Tectonically fragmented and dismembered ophiolite assemblages, in the eastern part of Iapetus, or the Ægir Sea. This involved an anticlockwise-rotating Baltican plate facing first Siberia, in Cambrian time, the Caledonides (Furnes et al. 1980, 1985, Pedersen et al. 1988). Their importance to our understanding of the polyphase evolution of the mountain belt far outweighs their comparatively modest and restricted areal extent; yet their very existence, as ophiolites, was unknown barely thirty years ago. At that time, detailed field studies and geochemical investigations were initiated on several of the thick, Early Palaeozoic, 'greenstone' units occurring in the nappes of the Central Norwegian Caledonides (Gale & Roberts 1972); and it was soon realised that we were dealing not with simple, homogeneous, metabasaltic greenstones, but rather with a spectrum of mostly mafic, magmatic rocks of diverse palaeogeographic settings, including fragmented ophiolitic assemblages (Gale & Roberts 1974, Prestvik 1974, 1980, Grenne et al. 1980, Roberts 1980, Heim et al. 1987, Grenne 1989, Slagstad 1998 and in prep.). Similar studies in other parts of Norway led to comparable discoveries (Furnes et al. 1976, 1985, 1988, Sturt et al. 1979, 1991, Boyle 1980, Boyd 1983, Thon 1985, Fuller 1986, Pedersen et al. 1988).

While the lithological variations, pseudostratigraphies and geochemical signatures of these ophiolite fragments are now fairly extensively documented, their crystallisation ages are comparatively poorly constrained, with a few exceptions. For obtaining precise isotopic ages, reliance has been placed on U-Pb dating of zircons from the few felsic rock types that occur within the ophiolitic assemblages, namely plagiogranites and their effusive or near-surface counterparts. The Norwegian Caledonide ophiolites were initially divided into two age groups based on geological and biostratigraphic relationships (Furnes et al. 1985). Although this early subdivision is partly incorrect, U-Pb geochronology has, in fact, demonstrated the presence of two generations of ophiolites, i.e., a Late Cambrian-Early Ordovician grouping of ophiolites generated in a general suprasubduction-zone oceanic setting; and a smaller group of Late Ordovician-Early Silurian age, of mainly marginal basin origin (Dunning & Pedersen 1988, Pedersen et al. 1988). The early ophiolites generally fall within the time interval 495-475 Ma, whereas the younger ones were generated at around 445-437 Ma.
In this contribution we present the results of a U-Pb dating study on zircons separated from two felsic magmatic rocks – a gneissic trondhjemite and a low-K rhyodacite – which form a small part of the Bymarka ophiolite (Slagstad 1998, in prep.), in the environs of Trondheim (Fig. 1). A secondary thrust or slide, corresponding to the Horg Fault of Walsh (1986) and marked by a line with single tags, dissects the Hovin and Horg Groups and parts of the ophiolitic assemblages.

Geological setting

The dismembered and fragmented ophiolites of central Norway occur within the exotic terranes of the Koli Nappes, which form the higher parts of the Upper Allochthon of Scandinavian Caledonide tectonostratigraphy. In the western Trondheim Region, and in the case of the Bymarka ophiolite, we are dealing specifically with the Støren Nappe (Fig. 1), at the base of which lies the fragmented Støren ophiolite (Gale & Roberts 1974). There is good evidence to suggest that the Støren ophiolite and the tectonically subjacent rocks of the Gula Complex (Gula Nappe) were initially deformed and metamorphosed in Early Ordovician time (Guezou 1978, Lagerblad 1983, Grenne & Lagerblad 1985) (the ‘Trondheim event’), broadly coevally with obduction (Furnes et al. 1980), prior to uplift, erosion and accumulation of the Mid Arenig and younger, volcanosedimentary successions of the Hovin and Horg Groups (Vogt 1945, Bruton & Bockelie 1980) (Fig. 1). The Lower and Upper Hovin Groups, in particular, are richly fossiliferous, especially in the Hølonda-Løkken district where the Arenig-Llanvirn faunas are largely of Laurentian affinity (Bruton & Bockelie 1980, Neuman & Bruton 1989). In southeastern parts of the Lower Hovin Group, however, fossils of this age are less common and of Baltic aspect (Spjeldnaes 1985, pers. comm. 2002). The Ordovician Hovin and Horg Group successions were subsequently deformed and metamorphosed during the Late Silurian-Early Devonian, Scandian orogeny. Along the eastern side of the Gula Complex, black phyllites intercalated with pillowed basalts of the Fundsjø Group contain the European dendroid graptolite Rhabdinapora flabelliforme, of Tremadoc age. These U- and V-rich, black phyllites are a characteristic feature of the epicontinental, Baltoscandian margin of Baltica (Gee 1981, Andersson et al. 1985).

Ophiolites, or fragmented ophiolites, have been described from areas northwest of the Støren ophiolite, at Vassfjellet (Grenne et al. 1980), Løkken (Grenne 1989), Resfjell (Heim et al. 1987) and Grefstadfjell (Grenne et al. 1980) (Fig. 1). In reality, and judging from the map pattern (Fig. 1), these all appear to belong to one and the same ophiolitic complex. The Bymarka ophiolite of the present study has, indeed, been correlated with the Vassfjellet and Løkken ophiolites (Grenne et al. 1980). In his memoir on the geology of the Trondheim Region, Carstens (1920) introduced the term ‘Bymark Group’ (named after the forested and moorland area ‘Bymarka’ W and SW of Trondheim) for the major part of the thick, volcanic, ‘greenstone’ units occurring in the western Trondheim district. In this definition, the Bymark Group embraced not only the magmatic rocks but also the overlying polymict ‘greenstone conglomerate’ which occurs in many places in the region (Figs. 1 & 2). Interestingly, Carstens (1920) considered that all the ‘greenstone’ units mentioned above (now described as fragmented ophiolites) formed part of his Bymark Group. Of the aforementioned ophiolites, only those at Løkken and Vassfjellet have hitherto been dated (U-Pb, zircons) (Dunning 1987, Dunning & Grenne, in prep; also Dunning, pers. comm., cited in Sturt & Roberts 1991, p. 751), yielding ages for plagiogranite dykes of 493 ± 10 and 487 ± 5 Ma for Løkken, and 480 ± 4 Ma for Vassfjellet.

A new and exciting aspect of the Caledonian geology of this region, and indeed of the Bymarka district, has been the discovery of relics of a blueschist-facies assemblage (primarily glaucophane, zoisite, rutile and albite) within the highly deformed, very basal part of the Bymarka ophiolite (Eide & Lardeaux 2002) (Fig. 2). The high-pressure paragenesis is strongly retrogressed, firstly at amphibolite facies and subsequently under greenschist-facies conditions. We will come back briefly to the consequences of this new development in the discussion.
Bymarka ophiolite

The constituent lithologies of the Bymarka ophiolite, and their petrography and geochemistry, have been described in detail by Slagstad (1998 and in prep.). Accordingly, we here present just a summary, concentrating on the geochemical features and in particular on the felsic rocks which are the subject of this isotopic dating investigation.

Although the rocks were moderately to strongly deformed during the Early Ordovician and later, Scandinavian orogenic events, large parts of an original ophiolite pseudostratigraphy appear to be preserved. In western areas, gabbric rocks are predominant, but these give way eastwards first to a unit dominated by mafic dykes and then to a thick sequence of pillow lavas with sporadic hyaloclastites, tuffites, quartz keratophyres (metarhyodacites) and layers of jasper and rare coticules. Overlying the pillow lavas is a unit dominated by felsic and mafic agglomerates with thinner layers of magnetite-rich metachert and felsic volcanites. Locally, the agglomerates grade into greenstones and pillow lavas. This unit is interpreted to represent ocean-floor sediments and exhalatives.

Geochemically, the mafic dykes and lavas have ocean-floor affinities, similar to correlatable ophiolitic rocks in the Løkken area (e.g., Gale & Roberts 1974, Grenne 1989). Three main types of felsic intrusive or extrusive rocks have been identified in the Bymarka ophiolite. These are termed the Klemetsaunet rhyodacite, Fagervika trondhjemite and Byneset trondhjemite (Fig. 2). The rocks are, in fact, metarhyodacites and meta-trondhjemites, but the prefix 'meta' has been omitted for simplicity. Of these, only the first two have been the subject of isotopic dating and are described briefly below (for details, see Slagstad 1998 and in prep.).

The Klemetsaunet rhyodacite (sample KL1; Figs. 2 and 3a) occurs as sheets and dykes ranging from a few dm to several tens of metres in thickness, and up to a few hundred metres in length. Similar rhyodacites are associated with the sedimentary unit described above, suggesting that it formed in an ocean-floor setting, coevally with the ophiolite. It is composed of microcrystalline quartz and albite, and is characterised by abundant, euhedral garnet and amphibole porphyroblasts interpreted to be metamorphic based on field observations such as bleached haloes and transection of foliation. The Fagervika trondhjemite (sample TR97K; Figs. 2 and 3b) is a medium- to coarse-grained rock of transitional trondhjemitic to granitic composition and gneissic character, forming an intrusive body ca. 10 km in length and up to hundreds of metres thick (Fig. 2). Along its margins and in thin dyke apophyses, the Fagervika trondhjemite is chilled against the host greenstones. In addition to quartz, plagioclase and K-feldspar, muscovite is locally important, constituting up to 15-20 vol. % of the mode.
Based on the geochemical data, the two rock types are clearly distinguished (Table 1). The Klemetsaunet rhyodacite has a flat rare earth element (REE) pattern with a slight light REE (LREE) depletion and a negative Eu anomaly, similar to that of plagiogranites in the correlatable Løkken ophiolite (Grenne 1989) (Fig. 4). A small, negative Nb-Ta anomaly may suggest formation in a suprasubduction-zone setting (Elthon 1991), as at Løkken (Grenne 1989). Alternatively, this anomaly may derive from fractionation. In contrast, the Fagervika trondhjemite has a strongly enriched LREE and slightly enriched heavy REE (HREE) pattern, also with a negative Eu anomaly (Fig. 4), features which are comparable to those of quartz-porphyritic rhyolites in the upper member of the Løkken ophiolite (Grenne 1989). In diverse, tectonic discrimination diagrams (Wood 1980, Pearce et al. 1984), the Klemetsaunet rhyodacite plots in the field of ocean ridge plagiogranite; and its oceanic origin is also shown by high Yb/Al$_2$O$_3$ ratios. The geochemistry is consistent with the field relationships suggesting that the Klemetsaunet rhyodacite formed in an ocean-floor setting, coeval with generation of the ophiolite. The Fagervika trondhjemite, on the other hand, shows clear subduction-related signatures indicative of an increasing arc influence at the time of its generation.

Finally, it should be noted that overlying the pillow lavas in the east (Fig. 2), above a strongly sheared, inferred unconformable contact, is a polymict conglomerate dominated by clasts of metabasalt, gabbro, trondhjemite and rhyodacite with minor jasper, set in a green-schist matrix. This constitutes the basal conglomerate of the Lower Hovin Group.

**U-Pb analytical methods and results**

*Analytical methods*

The samples were crushed and separated into zircon concentrates at the Geological Survey of Norway, Trondheim. The chemical separations and zircon analyses were carried out at the Department of Geological Sciences, Brown University, Rhode Island, USA (in 1998 for sample TR97-K and in 2000 for sample KL-1). Zircon concentrates were cleaned in successive solutions of 2N HNO$_3$, 2N HCl and distilled H$_2$O, and then split by magnetic character using a Frantz Isodynamic Barrier separator. The zircons were then sieved and/or hand sorted into representative subpopulations on the basis of morpho-
logical similarity, employing visual discriminants such as size, form, colour, clarity and aspect ratio. Zircons from sample KL-1 were characterised by bulbous, epitaxial overgrowths of inferred metamorphic zircon up to 20 μm thick that armoured the entire grain. When viewed in cross-section parallel to the c-crystallographic axis, it is apparent that the armouring zircon has grown on otherwise euhedral, zircon crystals. These grains were therefore subjected to a vigorous, 24-hour mechanical abrasion to remove the overgrowths and the outermost few microns of the underlying euhedral zircon.

Zircon dissolution and ion-exchange procedures were comparable to those described by Krogh (1973) and Parrish et al. (1987). A mixed $^{205}$Pb-$^{233}$U-$^{235}$U tracer was employed. Lead was loaded to W filaments and U to Re filaments and analysed on the Finnigan MAT 261 multicollector mass spectrometer at Brown University. Pb was analysed in static multicollector mode employing Faraday cup collection of masses 208, 207, 206 and 205 while simultaneously collecting mass 204 in a secondary electron multiplier (SEM). Uranium was analysed by peak hopping into a SEM. Additional analytical details are given in the footnotes to Table 2.

**Data and age interpretation**

Four fractions were analysed from the sample of trondhjemite from Fagervika (TR97-K) and three from the Klemetsaunet rhyodacite (KL-1). The data are presented in Table 2 and graphically displayed in Figs. 5 and 6.

Zircons from sample TR97-K are all moderately discordant and the data plot in a cluster, despite having analysed different morphological crystal types. We interpret this discordance to result from Pb-loss rather than from inheritance of older radiogenic Pb. This interpretation seems justified by the very narrow range of $^{207}$Pb/$^{206}$Pb ages of these zircon fractions. If inheritance was the mechanism of discordance, we would expect a much greater range of $^{207}$Pb/$^{206}$Pb ages. A discordia trajectory anchored at 0 Ma through the data defines an upper intercept of 481.4 ± 4.7 Ma (inclusive of decay constant uncertainties). Although we have no independent data regarding when these zircons lost Pb, they are not high-U zircons and the initial, Arenig, amphibolite-facies metamorphism is believed to have taken place shortly after zircon crystallisation. Hence, there would have been very little radiation damage in these grains prior to metamorphism; and the Arenig metamorphism is therefore unlikely to have resulted in significant loss of radiogenic Pb. Another way of assessing the age of this sample is by recourse to the mean weighted $^{207}$Pb/$^{206}$Pb age of the data set, which is 481.3 ± 2.6 Ma. We therefore interpret the crystallisation age of this rock to be c. 481 Ma, but acknowledge that additional zircon analyses are required to refine the age and uncertainty of this sample.

As described above, zircons from sample KL-1 are composite grains comprising nearly euhedral crystals armoured with an epitaxial, bulbous overgrowth of new zircon that is almost certainly of metamorphic origin. The overgrowths probably developed during the time of garnet and amphibole porphyroblastesis. In order to determine the igneous emplacement age of this rock, we therefore analysed 3 fractions of strongly abraded zircons. The crystals were examined several times during the abrasion cycle and extracted only after the overgrowths had been eradicated and the outermost few microns of the pre-existing crystal removed (to lessen the likelihood of analysing those parts of the igneous

| Table 1. Average composition of investigated rocks, from Slogsted (in prep.). n=number of samples. Average REE, Th, Nb, Hf, and Ta based on three representative samples from each rock type. |
|---|---|---|
| Klemetsaunet rhyodacite | Fagervika trondhjemite |
| n | 19 | 15 |
| SiO$_2$ | 76.47 | 74.56 |
| TiO$_2$ | 0.13 | 0.22 |
| Al$_2$O$_3$ | 12.12 | 12.91 |
| Fe$_2$O$_3$ | 1.84 | 1.65 |
| MnO | 0.03 | 0.05 |
| MgO | 0.24 | 0.60 |
| CaO | 1.07 | 0.93 |
| Na$_2$O | 6.21 | 4.94 |
| K$_2$O | 0.22 | 1.96 |
| P$_2$O$_5$ | 0.02 | 0.04 |
| LOI | 0.57 | 1.42 |
| Total | 98.88 | 98.86 |
| Ba | 39 | 734 |
| Rb | 5 | 54 |
| Sr | 51 | 80 |
| Y | 100 | 35 |
| Zr | 273 | 136 |
| Nb | 7 | 5 |
| Th | 1.46 | 17.68 |
| Hf | 9.55 | 4.06 |
| Ta | 0.46 | 0.36 |
| La | 12.13 | 34.28 |
| Ce | 35.97 | 63.64 |
| Pr | 5.87 | 6.93 |
| Nd | 30.56 | 24.37 |
| Sm | 11.45 | 5.32 |
| Eu | 2.40 | 0.92 |
| Gd | 13.70 | 4.56 |
| Tb | 2.40 | 0.75 |
| Dy | 16.62 | 5.14 |
| Ho | 3.48 | 1.09 |
| Er | 11.67 | 3.79 |
| Tm | 1.67 | 0.56 |
| Yb | 11.11 | 3.86 |
| Lu | 1.90 | 0.68 |
| ($\text{Eu}/\text{Eu}^*)_N$ | 0.58 | 0.55 |
| ($\text{La}/\text{Yb})_N$ | 0.79 | 6.28 |

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Table 2. Zircon analytical data, Klemetsaunet (KL-1) and Fagervika (TR97-K)

<table>
<thead>
<tr>
<th>Sample and fraction properties#</th>
<th>Amount analysed (grains)</th>
<th>Concentration + (ppm)</th>
<th>Pb Isotopic composition$</th>
<th>Radiogenic ratio @</th>
<th>Age and Uncertainty**</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Pb</td>
<td>U</td>
<td>206/204</td>
<td>206/207</td>
</tr>
<tr>
<td>KL-1</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>200 μm prism fragments with tips, 24 hr. ab., colourless</td>
<td>2</td>
<td>4.05</td>
<td>460</td>
<td>4.218 (0.10%)</td>
<td>14.756 (0.13%)</td>
</tr>
<tr>
<td>200 μm entire x1, 24 hr. ab., colourless</td>
<td>2</td>
<td>6.21</td>
<td>686</td>
<td>3.900 (0.16%)</td>
<td>15.614 (0.14%)</td>
</tr>
<tr>
<td>125 μm entire x1, 24 hr. ab., colourless</td>
<td>2</td>
<td>6.42</td>
<td>40.0</td>
<td>1.530 (0.26%)</td>
<td>4.381 (0.26%)</td>
</tr>
<tr>
<td>TR97-K</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>130 μm, stumpy prisms, pale brown</td>
<td>6</td>
<td>20.5</td>
<td>252</td>
<td>4.943 (0.01%)</td>
<td>17.244 (0.023%)</td>
</tr>
<tr>
<td>150 μm, prisms, pale tan</td>
<td>8</td>
<td>22.7</td>
<td>276</td>
<td>4.683 (0.01%)</td>
<td>17.383 (0.28%)</td>
</tr>
<tr>
<td>100 μm prisms, pale tan</td>
<td>8</td>
<td>24.8</td>
<td>302</td>
<td>4.709 (0.01%)</td>
<td>17.419 (0.028%)</td>
</tr>
<tr>
<td>80 μm prisms, colourless</td>
<td>20</td>
<td>20.4</td>
<td>236</td>
<td>3.988 (0.01%)</td>
<td>15.032 (0.03%)</td>
</tr>
</tbody>
</table>

# ab=zircons mechanically abraded for indicated number of hours
+Concentration is total Pb and includes blank Pb, common Pb in zircon, and radiogenic Pb. Total procedural blanks are ~2 picograms for U and ~10 picograms for Pb. Because weights of grains are estimated and a mixed U-Pb tracer was employed, uncertainty in estimates of grain weight affects only concentration data, not calculated U-Pb or Pb/Pb ages.
$Measured isotopic ratios corrected for mass fractionation of ~0.11% per atomic mass unit based on replicate analyses of NIST SRM 981 and 982 and adjusted for small amount of 206Pb in tracer.
@Value in parentheses is percent uncertainty in the calculated ratio, stated at the 2-sigma level.
**Decay constants: 238U = 1.5513 E-10/yr; 235U = 9.8485 E-10/yr. Atom ratio 238U/235U = 137.88. Uncertainty in the calculated ages is stated at the two-sigma level and estimated from combined uncertainties in calibrations of mixed 205Pb - 233U - 235U tracer, measurement of isotopic ratios of Pb and U, common and laboratory blank Pb isotopic ratios, Pb and U mass fractionation corrections, and reproducibility in measurement of NIST Pb and U standards.
crystal that had undergone Pb-loss during the event that produced the overgrowths). The data define a precise concordia age (Ludwig 1998) of 481.5 ± 4.5 Ma (inclusive of decay constant uncertainties) that we interpret as the time of emplacement of these sheets and dykes. For purposes of convenience in the discussion, the age of sample KL-1 is hereafter cited as 482 Ma and, likewise, the age for TK97-K is given as c. 481 Ma.

Discussion

Age
The crystallisation ages presented here - 482 Ma for the Klemetsaunet rhyodacite and c.481 Ma for the Fagervika trondhjemite - are mutually indistinguishable, even though the two rock-types show different geochemical signatures. The field criteria clearly denote that the rhyodacite and trondhjemite form part of the deformed and metamorphosed Bymarka ophiolite, although the latter probably intruded at a comparatively late stage. Their present distributions within the disrupted and sheared mafic rocks of the ophiolite probably owe much to tectonic imbrication, firstly during the Early Ordovician obduction event and later at the time of polyphase Scandian (Late Silurian-Early Devonian) orogenic deformation.

Based on the current time-scale for the Early Palaeozoic (Tucker & McKerrow 1995, Davidek et al. 1998, Landing et al. 2000), the dated felsic rocks from Fagervika and Klemetsaunet range from Late Tremadoc to Early Arenig, taking into account the error bars (Fig. 7). Comparing these ages with the precise U-Pb dates generated from zircons extracted from plagiogranite dykes in the Løkken and Vassfjellet ophiolites (Dunning 1987, Dunning & Grenne, in prep.) (Fig. 7), we see that there is a complete overlap in the data. The trondhjemite from Vassfjellet (480 ± 4 Ma) is closest in age to the rocks from Bymarka, whereas the felsic rocks from the Løkken ophiolite are largely Tremadoc, extending into the Late Cambrian. Here, we accept that the revised boundary between the Cambrian and the Ordovician is now placed at circa 490 Ma (Davidek et al. 1998).

Whether or not there is any real difference in general age between the Løkken ophiolite on the one hand, and the Bymarka and Vassfjellet ophiolites, is difficult to say, even though one might expect a broad contemporaneity, judging from the geological map (Fig. 1). It is not uncommon, however, for oceanic magmatic rocks in ophiolite assemblages to have been generated over a time interval of 10-15 million years. In the case of the ophiolites in the Karmøy-Bømlo region of SW Norway, a period of some 25 million years separates the oldest and youngest members of the fragmented assemblages (Pedersen 1992).

On the island of Ytterøya, some 50 km northeast of Bymarka, an oceanic metatrdnhjemite interbanded with metabasalts, has yielded a U-Pb zircon crystallisation age of 495 ± 3 Ma (Roberts & Tucker 1998). On Leka, in the Leka Ophiolite Complex, a trondhjemite in the intermediate-level gabbro/dolerite dyke unit has been dated (U-Pb zircon) to 497 ± 2 Ma (Dunning & Pedersen 1988). Although these two datings point to similar, Late Cambrian ages, they derive from felsic rocks that are considered to coexist with different parts of different mafic, oceanic assemblages. Moreover, they overlap in time with at least one of the dated samples from Løkken (Fig. 7) — and both of the Løkken samples overlap in age with our two samples dated from the Bymarka ophiolite. The logical inference from these comparisons, therefore, is that although we cannot easily separate these various ophiolite fragments in terms of age, the data do suggest that the Vassfjellet and Bymarka rocks...
are younger than those on Leka and Ytterøya. The dated zircons are, in all cases, extracted from felsic rocks inferred to be broadly coeval with different parts of ocean floor magmatic complexes, spanning a time interval of some 15-20 million years, from the Late, or possibly late Mid Cambrian up to earliest Arenig time. This also accords with the upper constraint on their age, imposed by the Mid to Late Arenig faunas in the sedimentary successions of the area southwest of Trondheim — in a basin accumulating debris shedded from the rapidly eroding, obducted and metamorphosed ophiolite.

**Tectonic model**

As noted earlier, the early Caledonian phase of deformation and metamorphism in this region involved obduction of the Støren ophiolite upon its Gula Complex substrate. The weight of isotopic and biostratigraphic evidence points to an Early Arenig temporal slot for this major tectonothermal event, which also evidently involved the other, neighbouring ophiolite fragments, including that of Bymarka. While the timing of this Early Ordovician ‘Trondheim event’ seems secure, the palaeogeographic and plate-kinematic scenario is more speculative (cf., Gale & Roberts 1974, Gee 1975, Roberts et al. 1985, Stephens & Gee 1985, Grenne & Roberts 1998, Grenne et al. 1999). Models relying specifically on faunal province affilition in Early Ordovician time and the inference that faunas of, e.g., largely North American affinity occur only in sedimentary rocks deposited on or along the margins of Laurentia, have naturally favoured reconstructions involving the early Caledonian obduction of most of the Norwegian ophiolites upon the margin of the Laurentian palaeocontinent (Gee 1975, Bruton & Bockelie 1980, Stephens & Gee 1985). Reliance on other criteria, such as general geological and structural relationships and geochemistry, led some workers (Roberts & Gale 1978, Ryan et al. 1980, Sturt & Roberts 1991) to question the too simplistic view of faunal provincialism. In this case, palaeo reconstructions favoured initial ophiolite emplacement upon either the margin of Baltica or an arc-related microcontinent (Roberts 1980, Roberts et al. 1985). This view has received support from the discovery of early Caledonian, HP eclogite-facies, metamorphic parageneses in the Baltic-Iapetus transition zone mafic rocks of the Seve Nappes (lower part of the Upper Allochthon) in northern Sweden (Stephens & van Roermund 1984, Andréasson et al. 1985, Mørk et al. 1988) — eclogites generated in a subduction zone facing away from the Baltic margin — and some support has also come from lead isotope data collected from the nappes in Central Norway (Bjørlykke et al. 1993).

Refinements to the Early Palaeozoic, palaeomagnetic database for Baltica, Laurentia and Siberia over the last few years have also had an important bearing on interpretations of the palaeotectonic settings for Norwegian Caledonide ophiolite generation, metamorphism and obduction. From Late Vendian time and through much of the Cambrian period, Baltica lay in an inverted position and reasonably close to the Siberian plate rather than Laurentia (Smethurst 1992, Torsvik et al. 1992, 1995, Torsvik & Rehnström 2001), but by Late Cambrian time it had begun to rotate quite rapidly in an anticlockwise sense. Collision of Baltica with a cryptic arc (Dallmeyer & Gee 1986), or arc-affected microcontinent (Sturt & Roberts 1991), above the eclogite-generating subduction zone (and, thus, broadly coeval with Late Cambrian-Early Ordovician, Finnmarkian deformation), was then reinterpreted as relating to interaction between Baltica and Siberia and the intervening arc (Smethurst 1992, Torsvik et al. 1992, 1995, 1996, Torsvik 1998), or arcs. This is in the oceanic tract which has been termed the ‘Eastern Iapetus Ocean’ (Pickering & Smith 1995) or Ægir Sea (Torsvik & Rehnström 2001) (Fig. 8a).

Farther south, into Central Norway, the Seve Nappes are succeeded by the more exotic Koli Nappes, including the Gula Complex and Støren ophiolite, and Fundsjø Group in the east, units which were initially deformed and metamorphosed in Early Ordovician time. The epicontinental Gula differs lithologically, in many ways,
Arenig, platformal and arc-fringing, Bathyurid trilobite occur-
500-480
1n the early part of Arenig time, the ophiolites are believed to have

Fig. 1998,
pressive scenario, leading to non-synchronous obduction!thrusting
situation, with Baltica then facing Siberia. With an anticlockwise
rotation Baltica, the Ægir Sea rapidly contracted in a dextral trans­
pressive scenario, leading to non-synchronous obduction/thrusting
of ophiolites. G-- Gula microcontinental block; NCB - North
China Block; SCB - South China Block; Encircled B - Tremadoc­
Cynagen, platformal and arc-fringing, bathyurid trilobite occu­
rences. Encircled P - Tremadoc-Arenig, platformal Ptychopyg­
eine/Megalaspis trilobite occurrences. (b) N-S schematic section at
the earlier stages of oceanic contraction, in Late Cambrian time,
with eclogites developed at depth in the earlier, Siberia-facing, sub­
duction zone. These eclogite-facies rocks were accreted onto the
Baltscandian margin (as part of the Seve Nappes) in Late Cam­
brian to earliest Tremadoc time (the Finnmarkian orogenic event).
(c) Schematic section showing the ophiolite assemblages and the
blueschist parageneses of the Tornheim Region originating from the
suprasubduction-zone domain and actual subduction zone,
respectively, to the north of the «Gula microcontinents», in a Lau­
rentia-facing situation, following anticlockwise rotation of Baltica.
In the early part of Arenig time, the ophiolites are believed to have
been obducted onto the microcontinent during this Early Ordo­
vician, transpressive, tectonothermal event.

from the Seve, and along its western edge contains a tec­
tonie mélange in deep-marine pelites directly beneath the
obducted, mylonitised Støren ophiolite (Horne 1979). In the east, the Baltic graptolite Rhabdinopora
occurs in phyllites intercalated with metabasalts of the
Fundsje Group. Accepting that the Støren, Løkkene, Vass­
fjellet and Bymarka ophiolite fragments did form part of
a common, subduction-influenced, ocean floor in Late
Cambrian to Tremadoc time, then the HP blueschist-
facies parageneses described by Eide & Lardeau (2002)
from the base of the Bymarka ophiolite occupy a compara­
able position to that of the mélange; effectively record­
evidence of both subduction and obduction at roughly the same structural level.

Incorporating these various elements into a credible
paleotectonic setting for this segment of the Baltosca­
dian margin in Late Cambrian-Early Ordovician time,
we build on the notion that the ophiolites of the Trond­
heim district, including Bymarka, were generated in a
suprasubduction-zone setting outboard of a microconti­
nent (Roberts 1980, Ihlen et al. 1997, Grenne et al.1999),
represented today by the Gula Complex (Fig. 8b), in Late
Cambrian to Tremadoc time. Their obduction, in
approximately Early Arenig time, was a form of large-
scale tectonic inversion, utilising the pre-existing, ocean­
facing subduction zone, down which the HP blueschists
were generated (Fig. 8c). The 'Gula microcontinent' is
envisaged as an elongate block or sliver of Baltic crust
that rifted off and drifted away from the paleocontinent
Baltica in Vendian-Early Cambrian time, during the ini­
tiation of Iapetus opening. Farther north, this microcon­
tinent may have narrowed and possibly merged with the
cryptic arc system that is inferred to have been involved
in the earlier, Finnmarkian, arc/continent collision,
which peaked in Late Cambrian to Tremadoc time –
again, a case of mega-inversion following the oceanward
subduction that had generated eclogite-facies rocks
(Dallmeyer & Gee 1986, Mørk et al. 1988) (Fig. 8b).

Our composite reconstruction thus involves two sepa­
rate subduction systems (Fig. 8), with obduction-
related, accretionary deformation at different times, first
(Finnmarkian; Fig. 8b) within the contracting Ægir
oceanic tract between Baltica and Siberia (cf., Eide &
Lardeau 2002). As Baltica continued to rotate anticlock­
wise, its then northern margin and outboard microcon­
tinent passed from the Ægir Sea into a contracting Iape­
tus Ocean, facing Laurentia, by Early Ordovician time;
thus explaining the influx of North American faunas in
basinal and arc-fringing successions which were later to
become parts of the Støren Nappe. In this framework,
the Bymarka and related ophiolite fragments of the
Tornheim district are representative of the most out­
board and younger, subduction/obduction couplet (Fig.
8c). In central Norway, there are no eclogite-facies rocks
involved in the inboard, older, 'Seve subduction' system.
Either subduction died out southwards, or HP Finn­
markian assemblages, if they existed, have been tectoni­
cally excised during Scandian thrusting.

Fig. 8. Palaeogeographic reconstruction and suggested plate tec­
tonie scenario for the Late Cambrian to Tremadoc time interval, c.
500-480 Ma. Prior to this, in Early Cambrian time, Baltica had
been in a more inverted position and rotated fairly rapidly anti­
clockwise into comparative proximity with Siberia, across the East­
ern Iapetus Ocean or Ægir Sea, by Late Cambrian time (Torsvik
1998, Torsvik & Rehnström 2001). (a) General palaeomagnetic
reconstruction for this broad Late Cambrian-Tremadoc period,
modified from Torsvik (1998). This emphasizes the Late Cambrian
situation, with Baltica then facing Siberia. With an anticlockwise-
rotating Baltica, the Ægir Sea rapidly contracted in a dextral trans­
pressive scenario, leading to non-synchronous obduction/thrusting
of ophiolites. G-- Gula microcontinental block; NCB – North
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vician, transpressive, tectonothermal event.
Conclusions

Zircons extracted from an oceanic rhyodacite sheet and an arc-related trondjemite body in the Bymarka ophiolite, close to Trondheim, Central Norwegian Caledonides, have yielded U-Pb ages of 482 ± 5 and c. 481 Ma, respectively. The dates, straddling the Tremadoc-Arenig boundary on the currently accepted time-scale, are interpreted to represent the crystallisation ages of these felsic rocks. Geochemical data indicate that the rhyodacite is comparable to plagiogranite dykes occurring in the nearby Lokken ophiolite, and is considered to be co-genetic with the metabasaltic greenstones of the Bymarka ophiolitic assemblage. The trondjemite, on the other hand, transects the greenstone layering and shows a clear, subduction-related, geochemical signature.

In common with neighbouring, dismembered and tectonically disrupted ophiolites, the Bymarka ophiolite is considered to have been obducted upon a microcontinent represented by rocks of the Gula Complex, in Early Arenig time. The oldest fossil-bearing rocks overlying the deformed, metamorphosed and uplifted ophiolites of this district are of Mid Arenig age. Prior to the obduction event, seaward subduction of oceanic crust led to the generation of late-stage arc products and provided the appropriate conditions for the formation of high-pressure blueschist-facies metamorphic assemblages (Eide & Lardeaux 2002). This outboard, latest Cambrian-Early Ordovician, subduction/obduction system is considered to have been located in the easternmost part of the Iapetus Ocean, merging into the Ægir Sea. After facing Siberia in Cambrian time, an anticlockwise-rotating Baltica passed into the contracting Iapetan oceanic realm in the Early Ordovician, repositioning to face Laurentia. Plate rotation in a contracting ocean thus provided ideal conditions for intra-oceanic subduction, arc/back-arc systems and transpressive accretionary events involving ophiolite/arc obduction.

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