

# Exhumation of high-pressure rocks beneath the Solund Basin, Western Gneiss Region of Norway

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**ABSTRACT** The Solund–Hyllestad–Lavik area affords an excellent opportunity to understand the ultrahigh-pressure Scandian orogeny because it contains a near-complete record of ophiolite emplacement, high-pressure metamorphism and large-scale extension. In this area, the Upper Allochthon was intruded by the *c.* 434 Ma Sogneskollen granodiorite and thrust eastward over the Middle/Lower Allochthon, probably in the Wenlockian. The Middle/Lower Allochthon was subducted to *c.* 50 km depth and the structurally lower Western Gneiss Complex was subducted to eclogite facies conditions at *c.* 80 km depth by *c.* 410–400 Ma. Within < 5–10 Myr, all these units were exhumed by the Nordfjord–Sogn detachment zone, producing shear strains > 100. Exhumation to upper crustal levels was complete by *c.* 403 Ma. The Solund fault produced the last few km of tectonic exhumation, bringing the near-ultrahigh-pressure rocks to within *c.* 3 km vertical distance from the low-grade Solund Conglomerate.

**Key words:** eclogite; exhumation; Norway; Solund; ultrahigh pressure.

## INTRODUCTION

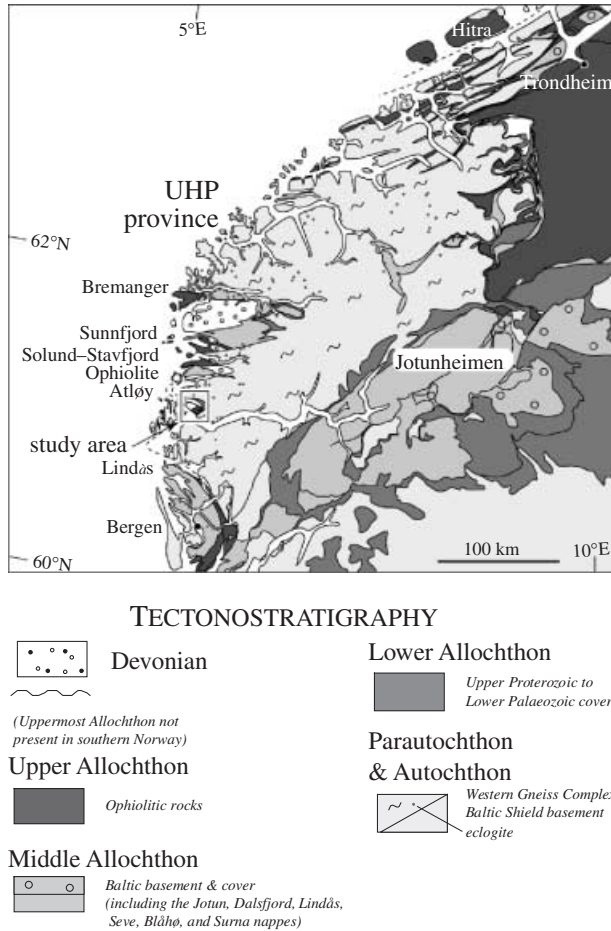
The Norwegian Caledonides are one of the best places to study the exhumation of high-pressure (HP) and ultrahigh-pressure (UHP) rocks because of excellent exposure across the vast Western Gneiss Region (WGR), which contains spectacular coesite-bearing eclogites, garnet peridotites, and coesite- and diamond-bearing gneisses (Cuthbert *et al.*, 2000; Wain *et al.*, 2000) (Fig. 1). Salient incompletely answered questions about the Norwegian Caledonian UHP rocks include the following.

(1) How and when did the HP–UHP metamorphism occur? Was there more than one HP or UHP event? Was it the result of an arc–continent or continent–continent collision? For many years, the palaeomagnetic record of collision at *c.* 425 Ma (Torsvik, 1998), the pioneering Sm/Nd geochronology that dated HP metamorphism at 425 Ma (Griffin & Brueckner, 1980), and the stratigraphic record on the island of Atløy implying ophiolite emplacement in the Wenlockian (Andersen *et al.*, 1990), were interpreted as indicating that the UHP metamorphism occurred during a continental collision. The subsequent realization that the age of the UHP event is 410–400 Ma (Tucker, in Lutro *et al.*, 1997; Mearns, 1986; Mørk & Mearns, 1986; Terry *et al.*, 2000) indicates that the orogeny spanning collision and HP–UHP metamorphism may have lasted for 25 Myr. Moreover, the material emplaced onto the Caledonian margin of Baltica includes not only

telescoped continental margin rocks, but also ophiolites, outboard Baltica crystalline and sedimentary rocks, and Laurentian continental rocks. Constraining the time that these various units were emplaced relative to the time of UHP metamorphism is still an important issue.

(2) How did the exhumation occur? Was it syn-collisional or post-collisional? Did the subducted plate simply roll back as a semirigid sheet, did it rise buoyantly through the upper plate, or was it thrust onto the Baltica margin accompanied by higher-level extension? A significant part of the exhumation of the HP–UHP rocks can be attributed to an array of regionally extensive normal faults and extensional detachments, of which the Nordfjord–Sogn detachment zone (NSDZ) is the most important (Andersen & Jamtveit, 1990). The NSDZ is a Caledonian amphibolite- to greenschist facies shear zone commonly capped by younger brittle faults that were reactivated in the Permian and Late Jurassic (Eide *et al.*, 1997; Norton, 1986). The age of the ductile shearing is not well known because, despite concerted <sup>40</sup>Ar/<sup>39</sup>Ar dating (e.g. Andersen, 1998), most of the movement on the ductile zone occurred at temperatures in excess of hornblende closure to Ar diffusion. Thus, constraining the time of motion along the ductile NSDZ is important to constraining the exhumation history.

Both of these two general problems can be addressed in the Solund–Hyllestad–Lavik area of the WGR. The Solund–Hyllestad–Lavik area is noteworthy because it

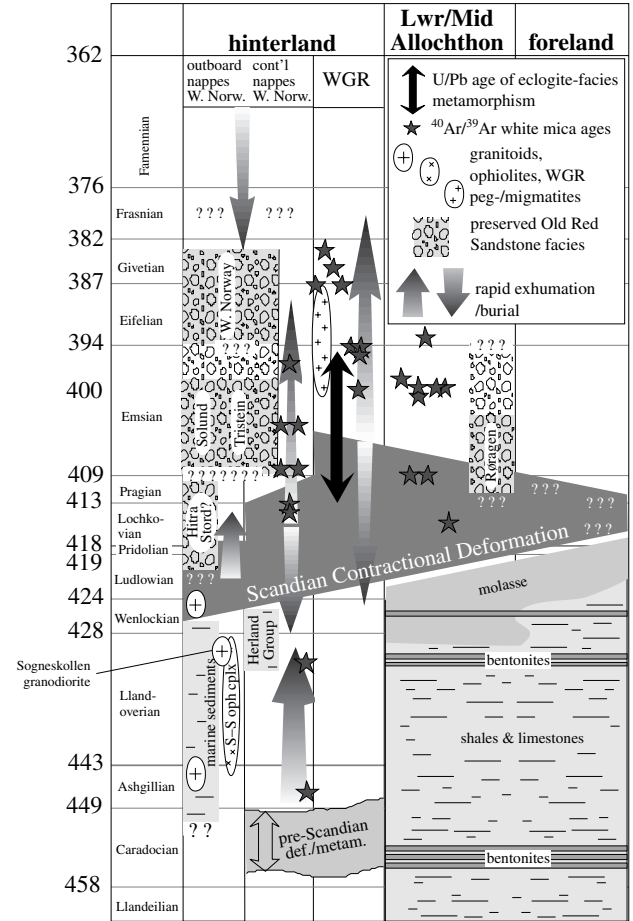


**Fig. 1.** Geology of southern Norway. Heavy dark lines show major extensional structures and UHP area is ruled.

contains a record of ophiolite emplacement, near-UHP metamorphism, and large-scale displacement on the NSDZ and the overlying Solund fault. The purpose of this paper is to document the emplacement and exhumation histories of high-pressure rocks in this key area.

**SCANDINAVIAN CALEDONIDES**

The Scandinavian Caledonides are 1700 km long, comparable to the length of the Himalaya. They are a type example of an orogen composed of thin (= 10 km), far-traveled (= 300 km), and areally extensive (= 50 000 km<sup>2</sup>) thrust sheets (Andersen, 1998). The SE-directed emplacement of these thrust sheets is conventionally explained as the result of continent–continent collision between Baltica (including Norway) and Laurentia (including Greenland). The nappes and thrust sheets are conventionally lumped into structurally distinct entities separated by regionally extensive faults: autochthon/parautochthon, Lower Allochthon, Middle Allochthon, Upper Allochthon and Uppermost Allochthon (Roberts & Gee, 1985).



**Fig. 2.** Timeline of tectonic events in southern Norway. Time-scale based on (Tucker *et al.*, 1998; Tucker & McKerrow, 1995). Beginning of Scandian deformation signaled by emplacement of Solund–Stavfjord ophiolite complex (S–S oph cplx), deposition of Herland Group, and molasse sedimentation in foreland. Switch from contraction to extension at deep levels heralded by white-mica ages in the Western Gneiss Region (WGR).

**Geological Units**

The autochthon/parautochthon is the west-facing passive margin of Baltica, which consists of Archean to Proterozoic crystalline basement overlain by Upper Proterozoic rift sediments, Cambrian to Wenlockian shelf sediments, and Ludlovian–Pridolian molasse (Fig. 2) (Bockelie & Nystuen, 1985). West of the autochthon, and separated from it horizontally by as little as 20 km of overlying allochthonous units, is the Western Gneiss Complex (WGC) (Milnes *et al.*, 1997), which contains HP and UHP eclogites; the UHP eclogites are found along the coast north of 62°N, whereas HP eclogites crop out NE, E, SE and S. The S and SE part of the WGC is correlated with the Baltic shield because both have basement and sedimentary cover of similar rock types and ages (Gee *et al.*, 1994; Milnes *et al.*, 1997; Skår, 1998), but whether the WGC is nearly autochthonous or has been displaced horizontally a significant distance is unknown.

The Lower Allochthon is composed chiefly of sedimentary rocks correlated with the sedimentary cover of the autochthon that was thrust eastward over the autochthon (Hossack *et al.*, 1985). Orthogneisses overlain by unfossiliferous feldspathic sandstones are characteristic of the Middle Allochthon (including the Jotun, Dalsfjord, Särvi and Sætra nappes) and the Seve/Blåhø/Surna

nappes of the Upper Allochthon. Strong similarities between these sialic nappes and the Lower Allochthon and autochthon imply original contiguity (Gee & Zachrisson, 1979), hence the sialic nappes are generally interpreted as Baltica crust or transitional continental/oceanic crust (Seve/Blåhø/Surna) that lay outboard of the WGC (Milnes *et al.*, 1997) or as a microcontinent (Andersen & Andresen, 1994).

The Upper Allochthon includes predominantly intraoceanic arc and marginal basin assemblages (Stephens & Gee, 1985). Its oldest plutonic and volcanic rocks have zircon ages in the range of 497–472 Ma. Intercalated and overlying sediments have late Arenigian–early Llanvirnian (*c.* 470 Ma) fossils of Baltica, Laurentian, or mixed Laurentian–Baltica affinity, implying local geographic separation from Baltica (Pedersen *et al.*, 1992; Sturt *et al.*, 1991). Sr isotopes in stitching plutons reveal that the Karmøy and Lykling ophiolites were emplaced onto continental crust by 474 Ma, and the Vågåmo Ophiolite was faulted onto the arenaceous Heidal Series before deposition of the overlapping late Arenigian–early Llanvirnian Otta Conglomerate (Sturt *et al.*, 1991). The Heidal Series rests depositionally on crystalline Baltica basement, mandating ophiolite emplacement onto the Baltica margin (Sturt *et al.*, 1991).

These Late Cambrian–Early Ordovician ophiolites are intruded/overlain by Ordovician to Silurian arc plutonic and volcanic rocks, Ashgillian (449–443 Ma) limestones, Llandoveryan (443–428 Ma) black shales and turbidites and volcanic rocks of unknown age (cf. Andersen & Andresen, 1994). The Ashgillian–Llandoveryan sedimentary rocks are similar to the Lower Allochthon, implying proximity to Baltica (Stephens & Gee, 1985). The Sulitjelma ( $437 \pm 2$  Ma) and Solund–Stavfjord ( $443 \pm 3$  Ma) ophiolites may have formed exclusively at this time. Many of these oceanic rocks, plus the Uppermost Allochthon (ascribed to Laurentia) and the westernmost edge of the WGC, are intruded by a suite of gabbroic through granitic plutons and dykes ranging in age from 443 to 432 Ma. Two features suggest that this intrusive event predated emplacement of the oceanic nappes onto the Baltica margin: (i) intrusions of this age are absent within all units structurally below the oceanic nappes and above the parautochthon; and (ii) plutons piercing the Uppermost Allochthon have Sr initial ratios = 0.706, implying crustal contamination, whereas plutons piercing the oceanic nappes have Sr initial ratios < 0.706, implying that they did not interact with the continental rocks that currently underlie the oceanic nappes (references listed below).

### Allochthon emplacement and extension history

Initial emplacement of the allochthons onto the Baltica craton can be constrained as Wenlockian (428–423 Ma) by several criteria (Fig. 2) (Greiling *et al.*, 1998). (i) On the island of Atløy, the Wenlockian Herland Group was deposited during emplacement of the Solund–Stavfjord ophiolite on Middle Allochthon basement and cover (Dalsfjord Suite and Høyvik Group) (Andersen *et al.*, 1998; Andersen & Jamtveit, 1990; Andersen *et al.*, 1990). (ii) As noted above, 443–432 Ma plutons have Sr isotopic ratios and a spatial distribution that suggest that their intrusion predates allochthon emplacement. (iii) The youngest sedimentary rocks in the Lower Allochthon are Wenlockian (Bassett, 1985), requiring that the faults bounding the Lower Allochthon be no older than Wenlockian. (iv) The youngest sedimentary rocks in the Upper Allochthon are upper Llandoveryan (*c.* 430 Ma), and these are overlain by a thick turbiditic succession that may stretch into the Wenlockian (Bassett, 1985), requiring that the faults bounding the Upper Allochthon be no older than upper Llandoveryan. (v) The absence of post-Wenlockian strata throughout the allochthons, with the exception of the extension-related Late Silurian (?) to Middle Devonian continental basins, implies ongoing tectonism. (vi) The transition from marine carbonate platform to continental fluvial molasse sedimentation in the foreland took place in the latest Wenlockian (Fig. 2). (vii) Similarities in the 420–430 Ma palaeomagnetic poles for Baltica, Scotland and North America imply that Baltica collided, probably obliquely, with Laurentia at this time (Torsvik, 1998). (viii) The southward

transgression of flysch from Laurentia (Scotland) onto Avalonia (S Ireland) also indicates a late Llandoveryan–early Wenlockian continental collision (Soper *et al.*, 1992).

Toward the end of the Scandian orogeny (Fig. 2), Devonian sedimentary rocks were deposited unconformably on the Lower, Middle and Upper Allochthons (Norton *et al.*, 1987; Séguret *et al.*, 1989; Séranne & Séguret, 1987). Fossils within these basins are Ludlovian(?)–Pridolian through Emsian–Eifelian (Siedlecka, 1975) or possibly upper Llandoveryan–lower Wenlockian on Hitra, Lower Devonian in the Røragen basin, and Ludlovian through Pridolian(?) in the Oslo graben (see summaries in Bockelie & Nystuen, 1985; Steel *et al.*, 1985). The sedimentation in and structure of the western Norway basins were controlled by an early system of normal and strike-slip faults (Andersen, 1998; Cuthbert, 1991; Osmundsen & Andersen, 2001; Osmundsen *et al.*, 1998).

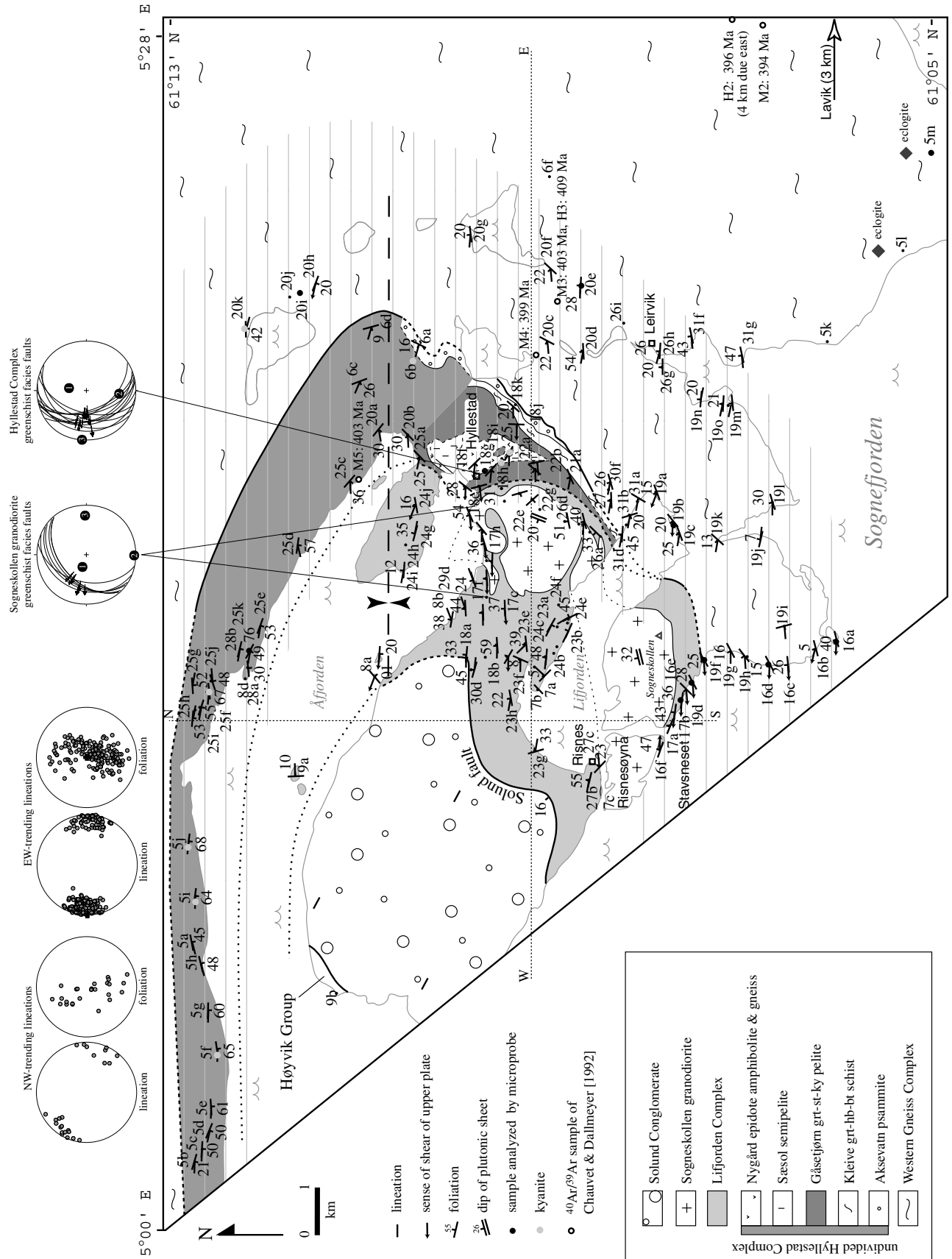
Following their emplacement onto the Baltica continental margin, the Caledonide thrust sheets were reworked by a major extensional event that resulted most significantly in collapse of the thrust sheets and exhumation of eclogites in the WGC. The best-known extensional structure is the Nordfjord–Sogn detachment zone (NSDZ). The NSDZ carried the Middle and Upper Allochthons and overlying Devonian sedimentary rocks in the hanging wall down to the west over eclogite-bearing gneisses of the WGC in the footwall (Norton, 1987). The horizontal displacement on the NSDZ is at least 70 km, but even allowing for a considerable vertical shortening component in the footwall (Andersen *et al.*, 1994; Dewey *et al.*, 1993), this is probably insufficient to account for the juxtaposition of *c.* 3 GPa eclogites with Devonian sedimentary rocks in the hanging wall, indicating earlier or coeval contributions from another exhumation process.

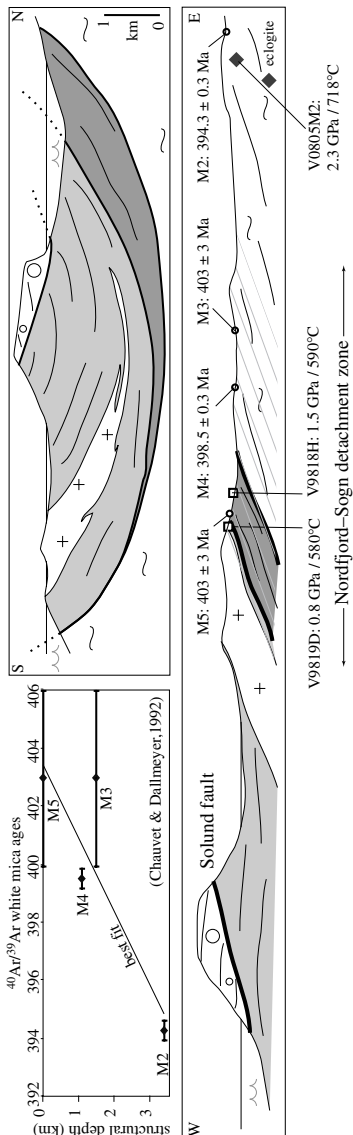
### SOLUND–HYLLESTAD–LAVIK AREA GEOLOGY

The significance of the Solund–Hyllestad–Lavik area (Fig. 3) stems from its exposure of (i) one of the rare syn-kinematic plutons in the Norwegian Caledonides, the Sogneskollen granodiorite (Skjerlie *et al.*, 2000); (ii) a nearly complete stack of the Caledonian nappes and overlying sediments; and (iii) an excellent structural section from the WGC up through the Upper Allochthon, the Solund fault, and into the Solund Devonian Basin.

The Western Gneiss Complex (WGC) is the structurally lowest unit in the area. It is chiefly a granitic to gabbroic orthogneiss with sparse lenses of eclogite. Except for the eclogite blocks, the WGC in the Solund–Hyllestad–Lavik area is dominantly amphibolite facies and has a weak retrograde greenschist facies overprint. Structurally above the WGC is the Hyllestad Complex, named and subdivided by Tillung (1999) into the Aksevatn psammite, Kleive mafic schist, Gåsetjørn pelite, Sæsol semipelite and Nygård amphibolite. The lowest of these units, the Aksevatn psammite, is a quartzose meta-sandstone. The WGC contains structurally interleaved layers of similar rocks with local pebbly layers, suggesting that this unit *may* have been deposited on the WGC, analogous to basal conglomerates elsewhere in the WGR (Tillung, 1999). The Kleive mafic schist is characterized by hornblende + biotite + garnet + plagioclase assemblages, the Gåsetjørn pelite contains local highly aluminous horizons with kyanite + staurolite + garnet + chloritoid, and the Sæsol semipelite includes marble to calc-silicate layers. The Nygård unit comprises epidote + biotite amphibolites and minor felsic gneisses and may represent a highly attenuated sheet of allochthonous basement (Tillung, 1999).

The overlying Hyllestad Complex is *c.* 1 km thick and has been correlated with the Lower Allochthon (Chauvet & Dallmeyer, 1992; Swenson & Anderson, 1991) or the Høyvik Group of the Middle Allochthon on Atløy (Tillung, 1999); all three units contain similar highly aluminous pelites. If correlative with the Lower Allochthon, which elsewhere contains sedimentary rocks as young as Wenlockian, this places an older limit on the metamorphic age of the Hyllestad Complex. If correlative with the Høyvik Group, which (i) lies depositionally on top of the Jotun-like Dalsfjord Suite (Brekke & Solberg, 1987) (ii) is depositionally overlain by the Wenlockian Herland Group (Andersen *et al.*, 1990; Brekke &





**Fig. 3.** Geology of the Solund–Hyllestad–Lavik area, from our own mapping and Tillung (1999). Low-grade Solund Conglomerate and high-grade WGR eclogite are separated by only a few km of the medium-pressure ophiolitic Lijfjorden Complex and high-pressure passive margin(?) Hyllestad Complex. ‘M’, ‘H’ refer to muscovite and hornblende  $^{40}\text{Ar}/^{39}\text{Ar}$  ages of Chauvet & Dallmeyer (1992); some ages have been interpreted slightly differently than in the original paper. The three lineations from the Solund Conglomerate are from Chauvet & Séranne (1989). Horizontal ruling shows extent of NSDZ.

Solberg, 1987), and (iii) contains muscovite with *c.* 447 Ma  $^{40}\text{Ar}/^{39}\text{Ar}$  ages (Andersen *et al.*, 1998), the Hyllestad Complex is older than 447 Ma.

Structurally overlying the Hyllestad Complex is the Lijfjorden Complex (Tillung, 1999), which consists of *c.* 3 km structural thickness (Fig. 3) of mostly metagraywacke, greenschist and greenstone, with minor serpentinite, metagabbro, chert, quartzose sandstones, marble and volcanogenic conglomerates. The greenstones, serpentinites and metagabbros occur chiefly toward the bottom of the section, whereas the top of the section is dominantly clastic (Tillung, 1999). The volcanogenic conglomerates, which appear in the middle of the section, include pebbles to cobbles of chiefly greenstone, plus sandstone, metagabbro, granodiorite and quartzose rock in a volcanogenic matrix (Tillung, 1999). The rock types and deformation of the Lijfjorden Complex are similar to, and probably correlative with, the Staveneset Group (Furnes *et al.*, 1990) metagraywacke and metavolcanic rocks that overlie the 443-Ma Solund–Stavfjord Ophiolite just west of the study area (Fig. 1, Skjerlie *et al.*, 2000).

The youngest unit in the area is the Devonian conglomerates and minor sandstones of the Solund Basin (Nilsen, 1968), which crops out in the Solund, Bulandet and Værlandet areas (Steel *et al.*, 1985). The unit is > 6 km thick and contains probable Early Devonian fossils (Steel *et al.*, 1985), a broad spectrum of proximally derived graywacke, quartzite, greenstone and gabbro cobbles, and giant landslide deposits. The landslides contain an Early Silurian rhyolite with a U/Pb zircon age of  $439 \pm 1$  Ma (Hartz *et al.*, in press). Sedimentary structures indicate north-westward flow of braided streams on alluvial fans built against a NW-dipping fault scarp (Nilsen, 1968). While most of the Devonian basin in the study area are underlain by the Solund fault, it rests positionally on the Solund–Stavfjord Ophiolite, and on the Høyvik Group at the west end of the study area (Fig. 3).

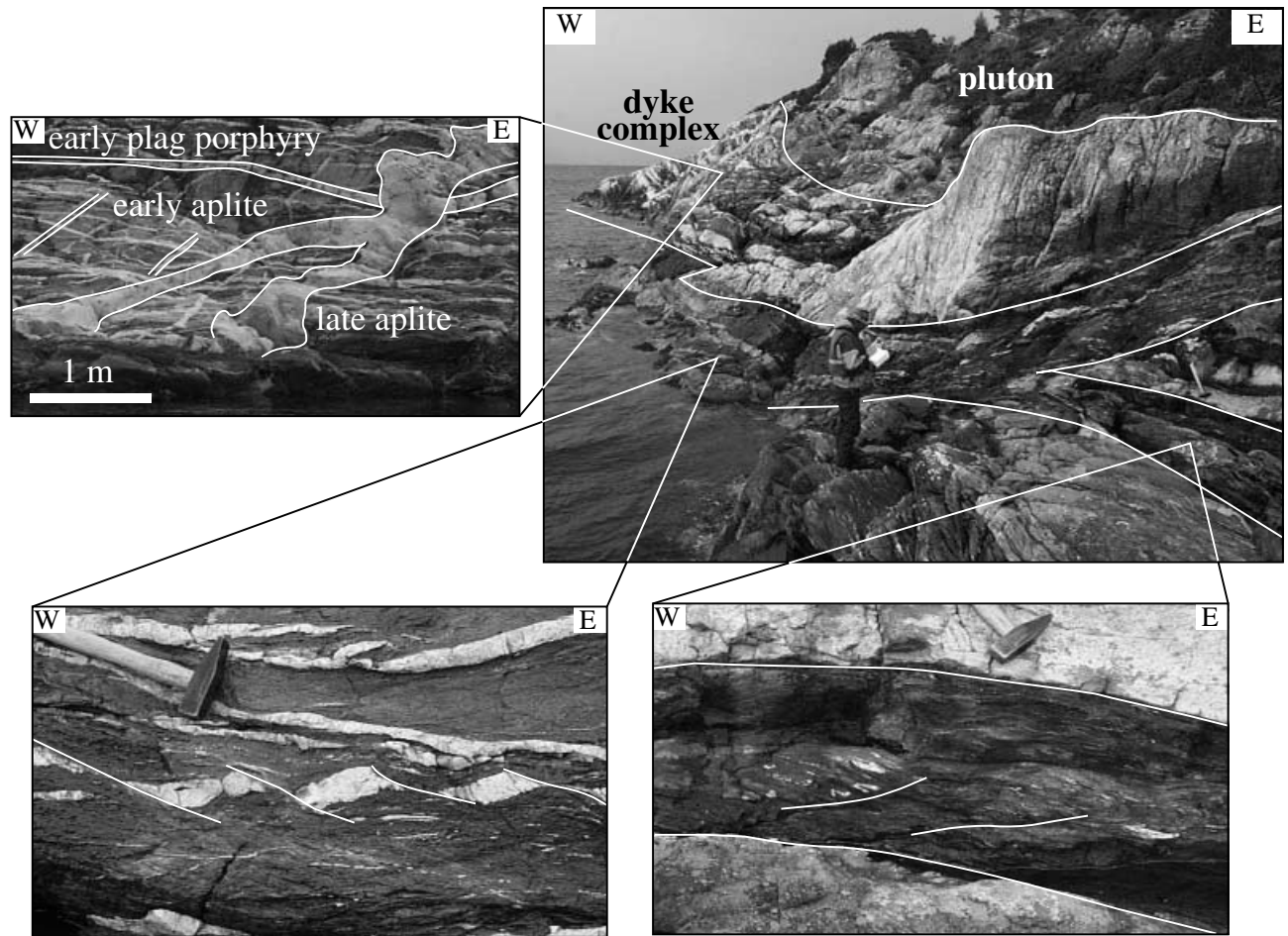
### SOGNESKOLLEN GRANODIORITE

The Sogneskollen granodiorite is a medium- to fine-grained leucocratic granodiorite to quartz monzonite with < 5 vol% biotite and epidote (Skjerlie *et al.*, 2000; Tillung, 1999). The epidote occurs in clusters of crystals commonly associated with biotite and is interpreted as magmatic (Tillung, 1999).

The granodiorite forms a sheetlike body intruding the Lijfjorden Complex that is > 300 m thick and dips *c.* 20° west. Skjerlie *et al.* (2000) interpreted the high Ba, Sr, and Na/K, and low Y,  $\epsilon_{\text{Nd}}$ ,  $^{87}\text{Sr}/^{86}\text{Sr}$  and HREE concentrations to indicate formation of the Sogneskollen granodiorite by melting of hydrous sediments outside the stability field of calcic plagioclase and within the stability field of garnet. They specifically proposed that the Sogneskollen granodiorite formed from melting of graywackes like the Lijfjorden Complex during thrusting beneath the Solund–Stavfjord Ophiolite. Granite veins and dykes with similar compositions are common in the upper parts of the obduction melange thrust on top of the Herland Group in the Sunnfjord area (Osmundsen & Andersen, 1994; Skjerlie *et al.*, 2000). Similar high-Sr and Ba, biotite–epidote granites with Rb/Sr whole-rock ages of  $430 \pm 10$  and  $430 \pm 6$  Ma and Sr initial ratios of 0.7056–0.7066 intrude ophiolites and metagreywackes of the Upper Allochthon south of Bergen (Andersen & Jansen, 1987; Fossen & Austrheim, 1988).

In general the pluton is strongly affected by the extensional fabrics described below, but locally, the base of the pluton has been sheltered from the deformation and details of its original intrusive relationships may be discerned. The basal contact is a *c.* 10 m thick swarm of aplite and plagioclase–porphyry dykes (‘grey dykes’ of Skjerlie *et al.*, 2000) (Fig. 4). Both types of dyke are mutually intrusive, although the aplites appear to record a slightly more complex deformation history, implying that they began intruding first. The aplite dykes are < 1 m in width and have a mutually intrusive relationship with the pluton. The plagioclase–porphyry dykes cut the pluton; they consist of plagioclase in a fine-grained matrix of quartz, epidote, feldspar and biotite, and are slightly elevated in Fe + Ti relative to the pluton (Tillung, 1999).





**Fig. 4.** Ductile shear zone at the base of the Sogneskollen granodiorite shows coaxial to weak top-E fabrics in wallrock graywacke and early aplite dykes (lower left), and top-W simple shear in younger plagioclase porphyries (lower right).

## METAMORPHIC PETROLOGY

Metamorphic grade increases downsection in the Solund–Hyllestad–Lavik area. The Solund Conglomerate reached prehnite–pumpellyite facies (Norton, 1987) during a metamorphic episode that must post-date the Early Devonian depositional age of the sediments. The Liffjorden Complex above the Sogneskollen granodiorite is generally of greenschist facies, with quartz + plagioclase + muscovite + biotite stable in metasedimentary rocks and the same minerals plus epidote and chlorite in volcanogenic rocks. Garnet was found at two localities above the Sogneskollen granodiorite, in metasedimentary rock at V0807a and in metamafic rock at V0807c.

The portion of the Liffjorden Complex beneath the Sogneskollen granodiorite, the bulk of the Hyllestad Complex and the non-eclogitic rocks of the WGC, contain amphibolite facies assemblages of biotite + muscovite + plagioclase + quartz ± garnet in felsic rocks and biotite + hornblende + plagioclase + quartz ± garnet in mafic varieties (mineral

compositions in Table S1). Most of the garnet-bearing rocks are in the Hyllestad Complex. All the garnet show bell-shaped Mn profiles, implying that temperatures remained too low for volume diffusion (< 600 °C, Florence & Spear, 1991). All the aluminous samples contain garnet with core-to-rim increases in Mg# and decreases in Ca. The increase in Mg# implies growth during increasing temperature for these mineral assemblages, but the decrease in Ca is more difficult to assess because of the consumption of plagioclase and the growth of paragonite. Using THERMOCALC, we assessed pressures and temperatures for garnet-bearing samples, and recalculated pressures and temperatures for the ‘micaschist’ and ‘gneiss’ samples of Chauvet *et al.* (1992) (Table 1). Metamorphic conditions for the felsic rocks, assessed chiefly with garnet–biotite Fe–Mg thermometry and garnet–biotite–muscovite–plagioclase barometry, fall mostly in the range 500–600 °C, 0.7–0.9 GPa, implying metamorphism at depths of *c.* 30 km. The two mafic rocks, evaluated with garnet–hornblende Fe–Mg thermometry and garnet–hornblende–plagioclase–quartz barometry, fall in the same range.

Higher peak pressures, indicating metamorphism at depths of *c.* 50 km, are recorded by highly aluminous and ferric rocks in the Gåsetjørn unit that contain subassemblages of garnet + staurolite + kyanite + chloritoid + paragonite + muscovite + quartz without biotite or talc (Chauvet *et al.*, 1992). Textural relationships indicate that garnet, staurolite, chloritoid and kyanite constitute the highest-pressure equilibrium assemblage. This aluminous paragenesis defines a reaction (575–600 °C and 1.4–1.6 GPa) in the KFMASH system where staurolite + chloritoid change to kyanite + garnet with increasing pressure or temperature (program GIBBS v. 2/01, Spear & Menard, 1989) (Fig. 4; Table 1). Two Hyllestad Complex samples (V9818I & V9825F) contain kyanite + staurolite + garnet, which defines a narrow *P–T* region from 575 °C, 1.6 GPa to 670 °C, 0.7 GPa; with THERMOCALC, *P–T* conditions were calculated near the high-*P* end of this divariant field for sample V9818I.

Textures and mineral zoning reveal that the *P–T* path leading to the peak recorded conditions involved

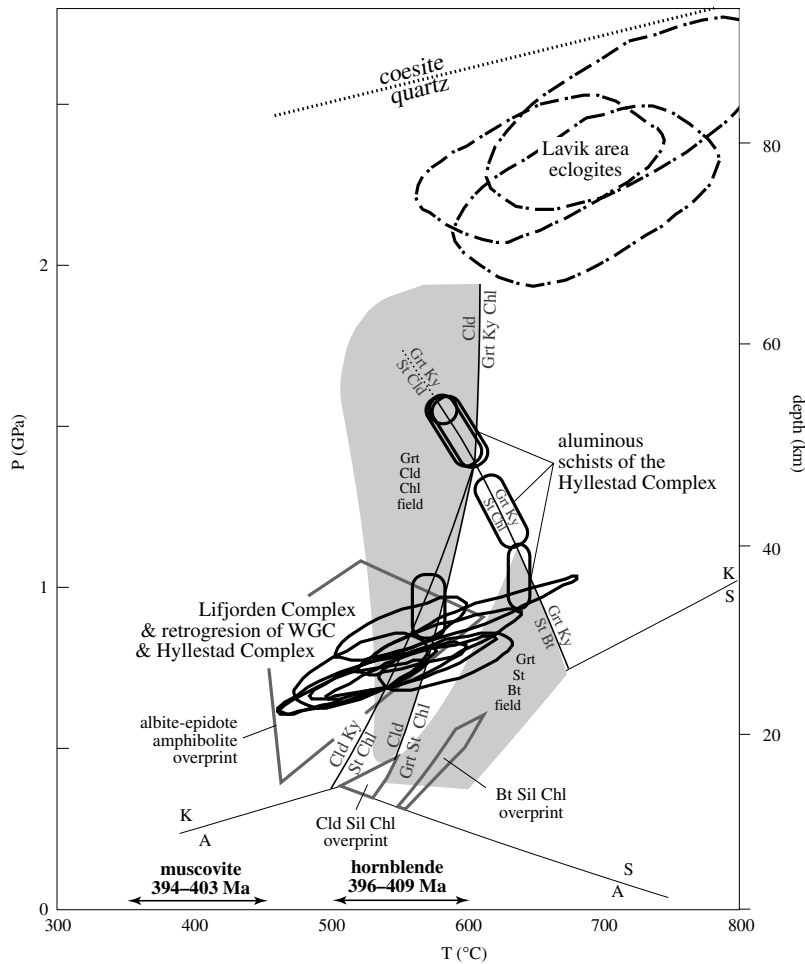
heating. (i) Aluminous sample V9819H2 contains chloritoid inclusions in garnet, indicating that chloritoid-consuming reactions were active during the prograde path. (ii) Sample V9820I has the Fe-rich assemblage garnet + staurolite + biotite; the staurolite is small and rare, and garnet contains chloritoid and chlorite inclusions, suggesting progression from the garnet–chloritoid–chlorite field to higher temperatures, across the chloritoid = garnet + biotite + chlorite and garnet + chlorite = staurolite + biotite reactions. Using the program GIBBS (Spear & Menard, 1989), inverse modelling of zoning in Fe–Mg phases, and forward *P–T* path modelling to replicate the zoning observed in garnet, we calculate that the prograde path for sample V9820I involved a pressure increase of 450 MPa and a temperature increase of 80 °C. (iii) All nonclogite garnet exhibit core-to-rim increases in Mg#, and plagioclase in garnet–epidote bearing rocks shows core-to-rim increases in Ca.

Retrograde metamorphism of the aluminous rocks involved nearly isothermal decompression. Two distinctive mineral assemblages developed in the

**Table 1.** Thermobarometry results.

Sample	Minerals	Thermometer	Barometer	T (°C)	P (kbar)	cor
V9816A1	Grt Ms Bt Pl Qtz Chl Rt	KFMASH grid	Grt Bt Ms Pl	540–610	9.4 ± 1.0	n/a
V9816D2	Grt Hbl Bt Ms Pl Qtz Rt Ep	Grt–Bt	Grt Bt Ms Pl	560 ± 50	7.5 ± 0.8	0.917
V9818G	Grt Hbl Ms Pl Qtz Chl Rt Ep	Grt–Hbl	Grt Hbl Pl Qtz	523 ± 39	7.4 ± 0.9	0.651
V9818H	Ky St Cld Grt Ms Qtz Rt Sil	KFMASH grid		575–600	14–16	n/a
V9818I1	Ky St Grt Ms Pg Qtz Chl Rt Sil	Ky Grt St Ms Qtz intersection KFMASH grid KFMASH grid	Ky Grt St Ms Qtz reaction	616 ± 201 575–670 570–585	18.4 ± 6.6 7.4–16 15–16	0.959 n/a n/a
V9818I9B	Ky St Cld Grt Ms Pg Rt Sil	Ky St Grt Ms Qtz intersection KFMASH grid		595 ± 175 575–600	14.5 ± 5.0 14–16	0.930
V9819D	Grt Hbl Ms Bt Pl Qtz Chl Rt Ep Ilm	Grt–Bt Grt–Hbl	Grt Bt Ms Pl Grt Hbl Pl Qtz	580 ± 48 577 ± 38	8.3 ± 0.8 5.0 ± 1.2	0.845 0.405
V9819E	Grt Hbl Ms Grt Pl Qtz Chl Rt Ep	Grt–Bt Grt–Hbl	Grt Bt Ms Pl Grt Hbl Pl Qtz	520 ± 72 521 ± 36	7.2 ± 0.8 5.4 ± 0.8	0.875 0.583
V9819F	Grt Hbl Ms Bt Pl Qtz Chl Rt Ep Ilm	Grt–Bt Grt–Hbl	Grt Bt Ms Pl Grt Hbl Pl Qtz	543 ± 46 513 ± 34	7.4 ± 0.7 4.6 ± 0.9	0.850 0.430
V9820E	Grt Hbl Bt Pl Qtz Rt Ep	Grt–Hbl	Grt Hbl Pl Qtz	552 ± 38	8.7 ± 0.8	0.755
V9820I	Grt St Ms Bt Pl Qtz Ep	KFMASH grid	Grt Bt Ms Pl	600–630	10.3 ± 1.1	n/a
V9820K	Grt Hbl Bt Pl Qtz Chl Rt Ilm	Grt–Hbl	Grt Hbl Pl Qtz	585 ± 39	7.7 ± 0.7	0.651
V9825F	Ky St Grt Ms Bt Pl Qtz Chl Rt	Ky St Grt Bt Ms KFMASH grid	Qtz intersection Grt ASt Pl	614 ± 56 620 ± 10	11.4 ± 2.2 11.8 ± 1.1	–0.41 n/a
V0805M 1	Grt Om Pg Ms	Grt Om Pg	Grt Om Pg Ms	687 ± 77	22.1 ± 2.2	0.515
V0805M 2	Grt Om Pg Ms	Grt Om Pg	Grt Om Pg Ms	718 ± 83	24.7 ± 2.4	0.670
EC§	Grt Om Pg	Grt Om Pg Ms		654 ± 72	23.0 ± 1.8	0.475
MS§		Grt–Bt	Grt Bt Ms Pl	516 ± 44	6.9 ± 0.7	0.881
GN§		Grt–Bt	Grt Bt Ms Pl	607 ± 59	9.1 ± 1.0	0.953

Note: 'KFMASH grid' refers to pelite phase diagram produced with Gibbs (Spear & Menard, 1989) from Powell & Holland 1998 database; 'reaction' and 'intersection' refer to THERMOCALC v3.1 with May, 2001 database (Powell & Holland, 1988). Mineral formulae and activities were calculated with the program 'A-X', by T.J.B. Holland and R. Powell; A-X calculates Fe<sup>3+</sup> in clinopyroxene using charge balance considerations, which Carswell *et al.* (2000) demonstrated is a good approximation to Fe<sup>3+</sup> measured by Mössbauer spectrometry. Uncertainties are ± 1  $\sigma$ ; 'cor' is correlation coefficient from THERMOCALC. §: calculations based on mineral compositions reported by Chauvet *et al.* (1992).



**Fig. 5.** Calculated pressures and temperatures for the Solund-Hyllestad-Lavik area. Eclogites in the Lavik area record peak near-UHP conditions and albite-epidote amphibolite facies overprint. Aluminous schists of the Hyllestad Complex decompressed nearly isothermally from *c.* 1.5 GPa to 0.4 GPa at *c.* 600 °C. Lifjorden complex was buried to depths of *c.* 30 km.

aluminous rocks: (i) Fe-rich chloritoid between boudinaged staurolite grains and on kyanite rims, and late-stage top-W shear bands with sillimanite and chlorite (*c.* 500 °C and 400 MPa); and (ii) biotite + sillimanite + chlorite (550–600 °C and 300–600 MPa).

The Sogneskollen granodiorite predates the amphibolite facies metamorphism and associated top-W extension. Garnet atolls developed around plagioclase grains in the aplite dykes (Tillung, 1999) might have grown during at this time.

Eclogites crop out at a few localities in the Lavik area (Fig. 3). Chauvet *et al.* (1992) described phengite + epidote + rutile bearing eclogite variably retrogressed to amphibolite and then albite-epidote-amphibolite facies. The garnet in the examined samples are homogeneous  $\text{alm}_{58}\text{grs}_{21}\text{prp}_{19}\text{sps}_{01}$  and contain inclusions of  $(\text{K}_{0.1}\text{Na}_{0.5})(\text{Na}_{1.1}\text{Ca}_{0.9})(\text{Mg}_{2.1}\text{Fe}_{1.3})(\text{Fe}_{0.4}\text{Al}_{1.2})(\text{Al}_{1.5}\text{Si}_{6.5})\text{O}_{22}(\text{OH})_2$  hornblende, epidote and plagioclase, indicating prograde amphibolite facies metamorphism. Clinopyroxene is also nearly homogeneous, lacks the exsolved  $\text{SiO}_2$  rods common in eclogites farther north in the WGC and has core compositions of  $\text{jd}_{50-54}\text{di}_{31-33}\text{hed}_{11-12}\text{acm}_{07-11}$ ;

Chauvet *et al.* (1992) reported much lower jadeite contents of 25–30 mol% from their sample. The highest phengite content found in K-white mica was 3.32 Si atoms pfu, slightly less than the 3.36 reported by Chauvet *et al.* (1992). Following the logic outlined by Carswell *et al.* (2000), we used THERMOCALC to calculate  $P$ ,  $T_{\text{max}}$  using the lowest Fe/Mg garnet, highest Fe/Mg omphacite and highest Fe/Mg phengite compositions, and  $P_{\text{max}}$ ,  $T$  using the most jadeite-rich omphacite, most siliceous phengite, and the garnet with the highest  $a_{\text{prp}}a_{\text{grs}}^2$ .  $P$ – $T$  were calculated based on Fe–Mg exchange between garnet and clinopyroxene and the inverse Tschermak substitution in coexisting phengite, finding conditions of *c.* 700 °C, 2.3 GPa for our two eclogite samples and Chauvet *et al.*'s sample, regardless of whether the  $T_{\text{max}}$  or  $P_{\text{max}}$  criteria of Carswell *et al.* (2000) were used.

In summary, the metamorphic petrology places the following significant constraints on the evolution of the area. (1) The WGC experienced definitively higher pressures and temperatures than all overlying units, reaching *c.* 700 °C at *c.* 80 km depth. (2) The Hyllestad Complex experienced definitively higher pressures



and temperatures (*c.* 600 °C at *c.* 50 km depth) than the Lifjorden Complex (*c.* 550 °C at *c.* 30 km depth). (3) The similarity of the prograde amphibolite facies pressures and temperatures of the Lifjorden Complex with the retrograde amphibolite facies pressures and temperatures of the Hyllestad Complex and the WGC imply that all three units shared a common late-stage history. (4) All three units reached peak conditions via a heating path, which was then followed by decompression to *c.* 15 km depth and only minor cooling of *c.* 50 °C. (5) This metamorphism post-dates intrusion of the Sogneskollen granodiorite. (6) The pressure differences between the three major units implies considerable excision or thinning of the section. The 2.3 GPa eclogites and the 1.5 GPa rocks in the Hyllestad Complex are separated by only *c.* 2.7 km of structural section, implying thinning by a factor of *c.* 11. If this exhumation occurred along a normal-sense shear zone with a dip of 40–25°, this corresponds to a shear strain of 17–27. The 1.5 GPa rocks in the Hyllestad Complex and the 0.8 GPa rocks in the Lifjorden Complex are separated by only *c.* 0.3 km of structural section, implying thinning by a factor of *c.* 66. If this exhumation occurred along a normal-sense shear zone with a dip of 40–25°, this corresponds to a shear strain of 100–150.

## STRUCTURE

The large-scale structure of the study area is a > 20-km wavelength, W-plunging synform whose axis is parallel to the regional stretching lineation. The WGC, Hyllestad Complex and Lifjorden Complex contain a gently plunging E–W stretching lineation in most localities (Fig. 3). In the WGC and Hyllestad Complex, this lineation formed at amphibolite facies conditions, whereas the lineation in the part of the Lifjorden Complex above the Sogneskollen granodiorite formed at greenschist facies conditions. The E–W lineation is contemporaneous with the noncoaxial top-W extension (Chauvet & Séranne, 1989). The extensional fabric is fairly homogeneous throughout the WGC and Hyllestad Complex between Lavik and Hyllestad, but higher in the section, it is less penetrative and mafic orthogneisses along the shore south of Sogneskollen are highly phyllonitic. The portion of the Lifjorden Complex structurally beneath the Sogneskollen pluton is also strongly deformed and partly phyllonitic, except in a local strain shadow adjacent to the pluton at Risnesøyna (see above and Fig. 4). Structurally above the pluton, however, the extensional deformation is less penetrative. We interpret domains in the upper part of the Lifjorden Complex where the lineation plunges NW (Fig. 3), rather than E–W, as a younger fabric.

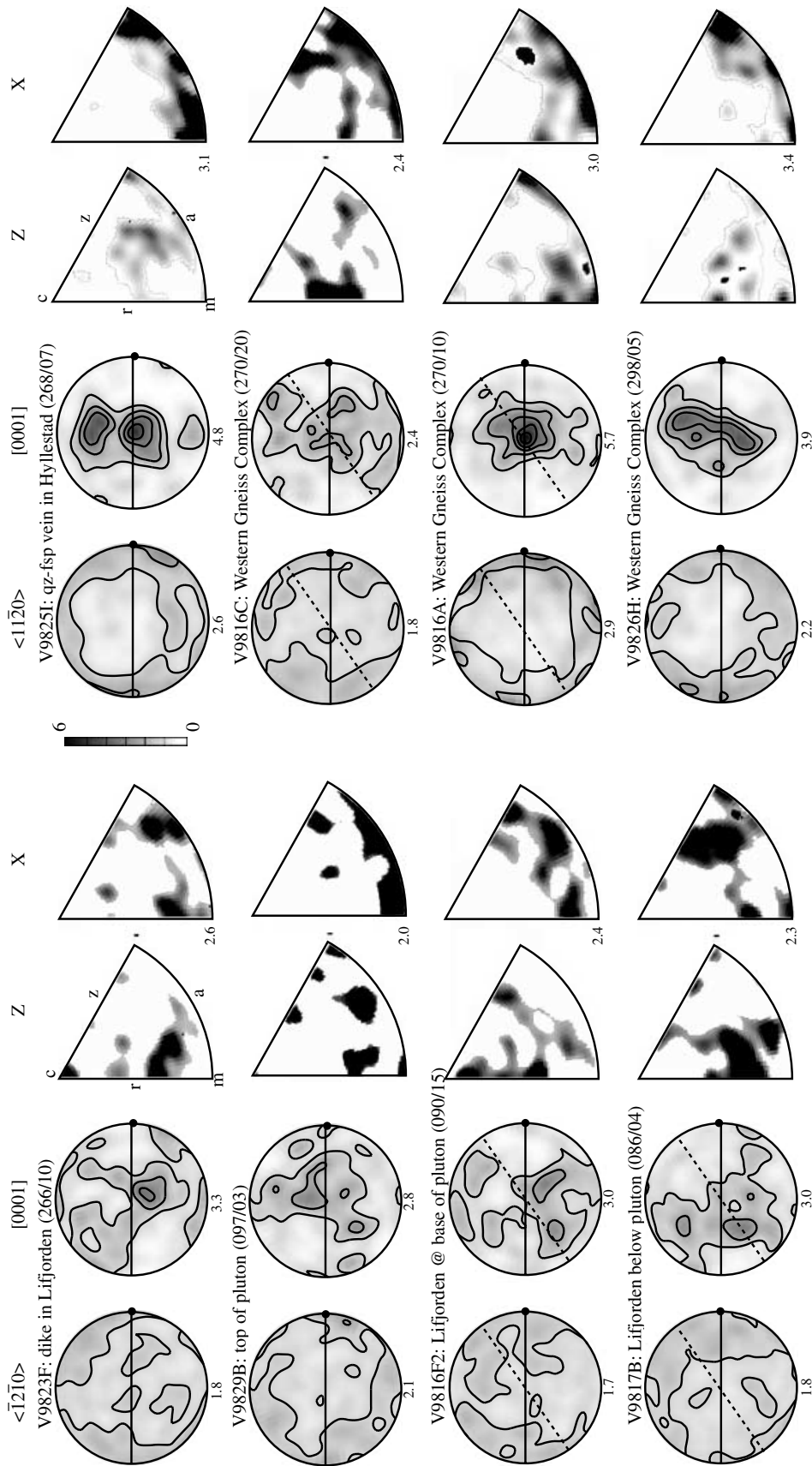
The most significant structural feature in the area is the ductile shear zone at the base of the Lifjorden Complex (Fig. 4). The first 10 m of the dyke swarm beneath the Sogneskollen granodiorite at Risnesøyna are weakly deformed, and contain important struc-

tural and intrusive relationships obliterated elsewhere. Here, the intrusive relationships between aplite dykes and their envelope show that the aplites post-date early contractional folds, cleavage formation and thrust-related top-east fabrics. The country-rock greywackes between the dykes contain an S1 cleavage parallel to compositional layering, S0, that is folded into isoclinal m-scale folds. An axial-planar cleavage, S2, is developed in the hinges of these folds. These folds and S1 and S2 cleavages are truncated by both the aplite and plagioclase-phyric dykes. The greywacke and aplite dykes in the low-strain zone have dominantly symmetrical structures implying dominantly coaxial deformation at the scale of the preserved low-strain zone, but locally the dykes have asymmetric boudins indicating thrust-related top-E shear. The pre to syn-granite emplacement contractional fabrics yield no evidence for HP metamorphism, supporting our interpretation of a major metamorphic pressure break below the Lifjorden complex. The plagioclase-phyric dykes truncate these fabrics and themselves locally exhibit top-W shear. At Stavsneset (Fig. 3, headland east of Risnesøyna) deformation intensifies about 50 m below the base of the pluton: the host greywacke develops strong top-W asymmetric fabrics, and the dykes become disrupted, asymmetrically boudinaged and transposed parallel to an intensified foliation. This zone is succeeded downward by a few metres of strongly deformed greywacke with rare boudins of the aplite and plagioclase-phyric dykes. Below this, the asymmetric top-W fabric becomes phyllonitic, strengthens abruptly and no further dykes are seen, suggesting very large shear strains and correspondingly large-scale displacement.

In aggregate, the structural and intrusive relationships reveal that the aplite dykes intruded at a late stage of the top-E thrusting, and that the top-W shear began after dyke intrusion. The plagioclase-phyric dykes exhibit variable degrees of deformation in the extensional shear bands; but we ascribe this variation to the inhomogeneous bulk strain preserved in the local strain shadow at Risnesøyna. Textbook-quality shear bands, asymmetric boudinage,  $\sigma$  and  $\delta$  clasts indicate pervasive top-W extension in all rocks exposed structurally beneath the Sogneskollen granodiorite (Fig. 4).

We measured quartz lattice preferred orientations in eight samples (Fig. 6). V9816A shows the strongest fabric, with a *c*-axis maximum parallel to Y and the prism planes aligned parallel to the foliation, implying prism-*a* slip, characteristic of amphibolite facies deformation, during coaxial deformation (Schmid & Casey, 1986). Two other samples in the deeper part of the section are similar, whereas the remaining samples have rather weak fabrics. Only the uppermost sample, V9823F exhibits any evidence of low-temperature *a* slip in the basal plane.

Deformation in the Solund Basin sedimentary rocks is different than that in lower units. Pebbles



**Fig. 6.** Quartz lattice preferred orientations (= 200 grains per sample), arranged from structurally highest (V9823F) to structurally lowest (V9826H) sample. Structurally lower samples show evidence for prism- $\langle a \rangle$  slip during coaxial deformation, whereas shallower samples have weak fabrics. Circles are equal-area, lower hemisphere, stereographic projections of  $\langle a \rangle$  and  $|c|$  axes (foliation shown by horizontal line, down-plunge direction of lineation shown by filled dot, and shear bands shown with dashed lines). Wedges are inverse pole Figs for foliation (Z) and lineation (X) (c, m, r and z refer to basal, prism, positive rhomb and negative rhomb planes). Contour intervals are multiples of mean uniform distribution (Kamb, 1959); number beneath each Figure is maximum density. Trend and plunge of lineation given in parentheses.

throughout the basin are oriented NW–SE (Séranne & Séguret, 1987), and have NW–SE trending strain shadows and NE-striking tension gashes developed during metamorphism (Chauvet & Séranne, 1989). The Solund fault, where it bounds the Solund Conglomerate, is a few-metre thick zone with pseudotachylite, ultracataclasite and ultramylonite derived mainly from the Solund Conglomerate. Outcrop/topography relations indicate a dip of *c.* 16° toward 320, implying that the Solund fault truncates structures within the underlying Liffjorden Complex; this is also evident in the discordant relationship between the NE-trending folds in the Solund–Bulandet–Værlandet Basin (Norton, 1987; Osmundsen & Andersen, 2001) and the E–W folds developed in all deeper units. The upper 7 m of the underlying Liffjorden Complex are mylonitized with a 302/18 stretching lineation and shear bands indicating extensional movement along the Solund fault (Norton, 1987). Less than 100 m beneath the Solund fault, the greenschist facies Liffjorden Complex is less deformed. It bears the folds and *S*<sub>1</sub> and *S*<sub>2</sub> foliations similar to those beneath the Sogneskollen granodiorite.

Chauvet & Séranne's (1989) reconnaissance investigation of brittle faults within the Solund Basin also showed NW–SE extension. We measured brittle-ductile faults with quartz + chlorite + epidote mineralization at two well-exposed localities within the Sogneskollen granodiorite and Hyllestad Complex (Fig. 3). At both localities, fault-striae analysis (Ratschbacher *et al.*, 1994), indicates roughly E–W extension after the local foliation is restored to pre-folding horizontal.

Thus, the section in the Hyllestad area is marked from top to bottom by (i) NW-directed extension within the weak prehnite-pumpellyite facies Solund Conglomerate; (ii) NW-directed normal slip along the discrete Solund fault; (iii) NW-directed extension within the weakly deformed greenschist facies section of the Liffjorden Complex; (iv) W-directed extension within the Sogneskollen granodiorite; (v) the profound top-W extensional shear zone, which began at amphibolite facies conditions (sillimanite stable) and which contains the HP rocks in its footwall; and (vi) penetrative top-W extensional fabric throughout the Hyllestad Complex and extending down into the WGC at least as far as Leirvik. Features iv, v and vi comprise the NSDZ, quite different from suggestions that the NSDZ is a homogeneous 2-km thick shear zone restricted to the Liffjorden Complex (e.g. Chauvet & Brunel, 1988).

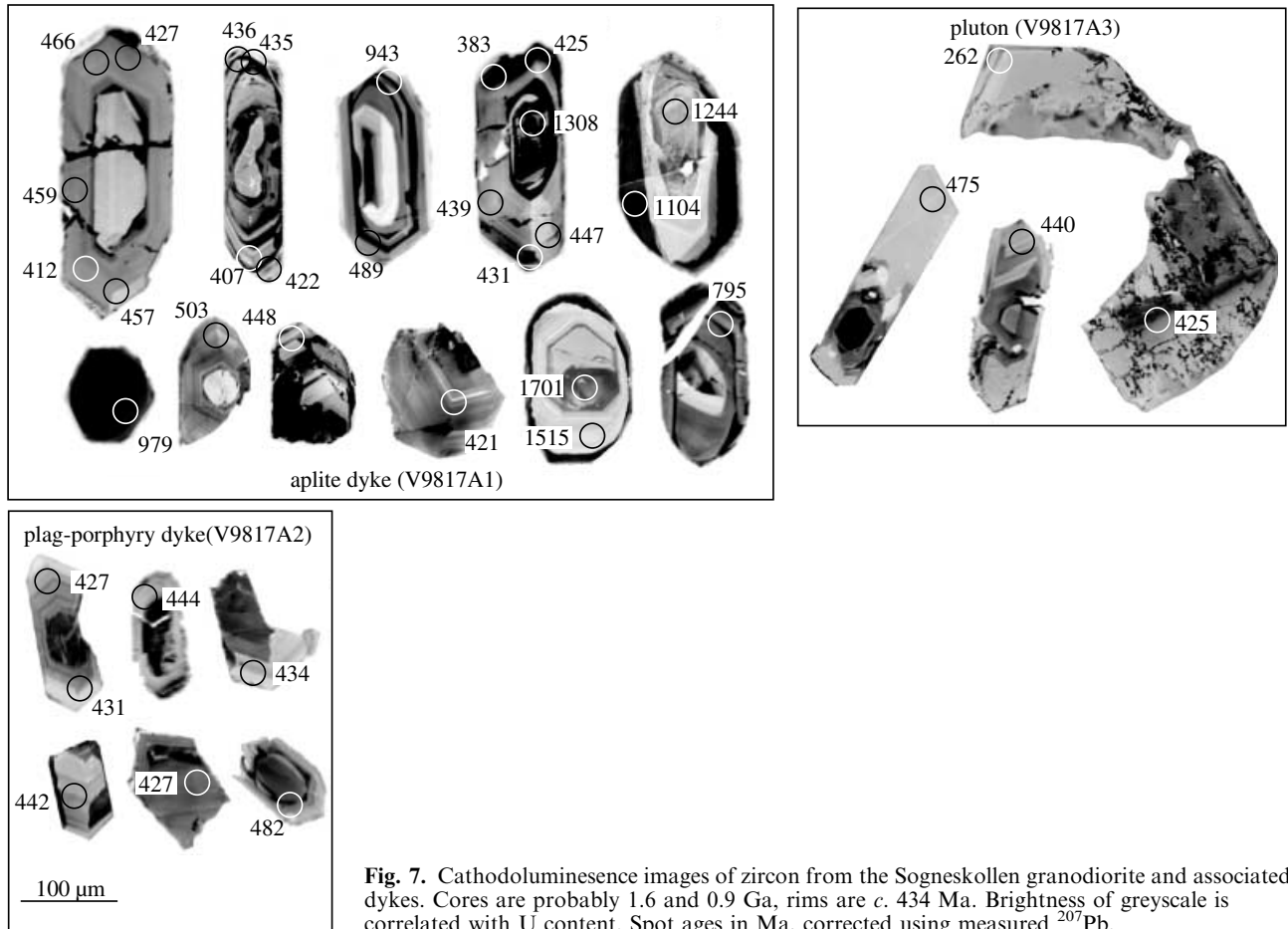
## GEOCHRONOLOGY

Previous attempts to date zircon and titanite from the Sogneskollen granodiorite have been unsuccessful, although Skjerlie *et al.* (2000) reported zircon rim ion-probe spot ages of *c.* 450–400 Ma. Skjerlie *et al.* (2000) argued that the Sogneskollen granodiorite is

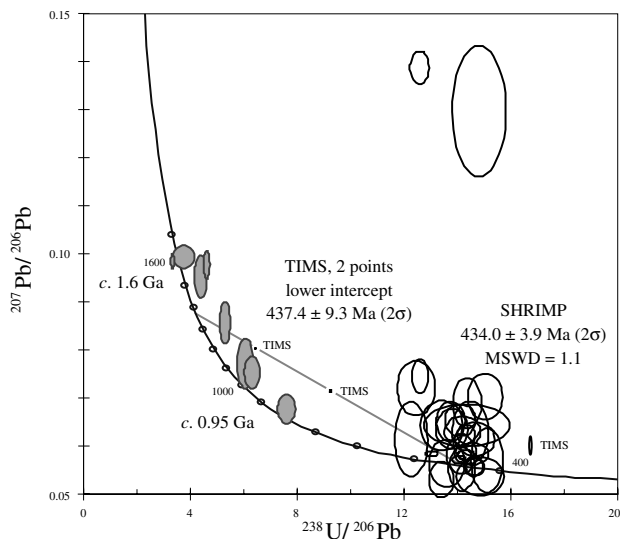
*c.* 420 Ma, based on compositional similarity to a 419 ± 9 Ma granitic dyke from the Lindås nappe (U/Pb zircon age, Austrheim, 1990) and the assumption that contractional structures within the Sogneskollen granodiorite developed during the Wenlockian emplacement of the Solund–Stavfjord Ophiolite.

We separated zircon from three samples for U/Pb dating: the Sogneskollen granodiorite (V9817A3), an aplite dyke (V9817A1) and a plagioclase–phyric dyke (V9817A2). All three samples were taken from the base of the pluton and its associated dyke swarm. The zircon exhibit obvious cathodoluminescent cores and rims (Fig. 7) that proved to have high and low Th/U, respectively. As a preliminary investigation, we analyzed one fraction of zircon from each of the three samples by single-step digestion and thermal-ionization mass spectrometry (TIMS) (Fig. 8; Table S2). All three fractions are discordant. Sample V9817A3 contains both clear and strongly cloudy zircon. A small population of clear grains was used for Sensitive High Resolution Ion Microprobe (SHRIMP) work, and a fraction of the cloudy grains analysed by TIMS. This zircon has a high U content (*c.* 1100 p.p.m.), and yields <sup>206</sup>Pb\*/<sup>238</sup>U and <sup>207</sup>Pb\*/<sup>206</sup>Pb\* ages of *c.* 373 and *c.* 610 ± 50 Ma, respectively (the large error on the <sup>207</sup>Pb\*/<sup>206</sup>Pb\* age stems from a high common Pb content). By comparison with the other TIMS data and the SHRIMP data (discussed below), these ages suggest significant Pb loss plus inheritance. The Pb loss is unsurprising in light of the high U content and cloudy appearance of the zircon. TIMS fractions V9817A1 and A2 consist of clear zircon, and are relatively low in U (*c.* 125 and 34 p.p.m., respectively). These fractions are strongly discordant, with <sup>206</sup>Pb\*/<sup>238</sup>U and <sup>207</sup>Pb\*/<sup>206</sup>Pb\* ages of 932 and 1208 ± 1.7 Ma, respectively, for V9817A1, and 662 and 972 ± 9 Ma, respectively, for V9817V2. For reference, these data define a lower intercept age of 437.4 ± 9.3 Ma, but we place little confidence in this two-point intercept. For the lower intercept to have rigorous age significance, the zircon would need to have differing proportions of an inherited component of identical age (or at least identical mean age), and no post-crystallization Pb loss (a distinct possibility given the moderate to low U concentrations). In any case, the lower intercept age is in agreement with the SHRIMP results reported below and thus tends to reinforce those results.

The Sensitive High Resolution Ion Microprobe (SHRIMP) at the Stanford/USGS Microanalytical Facility was used to obtain 37 spot ages from grain cores and rims (Fig. 8). Seven spot analyses of grain cores from the aplite dyke have ages that lie along a rough mixing line from *c.* 0.95 to *c.* 1.6 Ga, and Th/U ratios of 0.2–1.3, indicating an igneous provenance. Textural relations visible in cathodoluminescence images reveal that this range of ages is not the result of mixing of different age domains during ion-probe



**Fig. 7.** Cathodoluminescence images of zircon from the Sogneskollen granodiorite and associated dykes. Cores are probably 1.6 and 0.9 Ga, rims are *c.* 434 Ma. Brightness of greyscale is correlated with U content. Spot ages in Ma, corrected using measured  $^{207}\text{Pb}$ .



**Fig. 8.** Tera-Wasserburg diagram of U/Pb zircon analyses of the Sogneskollen granodiorite and associated dykes. SHRIMP ratios are uncorrected and shown as  $\pm 1\sigma$ . Core analyses shown with filled grey ellipses.

analysis; the range therefore most likely reflects Early Proterozoic grains that underwent Late Proterozoic Pb loss. Tectonics events of 1.75–1.45 Ga and 1.25–0.95 Ga are known in Scandinavia as the Gothian and Sveconorwegian, respectively, and have both been identified in the southern part of the WGC immediately east of the study area (Skår, 1998). Except for one spot age of 795 Ma, which cathodoluminescence imaging suggests is the result of simultaneous analysis of two age domains, the remainder of the spot ages from the zircon rims are Palaeozoic—503–262 Ma. The three oldest Palaeozoic spot ages, 503, 489 and 482 Ma also appear to be the result of mixing of two age domains. Twenty-one of the remaining spot ages can be pooled to form a  $^{207}\text{Pb}$ -corrected, weighted mean  $^{238}\text{U}/^{206}\text{Pb}$  age of  $434.0 \pm 3.9$  Ma ( $2\sigma$  internal error) with a good fit (MSWD = 1.1). The three separate samples have weighted mean  $^{238}\text{U}/^{206}\text{Pb}$  SHRIMP ages that are equivalent at the 95% confidence interval, and are therefore coeval at the current level of precision. Four younger Palaeozoic spot ages, 412, 407, 383 and 262 Ma, lie significantly (95% confidence interval) outside the  $434.0 \pm 3.9$  Ma population. The low 262 Ma result is from a spot with extremely high (*c.* 3300 ppm) U concentration and clearly reflects significant Pb loss. The 412, 407 and 383 Ma results

are from spots with low U concentrations: *c.* 78, 61 and 42 p.p.m., respectively. Pb loss associated with radiation-induced lattice damage is a less attractive option to explain these results and perhaps they reflect very thin overgrown rims at *c.* 400 Ma. Considering all the results, *c.* 434 Ma is interpreted as the crystallization age of the magma and the younger Palaeozoic spot ages as the result of Pb loss and/or later overgrowths. It is possible, however, that the 434 Ma age includes inherited grains.

The age of metamorphism in the Solund–Hyllestad–Lavik area has been partly constrained by Chauvet & Dallmeyer (1992), who dated four muscovite and two hornblende (Fig. 3a). Two of the four muscovite yielded well-behaved spectra with ages of  $398.5 \pm 0.3$  and  $394.3 \pm 0.3$  Ma (our recalculations); the other two spectra show serially increasing or decreasing step ages for which we take weighted mean ages of  $c. 403 \pm 3$  Ma. The muscovite ages may decrease with structural depth (Fig. 3b)—a progression typical of muscovite  $^{40}\text{Ar}/^{39}\text{Ar}$  ages within the NSDZ (Andersen, 1998; Boundy *et al.*, 1996; Chauvet & Dallmeyer, 1992; Eide *et al.*, 1997); they indicate that the amphibolite facies extension along the NSDZ was over by 403 Ma. The two hornblende ages of Chauvet & Dallmeyer (1992), *c.*  $395.9 \pm 1.1$  Ma and  $c. 409 \pm 3$  Ma, are systematically older than muscovite from the same localities by 2–4 Ma.

There are presently no radiometric ages on the eclogites in the Lavik area, so, we provisionally assume that the Lavik eclogites formed during the *c.* 410–400 (?) Ma HP event in the WGC (Lutro *et al.*, 1997; Mearns, 1986; Mørk & Mearns, 1986; Root *et al.*, 2000; Terry *et al.*, 2000) because the Lavik eclogites are part of this complex.

## DISCUSSION

An important finding from our study is that E-directed thrusting recorded in the dyke swarm at the base of the Sogneskollen granodiorite finished after  $434 \pm 4$  Ma, and that major top-W extension at amphibolite facies conditions developed subsequently. As implied above, igneous rocks similar in age to the Sogneskollen granodiorite are widespread in the Uppermost and Upper Allochthons. These bodies include, for example, the Bindal Batholith ( $430 \pm 7$ – $447 \pm 7$  Ma U/Pb zircon ages, Nordgulen *et al.*, 1993) and the Smøla–Hitra Batholith (*c.* 435 Ma U/Pb zircon, Nordgulen *et al.*, 1995). In western Norway, the Sunnhordland Batholith (Rb/Sr *c.* 430 Ma, Andersen & Jansen, 1987; Fossen & Austrheim, 1988), the Bremanger Granodiorite and the Gåsøy diorite (U/Pb zircon  $440 \pm 5$  Ma, Hansen *et al.*, 2001) are arc plutons intruding ophiolites, arc volcanic rocks and olistostromes. Thus, the relationships observed in the Hyllestad area—a granodiorite intruding oceanic rocks and their sedimentary cover, but not the structurally lower continental rocks—are

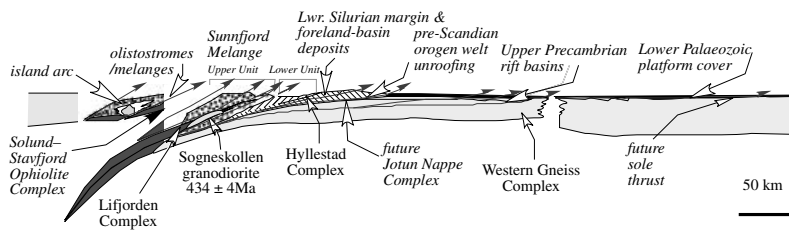
typical of this suite of broadly Llandoveryan to Wenlockian igneous rocks, and support a correlation of the Lifjorden Complex with the oceanic nappes of the Upper Allochthon. The nearest oceanic rocks of the Upper Allochthon are the Solund–Stavfjord ophiolite, which has a U/Pb zircon age of  $443 \pm 3$  Ma (Dunning & Pedersen, 1988) and is overlain by the Staveneset Group metagraywacke and arc-related metavolcanics (Furnes *et al.*, 1990). The ophiolite probably formed near a continent, because some of the MORB sills intruded greywackes containing metamorphic rock fragments, abundant clastic quartz grains and clastic zircon of Early Proterozoic to Middle Ordovician age (Pedersen & Dunning, 1993; Skjerlie *et al.*, 1989). Recently, an arc rhyolite landslide block in the Devonian Solund Basin has been dated at  $439 \pm 1$  Ma (Hartz *et al.*, in press), and unpublished U/Pb data from a quartz-diorite dyke in the ophiolite suggest that the terminal magmatic stage of the ophiolite is late Llandoveryan (Y. Dilek & H. Furnes pers. com. 2001). The final stages of the ophiolite emplacement onto the continental Middle Allochthon Høyvik Group took place during deposition of the Sunnfjord obduction melange and the Herland Group, a Wenlockian passive-margin sequence (Andersen *et al.*, 1990).

The presence of the  $434 \pm 4$  Ma Sogneskollen granodiorite within oceanic rocks correlated with the Solund–Stavfjord ophiolite does not conflict with the Wenlockian emplacement age of that ophiolite (Fig. 2). Whether the top-E deformation recorded in the Sogneskollen granodiorite can be correlated with the Wenlockian emplacement age of the Solund–Stavfjord ophiolite is, however, an open question. There are at least two possible scenarios that can explain the geological observations: (1) While the base of the Wenlockian is quoted as 428 Ma, the actual radiochronologic constraint is  $= 430.1 \pm 2.4$  Ma (Tucker & McKerrow, 1995). This age is indistinguishable from  $434 \pm 4$  Ma at the 95% confidence interval, and thus, the top-E displacement in the Solund–Hyllestad–Lavik area could be temporally and kinematically equivalent to emplacement of the Solund–Stavfjord ophiolite, but it must have taken place in the earliest Wenlockian. (2) The crystallization age we report for the Sogneskollen granodiorite may be wrong. As discussed above, at the current level of precision, nearly all of the SHRIMP rim ages form a single population but the spread of rim ages is large enough that higher precision analyses might reveal more than one population within the cluster.

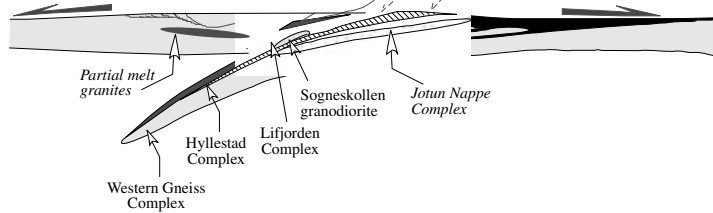
## BURIAL AND EXHUMATION OF THE HP ROCKS

We infer the following tectonic history for the study area (Fig. 9). (1) The Hyllestad Complex was deposited on the WGC; it could possibly be as young as Wenlockian if it is part of the Lower Allochthon,

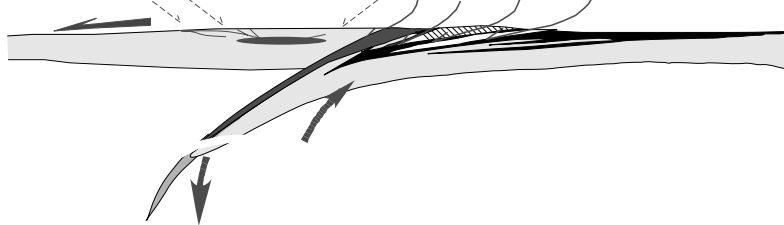
428 Ma (base of the Wenlockian): convergence rate 8–12 cm/a



410 Ma Early Devonian: subduction of WGC & onset of extension at high structural levels



400 Ma Middle Devonian: major extension across the belt



**Fig. 9.** Tectonic evolution of western Norway. Following crystallization of Sogneskollen granodiorite at *c.* 434 Ma, Upper Allochthon emplaced onto lower units around *c.* 428 Ma. Lower Allochthon and WGC subducted to high pressures at *c.* 410 Ma. Major exhumation finished by *c.* 400 Ma.

and likely experienced a *c.* 447 Ma metamorphism if it is part of the Middle Allochthon Høyvik Group. (2) The Lifjorden Complex formed as part of the Upper Allochthon (chiefly as sediments deposited on mafic crust) and then was intruded by the 434 Ma Sogneskollen granodiorite. (3) The Lifjorden Complex was thrust eastward over the Hyllestad Complex soon thereafter—likely extending into the Wenlockian—and imprinting minor top-E structures in both units. (4) At *c.* 410–400 (?) Ma, the WGC was subducted to 80 km depth, forming the Lavik eclogites; the Hyllestad Complex was buried/subducted to 50 km depth, metamorphosing the highly aluminous schists; and the Lifjorden Complex was buried to 30 km depth. (5) By 403 Ma, when muscovite within the WGC closed to Ar loss (Chauvet & Dallmeyer, 1992), the 0.7–0.9 GPa metamorphism common to all the units in the Solund–Hyllestad–Lavik area occurred. This implies that, by that time, the WGC and Hyllestad Complex were exhumed *c.* 50 km and *c.* 20 km vertically by the NSDZ, respectively. This exhumation corresponds to a normal displacement of 70–110 km depending on the synkinematic dip (40–25°) of the NSDZ (present dip *c.* 25°). While 70 km of normal displacement is compatible with the

integrated shear strain calculated from strain measurements in the *c.* 2–3 km thick NSDZ in the Sunnfjord area (Hveding, 1992), the 110 km estimate based on the present orientation of the shear zone may be too high, because of the asymmetrical uplift that accompanied the denudation of the mountain belt in western Norway. (6) At an unknown later time, the Solund fault juxtaposed the *c.* 0.4 GPa sillimanite-stable overprint of the Hyllestad Complex against the prehnite–pumpellyite facies Solund Conglomerate (= 0.3 GPa)—only a few km of exhumation.

## CONCLUSIONS

Geological field relations, combined with thermobarometry, structural petrology and geochronology, document that the geological history of the Solund–Hyllestad–Lavik area involved (i) intrusion of 434 Ma granodiorite into oceanic rocks; (ii) emplacement of those rocks over continental margin sediments; (iii) subduction of the continental margin sediments to 50 km and the underlying continental crystalline rocks to 80 km; and (iv) exhumation and attenuation of the entire sequence by normal-sense motion along the



Nordfjord–Sogn detachment zone by 403 Ma. These observations solidify our understanding of the timing and nature of the cycle of ophiolite emplacement, continental subduction and large-scale exhumation that appears to typify the Caledonian history of western Norway.

#### ACKNOWLEDGEMENTS

Physics of Geological Processes publication no. 13. Supported by NSF grant EAR-9814889 and the Physics of Geological Processes Center, University of Oslo. Helpful reviews were given by H. Brueckner and S. Cuthbert. Some of the samples and data were collected by A. Vasquez.

#### SUPPLEMENTARY MATERIAL

Tables S1 and S2 are available for downloading from <http://www.blackwellpublishing.com/products/journals/suppmat/JMG/JMG468/JMG468sm.htm>.

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*Received 14 February 2002; revision accepted 21 August 2002.*