographic distance between sampled outcrops, and the fact that seven of the samples are retrograded eclogite (see Stops 8-4, 8-5, 8-7). REE patterns (Fig. 8.2E) are mostly like Sætra dikes at Orkanger, as they are at most other localities regionally, but one (an amphibolite) from Sunnaland, near Brattvåg (Stop 8-11), has a REE pattern like those at Ystland on Trondheimsfjord. The three samples with the highest REE concentrations and negative Eu anomalies are evolved rocks with Mg' values of 19-34. Extensive plagioclase removal is indicated by negative Eu anomalies, and Fe-Ti oxide removal is indicated by large negative Ti anomalies in the two most evolved rocks (Fig. 8.2J). The positive P anomalies in these two rocks, however, shows that apatite saturation had not been reached, or had not yet removed significant apatite at the time of dike emplacement. Two eclogite samples from the Sætra in the Helleneset syncline (west from Stop 8-4) are almost identical to some in the Moldefjord syncline (Fig. 8.2B, F, K).
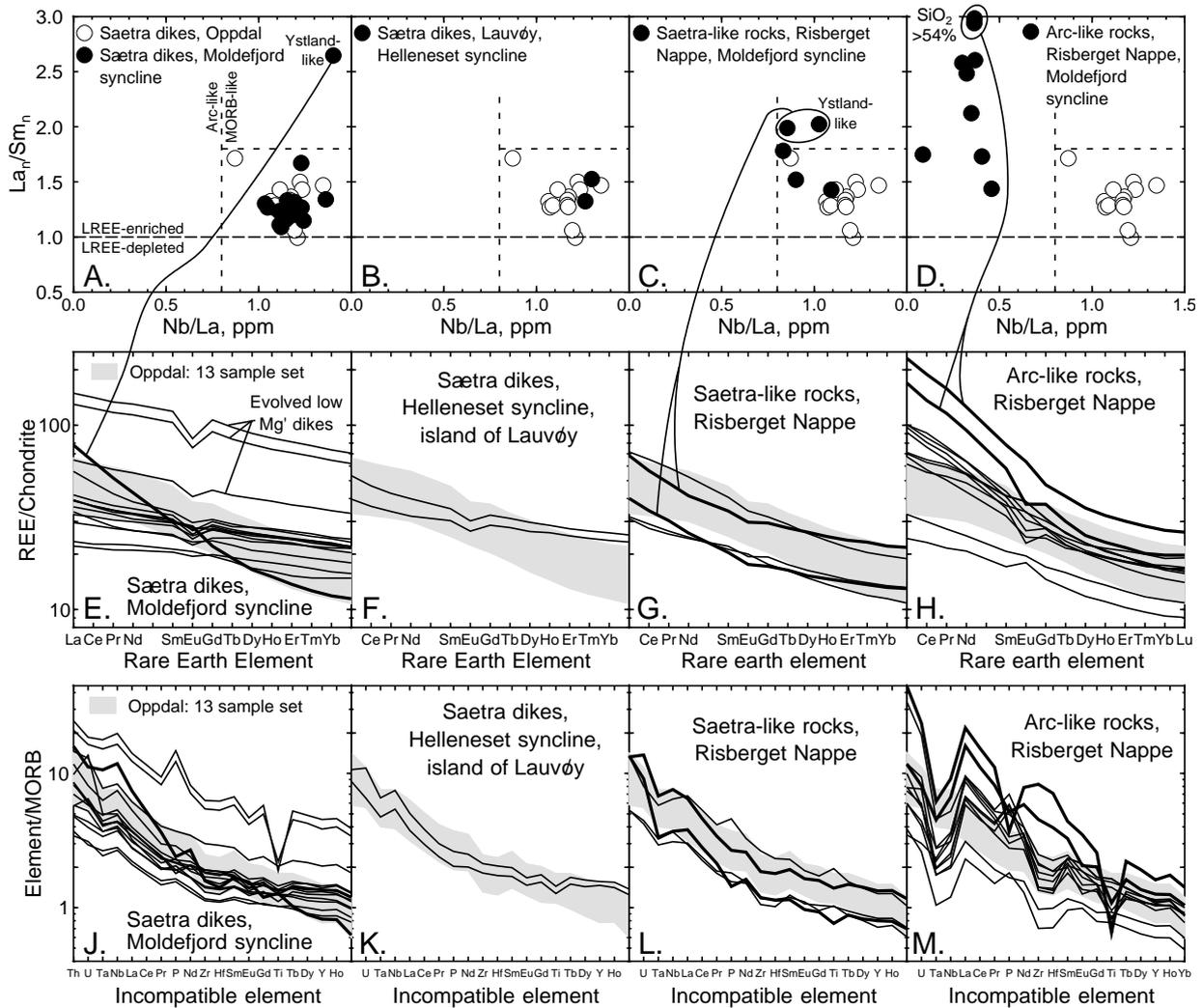
Fig. 8.2 Geochemical diagrams for Sætra Nappe dikes and mafic rocks in augen gneiss of the Risberget Nappe in the Moldefjord syncline, compared with Sætra dikes at Oppdal. A-D) Nb/La vs. La_n/Sm_n diagrams. E-H) REE diagrams for all of the same rocks as in A-D. J-M) Spider diagrams for all of the same rocks as in A-D. Data and diagrams (modified) are from Hollocher et al. (2007b). REE normalizing factors are from McDonough & Sun (1995), and MORB normalizing factors are from Pearce & Parkinson (1993).

A small subset of amphibolites sampled from augen gneiss of the Risberget Nappe (Fig. 8.2C, G, L) have compositions that are much like the Sætra Nappe dikes, possibly indicating that the Risberget Nappe contains dikes related to Iapetan rifting like those of the basement sausage rock (Days 7 and 9). Two of these samples have REE patterns that most closely resemble Sætra dikes at Ystland. In contrast, the majority of amphibolites sampled in the Risberget Nappe (Fig. 8.2D, H, M) plot in the arc-like field of Figure 8.2D and do not resemble the Sætra dikes. This is supported by Figure 8.2M, where all of these rocks have large negative Nb-Ta anomalies, and several also have larger negative Zr-Hf anomalies than any Sætra or basement sausage rock dikes. The Risberget Nappe in the Moldefjord syncline therefore has examples of both arc-like and Sætra-like rocks. The former are probably from pre-Iapetan magmatism, but the latter may represent another part of the Middle Allochthon with Late Neoproterozoic dikes related to opening of the Iapetus Ocean.

Blåhø Nappe. This nappe consists of porphyroblastic garnet-mica-feldspar schist, pyroxene-biotite schist, coarse gneissic garnet and pyroxene amphibolite, diopside calc-silicate rock, and pegmatite; interpreted as highly metamorphosed shales and volcanics (Stops 8-1, 8-4, 8-7, 8-8, 8-9, 8-10, 8-11, 9-7, 9-8, 9-10, 7-Y). In the upper part of the Blåhø Nappe there is a bed of impure marble and calcareous schist about 6m thick separated from the Støren contact by 15m or less of mica-garnet schist (Stop 8-10). The position of marble near or at the top of the Blåhø Nappe is critical to several fold interpretations in the area west of Brattvåg. Commonly there are also thin impure marbles or calcareous impure quartzites at or near the base (Stop 8-1, 8-8). In its best preserved portions, this nappe shows evidence of early high-temperature metamorphism with extensive development of migmatites unlike anything in the other nappes (Stop 8-8). In this respect it corresponds to part of the Seve Nappe in Sweden. Migmatitic features are particularly well shown in the lower part of the Blåhø Nappe, where there is a lesser amount of retrograde hydration. R. D. Tucker has obtained U-Pb zircon ages from two pegmatites in similar rocks near Orkanger that cut an older

metamorphic fabric but are themselves deformed in the late stages (See Stop 6-1). These have yielded ages of 430 and 422.7 Ma, considerably older than pegmatites in adjacent basement (Solli & Robinson 1997, Tucker et al. 2004).

Støren Nappe. This is composed of strongly laminated, fine-grained epidote amphibolite derived from metamorphosed basaltic volcanics, minor felsic schist derived from felsic volcanics with thin interbeds of fine-grained garnet-biotite-muscovite-tourmaline schist, and subordinate biotite-garnet-tourmaline schist representing intercalated sediments (Stops 8-8, 8-9, 8-10. Major layers of very coarse amphibolite and hornblendite, locally with relict cumulate textures, are metamorphosed massive to layered gabbro (Stop 8-9). There is also a lens of talc-anthophyllite-carbonate schist. These mafic and ultramafic rocks may be dismembered fragments of ophiolite such as the Bymarka and Løkken bodies near Trondheim. Their major- and trace-element geochemistry (see below) has been investigated by Hollocher et al. 2007c and in preparation.

Geochemistry of metamorphosed volcanic rocks and gabbros in the Støren Nappe. Figure 8.3 shows Støren Nappe rocks from the Moldefjord syncline. The Støren Nappe is dominated by fine-grained mafic rocks, probably volcanics, that are typically layered and boudinaged on a scale of centimeters. There are also coarse-grained, complexly layered rocks that are probably gabbros. Intermediate and felsic rocks were sampled preferentially, so their actual abundance in outcrop is considerably less than is indicated by the number of data points. Note that the diorites are all from Bolsøya, from the northeastern end of the Moldefjord syncline. Most of the mafic rocks plot on the tholeiitic side of the reference line, except for cumulate gabbros which plot on the calc-alkaline side of the reference line because of high Mg' values resulting from cumulus mafic minerals. In contrast, all of the intermediate and felsic rocks in the Moldefjord syncline are calc-alkaline.

Figure 8.3B shows that the Moldefjord syncline mafic volcanics occur only in the MORB depletion trend, with no representatives on the arc trend. The rocks with Nb/La ratios >1 have REE patterns (Fig. 8.3D) that resemble those of Sætra dikes at Oppdal, which are much like E-MORB volcanics (see Fig. 6.x3B). With incompatible element depletion, the rocks have progressively lower La_n/Sm_n ratios and resemble highly depleted N-MORB lavas. Spider patterns (Fig. 8.3G) indicate that the most depleted samples also have the largest negative Nb-Ta anomalies. It seems likely that this represents minor replenishment of the melting source by slab fluids, which may also explain the highly variable U anomalies. The compositional gap in Figure 8.3B (Nb/La ratios 0.9-1.3) probably represents sampling bias.

Figure 8.3C illustrates how coarse-grained gabbros differ compositionally from the fine-grained mafic volcanics. The volcanics span a factor of 100 range of Th concentrations, but vary little in Sc concentration which would have been partially buffered by most likely melting or crystallization processes. In contrast, the low-Th gabbros span a wide range of Sc concentrations that were probably caused by variable plagioclase-pyroxene ratios in cumulate rocks. This mineralogic variability is evident in outcrop (Stop 8-9). Calculations (Fig. 5.5C and related text) indicate that several of these gabbros probably represent adcumulates with only ~1% trapped liquid. Figure 8.3E shows that all of these gabbros are LREE-depleted and many have positive Eu anomalies, indicative of cumulus plagioclase. Because plagioclase tends to be LREE-enriched compared to parent magma, and clinopyroxene, orthopyroxene, and olivine tend to be LREE-depleted, cumulate gabbros tend to have REE patterns that resemble their parent magmas. The fact that all of these samples are LREE-depleted, many strongly so, and preliminary modeling, suggests that the parental magmas were also LREE-depleted. Spider patterns for the gabbros (Fig. 8.3H) are difficult to interpret without additional modeling, but it appears that the gabbros must have been derived from parental magmas that already had moderate negative Nb-Ta anomalies, probably like those at the lower left end of the MORB trend in Figure 8.3C.

Figure 8.3F, I show the more felsic rocks of the Støren Nappe in the Moldefjord syncline. The andesites and diorites have REE patterns and concentrations almost identical to the more REE-rich (E-MORB-like) mafic volcanics, but have moderate to large negative Nb-Ta anomalies which the REE-rich mafic volcanics do not. The dacitic rocks have REE and spider patterns like those of the Fagervika Trondhjemite, and so may represent partial melting of garnet-free source rocks in relatively shallow arc crust, inheriting the large negative Nb-Ta anomalies from the melting source. Further conclusions will have to await modeling.

Among the Støren greenstones at Myrakaia is a layer of ultramafic talc-anthophyllite rock, possibly representing a primitive picritic lava flow or sill, or a section of mantle rock. Although it is possible that this rock got its ~27% MgO content from hydrothermal alteration with seawater (e.g., Mottl & Holland 1978; Humphris & Thompson 1978), this seems an unnecessary complication considering the high Cr (~2300 ppm) and Ni (~1000 ppm) content of the rocks, and their relatively high 5.7-7.2% CaO content, which is exchanged for MgO during hydrothermal alteration with sea water. These rocks have REE and spider patterns (Fig. 8.3D, G) that are unique among the Støren samples: LREE-depleted like many of the gabbroic cumulates, but with large *negative* Eu anomalies. The origin of this ultramafic layer is not understood.

General Structure

The tectono-stratigraphic units are exposed in two tight to isoclinal synclines overturned to the north-northwest, with subhorizontal axes. The major Moldefjord syncline (Fig. 8.1) has been traced for 70 km and the narrower Helleneset syncline has been traced for 25 km. The structure east of Molde is a syncline, not fault controlled. This is shown by well exposed overturned north-facing sequences on Bolsøya and other islands near Molde, and on Grønnes Peninsula,
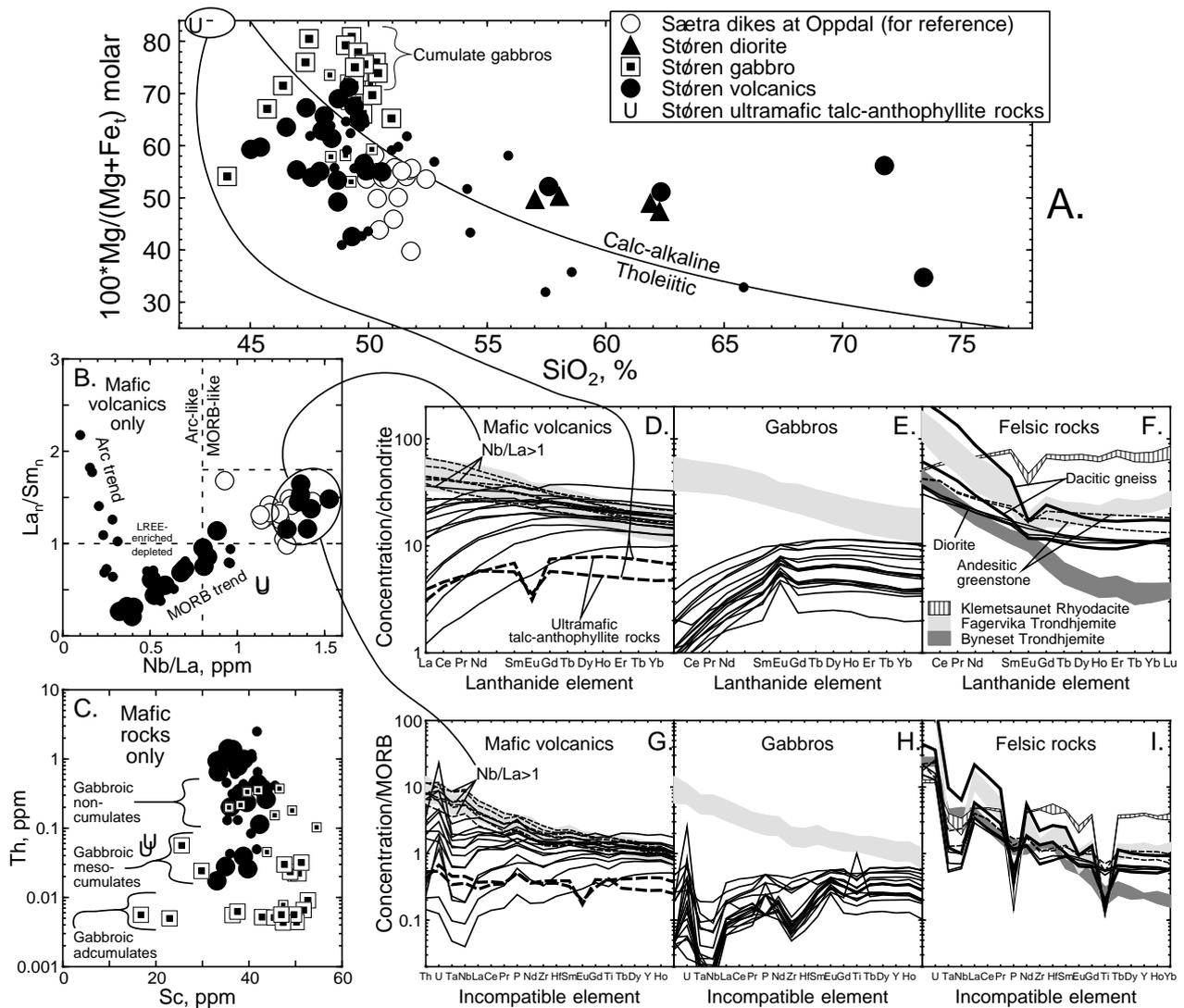
Fig. 8.3 Geochemical diagrams for the Støren Nappe in the Moldefjord syncline, compared with Sætra Nappe dikes at Oppdal and, for felsic and intermediate rocks, three felsic units from the Bymarka Ophiolite in Trondheim (see Fig. 5.5G and J and related text). A, B, and C use the same key, and small symbols represent the same units as identical large symbols, but from outside of the Moldefjord syncline. A) SiO_2 vs. Mg' diagram showing all Støren samples in the Moldefjord syncline. Discriminant line is after Miyashiro & Shido (1975). B) Nb/La vs. La_n/Sm_n diagram showing mafic volcanics (compare to Fig. 5.5A). C) Sc vs. Th diagram showing the relationships between gabbroic and volcanic units (compare to Fig. 5.5C). D-F) REE diagrams for mafic volcanics, gabbros, and felsic volcanics. G-I) Spider diagrams for mafic volcanics, gabbros, and felsic volcanics. All data are from Hollocher et al. (unpublished). Reference lines in B are from Hollocher et al. (2007b). REE normalizing factors are from McDonough & Sun (1995), and MORB normalizing factors are from Pearce & Parkinson (1993).

and by less well exposed right-side-up south-facing sequences on the north side of Grønnes Peninsula and in the Fannefjord Tunnel. The west-plunging synclinal hinge is mapped on the Grønnes Peninsula near Kortgarden. Near Brattvåg, on the south limb (Fig. 8.8), there are three tightly folded north-facing sequences separated by two fine-grained mylonite zones. Parts of the south limb are well exposed east of Brattvåg near Vatnefjord (Fig. 8.5), near Rekdal (Fig. 8.4), and on Tautra. The Helleneset syncline (Figs. 8.4, 8.5, 8.8) south of the Moldefjord syncline has been traced from west of Brattvåg to southeast of Rekdal. Its limbs and axial surface have been refolded about late open folds with horizontal or gently south-dipping axial surfaces (Fig. 8.7). Between these synclines (Fig. 8.4) is the complex Rekdalshesten anticline outlined by a thick unit of Åmotsdal feldspathic quartzite separating layers of lower and upper basement near Vatnefjorden and in mountains to the east (Robinson 1997).

Sequence of Structural Development

Although various nappes probably contain older structural features, the earliest recognizable event was the progressive development and emplacement of thrust nappes for several hundred km onto the Baltica margin in the **early part of the Scandian** orogeny. We believe (Robinson et al. 1996, 1997; Tucker et al. 1997a, 2004) that this carried rocks of the Støren Nappe far onto the then cool Baltica foreland, while contemporaneously there was subduction of the

Baltica margin with its heating and high pressure metamorphism, locally to eclogite facies, producing early transverse top-to-SE shear fabrics (see Stop 9-1), new metamorphic zircon variously at 415, 412, 410, 402 and 401 Ma (Krogh et al. 2003,2004), and progressive Pb loss in Proterozoic igneous titanite.

The **middle part of the Scandian** involved basement imbrication at deep levels, with contemporaneous extensional collapse at higher levels. Features of the middle Scandian included the following: Thrust imbrication of the subducted heated basement slab or slabs onto cooler basement toward the foreland. We suggest this accounts for the repetition of basements separated by quartzite in Trollheimen and in the Rekdalshesten anticline, where eclogite is found only in the upper basement. At the same time there was backsliding and extensional emplacement of the cool high-level nappes against the cooling basement. Cooling of the basement slab or slabs from above and below abruptly terminated the short period of Pb loss in basement titanite at 395±2 Ma, while titanites in overlying Ordovician igneous rocks of the Støren Nappe were never warm enough to be reset. This first extensional collapse was the main event that brought deep metamorphic rocks relatively close to the surface.

The **late Scandian** was characterized by further extensional deformation in a field of sinistral transtension. Early in the late Scandian there was development of northeast-trending ductile folds and stretching lineations within the previously thrust and extensionally juxtaposed package. During this deformation subhorizontal layers developed top-west shear features, whereas steep to overturned layers developed predominantly sinistral shear features that pervade the region. The later part of the late Scandian brought ductilely deformed rocks into contact with more brittle rocks from higher levels associated with Devonian sandstone basins deposited earlier in the period 403-394 Ma.

Note in the above scenario that we show no place for the eclogite-facies metamorphism of the Sætra Nappe near Brattvåg, nor for the early high-T metamorphism of the Blåhø Nappe. This is partly because we do not know the ages of these metamorphisms. The eclogites in the Sætra Nappe bear a very close resemblance to eclogites in the Seve Nappe of Arctic Sweden (Andréasson et al. 1985), also in late Proterozoic sandstone, which were metamorphosed at about 500 Ma, (Mørk et al. 1988) in an uncertain setting. The high-grade metamorphism and deformation of the Blåhø nappes seems to have begun earlier, but terminated at about the same time as intrusion of pegmatites dated at 430 and 422.7 Ma near Orkanger (Tucker et al. 1997a, 2004). These events might be ascribed to subduction of rocks on the Baltica margin that came earlier than the subduction of the presently exposed basement.

How do the details of the geology near Brattvåg and on Nordøyane fit into this picture? We suggest that most of the contacts between nappes and between nappes and basement are original thrust contacts on which there has been a major amount of extensional backsliding and thinning during the middle Scandian. Once these contacts had been developed they were folded, first into a series of tight to recumbent anticlines and synclines, and then in a series of more open folds with northeast-trending axes, strong lineation, and sinistral and top-west shear features. At this point it is not easy to tell whether the earliest isoclinal folds were produced during late stages of the first extensional collapse or early stages of the sinistral transtension (note features at Stop 9-3). This ductile transtension gradually evolved into a more focused and more brittle deformation which produced the major sinistral mylonite zones, including those at Brattvåg and at Åkre on Nordøyane. It seems likely that these were forming at the same time as the extensional emplacement of the earlier-deposited parts of the Devonian basins.

FIELD TRIP LOG FOR DAY 8

The logistics require beginning the day by taking the ferry from Lepsøya to Skjeltene, then driving on Route 659 to Brattvåg. The trip log begins where the road from Brattvåg Ferry Terminal intersects the main road.

From Brattvåg Terminal junction proceed south on Route 659 through center of Brattvåg and along west side of fjord. Near inner end of fjord note the turn off left for Helland to be used later. Stay straight, climb hill, and pass through tunnel to south slope of ridge with extensive view to Sunnmøre Alps. Descend eastward down mountain side nearly to sea level. At junction where Route 659 terminates against Route 661 that comes north from Ålesund and continues east toward Vatne, stay straight (east) on Route 661 toward Vatne. After crossing bridge to center of Vatne stay straight (east) away from Route 661 and the fjord on road with bends to the south onto mountain pass road to Fikksdal. The pass, above timberline, is reached after about 15 minutes of driving, and there is a view of the narrow peak, Mellen. Descend northeast to Fikksdal back near the shore, and turn left (toward Rekdal) on the coast road (again Route 661) which trends northwest (Fig. 8.4). After some kilometers the road turns moderately steeply upward toward the west before a road cut. Do not go around the bend but descend to right for 100 meters on a farm road to farm yard with walking access to the coast at Dragneset. Ask permission to park and pass through fields. Descend northeast through field as quickly as possible to north-facing shore. Walk east along shore to outcrops.

STOP 8-1: (30 minutes) Dragneset. Blåhø Nappe and Sætra Nappe, Moldefjord Syncline.
The first rocks encountered are porphyroclastic garnet-biotite schists and garnet amphibolites typical of the Blåhø Nappe with subhorizontal lineation and sinistral tails on porphyroclasts. Exercise care because rocks are slippery. The third or fourth large outcrop contains a buff weathering calcareous quartzose schist with a pronounced cleavage oblique to the principal foliation. If interpreted as an S-C relationship, dextral shear would be implied. However, close examination and thin sections show that the oblique cleavages are extensional shear bands completely consistent with sinistral shear.
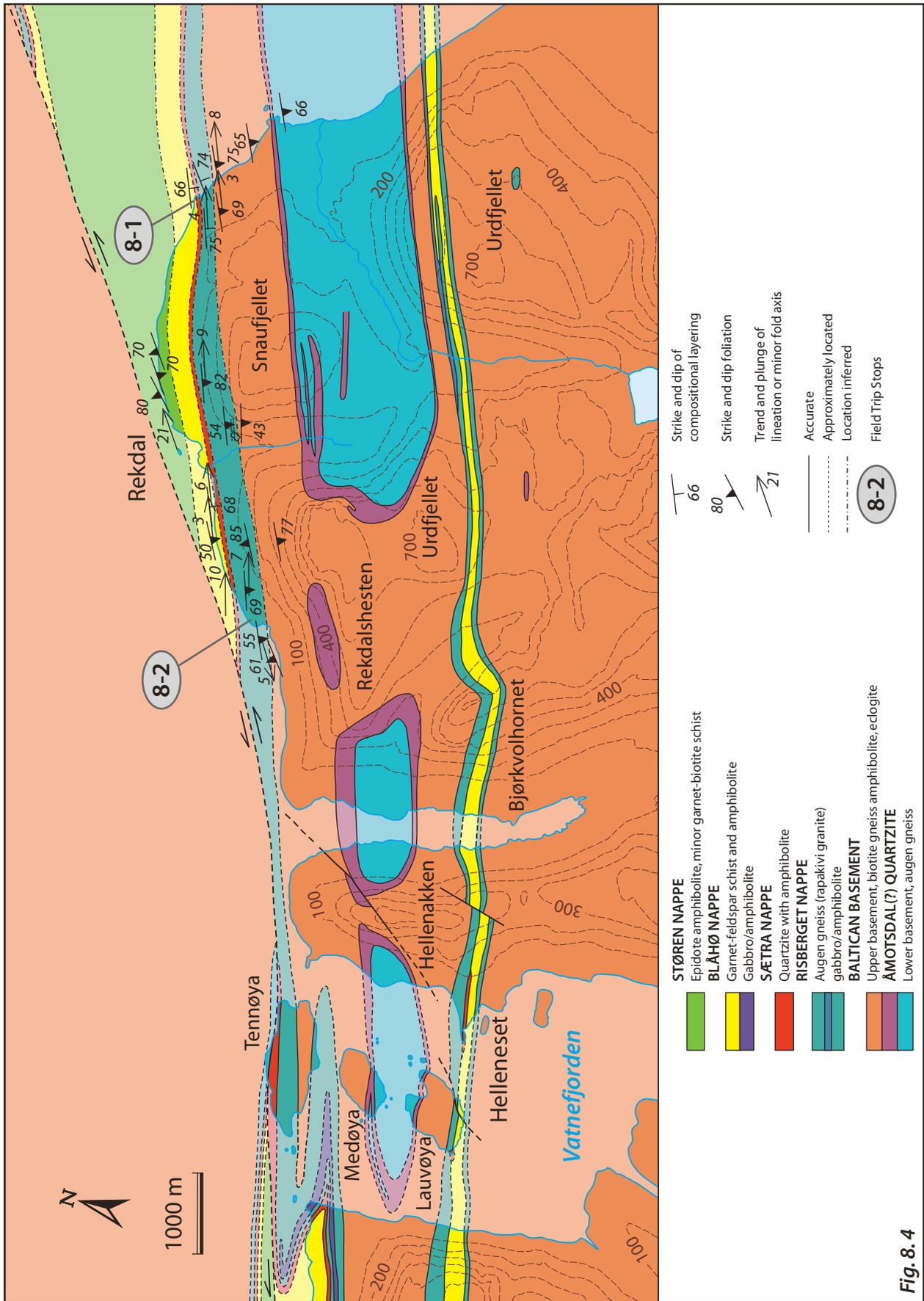
Fig. 8.4

   Beyond the last schist outcrop there is a narrow covered interval and then the first outcrops of quartz-rich quartzite and amphibolite typical of the top few meters of the Sætra Nappe. Beyond this around the corner is a pavement showing a lower part of the nappe with more biotite and feldspar in the quartzites, and with some parts consisting of more than 80% amphibolite. The amphibolite pods are varied and include some with scarce garnet, and others with plagioclase grains that are probably relict phenocrysts. Despite deliberate search, no eclogites have been found. A measured section of 34.6m from this outcrop is provided in Hollocher et al. (2007a) along with major- and trace-element analyses of 6 samples.

Return to vehicles at farm by the same route. Turn sharp right (northwest) and pass road cut in Sætra Nappe on left. Continue west over low hills and past farms to village of Rekdal (Fig. 8.4). Just offshore from the boat harbor is a large rock, which is composed of Støren amphibolite. From the mountains above, this rock has the shape of a shrimp (reke) giving the village its name, meaning "Shrimp Valley". The steep mountain, over 700 m high, southwest of the village contains the core of a huge anticline formed mainly of basement rocks, but outlined by a layer of late Proterozoic Åmotsdal Quartzite (see Stop 8-3). At the anticline crest the quartzite is over 100 m thick, but thins to as little as 1m on the south limb. Directly above Stop 8-2, on the north face, the quartzite in the core emerges as a local structural window. The steep mountain is named Rekdalshesten, which translates to "the horse of shrimp valley". From the village, a side road ascends Rekdal, passing a local soccer stadium along which there are outcrops of the Risberget augen gneiss and the underlying basement to the south. This was the training ground of Kjetild Rekdal, who scored for Norway in the 1991 World Cup in USA, and was recently the coach of a German football club. Continue west several kilometers from Rekdal to a region where the road enters a pine forest and begins descent. As road descends westward there is a rough dirt track descending steeply right toward the shore to a cultural monument. Park on the main road and walk down gravel road past ruined buildings to the gravel beach. The ruined buildings are what is left of a prison camp during 1940-1945 where prisoners were kept to work on a giant gun platform, the ruins of which can be seen on the mountain to the southeast. Local Norwegians say that the German commandant was kinder to his charges, mostly Russian and Yugoslavian, than was common in such camps. Walk north along the beach to the obvious outcrop and climb over to best part on the north side.

STOP 8-2: (20 minutes) Øygaardsneset. Augen gneiss of Risberget Nappe.
   This exposure (see Fig. 8.4) shows pink-weathering augen gneiss of the Risberget Nappe with typical strong subhorizontal stretching lineation and very weak steep foliation produced by constrictional strain during transtension associated with late Scandian exhumation. On the north side of the outcrop are adjacent horizontal and vertical surfaces. The horizontal surface shows both the lineation and sinistral shear tails around orthoclase porphyroclasts. On the vertical surface the rock gives an initial impression of being a little deformed porphyry and the foliation is detected with difficulty. Also on the northern side of the outcrop are some deformed mafic layers showing subhorizontal folds in foliation, one or two of which may be tubular. Some of the mafic layers have geochemistry identical to metamorphosed diabase dikes of the Sætra Nappe. Look up to exposures high on Rekdalshesten described earlier.

Walk back to vehicles by the same route. Drive west, passing signed boundary between Vestnes and Haram Kommunes. After several kilometers road turns left (south) around the west side of the mountain and into a long loop around the very shallow Vestrefjord. Some of the steep outcrops on the mountains on either side of Vestrefjord are composed of Åmotsdal Quartzite and steeply dipping nappe rocks in the Helleneset syncline. The road eventually comes back to the Moldefjord coast west of Vestrefjord and into the area of Figure 8.5 near Ørsnes. From here there is an excellent view of the islands of Vatnefjord, Tennøya, Medøya and Lauvøya, and Baraldsnes beyond (see Stop 8-5). The thick zone of Sætra Nappe on the north side of Tennøya is where Robinson first discovered eclogite pods in August 1990. These are now shown to be identical in geochemistry to amphibolites interpreted to be metamorphosed diabase dikes throughout the Sætra Nappe (Hollocher et al. 2007a). Also at the extreme north edge of Tennøya this Sætra is separated by mylonite from narrow exposures of Risberget and basement that are the only known exposures belonging to the north limb of the Moldefjord Syncline west of Fannefjord Tunnel (Robinson 1995c). Follow road south a short distance along west shore of Vatnefjord to parking space on right near blasted exposure of feldspathic quartzite on left.

STOP 8-3: (10 minutes) Åmotsdal Quartzite, north limb of Rekdalshesten anticline.
   This is the only easily accessible outcrop of the quartzite which outlines the core of the Reksdalshesten anticline (Figure 8.5). Here it is in a section about 100 m thick on the north limb. Unlike the fine-grained Sætra biotite quartzites, this contains quartz and feldspar fragments that were probably fine pebbles. The dominant mica is muscovite and small flakes of hematite are present, suggesting the protolith was a red sandstone and conglomerate. The basement overlying this quartzite, in probable thrust relationship, contains rare eclogites as seen at Stop 8-5, but none are known in the basement augen gneiss below. In addition, lenses of unusual hornblendite occur in this lower basement and the overlying quartzite. Some of these might be altered late Proterozoic alkaline mafic dikes.

Continue south along east side of Vatnefjord. Eventually there is a point on the road where two farms stand on high ground to the west (right) of the road and there are mailboxes also on the right. Just beyond the mail boxes is the steep access road to Helleneset Harbor and the main road turns left up a hill. Park as carefully as possible near the mail boxes. From the mail boxes walk directly up a private driveway toward the front door of the nearest house (beautiful stone construction) and ask permission. Walk west across field behind house (taking care with crops) to reach farm road and track that descends to shoreline on south side of outer point of Helleneset. Once on shore

outcrop walk west a short distance (don't look because you will be confused) to distinctive outcrop of Risberget augen gneiss where the group will gather..

STOP 8-4: (45 minutes) Helleneset syncline showing Risberget, Sætra and Blåhø Nappes

The stop begins on a pavement exposure of familiar augen gneiss of the Risberget Nappe, which lies on the north limb of the Helleneset syncline. Asymmetric tails on feldspar porphyroclasts indicate sinistral shear. Walking southward and slightly back toward the way we came, we cross grain-size reduced feldspathic rocks with rare vestiges of orthoclase porphyroclasts, that are interpreted as mylonitized equivalents of part of the Risberget Nappe. Eventually we reach superficially similar looking but more quartz-rich rocks with amphibolite layers and in the next meter or two these become even more quartz-rich. At the base of this thin unit, assigned to the Sætra Nappe, is a pod of fine-grained eclogite. Along strike from this at the southwest tip of Lauvøya there is an eclogite pod at least 5 meters across, where the rest of the nappe is only 1-2 meters thick!!! This pod has identical geochemistry (Fig. 8.2B,F,K) to other mafic layers in the Sætra Nappe interpreted as metamorphosed diabase dikes (Hollocher 2007a, 2007b). Here the quartz-rich quartzites give way to rusty-weathering garnet schist and garnet amphibolites of the Blåhø Nappe that forms the center of the Helleneset syncline. This laminated schist and amphibolite is beautifully exposed on a slanting surface just beyond where our trail emerged on the shore. At one point a few meters east from the small Sætra eclogite pod, one can stand with one foot on the top of the Risberget Nappe and one on the bottom of the Blåhø Nappe, with the Sætra just fitting between.

From the eclogite pod walk west through a bouldery area toward the outer point to a very clean photogenic waterline exposure showing the entire Sætra Nappe. Several amphibolite layers are broken into boudins with interesting shear features. One group can be interpreted as tipped "bookshelf style" in sinistral shear, or alternatively as tipped on extensional shear bands in dextral shear. "Experts" have given "definite" and totally conflicting opinions on this. A few other outcrops in the Helleneset Syncline are among the very rare places in the region where evidence of dextral shear has been found.

Re-enter trail and return to vehicles by the same route. Continue south along Vatnefjord to village of Vatne. Bear right there (still on Route 661), go west across river and reach T junction with Route 659. Stay straight (west) toward Brattvåg on Route 659. Pass through tunnel and descend nearly to sea level to junction for Helland. Turn right at junction (there is a small roadside quarry to right) and drive north along east side of Brattvåg inlet. Continue through village of Helland (see Fig. 8.8). Pass around corner and continue east (road narrows but pavement continues) to historic village of Skår (Fig. 8.5). Continue east from Skår across barren land past isolated building left from World War II to entrance to private driveway to Baraldnes. Park here without blocking driveway (it may be difficult with several vehicles).

STOP 8-5: (45 minutes) Traverse from Baltica basement with eclogite across Risberget and Sætra Nappes to Blåhø Nappe, south limb of Moldefjord syncline.

Walk northeast on Baraldnes Road then right (east) to prominent mound 75m east of road. Pavement exposures of biotite eclogite with 4 cm omphacite porphyroblasts, now mostly plagioclase - Ca-pyroxene symplectite. This eclogite also contains hornblende-plagioclase partially replacing garnet, and abundant plates of phengite rimmed by a coarse intergrowth of biotite and feldspar, the classic dehydration product of phengite. No probe analyses yet completed.

Walk north to road and farther northeast observing typical stripy gneiss of Baltican basement with thin amphibolite layers, eventually reaching sharp contact with base of Risberget augen gneiss. Contact with basement is perfectly exposed on southeast side of road. Leave road and walk west across pavement exposures of Risberget augen gneiss to contact with gabbro/amphibolite, observing abundant west plunging folds that are sympathetic to the Baraldnes syncline. Cross grassy area near west fence of Baraldnes farm with exposures of Risberget amphibolite and turn northwest toward crest of highest ridge. Observe upper augen gneiss member of Risberget Nappe just below ridge crest, then strongly sheared feldspathic gneiss in contact with interlayered impure quartzite-amphibolite-eclogite of the Sætra Nappe. At ridge crest are two eclogite boudins. Peeled off area displays folds in quartzite that apparently nucleated on walls of mafic dikes, now eclogite. Just west of crest are outcrops of Blåhø mica schist and amphibolite that lies close to the hinge of the Baraldsnes syncline. To the northeast at sea level it is possible to see large outcrops of the Risberget Nappe, particularly the gabbro/amphibolite member, on the northeast limb of the Baraldsnes syncline. The eclogites found in the Sætra Nappe, as well as one found in the Risberget Nappe, have the following characteristics (Robinson & Panish 1994). (1) All of the former omphacite is now a fine-grained reaction symplectite of pyroxene plus sodic plagioclase An 9-17 produced by unloading. The most sodic pyroxene in the symplectites, clearly modified from the original omphacite has the composition indicating a minimum 15% jadeite component. (2) Chemical zoning in the eclogite garnets shows high Mn cores and low Mn rims suggesting prograde eclogite-facies growth zoning with only minimal retrograde resorption. This is also demonstrated by abundant hornblende and zoisite inclusions in garnet, whereas relict omphacite occurs only in the matrix. Garnets in several samples show complex growth zoning with an abrupt increase in grossular content from Gr 17 Py 18 Sp 3 Alm 61 to Gr 31 Py 15 Sp 2 Alm 52 all with progressively decreasing spessartine and increasing ratio of Mg/(Mg+Fe). Preserved prograde growth zoning suggests these cover eclogites probably never reached the temperatures of 750°C and expected high diffusion rates reported for basement eclogites north of Moldefjord, including Nordøyane (Krogh 1980; Griffin & Carswell 1985; Mørk 1986; Terry et al. 2000a), a fact to be included in any tectonic reconstruction. 3) These eclogites contain relict coarse phengite with Si=3.25 comparable to phengites in eclogites from elsewhere in Norway. Phengite is always rimmed by biotite, even
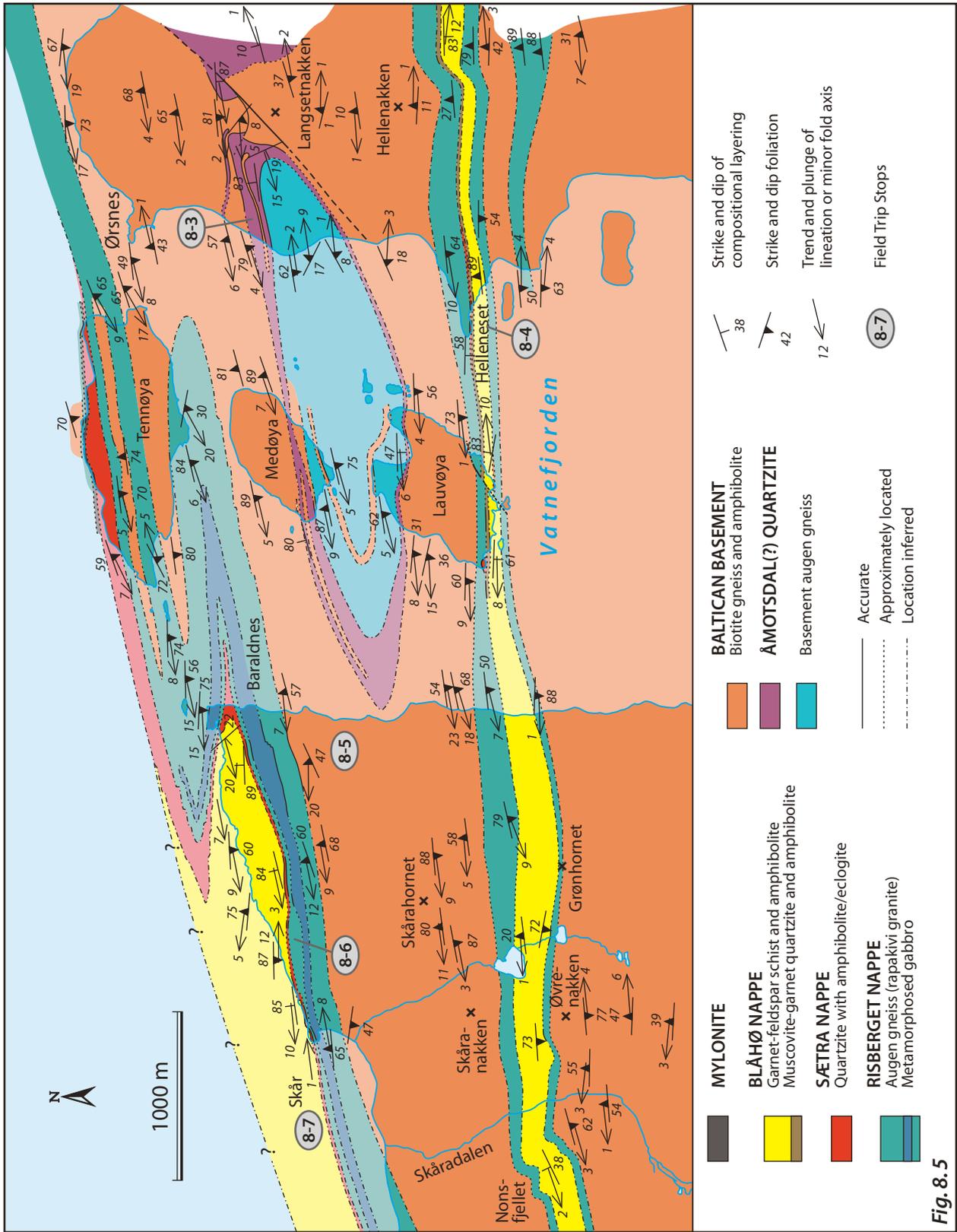
Fig. 8.5

where included in garnet. Detailed study of a garnet-phengite contact zone shows the biotite formed by an important <u>fluid-conservative</u> reaction: garnet + phengite = biotite + more anorthitic plagioclase, that can be calibrated for quantitative estimates of pressure. A critical problem in reconstructing the tectonic evolution of these eclogites and their enclosing rocks is to learn their age, but that will not be easy.

Return to vehicles by the same route. Drive west back toward Brattvåg. After 0.9 km park on gravel patch to left. This is just short of the first trees. The parking may be very tight. Cross into field on right and walk northwest to pavement exposures.

STOP 8-6: (20 minutes) Pavement outcrops of Rapakivi granite, augen gneiss.

Traverse begins in the gabbro/amphibolite member of the Risberget Nappe and proceeds north into strongly sheared augen gneiss of the upper member the same as in the outcrop at Skår (Fig. 8.5). The outcrop exposes 63.8 meters of this upper augen gneiss as compared to the total thickness of 15 m at Skår, as well as 7.6 m of the overlying Sætra Nappe with eclogite boudins. Follow foliation northwest and then walk southwest onto large red pavement. Within these exposures there is a lens, 24.3 m thick, of completely undeformed rapakivi granite with round pale red-brown orthoclase phenocrysts with preserved internal igneous zoning and with plagioclase mantles. Such a preservation in granitoid rocks is unusual across the Caledonides in the Risberget and Tännäs Nappes generally (Krill & Röshoff 1981), and is particularly surprising for this region where there have been many intense deformations. Although the rapakivi granite is undeformed, it is metamorphosed, with fine-grained matrix garnet, clinopyroxene and biotite. In thin section, the white "wiborgite" rims are composed of twinned plagioclase crammed with fine white needles that may be zoisite. Detailed petrographic examination of this rock will require electron microscopy. Does the matrix assemblage reflect an old granulite facies metamorphism as implied at Flatraket (see Selje Guidebook for 2003) or a partial eclogite-facies overprint reflected within the adjacent Sætra Nappe? Observe the extreme strain gradient with evidence of sinistral shear at the north edge of the granite boudin. You may also wish to walk to the north edge of the outcrops to see retrograded eclogite boudins in the lower part of the Sætra Nappe.

Return to vehicles by the same route. Continue west to parking space in view of inlet and point at Skår. There is very tight parking on both sides. Do not block driveways.

STOP 8-7: (40 minutes) Outcrop at Verpholmen, Skår. Upper part of the Risberget Nappe, the entire Sætra Nappe, and base of the Blåhø Nappe in continuous exposure.

The beach outcrop in the bay at Skår, lies in the lower part of the Risberget augen gneiss just north of the poorly exposed basement contact (Figure 8.5). The outcrop surface is eroded perpendicular to steep foliation and parallel to subhorizontal mineral lineation, and shows asymmetric tails on rounded relict orthoclase phenocrysts indicating sinistral shear parallel to the lineation. Thin and less common thick amphibolite layers are interpreted as deformed metamorphosed diabase dikes, possibly like those in the Sætra Nappe. In the high mountain valley directly east of Skår, a single boudin of fine-grained eclogite has been found on the south limb of the Helleneset syncline similar to the eclogites in the Sætra Nappe.

The southern part of the main outcrop consists of 44m of massive amphibolite, interpreted as metamorphosed gabbro, part of a continuous layer that was seen previously at Stops 8-5 and 8-6. Detailed geochemical study (Fig. 8.2) shows that most of these mafic rocks are geochemically unlike the Late Neoproterozoic diabase dikes of the Sætra Nappe, and are more likely Mesoproterozoic mafic rocks of similar age to the augen gneiss, but there is one distinctive boudin that is identical in geochemistry to the Sætra dikes (Hollocher et al. 2007a) This is succeeded northward by an additional 15m of Risberget rocks of which 10m near the base is rather massive phenocryst-rich augen gneiss in boudins in which the shear-elongated phenocrysts show strongly developed steep foliation and a lineation plunging 15° west.. This is the layer from which Handke et al. (1995) obtained a U-Pb zircon age indicative of igneous crystallization at 1190.3±2.8 Ma (Fig. 8.6). The upper 5m is fine-grained mylonitic gneiss with a variable abundance of relict phenocrysts and thin dark streaks that may represent mylonitized mafic dikes. North of the Risberget is 9-10m of the Sætra Nappe, of which the lower 7.5m is metamorphosed feldspathic sandstone with minor amphibolite and with typical subhorizontal folds. This lower part of the Sætra Nappe is typically more feldspathic than the upper part and is typically the host for retrograded eclogite boudins, two of which are exposed here, less than 1m north of the Risberget contact. The upper 1.5m of the Sætra is cleaner, more quartz-rich metamorphosed sandstone with more abundant amphibolite laminae near the base of the Blåhø Nappe. The base of the Blåhø Nappe against the Sætra consists of garnet-mica schist and amphibolite, some in boudins. Exactly on the contact in an eroded groove is 5-15 cm of calcite marble and diopside calc-silicate that commonly occurs in this position. A second calcareous layer is a thin, pitted, impure marble north of amphibolite boudins about 1.5 m above the contact.

Walk a short distance north to a "nest" of late steeply plunging sinistral folds that deform earlier subhorizontal folds and lineations. This exposure "freezes" a moment in the continuous process of fold development in a regime of extreme sinistral transtension. From the outer end of the point at Skår under good lighting conditions view the cliff of Skåradalen (Fig. 8.7) with its exposure of the refolded Helleneset Syncline. Return to vehicles by walking back around the point.

Leave Stop 8-7 and drive west toward Helland. After 3.8 km there is a big curve where the direction of the road changes from westward to southward. 0.6 km beyond curve there is a junction with mail boxes. Turn very sharp
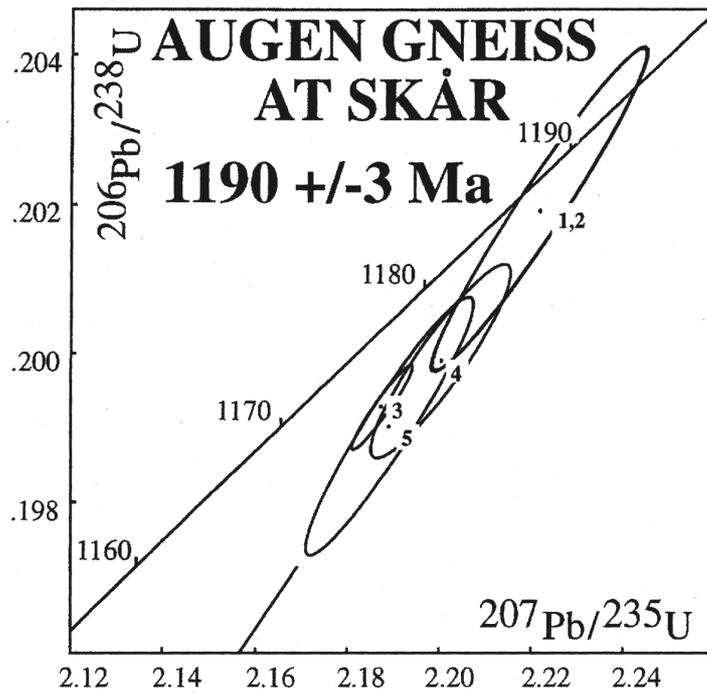
Fig. 8.6 Concordia diagram showing zircon U-Pb isotopic ratios and igneous crystallization age, determined by R. D. Tucker, for augen gneiss in the upper part of the Risberget Nappe exposed in the outcrop at Verpholmen point, Skår (Stop 8-7).
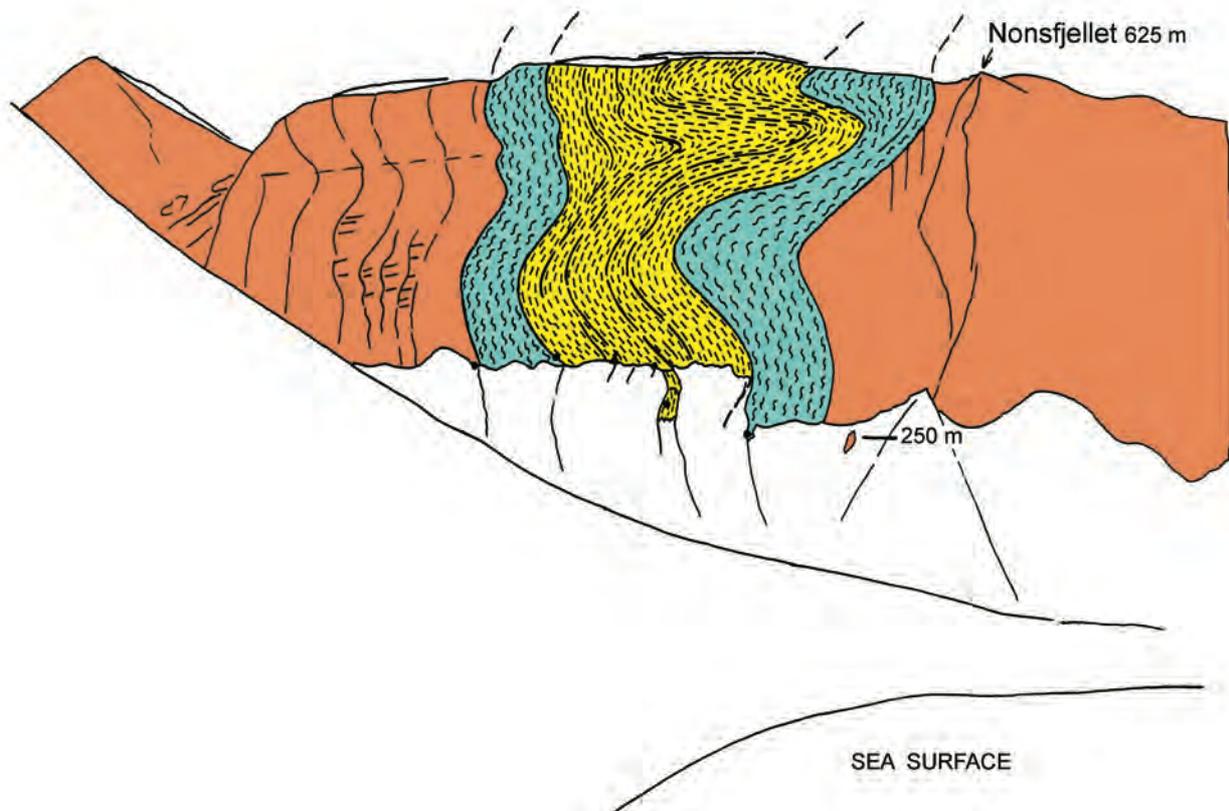

Figure 8.7 Field sketch made at the outcrop at Skår looking at the north-east facing cliff above Skåradalen. This shows the Helleneset syncline, a deep isoclinal fold of cover nappes into basement. The Blåhø Nappe along the axial surface of the fold is flanked by the Risberget Nappe. Elsewhere a very thin Sætra Nappe is present between (for example at Stop 8-3). The axial surface of the syncline is refolded about sub-horizontal axes and axial surfaces. Dots indicate locations that could be reached from structural measurements at the top of the steep talus slope below the cliff.

right by mail boxes and drive north on east coast of Brattvåg inlet. After 0.7 km there is a parking lot for a public beach. From beach walk south over outcrops of Blåhø Nappe to low pavement outcrop of Sætra Nappe.

STOP 8-8: (30 minutes) Helland Traverse: Sætra, Blåhø and Støren Nappes and southern ultramylonite belt.

The strata in the northern part of Figure 8.8 (and inset) belong to the main steep to northward-overturned south limb of the Moldefjord syncline, divided into three repeated sequences by two zones of ultramylonite formed during sinistral shearing. The sequences are complicated by local subhorizontal folds that are *en échelon* to the mylonite zones and truncated by them. The southern sequence contains all units from basement up to Støren, including the marble-bearing upper part of the Blåhø Nappe. The narrow middle sequence between the mylonites consists only of Sætra, Blåhø, and Støren. The northern sequence begins with sheared gabbro in the middle of the Risberget Nappe (see Stop 8-11) and extends well up into the Støren Nappe. In this traverse we will begin with a pavement exposure of the Sætra Nappe and walk across the Blåhø and Støren Nappes to the southern ultramylonite belt.

From the parking lot walk south along the coast to pavement exposure of Sætra Nappe consisting of laminated quartzite and amphibolite. The quartz-rich quartzite is typical of the uppermost meter of the Sætra nappe close to the Blåhø-Surna contact. A thin layer of marble is exposed at the north edge of the outcrop and is typical of the contact region between the Sætra and Blåhø nappes.

Walk back north past public beach to large exposures on peninsula. These are Blåhø mica-garnet schist and amphibolite with pegmatitic layers produced by intrusion and/or partial melting. This is typical of the lower part of this nappe where there is abundant shearing, but less late metamorphic reconstitution than close to the Støren Nappe.

Continue north along steep rock slope to complex contact with epidote amphibolites of the Støren Nappe. The last exposed parts of the Blåhø Nappe consist of mica schist and marble. The complexity of this folded contact zone is illustrated by the following measured section beginning with the first marble in the continuous outcrop from the peninsula: marble 1.8m, garnet schist 4.5 m, low-grade amphibolite 6m, schist 0.9m, marble with wild folds 2.4m, mica schist 4.5m, impure marble 1.5m, low-grade amphibolite 10.6m, garnet schist 3.6m and then low-grade amphibolite 7.6m to end of outcrop. The islet to the northwest appears to be all low-grade amphibolite of the Støren Nappe, plus one layer of metamorphosed gabbro, and was sampled extensively for geochemistry.

There is a gap in the outcrop of about 100 meters. Then walk east through industrial area with an artificial exposure of the full 27 m width of the southern belt of ultramylonite. Ultramylonite protoliths are impossible to determine without thin sections. Rock south of the ultramylonite is low-grade amphibolite. The rock to the north appears to be retrograded Blåhø mica schist and impure marble for about 5 m, then also low-grade amphibolite.

From this point walk south along industrial road back toward cars. At road junction walk northeast to another large man-made exposure showing Støren amphibolite to the south in contact with ultramylonite to the north with folds in mylonitic foliation. Elsewhere mylonites show tubular folds.

Drivers will pick up participants at this mylonite outcrop for the drive to Stop 8-9. Drive north on very narrow road. Stop near east end of jetty and find parking. Walk short distance east along north-facing coast.

STOP 8-9: (20 minutes) Layered cumulate gabbro and epidote amphibolite of the Støren Nappe north of ultramylonite belts.

The lower part of the Støren Nappe near Brattvåg (Fig. 8.8 and inset)) is dominated by fine-grained hornblende-epidote amphibolites, and these are overlain by mappable units of massive to layered hornblende gabbros locally with what appears to be relict cumulate layering. At one point 200 m east of Stop 8-9 there is a 1 m-thick layer of green amphibole that is probably a metamorphosed pyroxenite cumulate layer in the gabbro. These features suggest the possibility that these gabbros represent sheared fragments of an ophiolite like the Løkken and Bymarka units of the Støren Nappe near Trondheim (Grenne 1989a, 1989b, this guidebook Stops Days 4, 6). They also closely resemble in character and setting the gabbro slices studied by Boyd (1983) near Narvik, which he considers, on lithic and geochemical grounds, to be sheared fragments of the Lyngen Ophiolite. At Myrakaia (west edge of Fig. 8.8) within Støren amphibolites there is a narrow and poorly exposed lens of talc-anthophyllite-carbonate rock (see Fig. 8.3D,G) that might agree with the ophiolite hypothesis. Geochemical study of the Støren gabbros and the amphibolites, collected here and elsewhere in the Moldefjord Syncline (Hollocher 2007c and in preparation) indicate a cumulate origin for the gabbros and a likely MORB magmatic composition for the amphibolites.

North of the gabbro on the coasts north of Helland and Sunnaland there is an extremely narrow belt of garnet-mica schist. In coastal exposures northwest of Sunnaland this narrow belt contains garnets up to 3 cm in diameter, relict migmatite features, and calcareous schists, all suggesting that this is a very thin tectonic slice of Blåhø rocks. A similar very thin slice of Blåhø rocks has been mapped on Tautra and on Bolsøya over a total length of 40 km.

From parking at end of jetty walk east across boulders to pavement exposure of cumulate-layered but strongly sheared metamorphosed gabbro. Northeast beyond this is typical Støren epidote amphibolite, commonly with minor calcite. The extreme northernmost outcrops, accessible mainly at low tide, contain the narrow belt of garnet-mica schist interpreted as an early thrust slice of Blåhø rocks.

Fig. 8.8

MYLONITE

STØREN NAPPE
Thin slice of garnet-biotite schist (Blåhø)
Metamorphosed gabbro
Epidote amphibolite, minor garnet-biotite schist

BLÅHØ NAPPE
Marble and interbedded schist
Garnet-feldspar schist and amphibolite

SÆTRA NAPPE
Quartzite with amphibolite/eclogite

RISBERGET NAPPE
Augen gneiss (rapakivi granite)
Metamorphosed gabbro

BALTICAN BASEMENT
Biotite gneiss and amphibolite

38   Strike and dip of
     compositional layering

42   Strike and dip foliation

12   Trend and plunge of
     lineation or minor fold axis

———   Accurate
·······   Approximately located
- - -   Location inferred

8-7   Field Trip Stops

Reverse direction with difficulty and drive south to main road and turn right. After 5.6 km reach junction with Route 659. Turn sharp right toward Brattvåg. Pass through Brattvåg and past the junction for Brattvåg Ferry Terminal. After a long road cut in Risberget augen gneiss, the highway road bears left (west) through more road cuts in a by-pass of the coastal road. Stay on by-pass to next paved exit where old coast road joins up. Turn very sharp right and return east along old coast road. After a short distance there is ample roadside parking just west of a walking access road north to houses and the beach (muddy) near Stops 8-10 and 8-11 (Fig. 8.8 inset). Just near access road on the right (south) of old coast road is outcrop at base of telephone pole containing a layer of ultramylonite.

Walk north on access road and past cottage to muddy beach with stone wall. Walk north along beach and across pasture to outer coast, noting rusty Blåhø outcrops with pitted weathered layer to east. Stop on wave-washed outcrop on outer coast.

STOP 8-10: (20 minutes) Tectonic contact between Blåhø and Støren Nappes at Sunnaland.
The gray to white rock with vertical foliation exposed on the coast is a strongly mylonitized felsic volcanic rock (Hollocher et al. 2007c and in preparation) that constitutes of lowest rock in the Storen Nappe in this vicinity. To the south it is in contact with a mylonitic garnet schist of the underlying Blåhø Nappe. A few meters to the north at the water line it is overlain by normal fine-grained amphibolite. Walk west along the white rock to a "nest" of complex folds that refold older folds. The later folds are of diverse orientation, but all indicate a pattern of sinistral shear. The folds appear frozen in a process by which axial orientations were being progressively deformed toward parallelism with a subhorizontal transport direction during late amphibolite-facies exhumation. A few meters to the east where bedding is very well developed, one can observed brown layers of very fine-grained garnet-tourmaline schist that seems to be typical of some of the very few garnet-bearing rocks contained in the Støren Nappe in this vicinity and the tourmaline may have a volcanic exhalative origin. Along the northern contact of these felsic rocks and the overlying amphibolites there are several layers containing ultra-fine-grained garnet and finely divided magnetite that may have a similar origin.

With continued walking to the east the coast bends south slightly and one passes onto Blåhø rocks. In the central part of a broad depression the schist layer with pitted weathering is well exposed and proves to be highly calcareous, one of a group of persistent calcareous horizons near the top of the Blåhø Nappe in this vicinity. Walk north a few meters onto projecting shore rocks to the northernmost prominent exposure of well layered Blåhø schist. The south face of the outcrop shows the typical strong subhorizontal stretching lineation. The top of the outcrop shows countless small sinistral folds in foliation with steep axes, another example of folds forming during late deformation that have not been rotated into parallelism with the subhorizontal transport direction.

When finished with Stop 8-10 walk all the way back along the outcrop and then south to the south end of the muddy beach. From here walk the grassy and muddy shore west to the first shore outcrop.

STOP 8-11: (20 minutes) The Final Exam - Sunnaland.
The sequence of rocks exposed here is unique in the Brattvåg region because it contains the lower part of the north-facing tectono-stratigraphic sequence on the south limb of the Moldefjord syncline, but in an isolated repeated position on the north side of the Brattvåg mylonite zones. Coming after a long day of "training" on these rock units, it has served, since the visit of Simon Cuthbert in 1993, as the "Final Exam" (he passed!).

The first coastal exposures are of highly convoluted schist and amphibolite of the Blåhø Nappe. At the highest point of the outcrop the mica schist with sub-horizontal lineation contains canted muscovite plates in an S-C fabric indicative of sinistral shear (the famous "fish flash" can be observed on oriented hand specimens under favorable lighting conditions). Immediately south of this one can identify both pure and impure quartzite with amphibolite of the Sætra Nappe (recently confirmed to contain typical geochemistry of other Sætra dikes, Hollocher et al. 2007a), and then highly sheared augen gneiss and mylonitic gneiss of the Risberget Nappe. The southernmost outcrops in this extremely attenuated sequence are massive amphibolites assigned to the metamorphosed gabbro middle member of the Risberget Nappe (two analyses were done, one showing Sætra geochemistry, the other not).

Following Stop 8-11 walk back north along the grassy muddy shore to access road and back up slope to paved road and vehicles. Drive east about 500 m. En route pass a road cut in mylonite on the right. Here a tubular fold in mylonitic foliation was collected. Careful diamond saw work allowed it to be shown that the tubular fold was produced during sinistral shear. In the event of limited time, this road cut may be substituted for the official stop. Park opposite a driveway to a house that leads backward to the right (southwest). The owner has cleared an outcrop for a rock garden at the top of the driveway.

STOP 8-12: (10 minutes) Brattvåg mylonite zone.
The mylonite and mylonitic gneiss at this outcrop consists of intricately interleaved and interfolded augen gneiss and metamorphosed gabbro of the Risberget Nappe. It is illustrated in a photograph in Robinson (1995c), but the outcrop has become somewhat overgrown since then. In the top of the outcrop sinistral tails around feldspar porphyroclasts can be seen. In the east face of the outcrop a series of subhorizontal folds in mylonitic foliation was measured, and it can be shown that some of these folds are refolding the axial surfaces of earlier folds, all with the same subhorizontal orientation.. These results imply large and repeated constrictional strains, during which early folds, already parallel to the transport direction, were refolded, and the later folds were also rotated into this position.

Following Stop 8-12 or earlier, if needed, turn around and head west to junction with Route 659. Proceed west on
    Route 659 (8-9 minutes driving time) to Skjeltene ferry terminal and waiting line for ferry to Kjerstad, at 18:15, or
    19:15 and overnight at Lepsøya Misjonsenter.

Day 9: GEOLOGY OF HARAMSØY, FLEMSØY, AND LEPSØY

Saturday August 23 (potential shift to Friday in case of dire weather forecast)

by Mike Terry, Peter Robinson, Tom Krogh and Kurt Hollocher

Purpose: This day will bring into focus some of the features in the deepest part of the Scandian hinterland, on very large coastal exposures. The geology of the three islands is divided into three tectonic-stratigraphic segments. The northern segment is separated from the central and southern segments by the late Scandian amphibolite-facies Åkre Mylonite zone. A much more cryptic surface separates the central and southern segments. Our time on Haramsøy and Flemsøy will be devoted to the Åkre Mylonite and the northern segment, whereas the central and southern segments will be visited on Lepsøy.

General Route: From Lepsøya we go by ferry to Haramsøya, which is connected by bridge to Ullaholmen and Flemsøya. The ferry schedule is fairly convenient without complex connections and will allow 8-11 hours on these islands. However, the outcrops are large and intriguing, and time will pass quickly.   We will visit the Proterozoic Haram Gabbro (Stop 9-1) with superbly preserved primary igneous features and early Scandian eclogite-facies shear zones, the late Scandian Åkre Mylonite Zone (Stop 9-2), Ullaholmen (Stop 9-3) with its late Scandian subhorizontal tubular folds superimposed on earlier fabrics, the Flem olivine gabbro (Stop 9-4) intruding Mid-Proterozoic rapakivi granite and conversion of the gabbro to eclogite now well dated at 410 Ma, the Kvalvika garnet peridotite / pyroxenite (Stop 9-5) with early Scandian subduction fabric and evidence for a P-T path through 750°C- 40 kbar, Ulla Gneiss and augen gneiss (Stop 9-6) on northern Flemsøya with eclogite facies tubular folds, and the Nogva UHP kyanite-zoisite eclogite (Stop 9-7) with coesite pseudomorphs and retrograde sapphirine. There are four stops back on Lepsøya, which can be done either on a fine evening or on another morning, while awaiting a ferry. These include pavement exposures of the central segment near Hellevik lighthouse (Stop 9-8), northern Lepsøya, showing infolded Sætra and Blåhø Nappes in "Sausage Rock" basement, and three stops in the southern segment, a coarse hornblende eclogite (Stop 9-9) on an islet within walking distance of the Misjonsenter, now well dated at 412 Ma; kyanite-bearing migmatitic schists (Stop 9-10) of the infolded Blåhø Nappe and eclogites in adjacent basement of southwest Lepsøya, and a quarry in garnet corona gabbro (Stop 9-11) at Lauvsundholmen, southern Lepsøya.

Ferry Lepsøya to Haramsøya: Saturday 8:00-8.20 or 9:00-9:20 (Friday 8:00-8:15 or 9:00-9:15) (Will allow 7:00 A.M. breakfast. This ferry is rarely crowded so all vehicles can be taken. However, all vehicles must be backed onto the ferry, also on the return trip)
Ferry Haramsøya to Lepsøya: Saturday 17:35-17:50 or 20:35-20:50 (Friday 17:35-17:50, or 20:35-20:50. There is also a ferry by the long route 18:50-19:30.)

INTRODUCTION TO GEOLOGY OF NORDØYANE

Day 9 will be devoted entirely to three of the Nordøyane (literally THE NORTH ISLANDS) Haramsøy, Flemsøy, and Lepsøy (Figs. 8.1, 9.1) studied in detail by Terry (2000). The context of Nordøyane is best understood after viewing the rocks of the Moldefjord syncline near Brattvåg on Day 8, but in case of a dire weather forecast, these days may be switched. Nordøyane are the westernmost exposures of the belt of basement rocks with high-pressure high-temperature eclogites that lies to the north of the Moldefjord and Surnadal synclines with their exposures of low amphibolite-grade rocks. Fjørtoft is the site of recent reports of crustal microdiamonds (Dobrzhinetskaya et al., 1995) in a narrow belt we believe can be assigned to the Blåhø Nappe. It is also the site of microdiamonds in garnet peridotite (van Roermund et al. 2002) enclosed in basement.

Terry has divided the geology of Nordøyane into three segments (Fig. 9.1), **southern**, **central** and **northern**, based on the character of basement and nappe units in each. The **southern belt** (seen on Stops 9-9, 9-10 and 9-11) is dominated by granitoid Proterozoic basement gneiss with abundant boudins of eclogite and bodies of garnet-corona gabbro. Within this there are five narrow belts of mica-garnet±kyanite schist and garnet amphibolite that are interpreted as early isoclinal infolds of the Blåhø Nappe. Although eclogites have not been surely identified in this Blåhø on Lepsøya, eclogite has been found in a correlative synclinal belt on nearby Vigra. The belts of Blåhø provide markers that outline the late subhorizontal open folds superimposed on the early isoclines and also isoclinal refolds. These consist of an open anticline located in southwest Lepsøya (Fig. 9.2a) and a broad syncline centered on Goaldet, the high peak of Lepsøya, which is well displayed in views from Haramsøya or the ferry to Haramsøya. Asymmetric shear fabrics in the southern belt give a consistent top-to-west shear sense along the lineation that is parallel to these late folds, and is thus much like the features observed in ductile fabrics beneath the Devonian basins south of Nordfjord and below the Høybakken detachment north of Trondheim (see Fig. 5.1, and Stop 5-12).

The **central belt** is dominated by two basement types, granitoid gneiss and gray granitoid gneiss with amphibolites interpreted as deformed mafic dikes (Stop 9-8; for more about this see Stops 7-14 and 7-15). There is also a zone of relatively undeformed diorite with hornblende coronas and only minor garnet coronas. Throughout there is no evidence that these rocks ever went through eclogite-facies metamorphism, though that cannot be proved. The central belt also contains three extremely narrow isoclinal synclines dominated by mica-garnet±kyanite schists and garnet amphibolites of the Blåhø Nappe. The north limb of one of these synclines, at the basement-Blåhø contact, shows a consistent layer
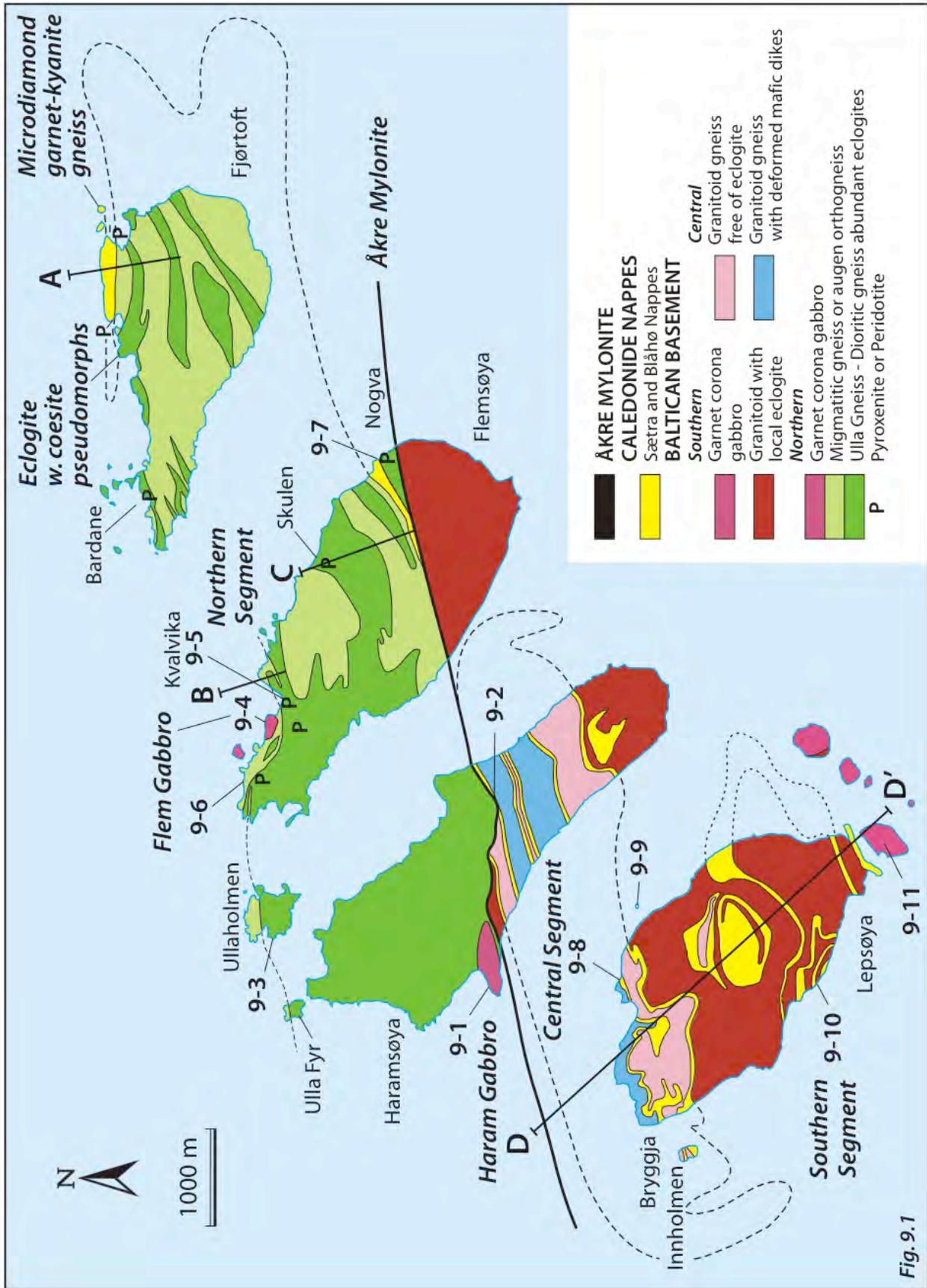
Fig. 9.1

Generalized geologic map of Nordøyane from Terry (2000) showing locations of stops for day 9.
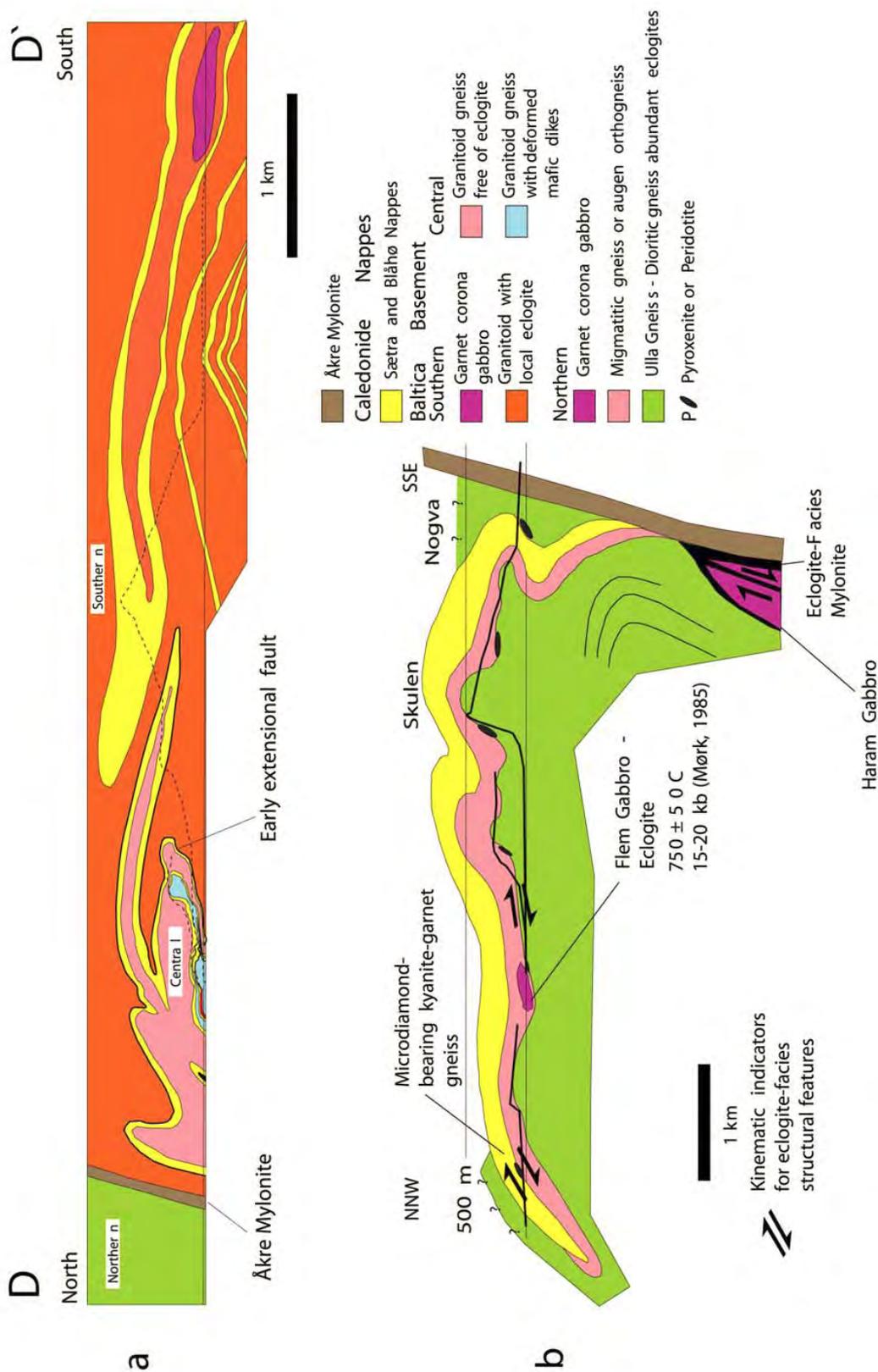
ÅKRE MYLONITE

CALEDONIDE NAPPES
Sætra and Blåhø Nappes

BALTICAN BASEMENT

Southern
Garnet corona gabbro
Granitoid with local eclogite

Central
Granitoid gneiss free of eclogite
Granitoid gneiss with deformed mafic dikes

Northern
Garnet corona gabbro
Migmatitic gneiss or augen orthogneiss
Ulla Gneiss - Dioritic gneiss abundant eclogites
Pyroxenite or Peridotite

P

Fig. 9.2 Generalized cross sections across Nordøyane. a) Section along D-D' of Fig. 9.1 showing the structure of the central and southern segments on Lepsøya and southern Haramsøya. Dashed line is a topographic profile. b) A composite section through the northern segment showing the approximate structural positions of sample locations. Heavy lines are topographic profiles along section lines A, B, and C on Figure 9.1.

about 1 m thick of interbedded quartzite and amphibolite that is assigned to the Sætra Nappe (Stop 9-8). Commonly at the Blåhø-Sætra contact there a few centimeters of marble and calc-silicate rock, a common feature at this contact as may be seen at Stops 8-7, 8-8 and 7-15. On Haramsøya limited outcrop of the central belt suggests it may be in a tight anticline so that gneiss of the southern belt lies to the north of it, close to the Haram Gabbro (Fig. 9.2a). The apparent metamorphic discontinuity between the southern and central belts must be an early feature because the contact shows no obvious fault features and is also complexly folded. Like the southern belt, the central belt is dominated by late Scandian features involving longitudinal extension, lateral constriction, and top-west shear, probably in a field of sinistral transtension (Terry & Robinson, 2003).

The **northern belt** is dominated by the Ulla Gneiss, a complex granitoid to tonalitic gneiss with very abundant boudins of eclogite and retrograded eclogite, which we affectionately call the "dog's breakfast" (Stops 9-3, 9-6). Subordinate rock types include augen orthogneiss (see Stop 9-4), garnet corona gabbro (Stops 9-1 and 9-4), garnet peridotite and garnet pyroxenite (Stop 9-5), and two belts of garnet and biotite gneiss and garnet amphibolite with eclogite, that we tentatively assign to the Blåhø Nappe, on the north coast of Fjørtoft and on eastern Flemsøya (Stop 9-7), the latter poorly exposed.   All of these rocks are arrayed in a large late subhorizontal anticline (Fig. 9.2b) with a steep southern limb and gentler undulating northern limb. Preliminary mapping of the ultramafic rocks (not shown in Fig. 9.1, but see Fig. 9.2a) suggest they may be aligned along ancient shear zones and may separate different slabs of basement that were brought into tectonic contact with sub-continental mantle within the subduction zone, before being ejected by later thrusting. The body at Kvalvika (Stop 9-5) contains a steep an eclogite-facies olivine fabric that can be directly related to similar eclogite-facies fabrics, dated by U-Pb zircon at 410 Ma, in and adjacent to the nearby Flem Gabbro. Detailed petrology of the Kvalvika body indicates a Scandian P-T path through 750°C, 40 kbar, in the diamond stability field. Like the southern and central belts, the northern belt is also dominated by late subhorizontal lineation and top-west shear fabrics, including subhorizontal tubular folds (Stop 9-3), however, it also preserves a variety of early transverse folds and lineations (Stops 9-1, 9-4), even vertical tubular folds (Stop 9-6), which we interpret as remnants of fabrics produced under eclogite-facies conditions during the subduction process. These fabrics are widely preserved in Ulla Gneiss, even where it is dominated by later subhorizontal folds and lineation (Terry, 2000; Terry & Robinson, 2003). It also appears in the "diamond-bearing" kyanite-garnet gneiss on Fjørtoft and in some of the garnet peridotites (Stop 9-5). It is best displayed in and near the Haram Gabbro (Stop 9-1) (Terry & Robinson, 1996, 2004), where early shearing has produced an essentially vertical lineation in mylonitic gneissic gabbro and fine-grained mylonite produced either from gabbro or from granitic gneiss. The gradual conversion of gabbro toward eclogite in the fine-grained gabbro mylonite implies this shearing was taking place under eclogite-facies conditions. Sense of shear observed in thin sections cut perpendicular to foliation and parallel to lineation indicate north-side-up consistent with formation during northwestward subduction of Baltica. The boundary between the northern and central/southern belts is unequivocally a zone of late subhorizontal sinistral shear with the development of about 100 meters of fine-grained mylonite including small boudins of eclogite. This Åkre mylonite (Stop 9-2), is superbly exposed in road cuts near the peak of Haramsøya, but obscures the original nature of this important tectonic contact.

FIELD TRIP LOG FOR DAY 9

Drive onto ferry for Haramsøya. Leave ferry at Haramsøya in village of Austnes. Turn left for Haram on main road.
     At Haramneset take driveways to outermost house and park. Ask owner permission to visit coastal outcrops.

STOP 9-1: (1 hour 45 minutes) Haram Gabbro
Structural setting and petrography. The Haram Gabbro (Fig. 9.1) lies in the northern belt dominated by Ulla Gneiss very close to the bounding late Scandian Åkre mylonite zone with its low amphibolite-facies mineralogy, subhorizontal lineation, and sinistral shear features.   It lies on the steep south limb of a major subhorizontal late anticline. The gabbro (Fig. 9.3) appears to be surrounded by zones of fine-grained eclogite-facies mylonite consisting mainly of gabbroic protolith, but also smaller amounts of granitoid country rock. The eclogite-facies mylonite zones also cut through parts of the gabbro (Fig. 9.4). Even though very fine grained, the eclogite-facies mylonite still contains relics of original igneous orthopyroxene, plagioclase and ilmenite in a matrix of omphacite (~ Jd 14), plagioclase (with kyanite inclusions), garnet (Pyr 22.6, Alm 45.5, Gr 30.4, Sp 1.3, FM 0.669), quartz, and minor rutile. The best estimate of P-T conditions for the mineral assemblage in the mylonite is 780°C and 18 kbar kbar (Terry et al., 2000b).

Mylonites and their fabric development. The unusual feature of the eclogite-facies mylonite (Fig. 9.4) is that it contains a vertical or near vertical foliation, a steep lineation, and indicators of north-side-up shear concordant to compositional layering/foliation in the surrounding gneisses (Terry & Robinson, 1996, 1997, 1998b, 2004; Terry 2000). When the foliation is rotated back to a subhorizontal position to cancel the effects of later folding, this gives a top-southeast sense of shear which could be compatible with fabrics produced during northwestward subduction of Baltica crust. Within the gabbro adjacent to the eclogite-facies mylonite and in some broader zones there are zones of mylonitic gabbro gneiss with the same kind of early shear features that are interpreted as segments of major high strain zones similar to those interpreted to have juxtaposed HP and UHP rocks.

Study of microstructures and textures using composition maps, orientation contrast imaging, and electron backscatter diffraction by Terry & Heidelbach (2004, 2006) in weakly deformed, lineated, and mylonitized samples indicate the following: 1) Garnet growth occurred during prograde metamorphism. 2) Both products and reactants involved in reactions forming garnet coronas show evidence for deformation in the form of a lattice-preferred
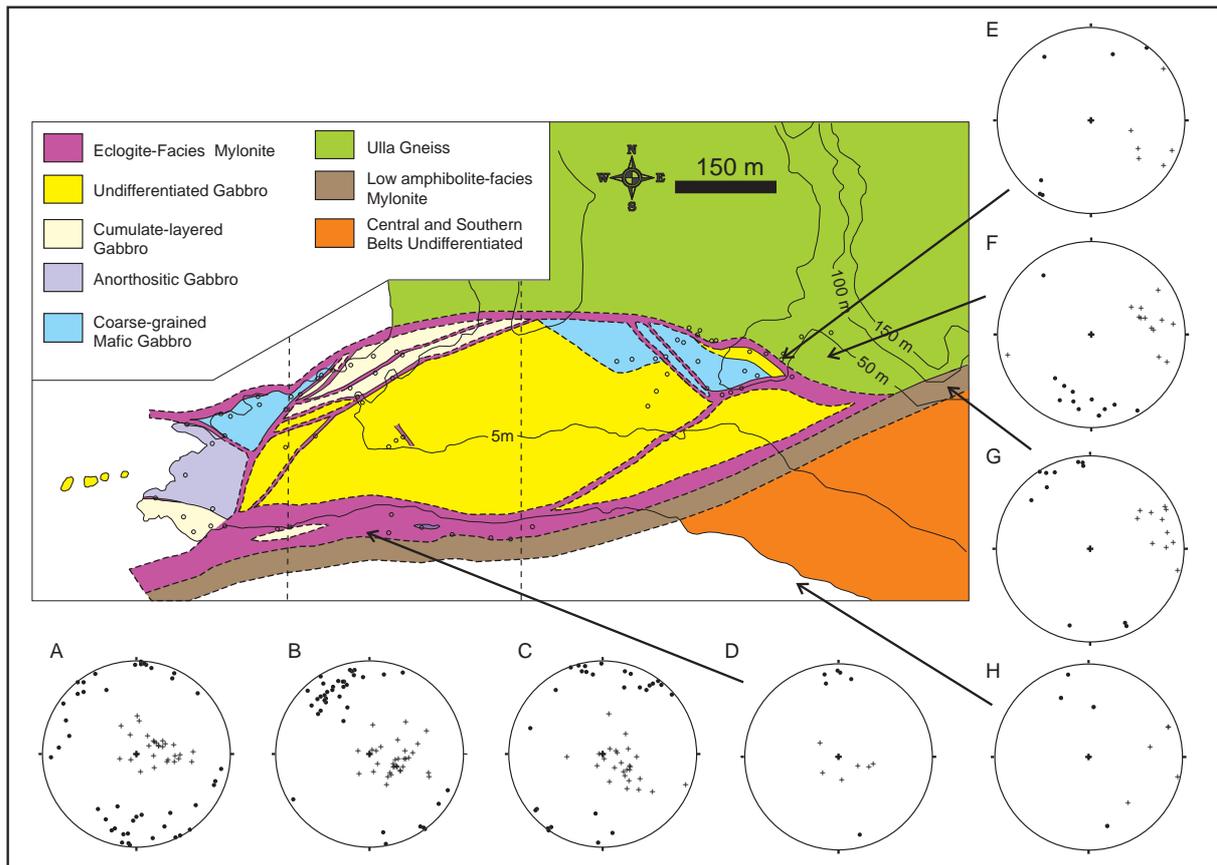
Fig. 9.3 Detailed geologic map of Haram Gabbro and vicinity (Stop 9-1) from Terry (2000). Small circles indicate areas of detailed structural measurements. Equal area diagrams show selected measurements of minor structural features: dots = foliation, crosses = lineation. Shear zones and eclogite-facies mylonites in the Gabbro (subareas A, B, C delineated by dashed lines) and the southern mylonite zone (subarea D) show near vertical lineation and, in detail, a north-side-up sense of shear consistent with early northwestward subduction. Subareas E, F, G, H, outside the gabbro and eclogite-facies mylonite, including Åkre mylonite in G, show typical gentle lineation trending E-W.

orientation (LPO) and/or preferred grain-shape orientation (Fig. 9.5 A, B, C, D) . This indicates that at least part of the garnet grew during deformation. 3). The weakly deformed sample shows classic corona structures and the systematic increase in grain size is interpreted to result from original diffusion-controlled growth (Fig. 9.5A). 4). The coarser inequigranular regions adjacent to plagioclase layers on the outside and retrograde plagioclase on the interior are interpreted to have resulted from surface area reduction during retrograde diffusion (Fig. 9.5 B). 5) Garnet layers in deformed samples show bimodal grain size (Fig. 9.5 C, D) that records continuous garnet growth during and after high-strain deformation. The combination of microstructures seen in the fine-grained part of garnet layers, and the general lack of a lattice-preferred orientation, indicate that grain-boundary sliding was the dominant deformation mechanism in garnet. The coarse-grained part of the garnet layers (Fig. 9.5 C, D) and the Ca-poor plagioclase grew after high-strain deformation (Fig. 9.5 D). Preservation of these fine early deformational features is thought to be related to little access of fluids during any later stages of deformation, and contrasts greatly with regions next to late pegmatite dikes where any of the mafic rock types in the complex may be converted completely to amphibolite.

Primary igneous features. The superb coastal exposures allow the original gabbro to be divided into three primary types (Fig. 9.3): Coarse-grained mafic gabbro, anorthositic gabbro, and cumulate-layered gabbro. These distinctions have not been successfully carried to the areas of smaller inland outcrops. The coarse-grained mafic gabbro occurs along the north margin of the complex and contacts with adjacent gabbros are mylonite zones. Of all the igneous types, this seems to have been most susceptible to deformation and formation of mylonitic gneiss with subvertical lineation. The anorthositic gabbro seems to have been least influenced by deformation and it is also host to the most spectacular gabbro pegmatites. Its contact with the cumulate-layered gabbro is locally primary. The cumulate layered gabbro contains sharp "inch-scale layering" of pyroxenes and plagioclase, which can be interpreted to indicate that the original top of the complex was to the south. From the overall aspect of the map, it seems likely that an unknown portion of the original complex may have been removed by early shearing. The fact that the total mass is so large gives some confidence that the early shear fabrics are in their approximately original orientation except for late folding, unlike the early fabrics in small eclogite boudins which commonly show highly diverse orientations (see Stop 9-3).

Mylonitic gneisses and mylonites of northern shore. From the house walk west into the pasture and then north and northeast across a fence to north-facing coastal exposures. The first rock that you will see approaching the coast is
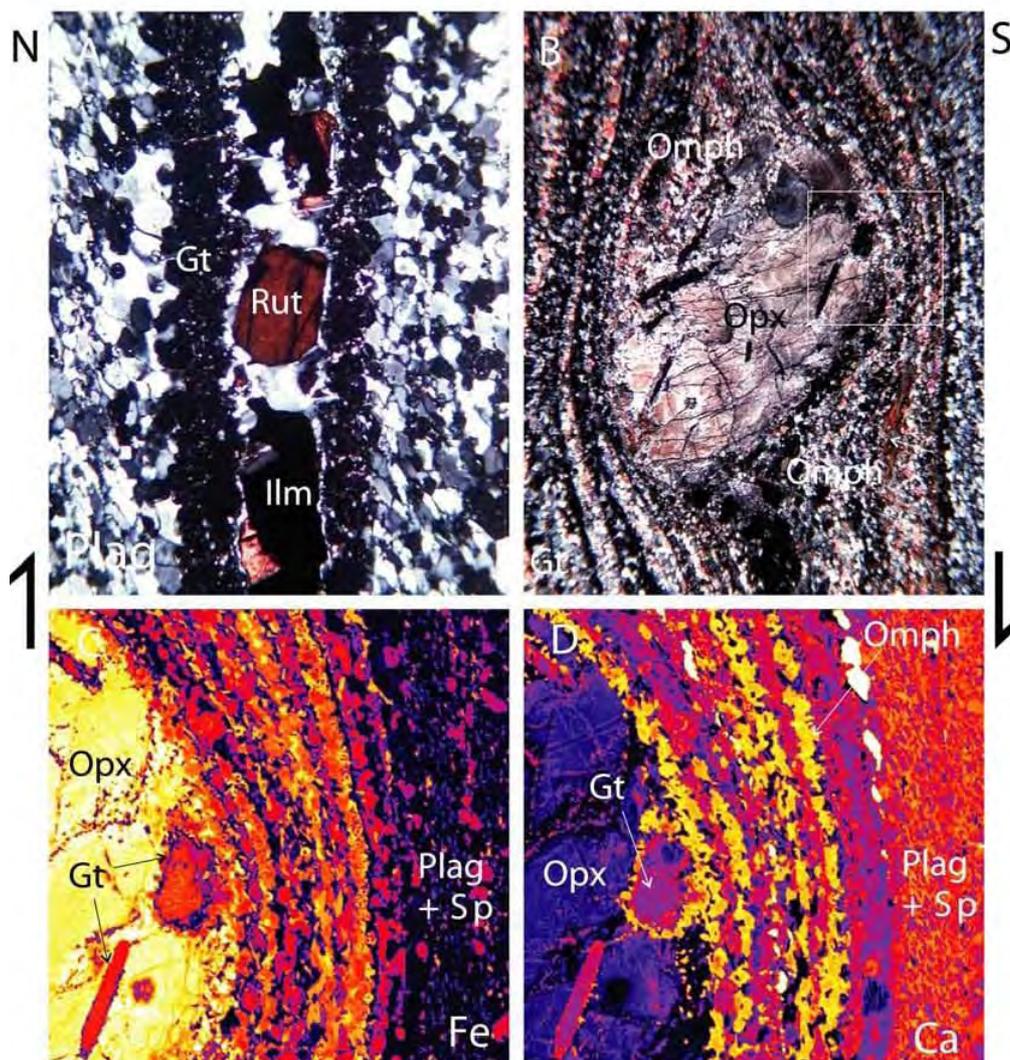
Fig. 9.4 Images of mylonitized Haram Gabbro. Evidence for deformation of some products and reactants can be seen in thin section (A, B) and in composition maps (C, D) from inset area of B, with the highest concentrations indicated by white/yellow and the lowest by black/blue. Ilmenite was deformed into layers and partially replaced by rutile (A) and plagioclase shows a strong grain-shape preferred orientation (A). Orthopyroxene is commonly deformed into layers and replaced by omphacite (Jd 18 -32). The original igneous pyroxene can occur as porphyroclasts (4 mm long) that show annealed fractures and replacement by omphacite (B, C, D). Plagioclase-rich layers show a strong grain-shape preferred orientation (A, B, C, D). Adjacent to garnet layers the plagioclase typically shows a decrease in Ca that suggests prograde growth zoning (D). Orthopyroxene porphyroclasts form sigma structures (B) and plagioclase layers (A, D)

coarse mylonitic gabbro gneiss with striking steeply plunging lineation. The fine-grained smooth outcrops near the waterline (we expect to be there at low tide) are fine-grained gabbro mylonite with a minor amount of felsic mylonite. These appear to form a carapace all along the north coast of Haramneset. Travel east a short distance (much slippery rock) to a location with unusual overhanging gabbro and a gully where one can walk south a short distance. Here one can see several shear zones, small gabbro pegmatites in various stages of deformation, and a zone of cumulate-layered gabbro. In the area of the gully there is a series of moderately south-dipping discontinuous shear zones that show a north-side-down shear sense. This opposite sense of shear is seen in a shear zone that juxtaposed the cumulate layered gabbro against gabbro. To the east of the north end of the gully there is a zone of mylonitic gneiss where there is extensive development of garnet and green omphacitic pyroxene. North-side-up shear sense has been observed in hand specimen here. Follow coast west observing mylonite where possible, passing first small headland on the shore side, to north side of last headland of coarse-grained mafic gabbro. Here one can observe the contact between eclogite-facies mylonite and gabbro and a curious mylonite zone that curves into the main mass of the gabbro. In this one can seen original garnet-rich eclogite and secondary amphibolite related to later fractures.

Mafic, anorthositic, cumulate-layered and pegmatitic gabbros, northwestern points. Cross over headland of coarse mafic gabbro (several routes possible) to southwest edge of coarse gabbro where it grades into coarse mylonitic gneiss with steep rodding and then into fine-grained mylonite. Across the gully is a large exposure of anorthositic
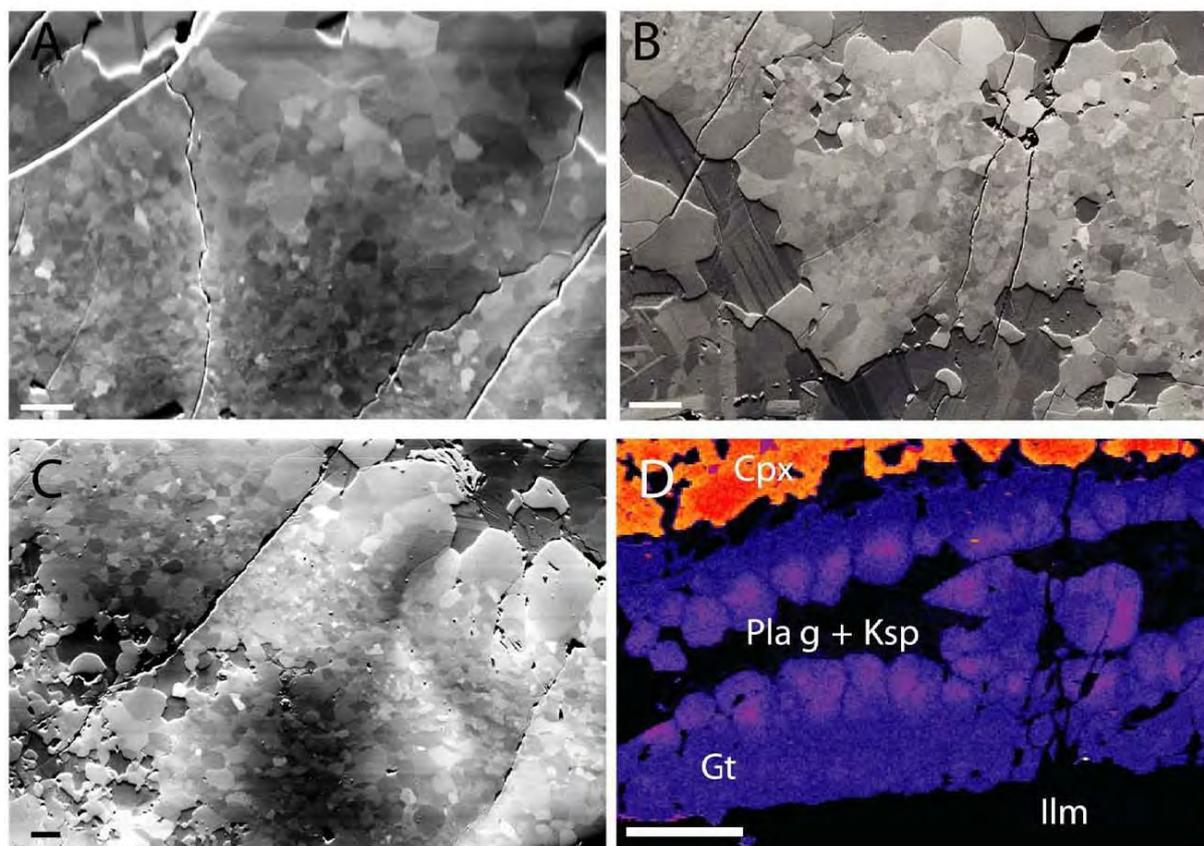
Fig. 9.5 Orientation contrast images of garnet in corona textures outside of the shear zone (A, B) and in mylonitized gabbro (C, D). D is a Ca composition map of garnet layers from the lineated gabbro (C). Scale bars are 10 µm in A, B, and C and 200 µm in D.

gabbro. From here the contrast between the two gabbros is striking. Walk west across anorthositIc gabbro to an area of where there appears to be a relict diabase dike and then climb onto the top surface of the headland northwest of the small automatic lighthouse. Here a pavement exposure shows spectacular zoned gabbro pegmatite pods with crystals plagioclase, pyroxenes, and magnetite 10-20 cm long. Zircon separated from one of these pegmatites by Tom Krogh gave a U-Pb age of 1466 ±2 Ma (see below) The only sign of the regional eclogite and amphibolite-facies metamorphisms are tiny garnet coronas between mafic minerals and plagioclase. In the broad gully north of the pegmatites is an eclogite-facies mylonite zone (not shown on Fig. 9.3) and then the contact with cumulate-layered gabbro.

Cumulate-layered gabbro, late pegmatite, and mylonite, southwest coast. After observing the layering walk southeast toward the lighthouse over quite rough ground to a large west-facing wall with a great display of inch-scale layering. Beyond this descend steeply down toward the waterline (big boulders nearby). Here is a special example of a late pegmatite cutting through the inch-scale layered gabbro with complete transformation to amphibolite, perfectly preserving the layering but none of the original mineralogy. U-Pb isotope measurements by Tom Krogh on zircons separated from this pegmatite indicate an age of 390±2 Ma (3 of 5 points, 60% probability of fit). Near here are outcrops and lose pieces where primary igneous layering is very photogenic. From here proceed southeast through a boulder field with many outcrops. These show the approximate contact between gabbro and gabbro-mylonite and examples of granitic mylonitic gneiss and one bit of calc-silicate rock with steep mylonite lineation. Tom Krogh has obtained a U-Pb age on zircon fractions from the granitic gneiss mylonite of 1663 ±3 Ma, a typical igneous protolith age of many tonalitic basement gneisses in the region (Tucker et al, 1990). In this example neither the eclogite-facies mylonitization nor the later amphibolite-facies metamorphism seems to have affected the zircons. Among the boulders are boudins of eclogite probably derived from outside the gabbro. Walk back northeast over raised beach gravels to house and vehicles.

Igneous age of the gabbro and comparisons with Flem Gabbro. According to Mørk (Mørk & Krogh, 1987) the chemical composition and the Sm-Nd isotopic composition distinguish the normally plagioclase-rich Haram gabbro from the more extensively studied Flem olivine gabbro (Stop 9-4). Sm-Nd dating of relict augite + orthopyroxene + 2 whole rocks gave an age of 926±70 Ma (Mørk & Mearns, 1986). However, Tom Krogh (Krogh et al., 2003) obtained a U-Pb age on zircon from a gabbro pegmatite of 1466 ±2 (4 points, 14% probability of fit) which is very close to the

1463 Ma age of the Selsnes Gabbro (Tucker et al. 1990) and may suggest metamorphic contamination of the Sm-Nd mineral fractions. Mork's report that metamorphic reactions involve only formation of garnet-hornblende coronas is not borne out by the present study, although the production of complete eclogite has not been observed, as it has at Flem (Stop 9-4). It is speculated that the driving force for corona development and transformation to eclogite is much stronger in the olivine-rich bulk composition at Flem than in the marginally quartz-bearing gabbro at Haram.

Leave Stop 2-1 and return toward Austnes. Back about 1 km in village of Åkre turn sharp left (east) on gravel road through fields leading to one-way Bomveg (toll road) over the mountains of Haramsøya. Pay toll by sign at base of steep slope and proceed southward up steep road to crest of ridge. On the left of the road are several artificial exposures of basement rocks and synclinal belts of Blåhø Nappe rocks of the central segment. In one exposure Blåhø mica schist contains a conspicuous bed of pink marble. At crest of ridge road turns sharply north and up a series of switchbacks blasted in the Åkre Mylonite. Park vehicles at a switchback on edge of west-facing cliff, with spectacular view of Lepsøya and overlooking the Misjonsenter. Walk up and down the road to outcrops of mylonite.

STOP 9-2: (30 minutes) Åkre Mylonite
The youngest major ductile structure of Nordøyane is the Åkre mylonite, which separates the northern and central structural segments of crust and is best exposed on Haramsøya (Fig. 9.1). This is a lower amphibolite-facies mylonite zone that cuts both eclogite-facies structures and amphibolite-facies structures and pegmatites formed earlier than the mylonite. The Ulla Gneiss is strongly mylonitized and contains fragments of partially amphibolitized eclogite. The steeply dipping mylonite zone shows well developed shear indicators that include asymmetric tails on porphyroclasts and shear bands that are consistent with sinistral shearing. Equal area projections of poles to foliations and lineations within the mylonite zone show considerable scatter, but the pattern is consistent with a steep north-dipping tabular zone that contains a subhorizontal lineation and strikes east-west. The Åkre mylonite is also exposed on the east coast of Flemsøya and shows the same complex deformation. The zone strikes N70°E based on the exposures on both islands.

Continue up toll road to grass lands on the highest part of Haramsøya. Cross toward northern end of island, then east down very steep section with exposures of Ulla Gneiss to main paved road in village of Ulla. Proceed north through Ulla and east across Ulla Bridge. At east end of Ulla Bridge turn sharp left onto rough causeway to Ullaholmen. Drive to field at the high point of the island. From there, walk through fields to prominent rocks on west coast.

STOP 9-3: (1 hour) Tubular and complex folding in Ulla Gneiss, Ullaholmen.
Introduction. This is a world-class exposure illustrating structural features related both to overprinting and progressive deformation that occurred during exhumation of HP and UHP rocks. While it is tempting to apologize to the hard core petrologist, it is important to emphasize that production and exhumation are thermal-**mechanical** processes. There is an amazing number of photogenic and important structural features and we will try to guide the participants to the best of these. Some of these are the locations on the prominent knob numbered 1-8 on Figure 9.6.

The dominant fabric in Ulla Gneiss is subhorizontal lineation and folds parallel to the east-northeast trending "stretching" direction during late Scandian sinistral transtension. However, the structure is somewhat chaotic because of the heterogeneous character of the rocks with abundant large and small eclogite and retrograded eclogite boudins. Furthermore there is abundant evidence of one or more early fold- and lineation-forming episodes that may have originally been formed parallel to the eclogite-facies mylonites near the Haram Gabbro.

South of the prominent knob. Walk directly west along north edge of the field to the first knob on the coast. Here the folds in layered gneiss are not dominantly tubular, but they show spectacular examples of folded lineations and refolded early fold systems, in some of which the early fold asymmetry is graphically displayed. From here, walk north around a broad inlet toward the most prominent coastal knob. En route you will encounter a large layered eclogite boudin and a curious intrusive gneiss, which Ole Lutro calls "ovalite" and which is widespread in the Romsdal near Trollvegen. Are the ovals deformed giant phenocrysts or are they oval xenoliths of another rock type?

On the prominent knob. The prominent knob (Fig. 9.6) is part of a major eastward-closing sheath fold that is bounded on either side by the Ulla Gneiss and has a core with complexly folded amphibolite to granitic gneiss with folded pegmatite. Although kinematic indicators are not abundant, the 3D exposure of the fold geometry leaves little doubt that the late folds were produced during top-west shearing with mineral assemblages indicating upper amphibolite-facies conditions. A speculative model to explain the origin of the major structure is that it developed in response to heterogeneity at the interface between the stronger Ulla Gneiss and the granitic gneiss that forms the core of the fold. Much of the complex refolding is a result of late folding though there is good evidence for early folding interpreted to have occurred at eclogite-facies conditions.

The south slope of the prominent knob (location 1 in Fig. 9.6) shows many places where early folds and lineations are seen to be folded in great circles around late folds with a transport direction that is only a few degrees away from the late fold axes. For those interested in having some of the more interesting features pointed out, we will proceed to the west to locations 2 to 4 on Figure 9.6 and then north to locations 5 to 8. The northern part of the knob is the location of the detailed geologic map, structural measurements, and interpreted cross section in Figure 9.7. The best tubular folds are to be seen on west-facing joint surfaces that do not show very well on the map. The nature of the outcrop makes it just possible to identify which of the closures are tubular "basins" and which are "domes" on the near
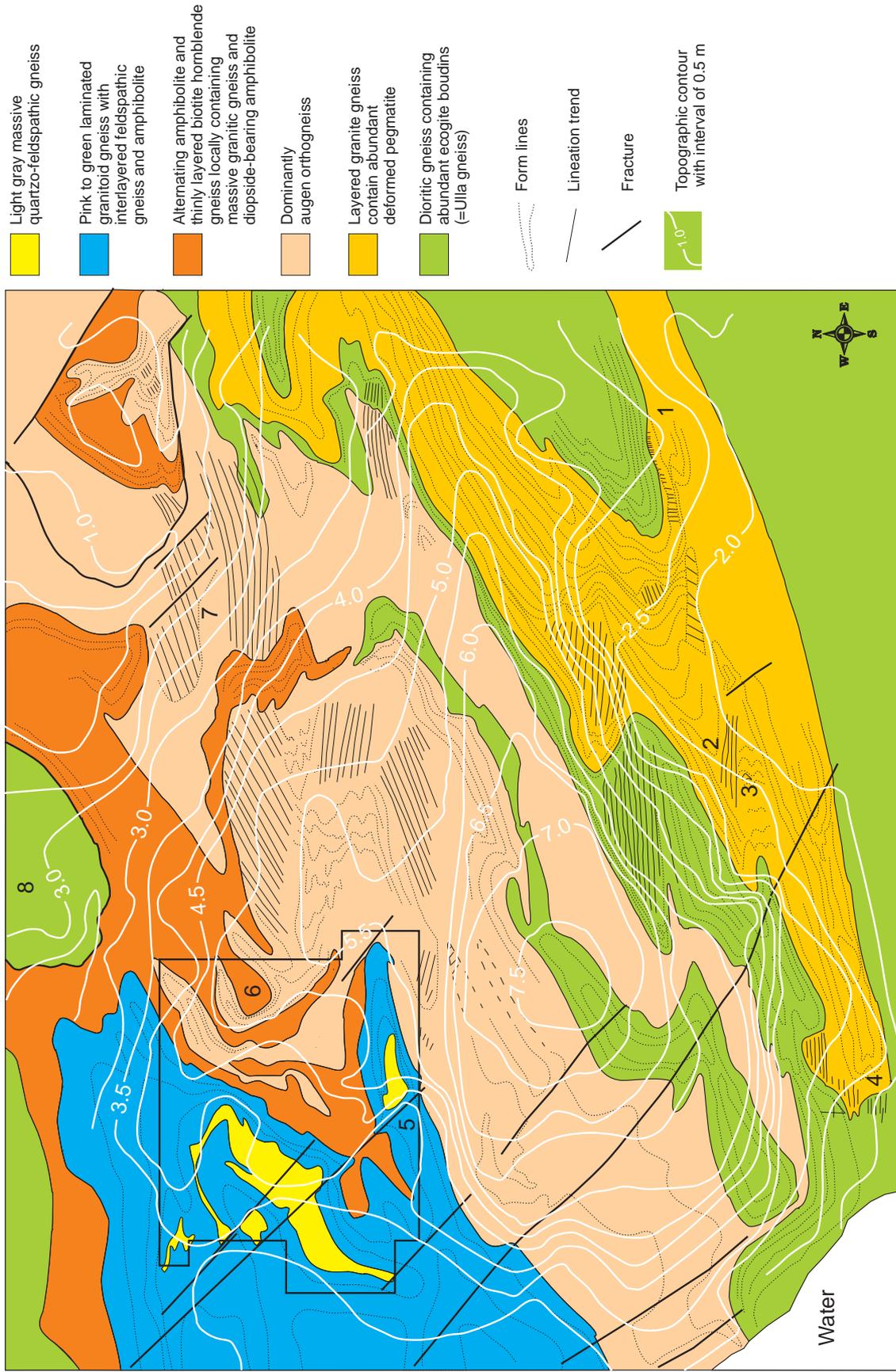
Fig. 9.6 Generalized geologic map for Stop 9-3 on Ullaholmen. Locations 1-8 are areas of interest. Box is the location of Fig. 9.7.

Legend:

- Light gray massive quartzo-feldspathic gneiss
- Pink to green laminated granitoid gneiss with interlayered feldspathic gneiss and amphibolite
- Alternating amphibolite and thinly layered biotite hornblende gneiss locally containing massive granitic gneiss and diopside-bearing amphibolite
- Dominantly augen orthogneiss
- Layered granite gneiss contain abundant deformed pegmatite
- Dioritic gneiss containing abundant ecogite boudins (=Ulla gneiss)
- Form lines
- Lineation trend
- Fracture
- Topographic contour with interval of 0.5 m

12 meters

Water

vertical surfaces. The tube axes along with most lineation and fold axes trend east-northeast and plunge gently east. On the nose of one "dome" (viewed from the west) it is possible to see earlier lineation bending in a great circle around the "dome" axis. According to Terry, many layers, which he traced in detail in the outcrop, turned out to be an isoclinal fold if traced far enough. Based on these observations, an important question is the extent of earlier tubular folding, and the extent to which tight isoclinal synclines identified in the region are actually only earlier manifestations of a prolonged episode of tubular folding. The best overall evaluation of these folds, as illustrated in the cross section, is that the tubes themselves reflect a sense of top-west and north-side-west (i.e. sinistral) shear during fold development.

Northern point of Ullaholmen. Persons with an additional half hour may wish to examine the large coastal exposures of complexly deformed augen-orthogneiss on the north point of Ullaholmen 250 m away. After the stop walk back to the field over boggy ground.

Leave Stop 2-3 and Ullaholmen. Turn left (east) onto main causeway to Flemsøya. At T junction at end of causeway turn left (north) toward Flem. Follow narrow road around NW corner of Flemsøya, and along north coast to small parking area at Sandvik sand beach.

STOP 9-4: (1 hour) Flem (Sandvikhaugane) Gabbro. Stops 9-4 and 9-5 within a short walk of the beach at Sandvik were both described by Mai Britt Mørk (Mørk & Krogh, 1987), and a study of the gabbro was part of her Ph. D. thesis at Oslo University (Mørk, 1985; Mork, 1986, Mork & Mearns, 1986).
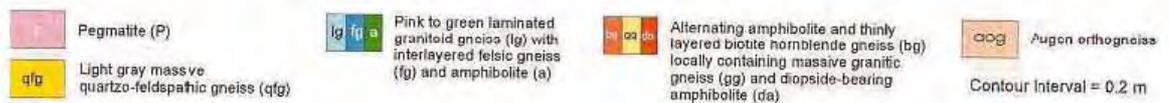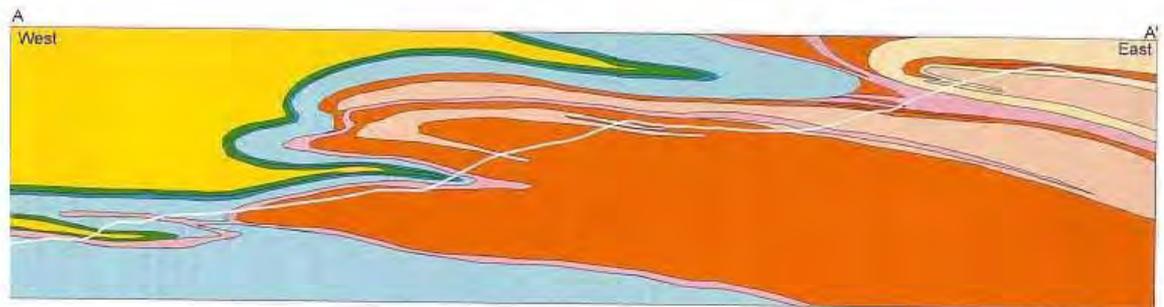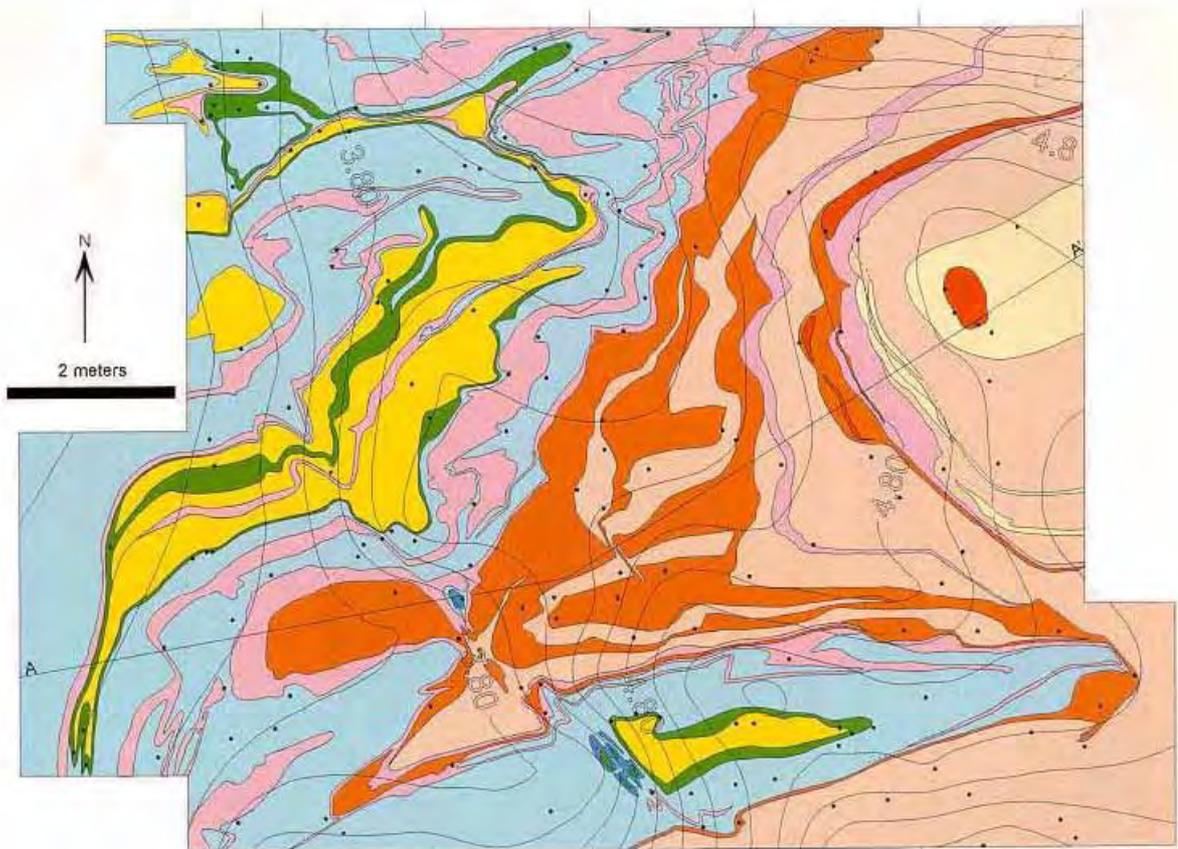
Eastern end of gabbro. Walk left (northwest) beyond the beach toward the point at Sandvikhaugene. At the east end of the big outcrop at low tide the original igneous intrusive contact of the gabbro with the augen orthogneiss can be seen, as well as a steep subduction related (?) linear fabric in the augen gneiss. Rare sugary textured granitic veins locally with coarse garnet and tourmaline in the gabbro may be backveins that went through eclogite-facies conditions. These contrast with coarse cross-cutting late pegmatite dikes that came in late in the Scandian exhumation and usually caused extensive amphibolitization in adjacent gabbro or eclogite. Walk northwest from the contact about 75 m to the northeast-facing point where the full transition from layered gabbro to eclogite can be examined in clean outcrop.

Gabbro transformation to eclogite. According to Mørk (Mørk & Krogh, 1987) the gabbro is a layered olivine gabbro with relict igneous mineralogy including olivine, plagioclase, augite, ilmenite and biotite. It shows all stages of transformation to an eclogite consisting of garnet, omphacite, biotite, amphibole and ilmenite. Three stages of transformation are recognized in the field: 1) Massive corona gabbro dominated by relict igneous mineralogy and cumulate textures. 2) Transitional gabbro-eclogite with relict igneous textures and some relict minerals but with garnet and omphacite clearly visible in the field. 3) Completely eclogitized gabbro, either with pseudomorphic igneous textures or as recrystallized and foliated eclogites. Pseudomorphic replacements involve formation of garnet and orthopyroxene coronas between plagioclase and olivine, replacement of igneous plagioclase by more sodic plagioclase with thin spinel needles, increase of Na content of intercumulus augite and exsolution of an opaque phase. With increasing degree of reaction omphacite forms both by continuous reactions with the original augite and by nucleation within olivine and transient orthopyroxene, while garnet always seems to replace plagioclase. (For element images of this transition stage see Figure 9.8) As modal % of pyroxene climbs from about 15% in the gabbro to about 40% in the eclogite, the composition changes from Acmite 14 Jadeite 0 to around Acmite 10 Jadeite 25-30, and the garnet composition from pyrope 29 to 42-45, almandine nearly constant at 44-48, grossular 27 to 8, and spessartine 1.2 to 0.8. With increasing degree of reaction the transient phases disappear and garnet becomes more homogeneous with lower Ca and slightly higher Mg/Fe than in the corona stage. It appears that complete eclogitization and mineral-chemical as well as isotopic homogenization are enhanced by factors like oxidation, fluid availability, and deformation, while the transient stages reflect restricted element exchange between the mafic and felsic original domains (Mørk, 1985; Mørk & Mearns, 1986).

Geochronologic data on eclogite. A sample of this contorted totally eclogitized north margin of the Flem Gabbro provided only a small amount of small rounded zircons to Tom Krogh. In fact, zircons were so scarce that two separate collecting efforts in 1997 and 1998 were required. Data for two concordant analyses gave a mean 207/206 age of 412 (411,412 Ma) and a mean 206/238 age of 409 Ma (408, 410 Ma). The age of 410 Ma on metamorphic zircon from the Flem eclogite represents the most precise metamorphic age yet determined for northern Nordøyane (Krogh et al., 2003, 2004). This new age is about 10 m.y. older than the 401 Ma age based on Ulsteinvik eclogite at 402±2 Ma and on the 400±16 Ma Sm-Nd mineral age of Mørk & Mearns (1986) on the Flem eclogite, assumed by Terry et al. (2000a) for the "Lower Plate of Nordøyane" in their exhumation model for Fjørtoft.

Gabbro pegmatite and granite pegmatite. Plagioclase-rich pegmatite dikes within the gabbro about 75 meters more to the west containing plagioclase, augite, orthopyroxene, and ilmenite have developed coronas with garnet and amphibole between plagioclase and pyroxene. A late Scandian pegmatite dike has introduced aquaeous fluid that produced bordering zones of hornblende-rich amphibolite about 0.5 m thick after either gabbro or eclogite.

Geochronologic data on gabbro. The gabbro is interpreted to have intruded into the augen orthogneiss (see above). Whole-rocks and relict igneous augite give a Sm-Nd isochron of 1289±48 Ma (Mørk & Mearns, 1986). This was interpreted as the age of the intrusion and similar to the age of other olivine gabbros from the Western Gneiss Region (Mearns, 1984). (By a strange quirk of fate this was also Robinson's field locality 1289 in July 1993!) Tom Krogh has reported a preliminary U-Pb age on zircon from the gabbro of 1252+/- 4 Ma (3 points, 54% probability of fit - refinement in progress).

Pegmatite (P)

lg fg a — Pink to green laminated granitoid gneiss (lg) with interlayered felsic gneiss (fg) and amphibolite (a)

qfg — Light gray massive quartzo-feldspathic gneiss (qfg)

bg gg da — Alternating amphibolite and thinly layered biotite hornblende gneiss (bg) locally containing massive granitic gneiss (gg) and diopside-bearing amphibolite (da)

aog — Augen orthogneiss

Contour Interval = 0.2 m

Summary of structural data for the Ullaholmen exposure. A. Contours of lineations and poles to foliation. B. Contours of fold axes.

Fig. 9.7 Detailed geologic map, cross section and equal area diagrams for part of the pavement exposure (see Fig. 9.6) at Stop 9-3 on Ullaholmen.

Fig. 9.8 Electron probe scanned element images of transitional olivine corona gabbro from Flem, Stop 9-4, for Mg (A), Mn (B), Ca (C) and Na (D). Color scheme relates directly to element abundance from abundant (white or red) to scarce or absent (violet or black. Relict igneous as well as metamorphic minerals are preserved in this stage as studied by Mørk (1985). The dominant feature in the center is a former igneous olivine, now completely replaced by orthopyroxene, rimmed by clinopyroxene and finally by garnet against matrix plagioclase. Mg (A) is richest in orthopyroxene (red) then in zoned clinopyroxene which is most magnesian (green) toward orthopyroxene and less magnesian outward, then in zoned garnet which is most magnesian (dark blue) toward clinopyroxene and less magnesian (violet) outward. The high-Mg phase (yellow) on the garnet-clinopyroxene boundary is not identified. Mn (B) is present in garnet, also detectable in orthopyroxene. Ca (C) is most abundant at the inner edge of clinopyroxene (white), decreasing irregularly outward (reds and yellows). It is also present in garnet where it is lowest (blue) against clinopyroxene and increasing outward (green-yellow red) against plagioclase (violet), and absent (black) in orthopyroxene and orthoclase in upper left. Na (D) is most abundant in plagioclase (red), shows strong zoning in clinopyroxene from low (violet) through intermediate (blue, green) to high (yellow, red) against garnet, shows a small amount (blue, violet) in orthoclase, and is absent in garnet and orthopyroxene (black). The figure illustrates the reality of mineral growth in a chemical potential gradient produced by disequilibrium between olivine and plagioclase, with production of jadeite-bearing clinopyroxene.

STOP 9-5: (20 minutes) Kvalvika garnet-peridotite

Depending on tide and sand conditions walk east along the beach from the parking space or along the road about 150 meters to a peculiar hollow in the gneissic rocks. This is eroded in a body of yellow-brown-weathering garnet peridotite with thin folded layers of garnet websterite consisting of garnet, Cr diopside, enstatite and rutile. The body is surrounded by typical Ulla Gneiss that was completely eclogitized near the contact with the peridotite. Most omphacite has been retrogressed to fine symplectite. In summer 1993 and 1996 the body was 90% covered by sand, whereas in spring 1997 storms had excavated most of the sand leaving about 90% exposure of ultramafic rock. The folds in garnet websterite are steeply plunging and probably parallel to an early lineation seen in a dunite layer near the margin of the body. This lineation has the same orientation as the relic lineation in the eclogitized Ulla Gneiss that is in contact with the unit. This implies: 1) Emplacement of the peridotite occurred during or prior to the formation of the eclogite-facies lineation. 2) Emplacement of the peridotite occurred at or before 410 Ma on the basis that the eclogitization of the Ulla Gneiss and the Flem Gabbro were synchronous. These same structural relationships can also be inferred for the large peridotite exposed at Skulen and regional study in the Ulla Gneiss indicates that this lineation was formed originally during top-SE shearing that is consistent with thrusting. However, these outcrops have yielded no information on the mode of emplacement of the body against the adjacent Ulla Gneiss, but may upon closer study.

Ulla Gneiss petrology. The eclogitized diorite gneiss contains a high-pressure assemblage of garnet + omphacite + rutile ± biotite ± apatite ± monazite. This assemblage is overprinted by a later assemblage of diopside + plagioclase + hornblende + ilmenite. Notably absent from both assemblages is quartz or any other SiO_2 polymorph. The high-pressure assemblage is mainly preserved as inclusions in garnet. Omphacite and rutile occur as inclusions in garnet and together these phases are interpreted to represent the HP assemblage. The matrix omphacite is nearly all replaced by undeformed diopside + plagioclase + hornblende symplectite. However, the original omphacite grain boundaries are decorated by coarser amphibole and had an original grain size of 1 mm. Garnet which has a grain size of 1.0-1.5 mm is elongate and generally rimmed by retrograde hornblende. Both garnet and relict omphacite in this eclogite have a strong shape-preferred orientation and were deformed at HP conditions. Rutile in the matrix is partly to completely replaced by ilmenite. The minerals that are interpreted to have formed during overprinting and fluid infiltration include hornblende, plagioclase, and ilmenite. Overprinting occurred under a range of conditions from amphibolite- to perhaps greenschist-facies.

Kvalvika peridotite petrology. Two samples were selected for detailed microstructural analysis (Terry et al. in revision). The location of the first one (1632) was less than 0.3 meters from the perfectly exposed contact with eclogitized Ulla Gneiss. Both peridotite and eclogitized gneiss contain the same lineation and foliation and clearly they were deformed together in the same event. The second sample (1632A) was collected in the center of the peridotite lens for comparison of the microstructural characteristics. The HP assemblage includes olivine + orthopyroxene + garnet + spinel, constituting a garnet-harzburgite that plots very near the dunite field. The sample from the center contains, in addition, a small amount of clinopyroxene (~6 %) making it a garnet-lherzolite. The grain size of olivine changes from about 1.5 mm in the center to about 1 mm at the edge of the peridotite body. Sample 1632A also has two generations of garnet. Grt-2 is partially preserved and contains inclusions of spinel. Elongate grains have a weak shape-preferred orientation that is parallel to the olivine lineation. Towards the rim of the peridotite in sample 1632, the grain shape of olivine becomes more ellipsoidal and the grains are more strongly aligned in the foliation. A CPO is already apparent in sample 1632 in the polarized light microscope with a 1λ plate. Olivine grains are fractured and serpentinized along both fractures and grain boundaries, with serpentinization being stronger at the edge of the peridotite body. Orthopyroxene in sample 1632 has a grain size similar to that of olivine and also a similar SPO in outcrop. It also helps to define a crude foliation and also grew around former garnet in corona structures. In the interior part of the corona, some orthopyroxene contains exsolution lamellae of spinel and is tentatively interpreted as an older generation of orthopyroxene (Opx-1). The original porphyroclastic garnet is completely replaced by amphibole + orthopyroxene + spinel. Element distribution maps of Cr show two generations of garnet (Grt-1 and Grt-2) with Grt-2 forming a corona rim on Grt 1. Grt-2 is partially to completely replaced by a symplectite of Opx + Spl and the original euhedral shape of the garnet is preserved by the symplectite.

Olivine microstructures. The combined EBSD and TEM results of Terry et al. (in revision) show that the deformation of olivine in the Kvalvika peridotite was achieved by intracrystalline dislocation creep with <001> as the direction of the Burgers vector. {010} and - to a lesser extent - {100} served as glide planes for dislocation motion. These slip systems are rather exceptional in upper mantle rocks (Ben Ismaïl & Mainprice, 1998) where the slip system (010)[100] generally dominates the texture development.

Metamorphic conditions and P-T evolution. Metamorphic conditions for olivine c-slip were determined by applying quantitative geothermobarometry and differential thermodynamic modeling in the NCFMAS system to constrain a P-T path during metamorphism. The geothermobarometry indicates conditions of 810-860 °C and 2.9-3.4 GPa for the garnet rim (Fig. 9.9). Backward modeling of the changes in Fe/(Fe+Mg) ratio and Al content in Opx yields a clockwise P-T path indicating a range conditions for olivine c-slip in the peridotite lens of ~740 to 850 °C and ~3.0 to 4.0 GPa. The P-T path is restricted to UHP conditions and captures the change
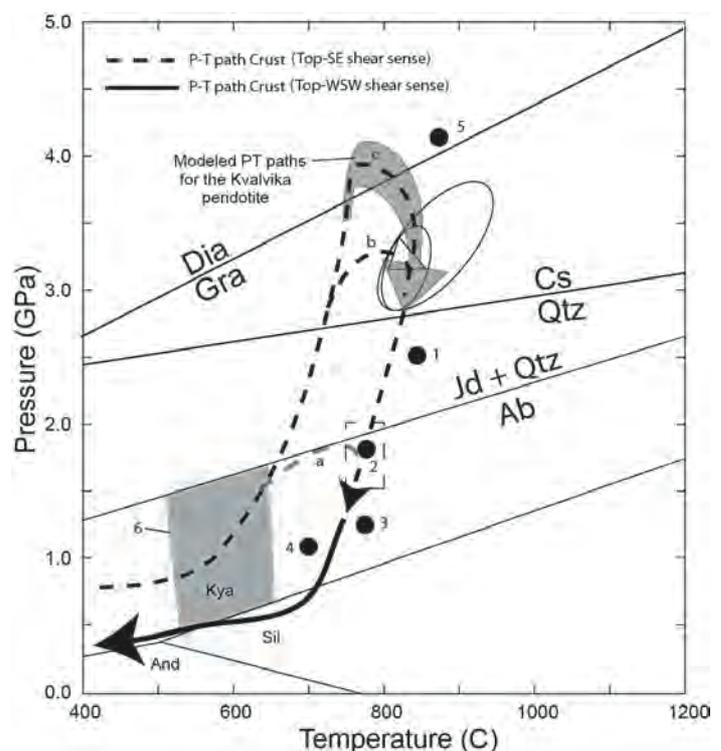

Fig. 9.9 P-T-t-D path for Ulla Gneiss and the Kvalvika peridotite lens. a, b, and c are alternative paths (Terry et al. (in revision). **a** is from Terry et al. (2000) and was the path then favored for the Haram Gabbro and the Ulla Gneiss. **b** and **c** are possible paths for the Ulla Gneiss interpreted on the basis of new P-T data from the Kvalvika peridotite lens. P,T estimates are from: (1,2) Terry et al. (2000), (3) Terry and Robinson (2004), (4) O'Conner and Terry (2002), (5) Carswell and van Roermund (2005), (6) Terry and Heidelbach (2006). Ellipses are P-T estimates from Terry et al. (in revision) using T estimates from the Kvalvika peridotite (small ellipse) and T estimates from the Ulla gneiss (large ellipse) combined with moderately pressure-dependent geobarometers from the Kvalvika peridotite. Triangle is the revised P-T conditions for the kyanite eclogite 1066b on Fjortoft.
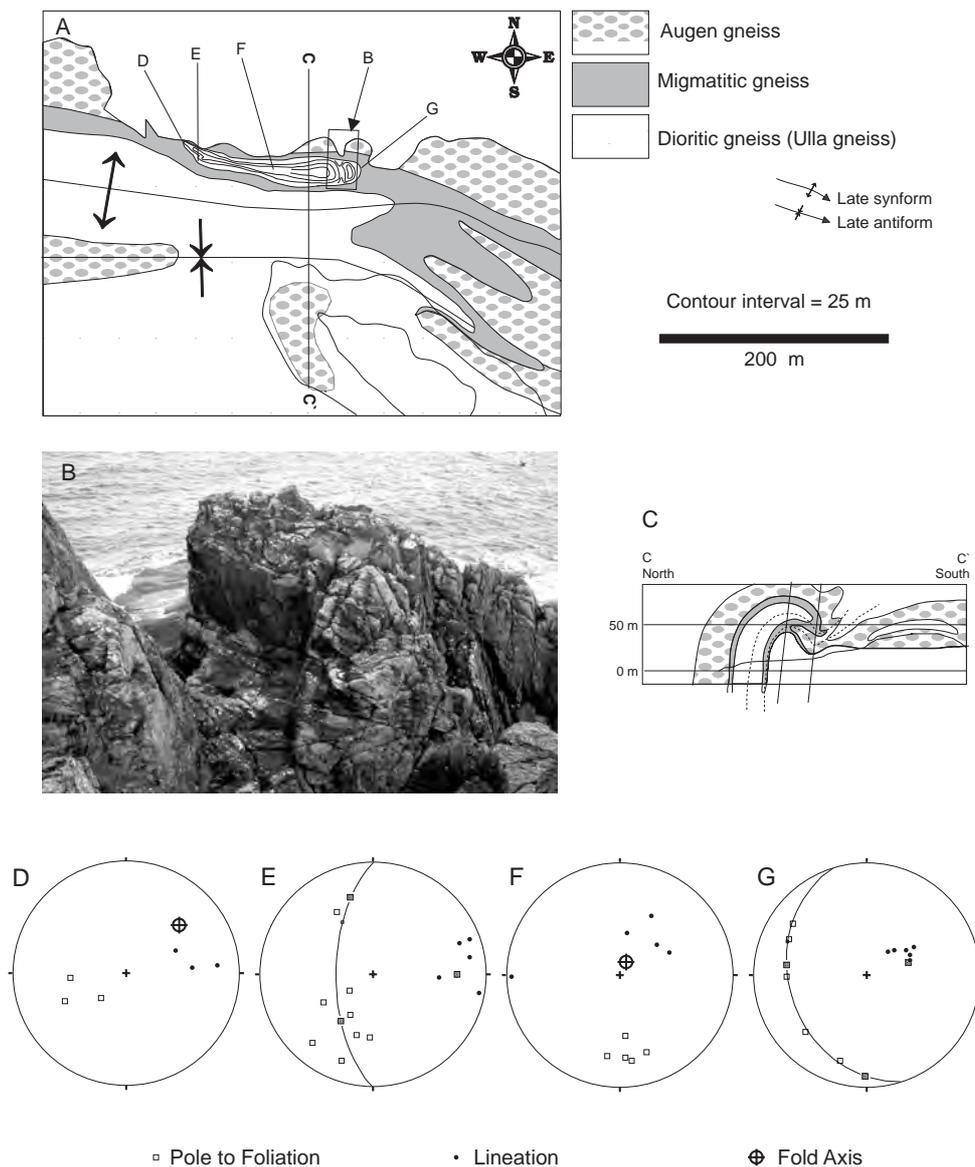
Fig. 9.10 Detailed map (A) in the area near a well preserved eclogite-facies fold on the northwest coast of Flemsøya B) Photograph of the eastern closure of the early fold where dark layers represent mafic compositions that locally preserve eclogite-facies assemblages. C) Cross section C-C' showing the geometry of the eclogite-facies fold being folded by late amphibolite-facies structures. Equal area projections (D, E, F, and G) show poles to foliations and lineations from west to east across the early fold structure with locations shown in A. See text for discussion.

from subduction to exhumation without a change in subduction kinematics. This favors entrainment of the mantle peridotite from the underlying mantle lithosphere and exhumation by ductile flow along weak shear zones at UHP conditions for the early stages of the Scandian orogeny (423-410 Ma).

Leave Stop 9-5 and drive back (west) toward the northwestern point of Flemsøya southwest of Kjellholmen. Just before a narrow pass at a high point of the road turn right (north) into a large parking space associated with a bedrock excavation.

STOP 9-6: (40 minutes) Eclogite-facies structural features in Ulla Gneiss and augen orthogneiss.
    From parking space walk northwest and down to sea level where there is huge exposure in a hollow of variably deformed augen orthogneiss with pegmatite intrusions. On a ridge to the northeast of the hollow there are three rafts of eclogite described by Mørk with hybrid boudin-neck pegmatites. One of these pegmatites studied by Tom Krogh contained abundant cored grains (Krogh et al., 2003). Data for 4 tips from these yield data that are 1.3, 1.5, 2.3 and 3.2 % discordant with a mean 207/206 age of 397 Ma. The two most concordant analyses are the youngest and hence are least likely to contain older growth components. These have a mean 207/206 age of 395 Ma (394, 395 Ma) remarkably consistent with ages from other pegmatites in the region as well as with the mean lower intercept chord of titanite ages reported by Tucker et al., 1990, 2004. One of the eclogite rafts at this location appears to have been the source of the Sm-Nd mineral isochron of 400 ±16 Ma reported by Mørk & Mearns (1986).

Walk northwest then west to steeper coast where there is an awkward ridge at the top of a sea cliff. A sharp projection at the east end of the sea cliff contains a steeply north-plunging tubular fold in well layered Ulla Gneiss (Fig. 9.10). This can be reached by careful scrambling, but can be viewed and photographed very well from the cliff top. The fold is outlined by an eclogite layer that contains an omphacite stretching lineation parallel to the tube axis. Like the subhorizontal tubular folds at Ullaholmen (Stop 9-3), this steep fold indicates deformation under constrictional strain conditions but during subduction at a much earlier time in the total strain history.

Return to parking place and rejoin road, turning right (west). Cross the small pass where there is a small outcrop of brown-weathering peridotite, and continue southwest to junction near causeway. Pass the junction and drive south toward Longva. Bear left at Longva and follow inland road around south end of Flemsøya to Nogva. Find parking beside a bus garage at Longva. Walk east down private driveway to small wharf and then north along shore to outcrop.

STOP 9-7: (25 minutes) Nogva kyanite-zoisite eclogite with rare polycrystalline quartz pseudomorphs after coesite and retrograde sapphirine and corundum.
Structural setting. The rocks exposed at and near Nogva are not easily mapped, but at present are interpreted to belong to the same horizon of Blåhø rocks as on the northeastern part of Fjørtoft. These rocks are folded across the crest of the large late anticline that dominates the northern structural segment of Nordøyane (Fig. 9.2). Here, we are only a few hundred meters north of the Åkre Mylonite Zone which is well exposed in a small harbor on the east coast of Flemsøya.

Petrography and petrology. There are two moderately large eclogite boudins in this coastal outcrop studied by Terry et al. (2000b, see also major and trace-element analyses of the rock by Hollocher et al. 2007b). They have compositional layering that is truncated by the boudin margins, and also mineral lineation that is transverse to the normal northeast trend. There are two dominant kinds of layers. The first are striking bright orange garnet-rich layers with very fresh omphacite partially replaced by fine plagioclase+pyroxene symplectite. According to Tony Carswell, personal communication 1996, such a color usually denotes a high Ca content in garnet, in this case Pyr 32.7, Alm 36.7, Gr 30.1, Sp 0.5. The other layers are lighter-colored and dominated by plagioclase-pyroxene symplectite after omphacite and lesser garnet. However, in different places they contain abundant quartz, zoisite, plagioclase (as a replacement of zoisite), hornblende, biotite, rutile, biotite+feldspar symplectite after phengite (?), and kyanite. At one point there is a boudin neck zone containing crystals of zoisite up to about 8 cm long. Commonly surrounding kyanite or replacing it completely are symplectites involving calcic plagioclase and one or two of the phases sapphirine (a very pale green variety), corundum, or spinel.

    Textural and rimming relationships suggest a variety of reactions indicative of decompression under relatively high-temperature conditions. Locally secondary calcic pyroxene is concentrated as a rim between quartz and pyroxene-plagioclase symplectite. The latter as well as garnet is commonly replaced partially by hornblende, but some samples have no hornblende. Large zoisite grains may contain plagioclase or zoisite-plagioclase symplectite and may be a partial result of the reaction quartz + kyanite + zoisite = anorthite. The symplectic rims on kyanite denote decompression reactions between kyanite and omphacite producing calcic plagioclase plus one or more of the aluminous phases sapphirine, corundum and spinel, dependent on the local Fe/Mg ratio. Although we have not found any completely fresh kyanite eclogite in this outcrop, a boudin on nearby Fjørtoft contains a completely fresh core, with no symplectite and with only traces of secondary plagioclase, showing all mutual contacts among the phases garnet, omphacite, zoisite, kyanite, quartz, hornblende, pale brown biotite and rutile. A garnet-kyanite-omphacite-coesite assemblage can be reconstructed easily from inclusions in the garnet-rich part of the present outcrop, and geothermobarometry indicates 820°C, 30-36 kbar, slightly lower P than indicated for the assemblage studied at Fjørtoft, but well within UHP conditions. When phase compositions are plotted together with those from Fjørtoft eclogite, the omphacite from here is much richer in jadeite component (Jd 41), the grossular content of garnet is similar (Gr 30), but both omphacite and garnet have a much higher ratio of Fe/(Fe+Mg). Overall these relations suggest equilibration under fairly similar conditions but here in a rock with a lower original anorthite content and higher Fe/(Fe+Mg) ratio.

Geochemical keys to radiometric dating of eclogite. Eclogites are favorite targets for radiometric dating because they represent metamorphic rocks derived from the most dynamic and rarely accessible deep parts of mountain belts. Metamorphic assemblages combined with radiometric systems that record metamorphic mineral growth can yield insight into P-T-t trajectories otherwise unavailable. There is, however, long-standing problem that zircons for U-Th-Pb dating are particularly hard to find in kyanite eclogites. The origin of this problem is illustrated in Figure 9.11.

    Kyanite in eclogite results from the metamorphic reactions of plagioclase at high pressure, for example:
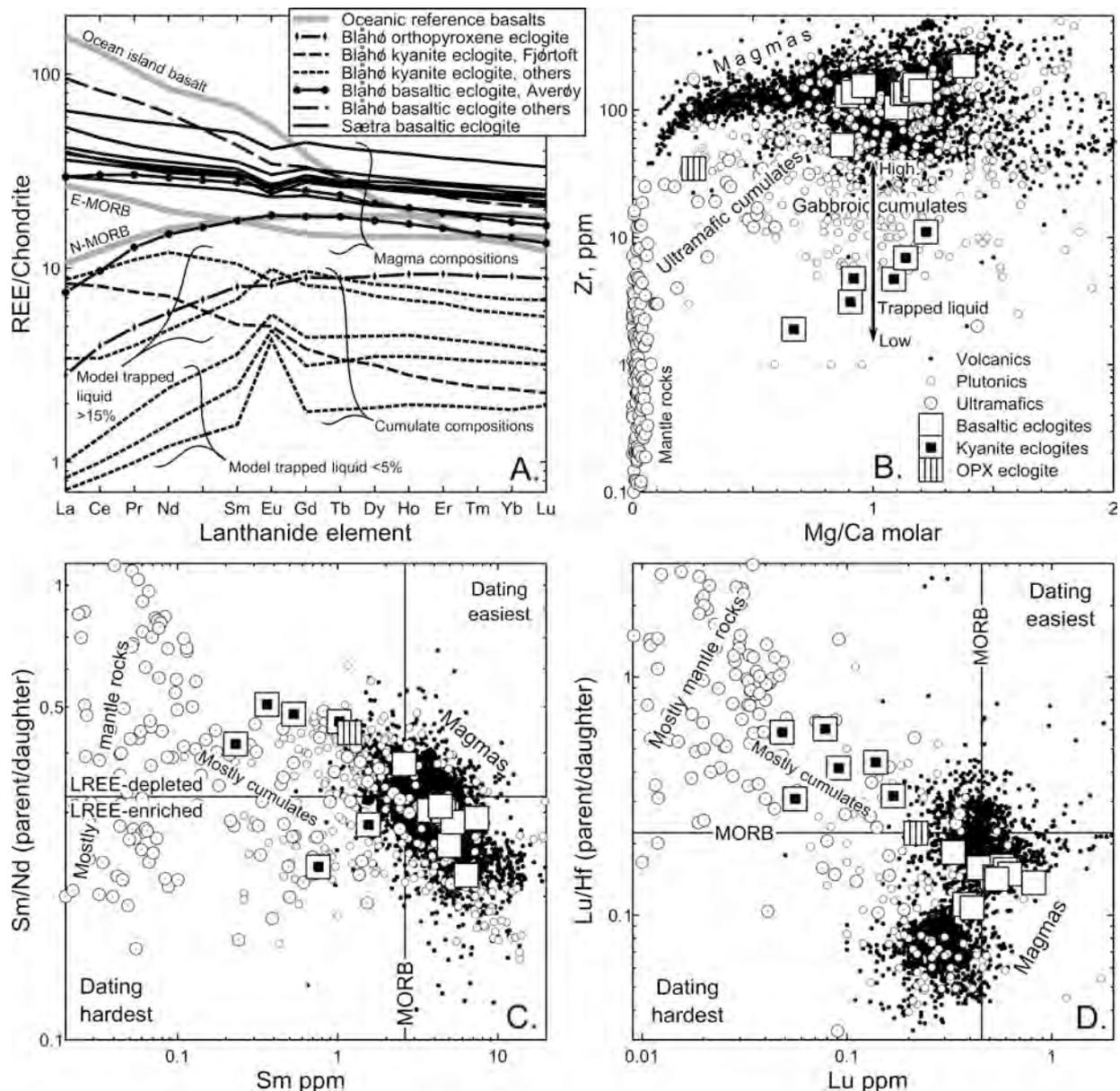
$$3CaAl_2Si_2O_8 \quad = \quad Ca_3Al_2Si_3O_{12} \quad + \quad 2Al_2SiO_5 \quad + \quad SiO_2$$
anorthite component   grossular component   kyanite   silica

$$4CaAl_2Si_2O_8 + H_2O \quad = \quad 2Ca_2Al_3Si_3O_{12}(OH) \quad + Al_2SiO_5 + \quad SiO_2$$
anorthite   zoisite   kyanite   silica

Fig. 9.11 This figure illustrates that kyanite eclogites in particular are probably gabbroic cumulates, and this has characteristic effects on their use for different radiometric dating systems. A) REE patterns of eclogites from the north coastal part of the Western Gneiss Region. MORB and ocean island reference lines are from Sun & McDonough (1989). B) Mg/Ca vs. Zr diagram showing important relationships for the Zircon U-Th-Pb radiometric system. Labels show approximate fields of the most common mantle cumulate, cumulates, and magmatic rocks. Gabbroic rocks having approximately magmatic compositions plot on top of the volcanics. C) Sm vs. Sm/Nd diagram showing relationships important for the Sm-Nd radiometric system. D) Lu vs. Lu/Hf diagram showing relationships important for the Lu-Hf radiometric system. All diagrams are adapted from Hollocher et al. (2007a). The volcanic, plutonic, and ultramafic reference set is 7548 whole rock and glass analyses from the PETDB, GEOROC, and NAVDAT igneous rock databases. Analyses were filtered to have 40-52% SiO_2, <4% Na_2O, <2% K_2O, sums of volatile components or LOI <2%, and major oxide sums excluding volatiles and LOI of 97-102% with all Fe as FeO. Chondrite normalizing factors are from McDonough & Sun (1995).

Eclogites from protoliths with abundant calcic plagioclase are therefore likely to have kyanite, though sufficient orthopyroxene or olivine can instead make more Fe-Mg garnet. It turns out that the most common protoliths for making kyanite eclogites are gabbroic cumulates which, because most Zr is originally in trapped liquid, can have very low Zr concentrations and thus rare or no zircons.

Figure 9.11A shows REE patterns for several eclogites from the coastal WGR of Norway, including one from this Nogva eclogite and two slightly different samples from the Averøy eclogite (Stop 10-2). Although the pattern shapes vary, some of the samples have REE concentrations <10 times chondrite, especially for middle and heavy REE, and the rest have higher REE concentrations. Figure 9.11B shows that rocks with Zr concentrations <20 ppm are mostly mantle rocks or ultramafic and gabbroic cumulates, where all of the kyanite eclogites also plot. The kyanite-free eclogites have Zr concentrations >40 ppm and plot on top of the reference volcanic rock set. The orthopyroxene eclogite plots in

between. This rock is normatively an OPX-rich olivine norite in which abundant olivine and orthopyroxene components prevented formation of metamorphic kyanite. The normative mineralogy and mineral proportions suggest that it is a cumulate, and that its relatively high Zr concentration may be the result of a relatively large proportion of trapped liquid.

Figures 9.11C and D illustrate a different situation for the Sm-Nd and Lu-Hf radiometric systems. The kyanite eclogites in general have higher parent/daughter ratios than common magmatic rocks, and therefore represent good targets in this respect for dating metamorphic mineral assemblages despite generally lower parent isotope concentrations. The moral of this story is that kyanite eclogites are difficult targets for zircon radiometric dating, giving the opportunity *in the field* to select more fertile, if less beautiful, radiometric dating prospects.

Walk back to vehicles at bus garage. Drive back on coast road around south end of Flemsøya and through Longva to junction for causeway. Turn left (west) onto causeway and over bridge to Haramsøya. Proceed south on main road through village of Austnes to ferry terminal. **Back** onto ferry for Lepsøya (Kjerstad).

Leave ferry at Lepsøya (Kjerstad). Turn right on main road above ferry pier toward Hellevik. Pass by Lepsøya Misjonsenter, our overnight location. Continue for 2.6 km drive over plateau and down steep switchbacks to Hellevik. Park in small parking place before driveway to the big lighthouse (now privately owned).

STOP 9-8: (45 minutes) Isoclinal infolds of Blåhø and Sætra Nappes in "sausage rock" basement.
Central segment of Nordøyane. The pavement exposure at Hellevik gives a superb taste of the geology of the central belt of Nordøyane (Figs. 9.2a, 9.12). The basement here is gray gneiss with amphibolite boudins interpreted as mafic dikes ("gray sausage rock"). The dominant cover nappe unit is the Blåhø Nappe dominated by mica schist and garnet amphibolite which occurs in three isoclinal synclinal belts on northern Lepsøya, of which we will see the northern two. The southern schist belt is in direct contact with eclogite-bearing basement of the southern belt (see Stop 9-9). At the north margin of the northern Blåhø belt there is a layer of Sætra quartzite and amphibolite about 1 m thick, which has been traced all along the north end of Lepsøya to Bryggja. There it is exposed on west-facing cliffs in the core of a large south-closing recumbent refold, itself refolded by upright folds (Figs. 9.2a, 9.13). Lineations and major and minor fold axes, shown by domains in Fig. 9.12, trend ENE and are subhorizontal, consistent with other constrictional strain features associated with late transtensional deformation in the region, such as the tubular folds at Stop 9-3. Is it possible that the pattern of major refolded folds in Figure 9.13 is no more than a manifestation of a prolonged episode of tubular folding?

Guided tour. From parking place walk east down grassy path to edge of the pavement. The first rock is kyanite-bearing schist in the middle belt of Blåhø. Walk north between large boulders a short distance to a high spot. This rusty schist consists of quartz, biotite, plagioclase, garnet, pleochroic gray gedrite and sulfides. It is considered to be a metamorphosed hydrothermally altered basalt (see Stop 7-15). Walk a few meters north to contact between amphibolite and very narrow belt of basement, here usefully termed "shredded sausage rock". This belt and this rock extends all the way across the north end of Lepsøya (southern belt with blue color in Figure 9.12) and forms the bulbous core of the big recumbent fold illustrated in Figure 9.13. To the north the basement is in contact with the northern Blåhø belt. About half way across this schist-amphibolite belt there is an isoclinal infold of quartzite and amphibolite that we believe is a tight infold of the Sætra Nappe. There is a 10 cm bed of marble along the south contact, as is common elsewhere in the region (see Stops 8-7, 8-8 and 7-15). At the north edge of the Blåhø belt there is about 1m of folded quartzite-amphibolite and then more "shredded sausage rock". A dip slope in this quartzite forms the south wall of the low bluff on which the small lighthouse is located. Traverse northwestward through the zone of shredded sausage rock to more substantial sausage rock. Here one gets the idea that the mafic layers may be dikes though the evidence is much weaker than in an outcrop on Otrøy (see Stop 7-14). If one walks south on the west side of the small lighthouse there is a pavement exposure in the middle of an area of pools and grass where one can see the Sætra overlying basement in a series of late east-plunging open folds. Here the Sætra has a bed of glassy quartzite 40 cm thick. From this point one may walk direct to the parking place in boggy ground or retrace the route around the point.

Geochemistry of mafic layers (dikes) in Sætra quartzite and gray sausage rock. Figure 9.14 shows data for Sætra Nappe and basement sausage rock dikes seen at Stop 9-8 on Lepsøy. In Figure 9.14A the Lepsøy Sætra dikes plot with data from other Sætra localities in the region. Remarkably, one sample has a REE pattern (Fig. 9.14C) identical to the slight Z-shaped patterns so prominent in the dikes at Oppdal, whereas the other five have patterns like those at Orkanger. This shows that the characteristic REE patterns at Oppdal are also found to the west in the WGR. Spider patterns for Lepsøy Sætra dikes (Fig. 9.14E) are like those seen elsewhere. The basement sausage rocks on Lepsøy (Fig. 9.14B) mostly plot in the same field with Sætra dikes and have Orkanger-type straight REE patterns that are so common in the Sætra dikes of this region (Fig. 9.14D). However, four of the sausage rock dikes plot at higher La_n/Sm_n ratios (1.8-2.5) and have slightly concave-upward REE patterns like Sætra dikes at Ystland. Spider patterns for the Lepsøy basement sausage rocks are just like the Sætra dikes, further indicating that they are part of the same dike swarms of the Middle Allochthon. Though the chemistry of Sætra and basement sausage rock dikes are very similar on Lepsøy and Midøy, the basement sausage rock dikes have slightly different REE pattern slopes (and shapes for some samples) from adjacent or nearby Sætra dikes. This suggests that, prior to Scandian thrust transport, at the time of dike emplacement, the Sætra sandstones and basement gneisses were not adjacent as they are now.
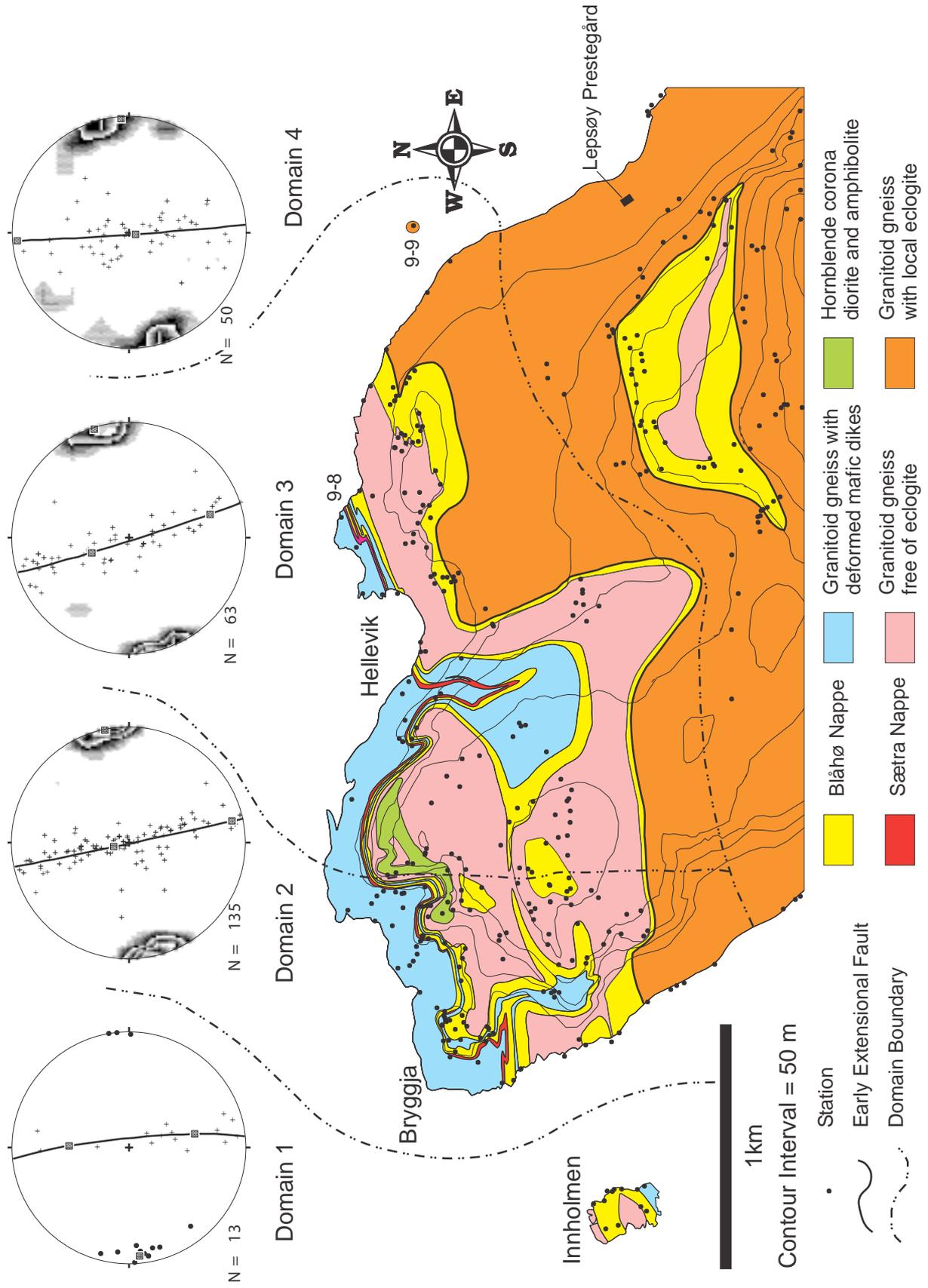
Fig. 9.12 Detailed geologic map of the northern part of Lepsøya from Terry (2000) showing the locations of Stops 9-8 and 9-9 as well as the Lepsøy Prestegård.
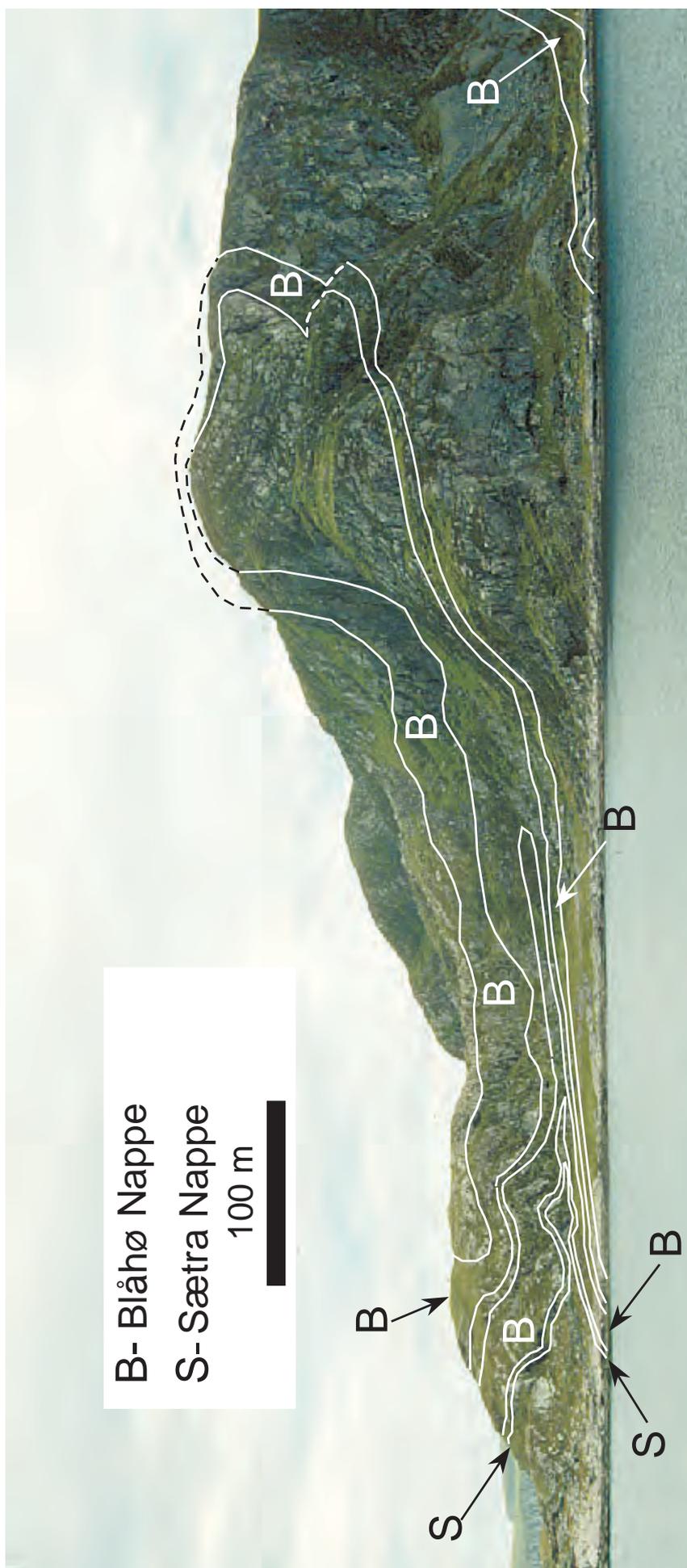
Fig. 9.13 Photograph modified from Terry & Robinson (2003b) taken from Innholmen showing the west-facing cliffs at Bryggja on the northern part of Lepsøya (Fig. 9.12). Shows an along-axis view of a refolded recumbent fold defined by rocks of the Blåhø and Sætra Nappes exposed in early isoclinal synclines. The nappes represent original thrust tectonostratigraphy in apparent normal sequence that was folded into isoclinal synclines within basement. These synclines were then refolded into a recumbent fold that was later folded again by upright folds. The bulbous core of the recumbent fold is occupied by the same thin belt of "shredded sausage rock" observed at Stop 9-8 and shown as the southernmost belt of blue color in Figure 9.12.

Fig. 9.14. This figure shows geochemical data for Sætra Nappe and basement sausage rock dikes on Lepsøy, compared to the same units elsewhere. A, C, E) Nb/La vs. La$_n$/Sm$_n$, REE, and spider diagrams for Sætra dikes on Lepsøy. B, D, F) Nb/La vs. La$_n$/Sm$_n$, REE, and spider diagrams for basement sausage rock dikes on Lepsøy. All data are from Hollocher et al. (2007a). REE normalizing factors are from McDonough & Sun (1995), and MORB normalizing factors are from Pearce & Parkinson (1993).

Leave Stop 9-8 and return toward Kjerstad. Turn left at Misjonsenter and either stop for overnight or continue around sharp left corner and along coastal road to closest parking for Stop 9-9. This stop may also be reached by walking from the Misjonsenter.

STOP 9-9: (30 minutes) Hornblende eclogite of southern segment basement north from Sæt.
This eclogite, in part very fresh, forms nearly all of an islet that is easily reached at low tide (Fig. 9.12). It contains the equilibrium assemblage omphacite-garnet-biotite-hornblende-quartz-rutile (Fig. 9.15). It is representative of eclogites found in the northernmost belt of Mid-Proterozoic basement on Lepsøya, which is in direct contact with Blåhø rocks of the southernmost of three belts in the central segment. Neither basement rocks nor Blåhø rocks of the central belt (Stop 9-8) contain evidence of eclogite-facies metamorphism. On this basis an early extensional fault has been postulated along the contact between the central and southern segments.
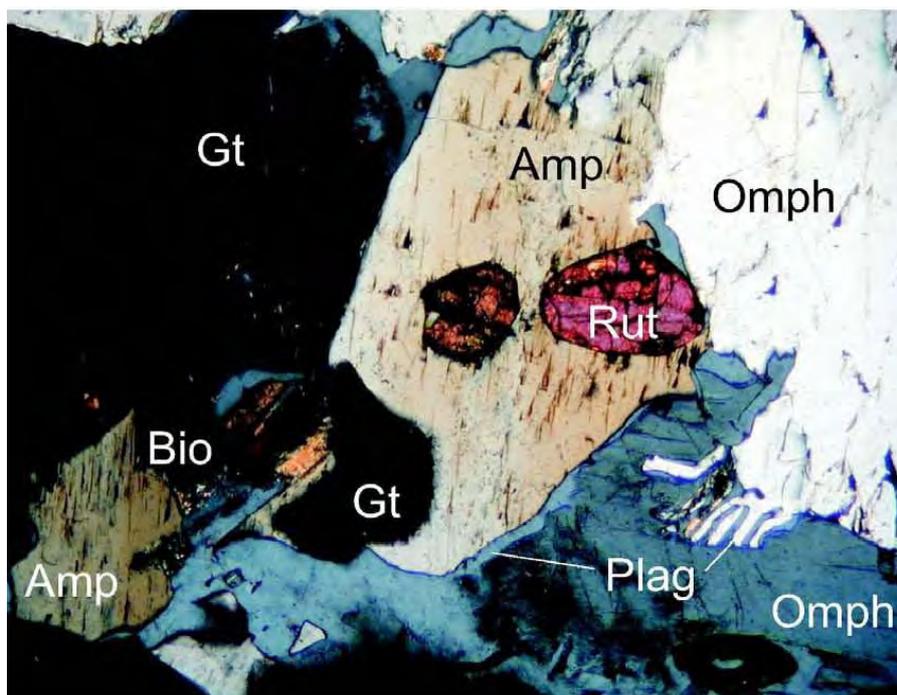
Fig. 9.15 Photomicrograph of the amphibole eclogite at Stop 9-9 in cross-polarized light containing the metamorphic zircons dated by Tom Krogh. The presence of a well equilibrated amphibole eclogite shows that rocks of the southern segment of Nordøyane could not have been to UHP conditions and that there was Scandian imbrication of the crust on Nordøyane. Differences in bulk composition could, however, explain the differences in the assemblages between eclogites in the northern and southern segments.

A sample of eclogite with very fresh coarse omphacite collected by Tom Krogh from this islet provided a small yield of small rounded zircon grains (Krogh et al., 2003, 2004). Data for two concordant fractions of these gave a mean 207/206 age of 414 Ma (414 and 413 ma) and a mean 206/238 age of 412 ma (412 and 412 Ma.) The age of 412 Ma is the only precise eclogite metamorphic age from the southern segment of Nordøyane and only slightly older than the eclogite at Flem in the northern segment (410 Ma). It is also important to note that hornblende is not stable in the UHP eclogites from the northern segment of Nordøyane, which indicates that the southern segment of crust also had a different metamorphic history although the ages are indistinguishable.

Return along coast to Misjonsenter. From Misjonsenter travel south to ferry junction in Kjerstad. Stay straight at junction and follow main road toward village of Rønstad until reaching the entrance to a large quarry on right on the southwest coast of the island. After examining exposures in the quarry, climb down over rubble pile to broad coastal exposures.

STOP 9-10: (45 minutes) Isoclinal infolds of Blåhø kyanite-garnet migmatite and amphibolite, eclogite-bearing basement , southwest Lepsøya.
Southern segment of Nordøyane. The extensive low exposures on this part of the coast are typical of the geology of the southern segment of Nordøyane as well as the adjacent northern part of Vigra. Here there are four isoclinal infolds of Blåhø rocks enclosed in eclogite-bearing basement. Although eclogites have not been found in the Blåhø rocks of Lepsøya, they are known at one location on northern Vigra. On southwestern Lepsøya five narrow Blåhø belts are deformed across a broad east-plunging late anticline (Fig. 9.2a). From the core of the anticline upward these are numbered 1, 2, 3, 4, 5. This stop is concerned with belts 2 and 3 on the northern limb of the anticline.

Outcrops in the quarry. The southernmost rocks in the quarry are well layered amphibolites including garnet amphibolites, and very scarce garnet-biotite schists, assigned to the Blåhø Nappe of belt 3. These are in sharp contact to the north with hornblende-bearing felsic gneisses of the basement, with a thick boudin of eclogite within 1 m of the contact. Most fabric features here equate with late amphibolite-facies transtension.

South contact of Blåhø belt 3. Climb down over rubble pile and walk west along strike to contact region at the southern edge of belt 3, where fine-grained eclogite occurs in basement just below the contact. The overlying Blåhø rocks are dominated by amphibolite, but kyanite schists have been found.

Southeast across basement and Blåhø belt 2. The outcrops just below the rubble pile are granitoid basement gneisses dipping north between Blåho belts 2 and 3. On outcrop surfaces facing south and parallel to the late extensional lineation there are many top-west shear indicators, including asymmetric tails on feldspar porphyroclasts and tail arrangements indicating counterclockwise porphyroclast rotation.

Walk southeast (down structural section) along coast. The contact between basement and the underlying Blåhø of belt 2 is subtle. The first Blåhø rocks are not very distinctive amphibolites, locally with garnet. Associated with these are hornblende gneisses, typically with unusual orange-colored garnets. Beneath these are coarse-grained kyanite-garnet schists containing coarse-grained kyanite-bearing leucosomes. The leucosomes appear significantly less deformed than the matrix that contains features related to late top-west or left-lateral shear. Possibly these rocks contain

an early migmatization recognized elsewhere the Blåhø Nappe or the migmatization occurred during decompression across the muscovite and/or biotite melting reaction(s) associated with late shearing. These hypotheses can be tested by the application of monazite geochronology. Below the kyanite schists are additional amphibolites and then basement gneisses between belts 2 and 1. It is possible in a few minutes to walk to the Blåhø rocks of belt 1 in the core of the anticline, but the exposure is not impressive. Walk back to coast road and vehicles in the quarry.

Drive back southeast, keeping to roads that stay closest to the coast to junction where branch road extends south toward Lauvsundholmen. Cross the very narrow causeway south to Lauvsundholmen. Turn left (east) at south end of causeway, then in a short distance right (south) up a steep rough road to quarry at the top of the island.

STOP 9-11: (10 minutes) Garnet corona gabbro quarry at Lauvsundholmen.
   The very fresh garnet corona gabbro exposed in this quarry appears to occupy the entire island of Lauvsundholmen, and is also exposed on Litle Lauka and Hestøya. The gabbro is structurally above Blåhø belt 5 as described for Stop 9-10 and illustrated in Figure 9.2a. The rock retains primary igneous features of an orthopyroxene-clinopyroxene gabbro, but contains abundant garnet coronas between plagioclase and mafic minerals, and development of secondary Na-enriched Ca-pyroxenes (Jd 12). The coronas are commonly most abundant near grains of ilmenite and magnetite, indicating the importance of Fe in promotion of garnet growth. Locally on the shore there are shear zones formed under high-T conditions in which the gabbro is mylonitic gneiss dominated by ribbons of garnet, Ca-pyroxene, plagioclase and oxides. The omphacitic pyroxene is not characteristic of an early granulite-faces metamorphism and might agree well with the hornblende-eclogite facies conditions observed in eclogites of the southern segment (see Stop 9-9). This idea can be being tested by application of quantitative geothermobarometry.

Leave quarry and cross causeway to Lepsøy. Stay straight (right) at junction and proceed to road to ferry terminal. Either turn right to ferry, or continue on coast road back to Misjonsenter.

Tom Krogh: Science, artistry and profound knowledge of zircons and rocks expressed in Concordia diagrams prepared in 2003 from collections made in person on Nordøyane in 1997 and 1998. All localities except Fjortøft are at stops of this field trip.
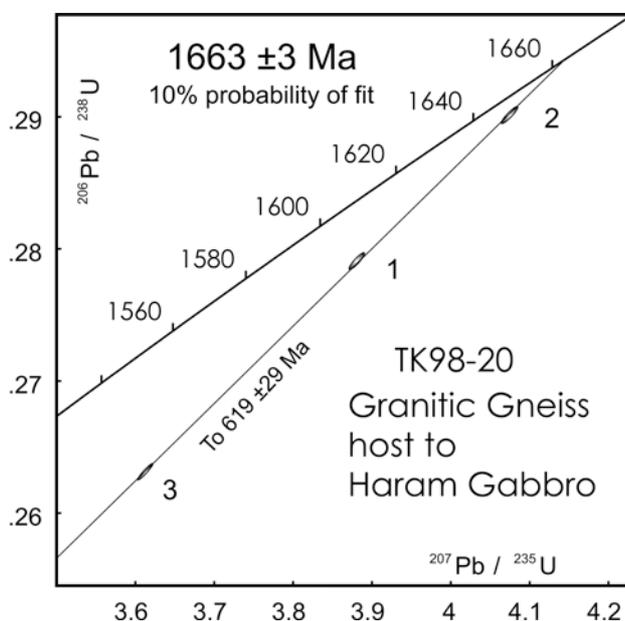

Fig. 9.16 Granitic gneiss mylonite within the eclogite-facies mylonite zone on S margin of the Haram Gabbro (Stop 9-1). Indicates a primary age of 1663 Ma of the granitoid rock intruded by the gabbro, with effects of all later events not fully defined.
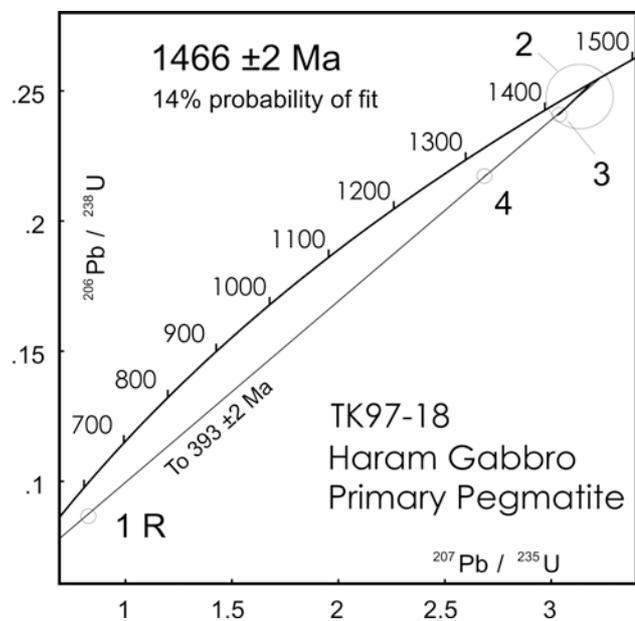
Fig. 9.17 Undeformed primary igneous gabbro pegmatite with a primary age of 1466 Ma. Shows some evidence of Scandian resetting. Collected at the spectacular exposures of pegmatite enclosed in anorthositic gabbro on the outer point of Haramneset (Stop 9-1).
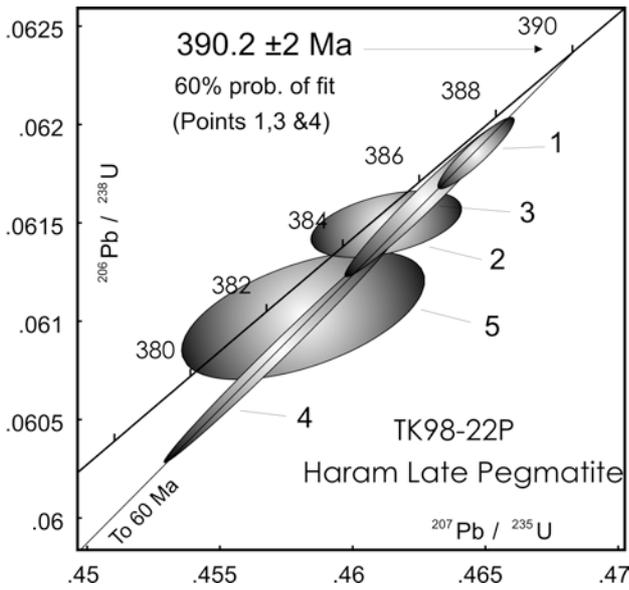
Fig. 9.18 Primary igneous age of 390 Ma for late granitic pegmatite dike cutting cumulate layered gabbro on outer SW point of Haramneset (Stop 9-1). The layered gabbro is converted to layered amphibolite along the dike contact.
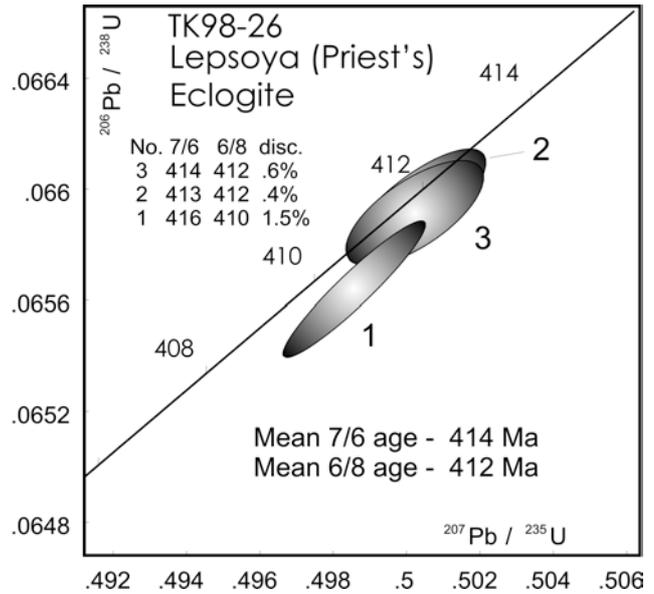

Fig. 9.19 Age of 412 Ma for metamorphic zircon in coarse omphacite-hornblende eclogite of the southern segment of Nordøyane exposed at low tide on an islet offshore from Lepsøy Prestegård (Stop 9-9).
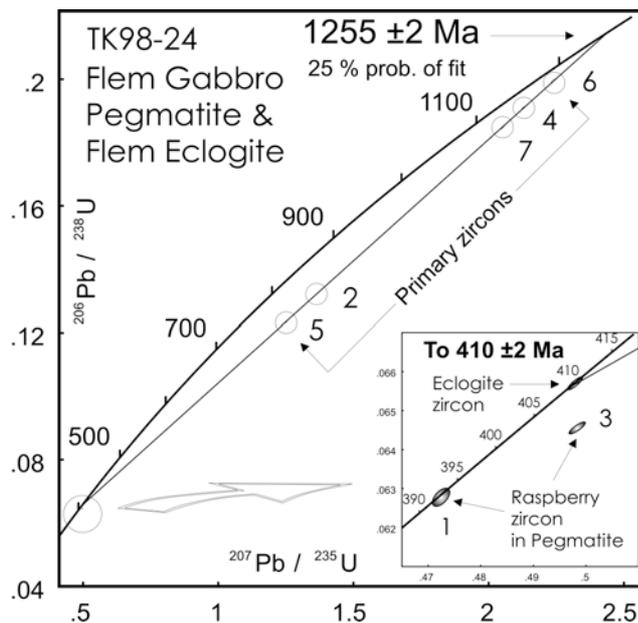

Fig. 9.20 Primary igneous age of 1255 Ma from zircon in gabbro pegmatite dike cutting Flem Gabbro (Stop 9-4). In the field the dike locally shows effects of both eclogite-facies and later amphibolite facies overprinting. Also shown are ages of metamorphic zircons from the eclogitized north margin of the gabbro.
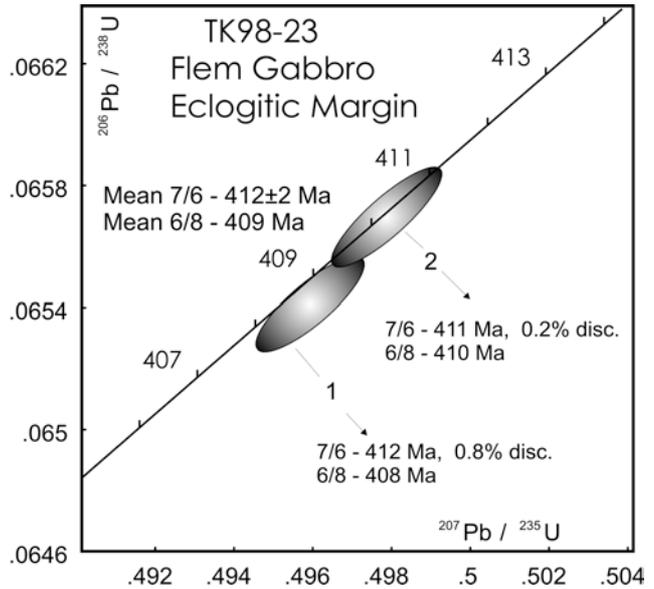

Fig. 9.21 Close-up view of concordant ages of scarce eclogite zircons from the eclogitized north margin of the Flem Gabbro (Stop 9-4). The age of 410 Ma dates the development of eclogite-facies minerals during deformation of the north margin of the gabbro, and, indirectly, the development of steeply plunging eclogite-facies strain fabrics in the augen orthogneiss country rock of the gabbro and the adjacent Ulla Gneiss with its enclosed Kvalvika garnet peridotite (Stop 9-5) with its modeled P-T path passing through 790°C, 40 Gpa (diamond conditions).
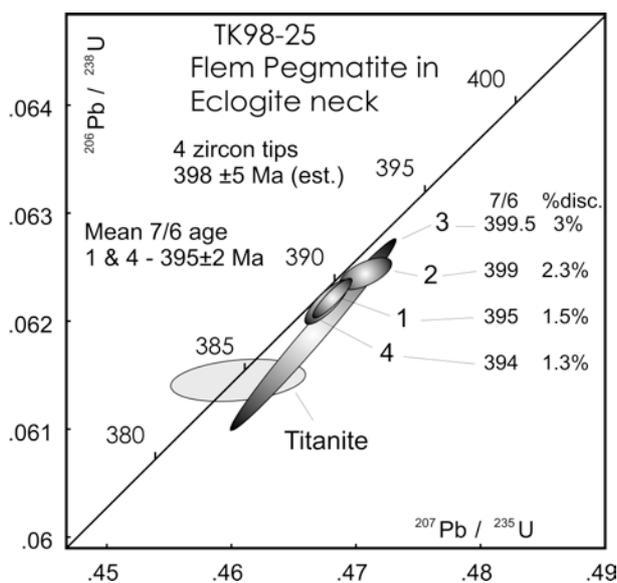
Fig. 9.22 Igneous zircon dated at 395 Ma from granitic pegmatite collected in a steep extensional neck line of an eclogite boudin related to subhorizontal stretching lineation near the northern tip of Flemsøy (Stop 9-6). This is a common age in the region for products of late melting during return from eclogite-facies to amphibolite-facies conditions.
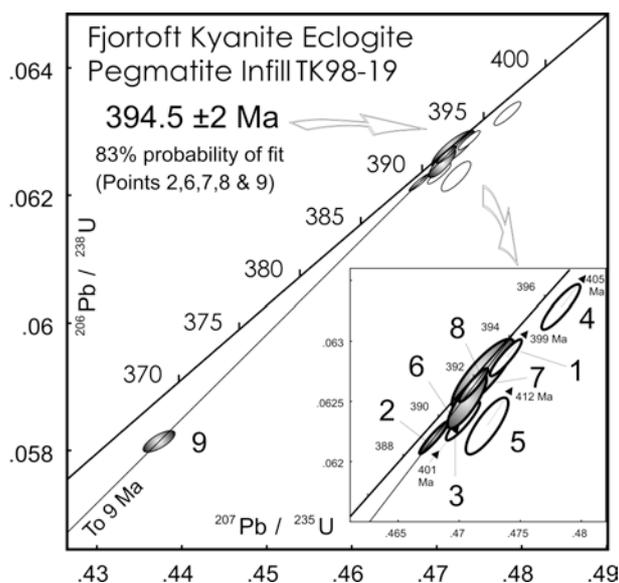
Fig. 9.23 Igneous zircon dated at 394.5 Ma from pegmatite in extensional neck line against a coarse kyanite-zoisite eclogite boudin with coesite pseudomorphs, on northern coast of Fjørtoft, that yielded estimated conditions 820°C, 3.4-3.9 GPa (diamond conditions). The eclogite itself yielded no zircon during intense search, probably explained by a very low Zr content (Fig. 9.11), because the eclogite protolith was a plagioclase-rich igneous cumulate.

Recollections of Tom Krogh's early days of robust invention

The secret to Tom's analytical success was a dedication to utter simplicity in every step of the process. I watched at the Geophysical Laboratory while he perfected the "Krogh bomb", a Teflon crucible with an interference taper for a lid inside a steel, screw-capped container. With this assembly, containing hot HF vapor at about a kilobar pressure, he could dissolve almost anything in a micro-environment with incredibly low blanks. And then at DTM, where the mass spectrometers lived, he perfected the micro-environment of the positive-pressure fume hood that, again, allowed digestion at low blank levels in a simple way. Micro ion-exchange columns followed.

Most entertaining, however, was his clever treatment and selection of the best zircons – those that might land on Concordia. First it was abrasion, using the miniature, emery-lined, air-driven, squirrel-cage sphere grinder developed for crystal structure analysis by W. L. Bond. This process wore away the fractured and otherwise compromised outer parts of the crystals, and produced beautiful pink spheres under the microscope, but of course wasted a lot of zircon and time. Then Tom got to manipulating grains under the microscope, either picking homogeneous and clear ones for digestion, or in the case of suspected multi-stage histories, physically breaking off rims and separating them from cores for analysis. In some samples, he could chose a certain color to get at the relevant age. In all this, of course, he would need only one crystal or one a few good rims for analysis, so refined were his methods. He was steeped in the lore of zircons.

At ROM he had a block of galena sitting outside the door, just to call attention to his successful shielding of samples and procedures from unwanted lead. He was fond of saying, Well, we don't wipe our shoes in the reagents, but with a little care we get satisfactory blanks [– i.e., the lowest in the business!].

On top of all this was an ironic, calm, self-deprecating guy who was comfortable inside his own skin, a bear for work, satisfied that he had done the best he could, and happy with the possibility that it was unsurpassed. He analysed two Kiglapait zircons and found them within a million years of each other and was willing to bet a case of beer he could put another point in the same group. Good thing for us he wasn't playing darts in a pub, or he'd have cleaned our clocks.

S. A. Morse, New England, USA, July 29, 2008 (exactly 978 years after Stiklestad)

Day 10: GEOLOGY ON THE ROUTE: LEPSØYA-BRATTVÅG-SOLHOLMEN-SVARTBERGET- ATLANTIC HIGHWAY-AVERØY-HALSA-TRONDHEIM

Sunday August 24

by Peter Robinson, Hans Vrijmoed, Tom Krogh and Kurt Hollocher

    General Route: There is the possibility to take a Sunday morning water taxi from Lepsøy to Roald and then a short taxi ride to Ålesund, Vigra Airport, for flights to other Norwegian cities and out of Norway. These people will not be involved further in Day 10. For the remainder there will be an early morning departure from Lepsøy Misjonsenter to the ferry Lepsøya-Skjeltene. We will then retrace our route of Day 7 without formal stops, including the ferries Brattvåg-Dryna and Solholmen-Mordalsvågen. From Mordalsvågen we take a sinuous coastal route involving a stop at Svartberget near Bud, a crossing of the Atlantic Highway to Averøy, two stops on Averøy, a return trip over the Atlantic Highway, thence through Eide and to Gjemnes Bridge where we join Route E39. We then follow E39 all the way to Trondheim, with one stop near a floating bridge, the ferry Kanestraum-Halsa, and a final stop in the mountains before descending to Trondheimsfjord at Orkanger, thence retracing part of our route of Day 5 back to Trondheim. Participants will be delivered back to the Vandrerhjem or hotel of choice for the evening and departures from Trondheim Værnes Airport Monday morning August 25.
    Ferry Lepsøya to Skjeltene: 8:15-8:35 (Driving time to next ferry is exactly 10 minutes!)
    Ferry Brattvåg to Dryna: 8:50-9:15. (Non-stop driving time Dryna to Solholmen ferry is 35 minutes. There are 25 minutes to spare for informal stops en route.)
    Ferry Solholmen to Mordalsvågen: 10:15-10:30.
    Ferry Kanestraum to Halsa: (20-minute crossing): **15:10, 15:30, 16:00, 16:30, 17:00, 17:30, 18:00, 18:30, 19:00**. (Driving time Halsa to Trondheim is about 2.5 hours.

Road Log Day 10

8:00 Depart Lepsøya Misjonsenter for Kjerstad ferry terminal.
8:15 Drive onto ferry for Skjeltene.
8:35 Leave ferry at Skjeltene. Turn left on Route 659 toward Brattvåg. After 11.9 km on Route 659 turn left (north) for Brattvåg ferry terminal (to Dryna).
8:50 Drive onto ferry for Dryna.
9:15 Drive off ferry at Dryna and proceed east along Route 668, passing location of Stop 7-15, which could be taken at this time if not earlier (15 minutes). Cross Dryna-Middøya bridge. Large roads cut on right showing pegmatites in Blåhø mica schist / amphibolite. From the western part of Midøy there is a view ahead of huge cliffs. which are mostly biotite schist and amphibolite of the Blåhø Nappe with pegmatite. There appear to be a few thin layers of basement gneiss. There is a view to left of Drynasund lighthouse resting on gabbro with recrystallized shear zones (Griffin & Råheim, 1973). We pass other locations described in the road log for Day 7, including the old stone wall at "Riksgrense".
Stay on 668 at north corner of Middøya and follow it south toward Midsund. At next junction for Ramsvik turn left toward Midsund and cross bridge to Otrøy. Turn sharp left still on 668 at junction on steep corner in Midsund. After 2.7 km pass the left turn to Ugelvik Quarry (Stop 7-2) Continue east toward Solholmen Ferry. After about 14 km road bends around northeast corner of Otrøy and south toward Solholmen. About 1.9 km further, one can park on right opposite factory road entrance and examine a road cut in the basement augen orthogneiss.

Informal Stop: (10 minutes) Augen orthogneiss.
    This rock and its highly deformed equivalents, also seen at Stops 9-4 and 9-6, which dominate the exposure on northern Midsund and part of the Molde peninsula, was studied extensively by Carswell & Harvey (1985). It is interpreted as an early Mesoproterozoic rapakivi granite that can be followed through progressive deformation and recrystallization until it can scarcely be differentiated from other more normal basement gneisses. Carswell & Harvey obtained a Rb-Sr whole rock isochron for these rocks of 1506±22 Ma., and more recently the same rock at Fanghol, across Julneset, within sight of this stop, has yielded a U-Pb zircon age of 1508 Ma (Tucker et al., 1990) similar to the U-Pb zircon age obtained on the porphyritic mangerite at Flatraket near Selje. In practice this basement augen gneiss is usually fine grained and the microcline phenocrysts are flattened and polycrystalline as a result of relatively homogeneous deformation, (Carswell & Harvey, 1985). Within this small road cut it is possible to see nearly every phase of progressive deformation of this porphyritic rock until the phenocrysts are no longer recognizable.

Continue south 1.6 km to Solholmen Ferry Terminal and enter waiting line. If time permits, a relict orthopyroxene eclogite is exposed on shore behind the no-parking sign on north side of terminal.

Informal Stop: (15 minutes) Coarse orthopyroxene eclogite with thick amphibolite rim.
    Most of this eclogite, surrounded by augen orthogneiss, was destroyed during construction of the ferry terminal. Carswell et al. (1985) completed detailed mineral analyses on this rock and Cuthbert & Carswell (1990) list estimated conditions of formation at 760°C and 18-20 kbar. Several other small eclogites occur several hundred meters south of the ferry terminal, but no eclogites have been found by Robinson anywhere on the south coast of Otrøy and one gabbro
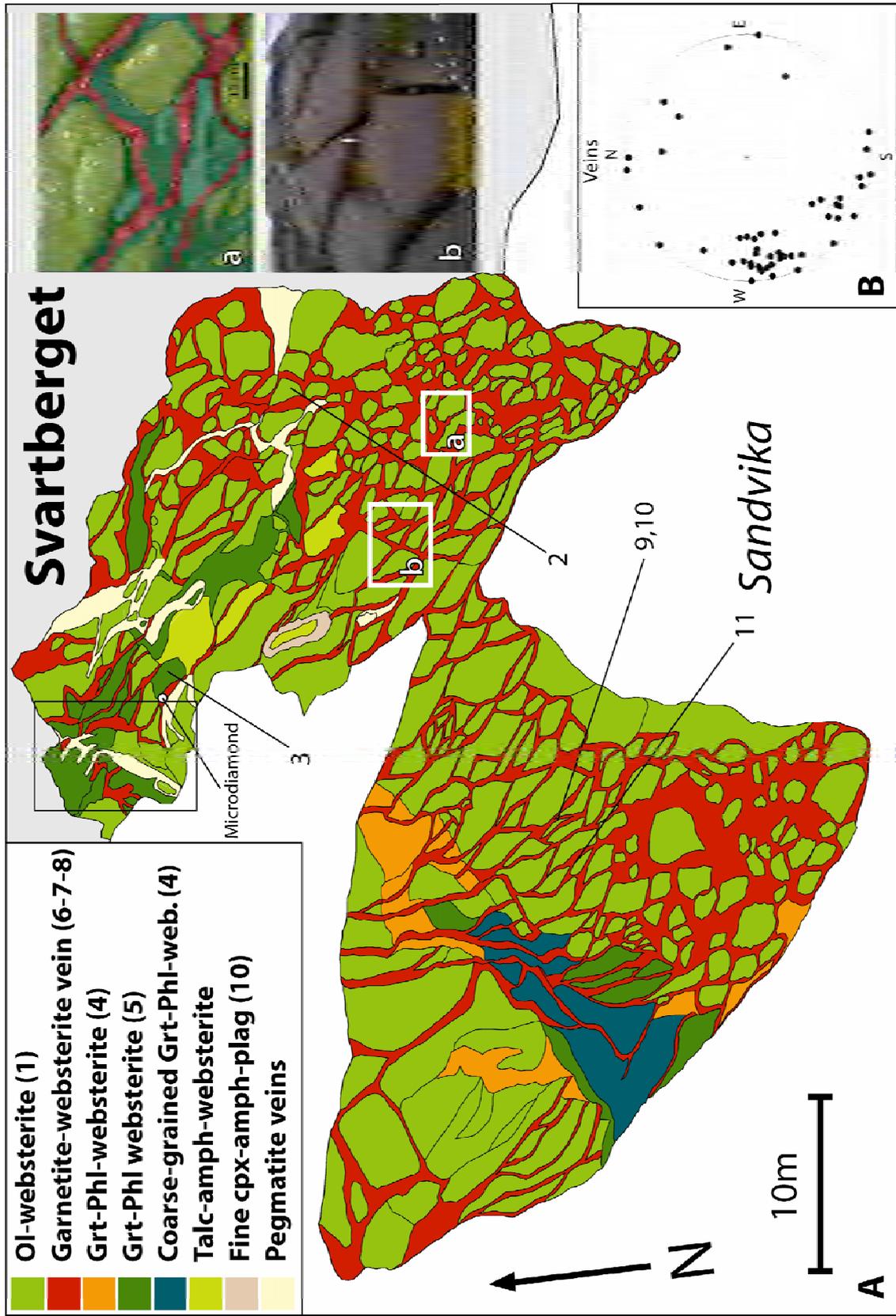
Fig. 10.1 Simplified geological map of the Svartberget olivine-websterite body. Numbers correspond to zones described in the text. Black rectangle in northwestern part of the body outlines the area in Fig.2.

Legend:
- Ol-websterite (1)
- Garnetite-websterite vein (6-7-8)
- Grt-Phl-websterite (4)
- Grt-Phl websterite (5)
- Coarse-grained Grt-Phl-web. (4)
- Talc-amph-websterite
- Fine cpx-amph-plag (10)
- Pegmatite veins

shows only very slight garnet coronas. The implication is that there are major metamorphic breaks running the length of the islands of Midsund.

On the next point to the north is a narrow belt of amphibolite, schist, calc-silicate and one 20 cm layer of buff-weathering pure marble. This is interpreted as an infold of the Blåhø Nappe. The same belt is exposed at Mordalvågen ferry terminal and the same buff marble appears in a road cut about 1 km to the east. This belt is likely the same schist/amphibolite belt that occurs just south of the Midsund mylonite zone further W (Figure 7.12).

10:15 Drive onto ferry Solholmen to Mordalsvågen.
10:30 Drive off ferry at Mordalsvågen. Note large road cuts of garnet-biotite schist and amphibolite of Blåhø Nappe on right. After driving 1.4 km east from the terminal (note thin marble on right) turn left (north) on Route 662 toward Hollingsholmen ferry terminal. This highway brings one past the augen orthogneiss outcrop yielding a U-Pb zircon age of 1508 Ma (Tucker et al. 1990). Do not turn off for the ferry but continue northeast and then east, eventually joining Route 64 north from Molde at Malmfjorden.

Drive north through Malmfjorden for several km to a traffic circle. At the circle take the exit on Route 663 for Elnesvågen. In center of Elnesvågen, do not turn right on Route 663 but stay straight on Route 664 toward Bud. After 13.7 km turn off on left into narrow road and parking for walk through fields to Stop 10-1 Svartberget, a prominent point on the coast.

STOP 10-1: (approximately 90 minutes) Diamond-bearing Svartberget olivine-websterite (Vrijmoed)

The Svartberget olivine-websterite body was described by Vrijmoed et al. (2006). The body is cut by a conjugate set of metasomatic fractures filled dominantly with coarse-grained garnet-phlogopite-websterite and garnetite. Standard thermobarometric techniques based on electron microprobe analyses yield pressure (P) – and temperatures (T) estimates around 4.0 GPa, and 800°C for the olivine-bearing body and 5.5 GPa, and 800°C for the websterite consistent with UHP conditions. The websterite contains microdiamond (Vrijmoed et al., in press) and confirms the UHP conditions estimated with thermobarometry.

A simplified field map (Fig. 10.1) shows the network of garnetite-websterite veins in the olivine-websterite body. The outcrop consists of a northern and a southern part. The southern part is a small island and is accessible when tides are low. The northern part consists of remnants of olivine-websterite rocks with garnetite-websterite between the remnant blocks. Erosion and vegetation obscured some parts of the outcrop but careful mapping reveals most of the lithology and structure. Aerial photographs and field measurements show a conjugate fracture pattern in the body.

Along these fractures a metasomatic column developed, dominated by garnet-websterite and garnetite. Based on detailed mapping eleven different metasomatic zones can be recognized. From wall rock to core these are: 1) olivine-garnet websterite 2) clinohumite bearing olivine-garnet websterite 3) coarse grained phlogopite-orthopyroxene-garnet websterite 4) coarse-grained phlogopite clinopyroxene-garnet websterite 5) phlogopite garnet-websterite 6) garnet-websterite 7) inclusion rich garnetite 8) garnetite 9) amphibolite 10) clinopyroxene-plagioclase rock 11) plagioclase-amphibole pegmatite.

1)   This is the wall rock of the veins/fractures. It is present as blocks of rock standing out in relief in the outcrop and it is the dominant rock type. The only fluid phase present in minor amounts is an unidentified yellow clay mineral, which has a composition close to serpentine.
2)   Along the borders of a number of blocks of the main body a 10cm wide rim exists. This rim is marked by cracks perpendicular to the edge of the block and by the appearance of Ti-clinohumite in the rock. These zones are only observed in the northern part of the body (Fig. 10.1.).
3)   Some of the original blocks of the body consist now only of orthopyroxene-rich-phlogopite websterite and this can be observed in at least two or three locations in the outcrop (Figs. 10.1, 10.2.).
4)   This zone is best observed in the southern part (the island, Fig. 10.1.). Here it occupies a large volume of the outcrop and does not follow the fracture pattern clearly. It varies in width from 30cm-4m. Grain size of the pyroxenes ranges from 10-40 cm in this zone.
5)   Observed generally in every fracture in the body we find this zone preserved usually in a 1-5cm zone along the fractures (Figs. 10.1,10.2,10.3). In some places the zone is more irregular and wider. Microdiamond is found in a sample of this zone in the northern part of the outcrop (Fig. 10.1).
6)   A narrow zone (<0.5cm) of phlogopite free (garnet-)websterite is present in hand specimens of the garnetite veins (zones 7/8) on both sides of the garnetite. It can be best observed in handspecimens of the garnetite (Fig.10.3a).
7)   In some garnetite veins a rim of garnets with inclusion rich cores is present. A garnetite vein in the southern part within the coarse-grained websterite (zone 4) shows this best (Fig. 10.3b).
8)   Generally observed in every fracture, the garnetite forms usually the core of the fractures (Fig. 10.2, Fig. 10.3a,b). The zone has an average width of 3-5 cm but has in many places very irregular shapes, which can be observed in several places of the outcrop in the north (Fig. 10.2).
9)   In a few cases the garnetite has developed a core of zone 10 which is rimmed by a thin zone of amphibolite (several mm's in thickness) (Fig.10.3c).
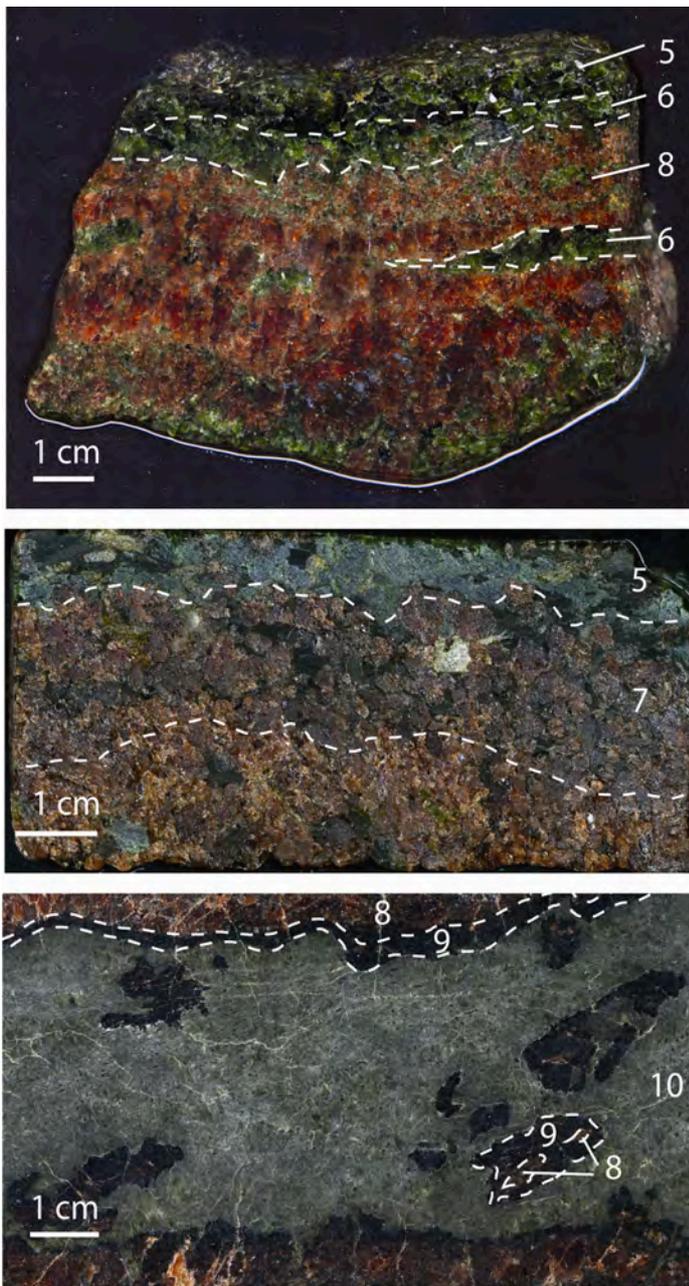
Fig. 10.3 Scans of hand specimen showing the detailed zonation of the garnetite-websterite veins. Numbers indicate the different zones described in the text. Dashed lines indicate a boundary between a zone. From top to bottom fig. a, b and c.

10)Rimmed by zone 9, the garnetite (zone 7/8) has in a few cases reacted to form a fine grained (~50mu) intergrowth of plagioclase, amphibole and clinopyroxene (Fig. 10.3c).
11)The core of zone 10 consist in some cases of coarse grained pegmatites (amph,cpx,plag) in some locations.

In the field we can observe the individual zones in several locations in the outcrop (Fig. 10.1). The northern part commonly shows a sequence of zones number 1, 2 5, 6, and 8. locally zone 3 is observed as well. The southern part commonly shows a sequence of zones 1, 5 and 8. Zone 3 and 4 and 7 are preserved at some localities in the southern part. Zone 9, 10 and 11 are present mainly in the southern part on a few localities.

The microdiamond is found in a sample from zone 5 taken from the northern part of the outcrop (Fig.10.1). A second set of fractures crosscuts all other features and metasomatism along these fractures can be observed along incoming melts/pegmatites from the gneiss. This can be studied best in the northernwestern part of the outcrop (Fig. 10.2.). The main minerals that form a result of this interaction are talc, chlorite and amphibole.

Walk back to vehicles. Return to Route 664 and turn left (north) toward Bud. After 2.2 km turn right inland onto Route 232 toward Hustad.
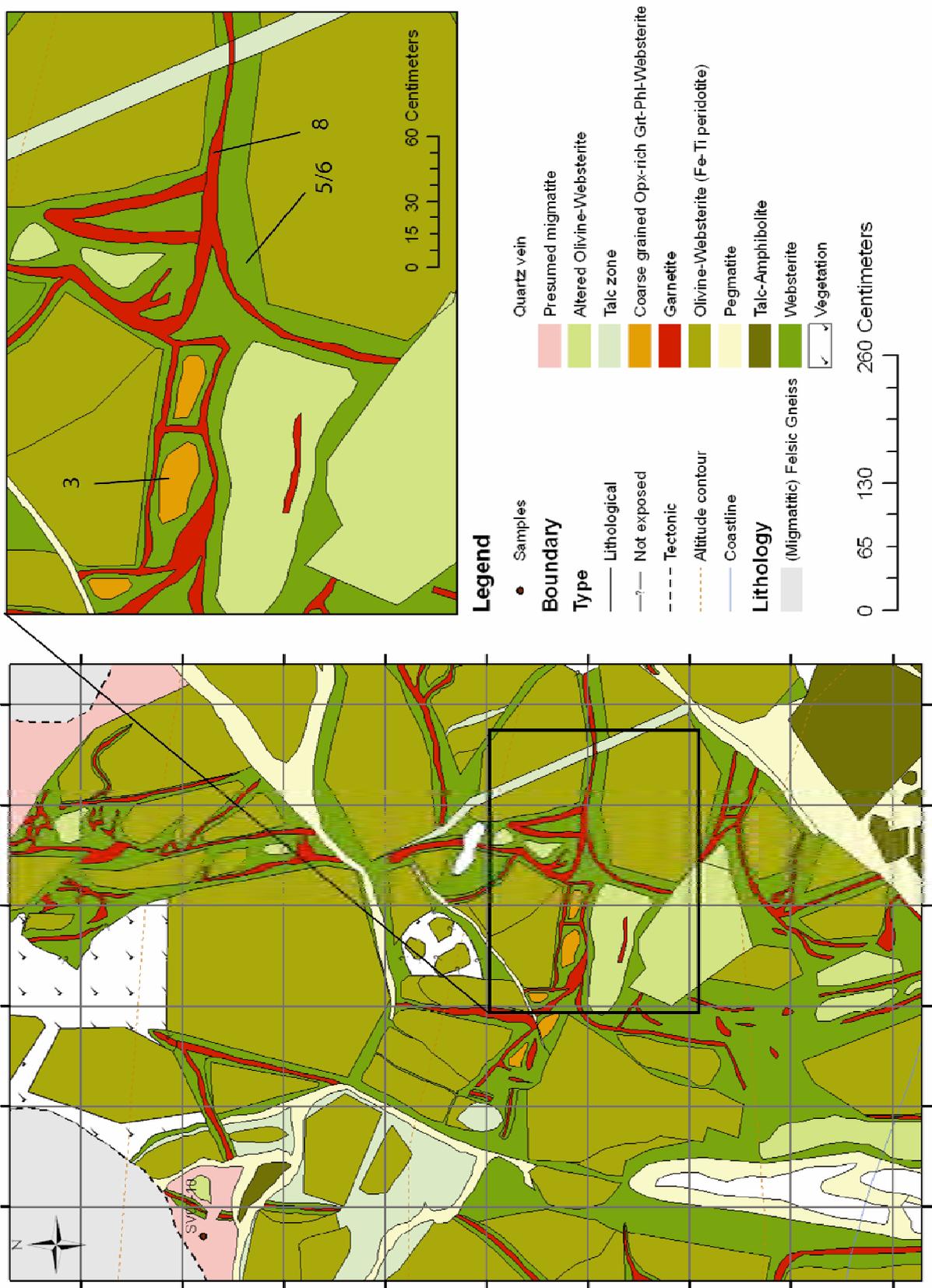
After 15.9 km reach junction with Route 664 and turn right toward Farstad. After another 4 km turn left on Route 663 which leads east to Route 64 at Vevang and beginning of the Atlantic Highway.

Drive east over bridges, causeways, and islands of the Atlantic Highway beside the stormy Hustadvika, until reaching the big wooded island of Averøya. At major T junction on Averøya turn right (west) toward the western end of the island. On a ridge crest in open country as the road bends toward the southwest, drive into a prominent roadside quarry close on the left of the highway.

STOP 10-2 (20 minutes) Layered eclogite in the interior of the Averøya body. (Robinson)
The Averøya eclogite, although still not mapped in detail, appears to be about 6 km long and about 3 km wide. Here and in many places the prominent layering is sub-horizontal. The layers alternate between garnet-, pyroxene- and quartz-rich. Within this large pervasively eclogitized mass there appear to be no textural relics suggesting the body was once a gabbro. A mafic volcanic protolith seems more likely, and a uniform water content could explain the uniform eclogitization. Based on field relations of other layered eclogites in this district, where there is an intimate association with garnet-mica schist and marble, it seems likely that this eclogite should be assigned to the Blåhø Nappe. Also limited C and Sr isotope data on one of the marbles indicate a Paleozoic rather than Proterozoic age.

Fig. 10.2 Detailed map of the area outlined in Fig. 1. It shows the textural relationships of the garnetite, websterite (together forming the red unit in Fig. 1) and the olivine websterite. Numbers indicate zones described in the text. Outlined area shows a detail of this map where the brown unit represents zone 3. Altered olivine-websterite, usually looks as blocks of the main body, but in many cases no olivine is present and the mineralogy resembles more the websterite in the fractures.

Eclogite petrology. In thin section the pyroxenes contain abundant oriented rods of plagioclase (Robinson et al. 2008). Sparse fresh omphacite has the composition $Jd_{38}Ts_5Wo_{25}En_{25}Fs_7$. In coarse 'graphic' plagioclase symplectite the pyroxene is $Jd_{11}Ts_{11}Wo_{35}En_{35}Fs_8$, with plagioclase An_{18-19}, and in fine symplectite $Jd_8Ts_6Wo_{41}En_{37}Fs$ with plagioclase An_{32}. Eclogite garnets are zoned from $Pyr_{39}Alm_{42}Grs_{18}Sps_1$ cores to $Pyr_{37}Alm_{47}Grs_{15}Sps_1$ rims. Garnets in rare quartz-rich layers are surrounded by zoned coronas consisting of moats of sodic plagioclase and necklaces of orthopyroxene ($Ts_1Wo_0En_{60}Fs_{38}$) and clinopyroxene ($Jd_9Ts_1Wo_{41}En_{36}Fs_{12}$). The moats are zoned from An_{26} near garnet to An_{12} near necklaces. These garnets are zoned differently with $Pyr_{42}Alm_{42}Grs_{15}Sps_1$ cores to $Pyr_{34}Alm_{49}Grs_{16}Sps_1$ rims. All garnets show a slight rim increase in Mn consistent with resorption and also an increase in Fe consistent with garnet/pyroxene exchange during cooling. This Fe effect is much greater in the garnet within quartz than within the eclogite, however Ca changes little in quartz, but decreases in eclogite garnet. Pseudomorphs of coesite have not been identified.

Several hypotheses emerged to explain this unusual moat-necklace structure, also reported by Julia Baldwin (ms) in the Snowbird high-P tectonic zone, Saskatchewan. One is that it is a prograde breakdown product of hornblende + quartz to give garnet + plagioclase + two pyroxenes. This, however, would not explain the unusual phase distribution. Another is that it is a breakdown product of UHP sodic garnet and SiO_2. Using probe analyses and modal estimates of the garnets, moats and necklaces, such a garnet would have had a composition 56% normal aluminous garnet (Ca,Mg)$_3{}^{VI}Al_2$ Si_3 O_{12}, 39% Na majorite garnet (Ca,Mg)$_2$ Na ^{VI}(Al, Si) Si_3O_{12}, and 5% majorite garnet (Ca,Mg)$_3{}^{VI}$ (Mg, Si) Si_3O_{12}. Limited experimental data of Gasparik (1989) indicates that sodic garnet can be stable in the system Pyrope-Enstatite-Jadeite at P 10 GPa and T 1450^oC, that seem well beyond what could be possible here. However, the experimental work was designed mainly to test the breakdown of pyroxene to garnet under mantle transition conditions, and did not include the Fe that is important here and could stabilize garnet to lower pressures. A third alternative is that during relatively high-T decompression, element mobility was sufficient to allow the garnet within the quartz layer to reach diffusional equilbrium with the eclogite, in particular providing Na released from omphacite to construct the plagioclase moat. If correct, this implies diffusion for at least 1-2 cm during re-equilibration and exhumation at constant or increasing T, which could explain the poor preservation of HP-UHP features in these eclogites.

Turn left out of quarry and continue southwest then south onto another low ridge. Turn sharp right and drive down hill to old Tuvik ferry wharf.

STOP 10-3 (20 minutes) Coarse pegmatite at contact between Averoya eclogite body and country rock gneisses on the shore. (Krogh, Robinson)

The pegmatite is exposed in the cut on the right next to the shore and the country rock is exposed at the shore. Pegmatite samples were collected by Arne Råheim on a formal eclogite field trip in 1987, when the party was waiting to take the ferry from Tuvik to the mainland. This ferry has been closed since construction of the Atlantic Highway. A coarse variety of eclogite with garnets up to 1 cm was sampled by Tom Krogh in 1997 (TK97-20) 30
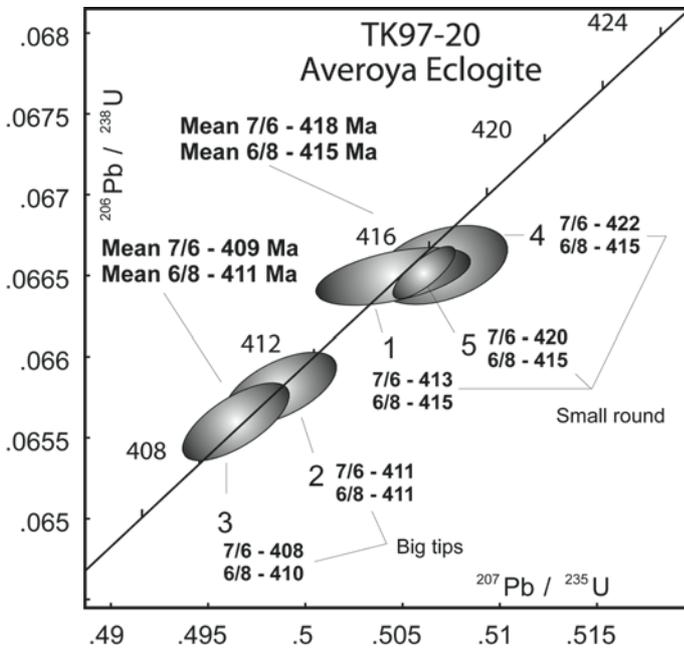

Fig. 10.4 Concordia diagram showing results of U-Pb isotopic measurements on zircons from Averøy eclogite at Stop 10-4. The older ages are considered to represent the time of initial eclogite crystallization. The younger ages are from two euhedral zircon tips and are considered to represent the time of grow of late coarse garnet, hornblende and plagioclase.
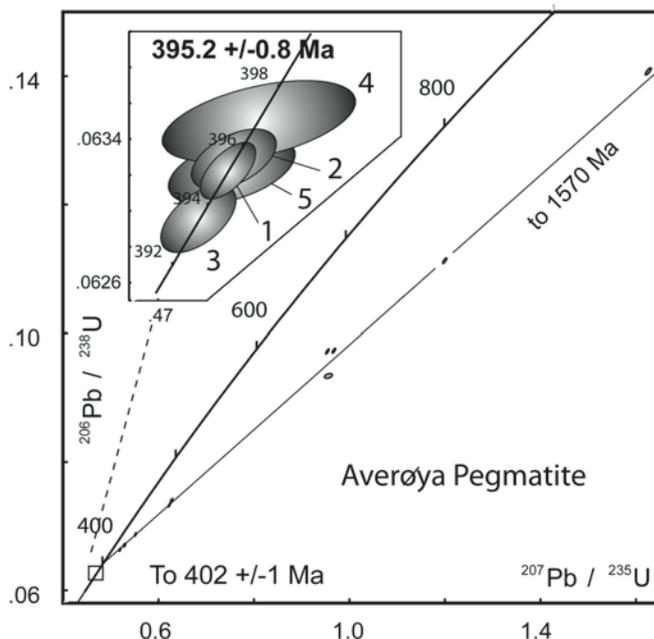

Fig. 10.5 Concordia diagram showing results of U-Pb isotopic measurements on zircons from an extensional pegmatite in Averøy eclogite at Stop 10-4.

meters north from the pegmatite and 2m up a 3m-high wall. Small 3x8cm pods of probable hornblende and plagioclase occur below the sample site. The resulting U-Pb zircon ages were first reported in a 2003 Eclogite Field Guidebook (and in a Conference Abstract). The locality is so-far unique in the WGR, because it is the only location where precise U-Pb zircon ages have been obtained both on eclogite and on a late extensional pegmatite in the same outcrop.

The eclogite is unique in having two stages of metamorphic zircon growth (Fig. 10.4). Three concordant fractions of the older, small rounded grains have a mean 207/206 age of 418 Ma (413,420,422 Ma) and a mean 206/238 age of 415±2 Ma (415, 415 ,415 Ma), and probably formed during initial eclogitization. Tips from two large euhedral grains have a mean 207/206 age of 410 Ma (408, 411 Ma) with 206/238 ages of 410 and 411 Ma that may date the late growth of hornblende and plagioclase. The precise mean 206/238 age is the best for the eclogitization because none of the five analyses here exhibit discordance and the 207/206 age has a large 2 sigma error of +/- 5 Ma.

The post-eclogite pegmatite (Fig. 10.5) gave 5 concordant analyses with 3 giving a mean 207/206 age of 395.9 and a mean 206/238 age of 395.2 ±1 Ma. These concordant analyses were only obtained after abundant U-rich inclusions in the low U host zircon were successfully avoided using an etch technique. Ten earlier analyses were contaminated to varying degrees by 1500-1600 m.y. inheritance projecting toward a lower intercept of 402 ±1 Ma. The upper intercept age may suggest that the source of granitic melt may have been in local basement rather than in Phanerozoic strata.

Return by same route across the Atlantic Highway, then turn south on Route 64 toward Eide. On this road we pass road cuts in eclogite at Lyngset, and see the huge marble quarry at Visnes, where the marble rests on layered eclogite.

Just beyond the center of Eide turn left on highway, which leads south across a stream and then east along the south shore of the fiord toward Gjemnes and junction with Route E39.

Junction with E39 is at west end of large suspension bridge connecting with Bergsøya. At the east end of the bridge pay toll and then take the exit for Trondheim (do not take tunnel to Kristiansund!).

Drive several kilometers on new fast road to the west end of Bergsøya and across the next channel on a floating bridge. At east end of floating bridge turn right into rest area and park by a closed kiosk. Walk east with care along the highway into a giant road cut. Total driving/stop time from Gjemnes Bridge to Kanestraum Ferry is 37 minutes including 15 minutes at Stop 10-4.

STOP 10-4: (15 minutes) Foliated granitic gneiss with a "train wreck" of eclogite boudins. (Krogh, Robinson)
The north wall of the cut shows layered eclogite mainly in section. On the south wall there are huge foliation surfaces showing the "porpoising" effect of boudinage on the sub-horizontal extensional lineation. Small pegmatites are well developed at some of the boudin neck lines. One of these was collected by Tom Krogh and zircons separated (Sample TK97-15, Figure 10-6). They gave a mean 207/206 age of 395.6 (5 points 0.2, 0.2, 0.9, 1.1, and 0.9 % discordant) identical to the post-eclogite pegmatite at Averøya 26 km to the west. By combining this and the Averøya data, an interval of 14 million years between final zircon growth in Averøya eclogite (410 Ma) and extension-related ductile flow (395.5 Ma) is defined, with first eclogitization at Averøya at least 5 million years earlier.

Leave Stop 6-8 and continue toward Kanestraum. After 5.6 km cross curved Straumsund Bridge and 0.9 km further turn sharp left on E39 toward Trondheim. Kanestraum ferry terminal and kiosk (Excellent polser!!) is 8.6 km further. Target times are **15:10, 15:30, 16:00, 16:30, 17:00, 17:30, 18:00, 18:30, 19:00** and on Sunday afternoon a wait is possible**.** When appropriate drive onto ferry for Halsa.

Drive off ferry at Halsa. Standard driving time from Halsa to Trondheim is 2.5 hours all on Route E39, except the final 19 km on E6. There will be two intermediate stops of 10 minutes each. At T-junction in Betna 7.2 km from Halsa turn left toward Trondheim. There is a 10-minute rest stop with toilet on Valsøya 18 km further and 1.3 km beyond we cross the curved Valsøy bridge.

Beyond this in 13.4 km is the Fylkke boundary of Sør Trondelag on a steep north-facing coast, and a further 12 km brings us to Vinjeora at the end of the fiord. Two km beyond Vinjeora there is a junction with yield sign, where a right turn should be made on E39 toward Trondheim and ascent toward a pass.

About 20 km up the valley on E39 we come to a large reservoir, Vasslivatnet, on the right, and eventually to a large turn out on the right with a former road quarry on the opposite side. Be careful when crossing to the outcrop.
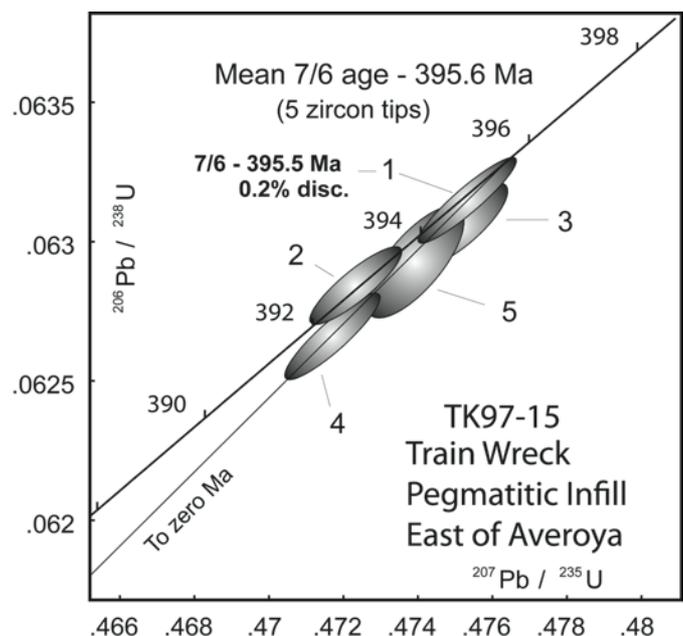

Fig. 10.6 Concordia diagram showing results of U-Pb isotopic measurements on zircons from an extensional pegmatite in eclogite at Stop 10-5

STOP 6-9: (10 minutes) Quartzite of Sætra Nappe with metamorphosed diabase dikes. (Hollocher)
    The geochemistry of four dike samples from the west end of this outcrop is presented in the introduction to Day 5. Their compositions closely overlap the reference suite collected near Oppdal (Day 6), but they are distinctly less enriched in LREE.

Beyond the reservoir, we ascend virtually to timberline with good views and a popular cross country skiing area, then descend over a winding road to a very dangerous junction with yield sign and very bad visibility beside the Orkla River, 42.6 km from Vinjeora. Just beyond this junction we cross the Orkla River to a traffic circle. Here is a very large gas station, where we may stop briefly for refreshments.
Leave the circle, and continue down the Orkla Valley on E39, soon passing Orkanger, and enter the Thamshavn/Øysand toll road and tunnel project described on the morning of Day 5 (with two cash toll payments, exact change needed). Soon beyond Øysand we again cross the Gaula River. The Klett traffic circle is 4 km east from Øysand where E39 meets E6 for the final 19 km run into Trondheim and delivery to overnight locations. END OF TRIP. WE ALL THANK YOU FOR ATTENDING AND THANK YOU FOR YOUR PATIENCE AND COOPERATION DURING THESE MANY DAYS!!!

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