Provenance and exhumation of an exotic eclogite-bearing nappe in the Caledonides: a U–Pb and Rb–Sr study of the Jæren nappe, SW Norway

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Abstract: The provenance and exhumation history of paragneisses that enclose the 471–458 Ma eclogites in the Caledonian Jæren nappe, SW Norway, were constrained by performing U–Pb analysis on detrital zircon and Rb–Sr dating of mica. Zircon grains from four samples were analysed by laser ablation inductively coupled plasma mass spectrometry, providing detrital U–Pb ages between 3130 and 620 Ma. These data suggest Laurentian provenance, implying the existence of an early Caledonian (Taconian) HP fragment of Laurentian crust in the southern Scandinavian Caledonides. Several rims and recrystallized zircon patches yielded U–Pb ages averaging at 469 ± 6 Ma (2σ). These are identical to the Lu–Hf ages of the eclogites and underscore the importance of Taconian metamorphism in this part of the Caledonides. The main phengite Rb–Sr age cluster at 424.4 ± 4.6 Ma provides time constraints on the amphibolite-facies recrystallization of former high-Si phengite. Five out of seven biotite Rb–Sr ages were identical, yielding a weighted mean age of 413.6 ± 2.6 Ma. Low-Si phengite and biotite in three samples gave c. 400 Ma ages. The Rb–Sr data record the mid-crustal emplacement of the exotic Jæren nappe and its subsequent unroofing along the western edge of Baltica during the backslliding and collapse of the Scandian orogenic front.

Supplementary material: The sample locations, U–Pb isotope data and ages for zircon, and the phengite and biotite major element data are available at http://www.geolsoc.org.uk/SUP18448.

The Scandinavian Caledonides are an archetypal continental suture zone that formed in response to the collision between the proto-continents Laurentia and Baltica, following the closure of the intervening Iapetus Ocean. Terminal collision forced the deep (>200 km) subduction of the Baltoscandian margin (e.g. Scambelluri et al. 2009; van Roermund 2009), fragments of which are exposed as one of the world’s largest and best-known ultrahigh-pressure (UHP) occurrences: the Western Gneiss Complex. Although most structural features in the Scandinavian Caledonides are related to the suturing stage (the Scandian Orogeny; 430–395 Ma), the orogenic belt includes several other (U)HP terranes that record pre-Scandian subduction (e.g. Mørk et al. 1988; Corfu et al. 2003; Brueckner & van Roermund 2007). Their petrological and chronological records signal the presence of multiple crustal fragments and independently operating subduction systems in the Iapetus domain (e.g. Roberts 2003; Brueckner & van Roermund 2004; Hacker & Gans 2005). Correlation of eclogite-bearing terranes and their integration into a consistent geodynamic model requires an improved understanding of events that pre- and postdate their eclogite-facies metamorphism.

This study aims to unravel the evolution of the Jæren nappe, where sporadic occurrences of eclogite (Majer et al. 1987) have recently been attributed to an early Caledonian HP event (Lu–Hf garnet age of 471 ± 1 Ma; Smit et al. 2010). This event is coeval with arc-type magmatism recorded in a structurally overlying accretionary complex (e.g. Dunning & Pedersen 1988), possibly signifying an upper- and lower-plate pair assembled during a previously unidentified subduction–eduction cycle. To clarify the geodynamic meaning of the Jæren rocks, we focused on the provenance and exhumation history of paragneisses surrounding the eclogites. Provenance was studied by performing U–Pb dating of detrital zircon using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS). This technique was chosen because its <35 μm spatial resolution allows the direct dating of single zones of zircon grains, and it allows for a high sample throughput, which is essential to extract significant age information from complex detrital-zircon age spectra in a reasonable time span (e.g. Andersen 2005). Using the Rb–Sr method, we dated phengite and biotite to place time constraints on the upper-crustal metamorphic history of the gneisses.

Geological setting

The Caledonides

The Scandinavian Caledonides are an orogenic belt that stretches along the entire length of peninsular Scandinavia (Fig. 1a). Most large-scale deformational and metamorphic features in this belt record processes that are related to the Scandian Orogeny (Gee 1975). This event involved (1) the deep subduction of parts of the Baltic Shield beneath Laurentia, causing (U)HP metamorphism of its primarily Proterozoic rocks (Tucker et al. 1990; Austrheim et al. 2003; Walsh et al. 2007), locally under diamond-stable P–T conditions (Dobrzhinetskaya et al. 1995; van Roermund et al. 2002; Carswell et al. 2003; Vrijmoed et al. 2008; Scambelluri et al. 2009) and (2) tectonometamorphic overprinting and long-distance transport of heterogeneous fragments from the Baltic and Laurentian margins. These fragments, which are exposed as thrust units (allochthons) that occur along the axis of the orogen (e.g. Brueckner & van Roermund 2004; Tucker et al. 2004; Hacker & Gans 2005), contain the geological record of the Iapetus Ocean. This ocean had opened during Late Proterozoic times, as evidenced by 650–580 Ma extensional dyke systems in Baltica-derived units of the thrust belt (Claesson & Roddick...
After reaching its maximum width of several thousand kilometres during the Late Cambrian, the basin contracted during the earliest Ordovician (e.g. Torsvik et al. 1992; 1996); this contraction was associated with the build-up of arc and accretionary complexes (e.g. Stephens & Gee 1985, 1989; Strachan 2000). Many of these complexes contain dismembered 440 Ma ophiolites (e.g. Dunning & Pedersen 1988; Pedersen et al. 1991; Oliver & Krogh 1995; Strachan 2000; Yoshinobu et al. 2002) that represent fragments of oceanic crust formed in short-lived back-arc or marginal-basin settings (e.g. Roberts 2003). The thrusting of the accretionary complexes and associated subduction-zone processes occurred both during and before Scandian time (Andersen & Andresen 1994; Andresen & Steltenpohl 1994; Dewey & Mange 1999; Soper et al. 1999; Yoshinobu et al. 2002; Roberts 2003; Hacker & Gans 2005). Subduction of oceanic lithosphere during Ordovician Iapetus-basin contraction formed the (U)HP crustal complexes in the Caledonides, such as the Seve nappe, the basement complex in Lofoten and the Tromsø nappe (Fig. 1a). These terranes record short-lived subduction–eduction cycles in at least three subduction systems operating independently close to the Baltic margin at c. 505 and 460–450 Ma and along the Laurentian margin at c. 450 Ma (e.g. Brueckner & van Roermund 2004; Roberts et al. 2007).

The Jæren area

The Jæren nappe is a small supracrustal unit in SW Norway (Fig. 1b), which hosts an eclogite occurrence that is somewhat enigmatic in the context of recent geodynamic models. Recently, Lu–Hf dating of garnet from sporadic eclogite outcrops yielded an age of 471 ± 1 Ma, which was interpreted to closely approximate the time when peak-metamorphic conditions of c. 2.6 GPa and 585–655 °C were reached (Smit et al. 2008, 2010). The data showed that HP metamorphism was protracted, lasting at least until c. 458 Ma. This result was supported by less precise Sm–Nd data, which additionally indicated younger (<430 Ma) age components. The provenance of these rocks is unknown and there are no age data available yet for the pre- and post-eclogite stages of the pressure–temperature–time–deformation (P–T–t–D) history. Preliminary U–Pb data from zircon in leucosome segregates indicate 1600 and 920 Ma age components (F. Corfu, pers. comm.). However, their geological meaning with respect to the pre-Caledonian history of the rocks is not known.

The eclogites occur in the SE part of the nappe (Fig. 1b) and are dispersed as small (<15 m) pods or as thin boudinaged layers in regionally foliated amphibolite-facies quartzite, metapelitic, calc-silicate rock, and mafic gneiss (e.g. Maijer et al. 1987; Smit et al. 2008, 2010). These lithological units dominate in the Jæren nappe, which additionally contains kilometre-scale, sheared granodiorite- and tonalite bodies, and marbles in the NW part (Birkeland 1981). In contrast to the mafic gneisses, which
contain omphacite and phengite inclusions in garnet and zoisite, no clear eclogite-facies mineral relics occur in the metasediments. Preliminary P–T estimates on garnet-bearing assemblages in the latter rocks are 1.1–1.4 GPa at c. 650–700 °C (M. A. Smit, unpubl. data), coinciding with early stage amphibolite-facies retrogression (Smit et al. 2008). The dominant lithological units of the Jæren nappe are similar to, and are correlated with, those in the nearby Hardangervidda area (Andresen & Færseth 1982). These record Proterozoic deformation structures and are considered to be Caledonian-reworked ('Caledonized') fragments of the domestic Baltic basement. This correlation appears to be supported by a single Proterozoic Rb–Sr whole-rock date for the NW Jæren nappe (995 ± 30 Ma; Roddick & Jorde 1981), but the geological significance of this result is ambiguous.

The Jæren nappe is overlain by the Karm Sund nappe (Fig. 1b), which is part of a rock suite containing a 490 Ma meta-ophiolite (Karmøy ophiolite) that was intruded by arc-type magmas between 490 and 470 Ma, and is discordantly overlain by sediments (Dunning & Pedersen 1988; Pedersen & Hertogen 1990; Pedersen & Dunning 1997). Faunal records and Archaean detrital zircon grains from the sediments were interpreted to suggest deposition in a peri-Laurentian environment (Pedersen et al. 1992), further emphasizing the exotic nature of the Karm Sund nappe. The Karm Sund and Jæren nappes are separated by an inferred thrust contact and are underlain by a relatively thin psammite (e.g. terrigenous turbidite, metagreywacke, and metasandstone; 08020 and 10017), which show a more or less pronounced compositional layering. Matrices exhibit a layering-parallel, spaced foliation that is defined by elongated (30-700 μm) elongated grains, and shows variable grain size (CSA that range from <10 μm to 100–600 μm, depending on the sample. The largest phengite crystals occur in sample 12001 (CSA up to 600 μm), whereas those in sample 08017 are smallest (CSA <100 μm). Crystals in the other samples are characterized by various maximum CSA values between 250 and 400 μm. Biotite is either greenish (08017, 09015 and 12001), brown (08007, 08008c and 09019), or both (12004). Biotite CSA vary between c. 10 μm and 100–650 μm. The largest biotite crystals occur in samples 08007, 09015 and 12004 (CSA up to 600 μm), whereas crystals in samples 08008c, 09019 and 12001 are particularly fine-grained (CSA <100 μm). Biotite occurs in textural contexts similar to those of phengite or as rims around it. The same textural relationship is observed for green biotite relative to brown biotite in sample 12004. Biotite in samples 09015 and 12004 is mildly affected by chloritization.

For U–Pb zircon analysis, three amphibolite-facies paragneiss samples were selected from the Garborg locality (08020, 10017 and 12001) and an additional sample was collected from the eclogite-facies Sola outcrops (09019; Fig. 1b). Details on samples 12001 and 09019 were given above. Sample 08020 is a plagioclase-bearing quartzite. The mineral consists of a bimodal distribution, comprising fine-grained (10–100 μm) and coarse-grained (300–500 μm) populations. The rock contains minor garnet, apatite, ilmenite, rutile, hematite, partially chloritized brown biotite, and zircon. Sample 10017 is a fine-grained (<250 μm), continuously foliated quartzite. The quartz matrix encloses partially chloritized biotite, garnet, staurolite, plagioclase and zircon. Ilmenite and hematite are common accessory phases. The sample includes thin (centimetre-scale) foliated layers that are dominated by quartz, amphibole, garnet and chlorite. Accessory phases in these layers largely comprise ilmenite and rutile. Transitions from these layers to the quartzitic segments of the sample are smooth and occur on a millimetre scale. Zircon in each of the selected samples occurs as round or elongated grains, and shows variable grain size (c. 30-700 μm) and colour (clear, pink, yellowish to brown).

**Analytical methods**

After crushing using a stainless-steel jaw crusher, the 2–6 kg samples were ground in a stainless-steel disc mill to further reduce grain size. For Rb–Sr analysis, the ground material was sieved into different grain size fractions. For U–Pb zircon studies, a heavy mineral fraction was pre-concentrated using a Wilfley gravity-separation table. The fraction was then separated by its magnetic content using a Frantz Inc. Isoodynamic® magnetic separator and further refined to a zircon-rich pre-concentrate by density separation in methylene iodide. Zircon grains were handpicked under a binocular microscope and separated into four grain size fractions (>400 μm, 300–400 μm, 200–300 μm and <200 μm). These were further subdivided on the basis of colour (clear, pink, and brown) and mounted in 1-inch epoxy discs. Cathodoluminescence (CL) imaging was performed using a JEOL 840 raster electron microscope, equipped with a K.E. Developments Ltd. Centaurus CL-detector system. A selection of zircon CL images is shown in Figure 2.

The U–Pb analyses were performed by LA-ICP-MS using a New Wave Research Inc. ArF Excimer (λ = 193 nm) laser system that was equipped with a small-volume teardrop-shaped ablation cell. This system was coupled to a magnetic sector-field single-collector ICP-MS system (Thermo Scientific Inc. Element2). Measurements and data reduction were carried out applying the approach of Kooijman (2010). Time-dependent fractionation was corrected using the intercept method (Sylvester...
whereas static fractionation, including mass bias, was corrected by applying external standard bracketing based on two measurements of the GJ-1 standard (Jackson et al. 2004) between sets of five unknowns. Corrections for common Pb content, which was estimated based on measured $^{204}\text{Pb}$, were applied using the two-stage evolution model for terrestrial Pb by Stacey & Kramers (1975). Spot diameters were 35, 25 or 12 μm, depending on grain characteristics. To monitor accuracy and external reproducibility, the 91500 zircon standard (Wiedenbeck et al. 1995) was measured as an unknown during the analytical sessions. The results for this standard indicate a 1.7% (2 SD) external reproducibility of $^{206}\text{Pb}/^{238}\text{U}$ and provided a concordia age of 1070 ± 3 Ma ($n = 39$), which agrees very well with the published $^{207}\text{Pb}/^{206}\text{Pb}$ age (1065 ± 1 Ma; Wiedenbeck et al. 1995). The U–Pb ages were calculated using $^{235}\text{U}$ and $^{238}\text{U}$ decay constants of Steiger & Jäger (1977). Concordia diagrams, constructed using IsoPlot (v. 3.72; Ludwig 2009), are shown in Figure 3 and age histograms are given in Figure 4. All uncertainties in the U–Pb data (i.e. on isotope ratios, ages and weighted means) are reported as 2 SD unless stated otherwise; cited ages are $^{206}\text{Pb}/^{238}\text{U}$ ages, except for ages older than 1800 Ma for which $^{207}\text{Pb}/^{206}\text{Pb}$ ages are given.

Micas, epidote, garnet, amphibole, and plagioclase were pre-concentrated from the 180–250 and 250–355 μm size fractions using a Frantz Inc. Isodynamic magnetic separator. The target phases were then handpicked under a binocular microscope. For electron probe micro-analysis (EPMA), c. 30 grains were isolated from the mica separates, mounted in epoxy, and subjected to major element analysis by wavelength-dispersive spectroscopy (WDS), using a JEOL Ltd. JXA-8900M Superprobe and applying a 15 nA electron beam at 20 kV. This was done to characterize mica composition; for example, to see whether the analysed crystals show zoning and to identify any mixed populations consisting of different mica generations. Phengite compositions are plotted in Figure 5.

Isotope measurements were performed by thermal ionization
Fig. 3. Concordia diagrams showing the results of U–Pb dating of detrital zircon for the age ranges 3500 Ma–present (a, c, e, g) and 1400–400 Ma (b, d, f, h) for samples 08020 (a, b), 10017 (c, d), 12001 (e, f), and 09019 (g, h). Ellipses indicate 2 SD analytical uncertainties on single spots. Listed 206Pb/238U ages represent concordant analyses (95–105% concordance).
mass spectrometry (TIMS) using a VG Instruments Ltd. Sector 54 system (Rb) and a Thermo Scientific Inc. Triton system (Sr). For sample preparation, mineral separates were washed repetitively using MilliQ H₂O and once using p.a.-grade ethanol. Brown mica separates were additionally ground in ethanol using an agate mortar and pestle and sieved using disposable 100 µm sieves to remove contaminants such as apatite inclusions. The separates were spiked using a mixed ⁸⁷Rb–⁸⁴Sr isotope tracer, digested in Savillex® screw-top PFA vials using HF:HNO₃ (5:1), dried down, re-dissolved in 6M HCl, and dried down again to ensure complete homogenization. Rb and Sr were isolated from the elemental matrix by ion exchange chemistry using quartz-glass columns containing Bio-Rad® 50W-X8 resin. Rubidium was loaded onto Ta filaments using H₂O, whereas Sr was loaded onto W filaments using 6M HCl and TaF₅. Total procedural blanks were c. 0.05 ng (Rb) and 0.1 ng (Sr). During the course of

Fig. 4. Histograms of concordant (95–105% concordance) detrital zircon in samples 08020 (a), 10017 (b), 12001 (c), and 09019 (d). For >1000 Ma zircon, ²⁰⁷Pb/²⁰⁶Pb ages are shown, whereas the ²⁰⁶Pb/²³⁸U ages are taken for younger zircon. Relative probability (Rel. prob.) is plotted as thick lines.

Fig. 5. Diagrams illustrating the Si contents of phengite cores and rims (a) and the Rb–Sr age (b) for each of the phengite-bearing samples. The compositional fields in (a) correspond to the age groups shown in (b). The paramagnetic (†) and less paramagnetic (*) fractions of sample 08008c are indicated in the legend, which applies to both diagrams.
this study, analyses of the NBS-987 Sr standard provided an $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.710189 ± 0.000026 (2 SD, n = 13). The Rb–Sr isotope data and the isochron ages that were calculated using the $^{87}\text{Rb}/^{86}\text{Sr}$ decay constant recommended by Steiger & Jäger (1977) are presented in Table 1. Isochron plots for the phengite and biotite analyses, which were constructed using IsoPlot (v. 3.72; Ludwig 2009), are presented in Figures 6 and 7, respectively. Uncertainties on $^{87}\text{Sr}/^{86}\text{Sr}$ are reported as 2 SE of single analyses and assumed to be 1% (2 SD) for $^{87}\text{Rb}/^{86}\text{Sr}$, whereas uncertainties on Rb–Sr isochron and weighted mean ages are given as 95% confidence limits.

Results

Zircon: textural characteristics and U–Pb dating

The CL characteristics of zircon indicate that grains consist of a large central domain and one or several rims (Fig. 2). In clear and pinkish grains, the central domains exhibit well-defined sector zoning or concentric oscillatory zoning made up of straight or irregular zones. Many of these domains have 2–3 concentric sub-domains that show truncation relationships. In brownish grains, which show only minor luminescence, most domains show relict oscillatory zoning and parts having patchy zoning or highly variable luminescence (e.g. Fig. 2d, grains 2.01 and 3.08). Rims show either homogeneous luminescence or patchy zoning. Luminescent zircon rims locally continue along fractures that invade internal zones. About 5% of the grains have large (>10 vol.%) rims that are characterized by faint oscillatory zoning. The U concentrations of patchy zircon in all samples are on average a factor of 2–4 higher than those of other zircon. The highest U contents were measured in zircon from samples 09019 and 10017 (2143 and 1989 ppm, respectively), which also show a relatively high abundance of zircon that shows patchy zoning.

Zircon analyses from quartzite sample 08020 are mostly concordant and dominated by a $^{206}\text{Pb}/^{238}\text{U}$ age population at c. 920 Ma (Figs 3b and 4a). This age represents central zircon domains and usually occurs in each of their sub-domains. Other concordant analyses include a single rim spot that yielded a $^{206}\text{Pb}/^{238}\text{U}$ age of 678 ± 16 Ma, and ages that are either 1400–1200 or c. 1050 Ma. Discordant analyses (c. 10 % of the total) are characterized by 1070–660 Ma $^{206}\text{Pb}/^{206}\text{Pb}$ ages and provided $^{206}\text{Pb}/^{238}\text{U}$ ages between 873 and 482 Ma. The discordant analyses correspond to fractured or patchy internal zones and luminescent rims, the latter providing exclusively ‘young’ (<500 Ma) $^{206}\text{Pb}/^{238}\text{U}$ ages.

Zircon from quartzite sample 10017 yielded (1) a main age population between 1360 and 910 Ma, showing an abundance peak at c. 1040 Ma, (2) an additional age range from 1850 to 1670 Ma, and (3) a minor Archaean population (2940–2620 Ma; Figs 3c, d and 4b). Relatively imprecise (±5–10%), discordant results of c. 600 and 460 Ma were acquired from an oscillatory-zoned grain and a luminescent rim, respectively. In contrast to Proterozoic ages, which occur in domains of variable geometry and zoning, Archaean ages were exclusively found in cloudy, oscillatory-zoned grains. Some analyses provided discordant results, yielding either Proterozoic or Archaean $^{207}\text{Pb}/^{206}\text{Pb}$ ages and indicating a lower intercept at <800 Ma. Most of these analyses correspond to brownish, weakly luminescent zircon showing cloudy zoning.

Metapelite sample 12001 provided a range of concordant results between 1860 and 960 Ma, showing abundance peaks at c. 1195 and 1070 Ma, and some older ages between 2770 and 2360 Ma (Figs 3c, f and 4c). Although different sub-domains within single grains usually provide an age from one of these populations, some grains gave both Archaean and 1700–1300 Ma ages. Two concordant c. 600 Ma ages were measured in luminescent oscillatory-zoned rims around c. 1100 Ma zircon. Although some discordant analyses indicate a Caledonian component, such ages were not found in zircon from this sample.

Zircon from metamorphosed sample 09019 (Sola locality) yielded mostly concordant results, which define a large age range roughly between 1800 and 930 Ma (Figs 3g, h and 4d). Abundance peaks are observed at c. 1650, 1180, 1070, and 1020 Ma. Five analyses yielded Archaean ages between 3130 and 2710 Ma. Two spots provided concordant c. 470 Ma results (±2%). Together with a single analysis in sample 10017, these are the only fully concordant Caledonian analyses (<5% discordant), defining a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 469.4 ± 5.9 Ma. Two spots yielded concordant c. 610 Ma results. These are identical to three concordant analyses in samples 10017 and 12001, and, together, they provide a weighted mean $^{206}\text{Pb}/^{238}\text{U}$ age of 607 ± 19. All Concordant >900 Ma ages were found in central domains or in luminescent rims whereas c. 470 Ma ages and one of the c. 615 Ma ages were measured in patchy, weakly luminescent central domains (e.g. Fig. 2d, grain 2.01). Analyses of patchy rims around Archaean domains or in luminescent rims around Proterozoic zircon domains yielded discordant ages. No grains having both Archaean and 1620–1030 Ma age components were found. No systematic relationship between age and the colour of grains, grain size, or U concentration was observed.

Mica composition

The average Si content of phengite varies significantly between samples and shows some within-sample variability (±1–6%, 2 SD; Fig. 5a). Lowest Si contents occur in phengite from sample 08017 (3.08 ± 0.04 p.f.u.). The Si contents of phengite in samples 08008c, 09015 and 12001 cluster at 3.2–3.3 p.f.u. Sample 08008c provided a bimodal distribution during magnetic separation, which is also recorded in different Si contents (Fig. 5a). Only phengite in samples 08007 and 09019 shows consistently high Si contents (up to 3.50 p.f.u.), averaging c. 3.33 (±5%) for the cores and 3.27–3.30 (±4–5%) for the rims of crystals. These are the only samples that show a common and consistent compositional variation between phengite cores and rims.

The average Mg-number of biotite varies slightly between most samples (0.51–0.61), but shows substantial intra-sample variability (3–9%), which occurs between, rather than within crystals. The average biotite Mg-number correlates with whole-rock Mg-number for all samples (M. A. Smit, unpubl. data). The K contents of biotite vary slightly between crystals in all samples (0.8–1.0 p.f.u.), translating to a slight interlayer non-stoichiometry. Brown biotite of sample 12004 is more Fe- and Ti-enriched (0.8–1.0 p.f.u.), translating to a slight interlayer non-stoichiometry.

Mica: Rb–Sr dating

Internal Rb–Sr isochrons constructed using phengite (2–3 grain size fractions) and epidote, plagioclase or garnet yield ages that overlap within analytical uncertainty for samples 08008c (422.9 ± 2.4 Ma), 09015 (425.7 ± 3.0 Ma) and 12001 (426.2 ± 3.8 Ma). The weighted mean age is 424.4 ± 4.6 Ma. No
Table 1. Rb–Sr isotope data and ages

| Sample, mineral* | Analysis | Size fraction (μm) | Rb (ppm) | Sr (ppm) | $^{87}$Rb/$^{86}$Sr | $^{87}$Sr/$^{86}$Sr | 2 SE† | Rb–Sr age (Ma)‡ |
|-----------------|----------|------------------|---------|---------|----------------|----------------|------|-----------------|
| 08007           | Phengite Phe.1 | 250–355  | 235   | 77.4   | 8.839           | 0.772203       |      | 20              |
| Phengite Phe.2  | 180–250   | 242    | 86.0  | 8.178  | 0.767283        |                |      | 37              |
| Plagioclase Pl.1| 250–355   | 1.41   | 12.7  | 0.3211 | 0.718392        |                |      | 72              |
| Brown biotite Bt.1 | 250–355 | 359    | 8.24  | 136.2  | 1.523730        |                |      | 443.5 ± 4.5     |
| Brown biotite Bt.2 | 180–250 | 387    | 12.3  | 96.19  | 1.284210        |                |      | 29              |
| Epidote Ep.1    | 250–355   | 34.7   | 851   | 0.1182 | 0.719605        |                |      | 46              |
| 08008c          | Phengite Phe.1§ | 250–355 | 249   | 18.1   | 40.72  | 0.97737        | 19       | 37              |
| Phengite Phe.2¶ | 250–355   | 247    | 17.8  | 41.23  | 0.961629        |                |      | 46              |
| Brown biotite Bt.1 | 250–355 | 527    | 4.47  | 403.0  | 2.55891        |                |      | 55              |
| Brown biotite Bt.2 | 250–355 | 526    | 6.03  | 284.7  | 2.032230        |                |      | 54              |
| Epidote Ep.1    | 250–355   | 2.35   | 1330  | 0.005123 | 0.731645 |                | 32      | 319.8 ± 2.2     |
| 08017           | Phengite Phe.1 | 250–355 | 88.9  | 160    | 1.615           | 0.727914       |      | 23              |
| Phengite Phe.2  | 250–355   | 103    | 190   | 1.562  | 0.727663        |                |      | 26              |
| Phengite Phe.3  | 180–250   | 101    | 188   | 1.566  | 0.727643        |                |      | 37              |
| Garnet Grt.1    | 250–355   | 1.62   | 3.77  | 1.249  | 0.735131        |                |      | 425.7 ± 3.0     |
| 09015           | Phengite Phe.1 | 250–355 | 300   | 108    | 8.074           | 0.781049       |      | 48              |
| Phengite Phe.2  | 250–355   | 302    | 110   | 7.990  | 0.780661        |                |      | 418.3 ± 4.1     |
| Garnet Grt.1    | 250–355   | 1.62   | 3.77  | 1.249  | 0.739623        |                |      | 411.7 ± 2.3     |
| 09019           | Phengite Phe.1 | 250–355 | 301   | 72.0   | 12.22           | 0.803814       |      | 40.7 ± 2.8      |
| Phengite Phe.2  | 180–250   | 292    | 84    | 10.12  | 0.789951        |                |      | 443.6 ± 3.6     |
| Garnet Grt.1    | 250–355   | 4.32   | 8.94  | 1.403  | 0.735131        |                |      | 46.4 ± 2.9      |
| 12001           | Phengite Phe.1 | 250–355 | 132   | 171    | 2.240           | 0.734978       |      | 18              |
| Phengite Phe.2  | 180–250   | 102    | 243   | 1.216  | 0.728786        |                |      | 27              |
| Garnet Grt.1    | 250–355   | 1.08   | 46.3  | 0.1302 | 0.722186        |                |      | 13              |
| Green biotite Bt.1 | 250–355 | 574    | 5.85  | 341.1  | 2.750650        |                |      | 426.2 ± 3.8     |
| Brown biotite Bt.2 | 250–355 | 565    | 6.39  | 299.3  | 2.43756         |                |      | 414.5 ± 2.9     |
| Garnet Grt.1    | 250–355   | 1.32   | 1.51  | 2.538  | 0.735385        |                |      | 25              |
| 12004           | Brown biotite Bt.1 | 250–355 | 272   | 2.50   | 385.3           | 2.98262        |      | 413.4 ± 2.9     |
| Brown biotite Bt.2 | 180–250 | 279    | 2.78  | 349.5  | 2.77728         |                |      | 60              |
| Garnet Grt.1    | 250–355   | 1.05   | 3.63  | 0.8408 | 0.722149        |                |      | 21              |
| Garnet Grt.1    | 250–355   | 1.52   | 43.0  | 0.1025 | 0.718283        |                |      | 14              |

* Analyses in italics were not included in isochron calculation.
† Absolute 2 SE internal error on $^{87}$Sr/$^{86}$Sr in the two least significant digits.
‡ Calculated assuming an external 2 SD uncertainty of 1% for $^{87}$Rb/$^{86}$Sr, and 0.005% for $^{87}$Sr/$^{86}$Sr, unless this is exceeded by the internal error, which was then used instead. Age uncertainty indicates 95% confidence limits.
§ Less paramagnetic fraction.
¶ Paramagnetic fraction.
statistically valid age could be resolved using all phengite analyses of sample 08007. Two-point isochrons for this sample, using only the data for coarse- and fine-grained fractions, respectively, provided ages of 443.5 ± 4.5 Ma (Fig. 6a) and 436.9 ± 4.5 Ma. The first result is identical to the Rb–Sr age of 09019 phengite (443.6 ± 3.6 Ma). Together, these samples yielded a weighted mean age of 443.6 ± 2.8 Ma. Phengite from sample 08017 provided a significantly younger age (407.9 ± 3.0 Ma). In all cases, the $^{87}$Rb/$^{86}$Sr value for the fine grain size fraction (180–250 μm) was equal to or lower than that of the 250–355 μm fraction. The three phengite age groups of c. 444, 424 and 408 Ma correspond to the high-, middle- and low-Si phengite clusters, respectively (Fig. 5).

Biotite Rb–Sr ages (Fig. 7) are largely identical within error.
Fig. 7. Internal Rb–Sr isochrons for biotite ages. Uncertainties on \(^{87}\text{Sr} / {^{86}\text{Sr}}\) are absolute 2 SD uncertainty in the least significant digits. The open circles in (c) and (h) represent data points that were excluded from the age calculations. The number of analyses used to calculate this age (n) is given with the total number of analyses in parentheses.
for samples 08007 (413.7 ± 2.9 Ma), 09015 (411.7 ± 2.3), 12001 (414.5 ± 2.9 Ma) and 12004 (brown biotite; 413.4 ± 2.9 Ma), and yield a weighted mean age of 413.1 ± 1.3 Ma. A two-point isochron for sample 08017 gave an age that is within error of this mean (418.3 ± 4.1 Ma). The above ages combined provide a weighted mean age of 413.6 ± 2.6 Ma. The analysis of the fine-grained biotite fraction (Bt.2) of sample 08017 plots far below the two-point isochron and would provide an apparent age of 168.8 ± 1.7 Ma. For biotite from sample 09019, an imprecise age of 406 ± 23 Ma was obtained (MSWD = 21). Excluding the analysis of Bt.1 from the age calculations yielded an age of 400.7 ± 2.8 Ma (MSWD = 0.15). The Rb–Sr age of green biotite from samples 12004 is significantly younger (396.6 ± 2.7 Ma). Biotite from sample 08008c provided an even younger age (319.8 ± 2.2 Ma). No apparent relation between apparent age and composition could be resolved for biotite.

Discussion

Zircon provenance

To understand the role of the Jæren nappe rocks in relation to the evolution of Iapetus and the pre-Scandinian Caledonides, detrital zircon provenance needed to be constrained. The youngest zircon that is of detrital origin is not common (Fig. 4b–d), but occurred consistently in three out of four samples (09019, 10017 and 12001), yielding a weighted mean age 206Pb/238U age of 607 ± 19 Ma. This age provides a maximum age estimate for the deposition of the sedimentary protolith in the Iapetus basin. Two potential source areas for such protolith must be considered: the Baltic- and the east Laurentian Shield. Because the Jæren nappe is in close proximity to the South Norwegian Baltic Shield and presumably correlates to ‘domestic’ nappes to the north (e.g. Andresen & Faerseth 1982), a Baltic provenance could reasonably be expected. However, reconstructions by Torsvik et al. (1992a, 1996) and Cocks & Torsvik (2002) indicate that during the late Caledonian the South Norwegian Baltic Shield was juxtaposed against the oceanic foreland of the eastern Laurentian basement. Similarly to the fragmented ophiolite complexes that occur in thrust nappes adjoining the study area (e.g. Pedersen et al. 1992), the Jæren metasediments may have been derived from this ‘exotic’ foreland.

Constraining provenance to either Baltic or Laurentia is not trivial because the geological histories of these proto-continents are largely similar. Dykes having c. 610 Ma ages are known from the margins of both the South Norwegian Baltic Shield and the Laurentian basement (Bingen et al. 1998). Zircon from most of our samples indicates ages between c. 1900 and 900 Ma, prominent abundance peaks at 1070–1020 and c. 920 Ma, and a minor peak at c. 1750 Ma. Most of the younger ages (<1650 Ma) are typical for the South Norwegian Baltic Shield (Bingen et al. 2005), whereas the older components may be attributed to the more distal Svecofennian orogen and the adjoining Transscandinavian Igneous Belt (e.g. Ahäll & Connelly 2008; Bingen & Solli 2009). However, these ages also represent the main age imprints in Laurentian terranes of North America (e.g. Schärer et al. 1986; Tucker & Gower 1994; Gower 1996; Rivers 1997), East Greenland (e.g. Taylor & Kalsbeek 1990; Kalsbeek et al. 1993, 2000; Watt & Thrane 2001), and the British Isles (e.g. Kinny & Friend 1997; Friend & Kinny 2001).

The situation is somewhat different in the case of Archaean ages. Three out of the four analysed samples yielded Archaean zircon grains, 19 in total (four slightly discordant). These represent a small, but significant detrital age component in the studied rocks. Archaean magmatic suites were extremely distal in Baltic (>1500 km away; Karelia and Murmansk cratons; e.g. Gail & Gorbatchev 1987). These are unlikely to have been the source for the Archaean grains, which are large (up to 0.4 mm), euhedral and U-rich (up to 400 ppm), and are therefore not expected to survive such long-distance transport (e.g. Goldstein et al. 1997). The suggested source candidate, the South Norwegian Baltic Shield, was subjected to extensive and thorough U–Pb studies (e.g. compilation by Bingen et al. 2005), but has not yielded concordant Archaean zircon that crystallized in situ. All Archaean zircon in this complex occur as detrital grains in some >1000 Ma sediments, which form a small part of the South Norwegian Baltic Shield. A single concordant Archaean analysis has been reported from detrital zircon in conglomerate pebbles in the Faurefell formation (de Haas et al. 1999). Several such analyses were reported by Bingen et al. (2001). All of these grains were from sediments that had extremely high abundance of 1900–1850 Ma zircon, which were not encountered in our study. The scarcity of Archaean zircon in the South Norwegian Baltic Shield and the presence of age components in the South Norwegian Baltic Shield that were not observed here do not preclude sediment sources in the South Norwegian Baltic Shield, but suggest that this part of the Baltic Shield is highly unlikely to have produced the observed Archaean detrital component.

The alternative source candidate, the east Laurentian basement of (south-)east Greenland and Scotland, contains abundant Archaean rocks (e.g. Friend et al. 1988; Whitehouse et al. 1997) that were reworked during Proterozoic magmatism and deformation (e.g. Heaman & Tarney 1989; Kalsbeek et al. 1993; Corfu et al. 1994; Kinny & Friend 1997; Friend & Kinny 2001; Thrane 2002; Nutman et al. 2008) and were subjected to Grenvillian overprinting (Strachan et al. 1995; Friend et al. 1997; Kalsbeek et al. 2000; Watt et al. 2000; Brewert et al. 2003). Detritus that was shed from this basement, occurring as sediments in NE Greenland, Ireland, Scotland and Norway, typically contains Archaean zircon grains and provides an additional Proterozoic age record that is identical to that found in our study (e.g. Strachan et al. 1995; Rainbird et al. 2001; Cawood et al. 2003; Friend et al. 2003; Barnes et al. 2007; Chew et al. 2008; Waldron et al. 2008). It is important to note that, as in the case of our samples (Figs 3 and 4), these Laurentian detrital records are continuous in the period of 1130–1060 Ma and often show abundance peaks in this range. Detailed age compilation has shown that this period marks a short amagmatic stage in the history of the South Norwegian Baltic Shield, which should therefore be absent in South Norwegian Baltic Shield detritus (‘Arendal stage’; Bingen & Solli 2009). Instead, zircon that was derived from the South Norwegian Baltic Shield should yield abundant ages between 1520 and 1480 Ma, which marked a period of widespread plutonism that formed voluminous igneous complexes in the western part of the South Norwegian Baltic Shield (e.g. Idefjorden terrane; Bingen et al. 2005). These ages are not common among Laurentian detrital records and are absent in our age record (Figs 3 and 4). The fact that the detrital age record established in this study (1) strongly resembles that of sediments shed from the Laurentian margin and (2) differs from the detrital age record of typical South Norwegian Baltic Shield detritus strongly suggests a Laurentian affinity of the Jæren metasediments.

Eclogite metamorphism and Pb loss

For three out of four samples, including 09019 from the northern Jæren nappe, U–Pb analysis revealed Caledonian age compo-
ments (Fig. 3). Some of these analyses were performed on rims around weakly luminescent cores, which show few internal structures and are separated from older and more U-rich cores by a sharp interface that crosscuts the CL zoning of the core (e.g. Fig. 2a and b). The textural and compositional characteristics suggest that these rims formed by fluid- or melt-mediated replacement of old, metamict or otherwise chemically unstable zircon (e.g. Geisler et al. 2003, 2007). As expected for such newly precipitated zircon, the U–Pb ages are concordant in several cases, reflecting fully re-equilibrated U–Pb systematics. Some analyses on such zircon rims provided discordant results (206Pb/238U age <550 Ma). Although no significant shifts in 206Pb intensity or 207Pb/206Pb were observed during these measurements, we cannot rule out that discordance resulted from age mixing owing to beam overlap between different age zones.

Caledonian age components were also provided by heterogeneous, patchy zircon (e.g. Fig. 2d, grain 2.01). These zircon grains additionally provided (1) >1000 Ma concordant analyses and (2) discordant data that, when using old concordant analyses from the same grain as the upper intercept, indicate Caledonian ages. The discordant ages are interpreted to have resulted from partial Pb loss, whereas the concordant c. 470 Ma ages presumably reflect total Pb loss during eclogite-facies overprinting. It is unlikely that this occurred by volume diffusion of Pb through a pristine zircon lattice because this process is characterized by slow kinetics (Lee et al. 1997; Cherniak & Watson 2001) and usually does not significantly affect U–Pb systematics of zircon during metamorphism (e.g. Guillon & Krogh 1973; Harrison et al. 1987; Chen & Williams 1990). The inferred Pb loss is presumably related to high U contents, which are at least a factor of two higher than those of most concordant zircon. Structural recovery (annealing) and partial recrystallization of their metamict patches during prograde metamorphism to >600 °C would presumably have led to higher Pb mobility and partial or complete Pb loss from zircon to the matrix (Mezger & Kroghst 1997).

Three Caledonian zircon analyses from sample 09019 (grains 2.01 and 2.09; e.g. Fig. 2d) and from sample 10017 (grain 3.58; Fig. 2b) were fully concordant and yielded a weighted mean 206Pb/238U age of 469 ± 6 Ma, which provides the best age estimate for the event that caused zircon recrystallization and Pb loss. This age is identical to the age of HP metamorphism established by Lu–Hf dating of eclogite garnet (471 ± 1 Ma, 2 SD; Smit et al. 2010), indicating that zircon crystallization in the paragneisses and garnet growth in the eclogites occurred at the same time. This is in accordance with the interpretation that the gneiss–eclogite association underwent subduction and eduction as a coherent unit.

The Middle Ordovician age of eclogite-facies metamorphism is consistent with the inferred Laurentian provenance. This epoch dates a brief time span (<10 Ma) during which the Laurentia–Iapetus plate margin was affected by arc magmatism, obduction of ophiolites, arc–continent collision, and crustal subduction (e.g. van Bremen & Bluck 1981; Hutton et al. 1985; Dewey & Mange 1999; Soper et al. 1999; Oliver 2001; Roberts et al. 2007; van Staal et al. 2007). Records of this ‘Taconian’ tectono-metamorphism are prominent in North America (e.g. Tucker & Robinson 1990; Karabinos et al. 1998; Robinson et al. 1998) and are well documented from the British Isles (Friedrich et al. 1999; Kinny et al. 1999; Friend et al. 2000; Oliver et al. 2000; Rogers et al. 2001; Chew et al. 2003; Flowerdew et al. 2005; Chew et al. 2008; Sawaki et al. 2010). In Scandinavia, these processes were identified only in the Laurentian metamorphic rocks in the North Norwegian mainland (Fig. 1a; e.g. Yoshinobu et al. 2002; Corifu et al. 2003; Steltenpohl et al. 2003; Barnes et al. 2007; Roberts et al. 2007) and on Spitsbergen, Svalbard (Labrouesse et al. 2008). Our study suggests the occurrence of Taconian-overprinted Laurentian crust in the southern Scandinavian Caledonides. The fact that the Jæren HP nappe structurally directly underlies an Early to Middle Ordovician accretionary complex (Karmsund nappe) of the same provenance leads us to suggest that they form an upper- and lower-plate pair that was assembled in a peri-Laurentian environment.

Exhumation timing and rates

The Rb–Sr data provide important constraints on the post-
eclogite part of the P–T–t–D history, as they may date (1) cooling through the closure temperature for Sr volume diffusion (Tc; Dodson 1973), or (2) the growth or recrystallization of micas at conditions below Tc. Meaningful application of the closure temperature concept requires the presence of a Sr sink and a medium to effectively remove Sr from mineral grain boundaries. In cases where mica crystals are free to exchange Sr with other Sr-bearing phases, it is assumed that the condition Tc,Bt < Tc,Phe is fulfilled for mica crystals of equal size (Scaillet 1998). However, the widely used absolute values of Tc,Phe (e. g. 500 °C; Jager 1979) and Tc,Bt (300–400 °C; e.g. von Blankenburg et al. 1989) must be considered to be subject to large uncertainty (e.g. Villa 1998). Factors such as the nature and modal abundance of Sr-bearing minerals in a rock strongly influence the effective Tc values (Jenkins et al. 1995; Jenkins 1997). Studies on a variety of metamorphic rocks have illustrated the importance of fluid ingress and rock deformation on Rb–Sr systems (e.g. George & Bartlett 1996; Kühn et al. 2000; Cliff & Meffan-Main 2003; Reddy et al. 2003; Gladyn et al. 2008a–c). These studies indicate (1) that the effective Tc values of micas and associated low-Rb/Sr phases often exceed the theoretical Tc, and thus (2) that Rb–Sr results often date periods of mica growth and recrystallization rather than static cooling.

Dodson (1973) showed that, in the case of slow cooling, small mica grains should provide consistently lower apparent ages than large grains. This effect is not reflected in our dataset, in which (1) the apparent ages of fine-grained mica fractions overlap within analytical uncertainty with those of coarse mineral fractions in most cases, and (2) no apparent relationship between the grain size of mica in the sample (CSA) and apparent Rb–Sr age was observed. These aspects are in accord with the following suggestions: (1) the apparent Rb–Sr ages record open-system behaviour, but cooling through the respective closure temperatures occurred too rapidly to allow the development of a resolvable dependence of apparent age on grain size; (2) the ages represent the formation or recrystallization of micas close to or below their effective Tc. The observation that the two oldest phengite Rb–Sr dates (averaging at 443.6 ± 2.8 Ma) were acquired for the two samples that contain abundant high-Si (>3.30 pfu) phengite, whereas the youngest phengite date was obtained from the sample that shows by far the lowest Si contents (Fig. 5a), suggests some correlation between apparent age and phengite compositions. Because all analysed samples are comparable in terms of bulk composition (M. A. Smit, unpubl. data) and contain abundant quartz and other silicates acting as Si-buffering assemblages (cf. Bröcker et al. 2004), the Si contents of phengite are likely to reflect P–T conditions during crystallization (Massonne & Schreyer 1987) rather than a bulk-rock compositional control. The relation between age and primary compositional features suggests that the Rb–Sr system
in phengite was largely undisturbed and thus provides ages of phengite (re-)crystallization (interpretation 2).

Besides yielding the highest Si values, phengite in samples 08007 and 09019 showed high-Si cores and low-Si rims (Fig. 5a). For 08007 and to a lesser extent for 09019, the fine-grained fraction plotted below a two-point isochron for the coarse fraction only. It is plausible that the rims formed by recrystallization and that the effects of this process are more pronounced for the smaller crystals owing to their lower flow strength and volume-to-surface ratio. We suspect that the high Si and celadonite content of the cores reflects relict eclogite-facies mica and that the rims record decomposition reactions under amphibolite-facies conditions (<1.4 GPa, c. 650–700 °C), consuming garnet and the celadonite component in phengite to form biotite and presumably staurolite (e.g. Keller et al. 2005). Linking the coinciding apparent Rb–Sr ages for samples 08007 and 09019 to any such petrological process is not possible. The data merely indicate that the degree of isotope re-equilibration, whether due to volume diffusion or to partial recrystallization, was similar for both cases and that the c. 444 Ma age provides a minimum estimate of the core age.

Phengite in samples 08008c, 09015 and 12001 yielded consistent results (424.4 ± 4.6 Ma), shows no significant core-to-rim zoning, and is compositionally similar to the rims in samples 08007 and 09019. This would be consistent with the interpretation that c. 424 Ma mica represents the fully recrystallized equivalent of the high-Si phengite relics that occur in samples 08007 and 09019. The average apparent age of 424.4 ± 4.6 Ma is likely to provide an accurate approximation of the age of this amphibolite-facies recrystallization process.

The plagioclase that was used to construct the isochron for 08017, which yielded a relatively young age (407.9 ± 3.0 Ma), was carefully screened but showed no signs of alteration. This suggests that the young age result is not due to excess 87Sr that could have been introduced by invading fluids. The result could be related to the small phengite crystal size in this rock (CSA < 100 μm), which may have caused grains to be more reactive or more susceptible to recrystallization as a result of low flow strength. Following constitutional flow laws that predict a relationship between physical conditions and recrystallized grain size (e.g. Kohlenstedt et al. 1995), the phengite of sample 08017 may record dynamic grain size reduction that was restricted to this sample and occurred at lower differential stress or temperature (or both) than the c. 425 Ma overprinting.

Biotite is expected to represent a major rock-forming mineral in former HP metapelites after decomposition into the amphibolite facies, which is consistent with the observed phengite zoning in samples 08007 and 09019. Nevertheless, the main biotite Rb–Sr age cluster (413.6 ± 2.6 Ma) does not overlap with the c. 424 Ma apparent ages of phengite in the same samples, which are interpreted to represent this part of the P–T–t path. Biotite that yielded the c. 414 Ma ages was of variable composition and occurred in different crystal sizes in the rocks. It is not clear whether the age cluster provides an age estimate for the rapid cooling of the rocks through a given Tc, or whether it represents the growth of the mineral through a retrograde reaction mechanism that could neither be identified nor linked to specific temperature conditions. In either case, biotite may have never been in isotope equilibrium with garnet, which was used to construct several isochrons (Table 1; Fig. 7). Nevertheless, the coincidence of Rb–Sr ages among biotite–garnet and biotite–plagioclase or biotite–epidote pairs would suggest that any potential disequilibrium had no significant influence on the biotite Rb–Sr isochrons and ages. The apparent biotite ages of samples 09019 and 12004 (400–395 Ma) presumably record variable degrees of isotope re-equilibration in response to localized, (tectono-)metamorphic overprinting at temperatures below Tc, during latest Scandian times. This may also be reflected by the young Rb–Sr date for 08017 phengite (407.9 ± 3.0 Ma), the slight age discrepancy being caused by the presence of relict ≥444 or 425 Ma phengite.

**Subduction in the Caledonides**

As also suggested for several ophiolite complexes in southwestern Norway (e.g. the Karmoy Ophiolite of the overlying Karmund nappe; Pedersen et al. 1992; Roberts 2003; Roberts et al. 2007), the Jæren nappe was removed from its presumed Laurentian context and stranded onto the South Norwegian Baltic Shield after the Iapetus closure. The youngest known eclogite Lu–Hf age (c. 458 Ma; Smit et al. 2010) and the inferred mid-crustal emplacement age based on the new Rb–Sr data (≥424 Ma) define the maximum time span during which exhumation from eclogite-facies to upper amphibolite-facies conditions (AP ≈ 1.4 Ga) was accommodated. This implies a minimum average exhumation rate of c. 1.3 mm a⁻¹. Similar or shorter exhumation time spans are deduced for all Ordovician-UHP complexes in the Caledonides, considering available 40Ar–39Ar and U–Pb data (e.g. Dallmeyer & Gee 1986; Williams & Claesson 1987; Dallmeyer et al. 1989, 1991; Krogh et al. 1990; Essex et al. 1997; Gromet & Gee 1998; Labrousse et al. 2008). They demonstrate that the different UHP terranes not only subducted at different times, but also underwent diachronous exhumation during separate short-lived subduction–eduction cycles. The data indicate that at least three separate Ordovician subduction systems independently consumed lithosphere on either side of the contracting Iapetus domain (Brückner & van Roermund 2004).

**Scandian terminal collision and aftermath**

The recrystallization of phengite at 424.5 ± 4.6 Ma was contemporaneous with the onset of eastward thrusting and metamorphic overprinting of heterogeneous nappes along the western margin of the South Norwegian Baltic Shield (c. 425 Ma; e.g. Hacker & Gans 2005; Walsh et al. 2007; Hacker et al. 2009). This consistency indicates that, by early Scandian time, the Jæren nappe had become part of the orogenetic nappe pile that was subsequently unroofed along the western edge of the South Norwegian Baltic Shield. The c. 414 Ma biotite Rb–Sr ages document continuing cooling and, potentially, associated retrogression during upward transfer, which was synchronous with development of top-to-the-(S)E fabrics in basement–cover contacts (Fossen & Dunlap 1998) and amphibolite-facies rehydration and retrogression of high-grade nappes in northerly regions (414.2 ± 2.8 Ma; Gldnyn et al. 2008b). Following c. 408 Ma, the Scandian orogenic front transgressed into an extensional regime, undergoing two-stage extension, initially involving continental eulaxation and backsliding, and subsequently crustal collapse (e.g. Fossen 1992, 2000; Andersen 1998). This occurred under the influence of the exhuming Western Gneiss Complex, the eastern parts of which had reached upper-crustal levels by c. 400 Ma (e.g. Walsh et al. 2007; Kylander-Clark et al. 2008; Hacker et al. 2009). During the collapse stage, extensional (top-to-the-west) fabric formation under greenshist- to amphibolite-facies conditions became widespread in the overlying crustal complexes (e.g. Fossen & Dallmeyer 1998; Fossen & Dunlap 1998).
400 Ma recrystallization of phengite and biotite in three of the studied samples is ascribed to this stage.

Post-orogenic extension of the Caledonides occurred from early Devonian to Mesozoic time (e.g. Torsvik et al. 1992b; Eide et al. 1997, 1999; Andersen et al. 1999). Rifting and associated heat flow in the Skagerrak, the North Sea, and the Oslo region occurred at 300 Ma and onwards (Sundvoll et al. 1990; Ferseth 1996; Neumann et al. 2004; Corfu & Dahlgren 2008). The time period between these rifting processes and latest Scandinavian orogenic collapse was bridged by a period of relative tectonic and thermal stability (e.g. Dunlap & Fossen 1998). As such, we have no direct geological context to explain the occurrence of the single Carboniferous Rb–Sr age acquired from statically grown brown biotite in sample 08008c. The significance of this age remains enigmatic.

Conclusions

Detrital zircon in paragneisses from the eclogite-bearing Jæren nappe, SW Norway, underwent Pb loss and recrystallization at 469 ± 6 Ma. This age is identical to the Lu–Hf garnet age of the eclogites, which supports a model suggesting that the gneisses and eclogites subducted as a coherent unit. Detrital zircon ages range from 3130 to 610 Ma. The youngest time constraints of this age spectrum suggest deposition of the sedimentary protolith after exhumation that occurred at an average rate of c. 1.3 mm a⁻¹. A generation of retrograde biotite that occurs in most samples yielded an age of 413.6 ± 2.6 Ma, whereas c. 400 Ma biotite ages were acquired from two samples. Taken together, the Scandinavian Rb–Sr record documents the exhumation of the exotic Jæren nappe along the western edge of Baltica during the complex backsliding and collapse of the Scandinavian orogenic front.

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