

Linking deformation, migmatite formation and zircon U–Pb geochronology in polymetamorphic orthogneisses, Sveconorwegian Province, Sweden

C. MÖLLER,¹ J. ANDERSSON,² I. LUNDQVIST³ AND F. HELLSTRÖM²

¹Geological Survey of Sweden, Kiliansgatan 10, SE-22350 Lund, Sweden (charlotte.moller@sgu.se)

²Geological Survey of Sweden, Box 670, SE-751 28 Uppsala, Sweden

³Geological Survey of Sweden, Geovetarcentrum, Guldhedsgatan 5A, SE-413 20 Göteborg, Sweden

ABSTRACT Absolute ages of migmatization in the polymetamorphic, parautochthonous basement of the Sveconorwegian Province, Sweden, have been determined using U–Pb ion probe analysis of zircon domains that formed in leucosome of migmatitic orthogneisses. Migmatite zircon was formed by recrystallization whereas dissolution–reprecipitation and neocrystallization were subordinate. The recrystallized migmatite zircon was identified by comparison of zircon in mesosomes and leucosomes. It is backscatter electron-bright, U-rich (800–4400 ppm) with low Th/U-ratios (generally 0.01–0.1), unzoned or ‘oscillatory ghost zoned’, and occurs as up to 100 µm-thick rims with transitional contacts to cores of protolith zircon. Protolith ages of 1686 ± 12 and 1668 ± 11 Ma were obtained from moderately resorbed, igneous zircon crystals (generally Th/U = 0.5–1.5, U < 300 ppm) in mesosomes; protolith zircon is also present as resorbed cores in the leucosomes. Linkage of folding, synchronous migmatization and formation of recrystallized zircon rims allowed direct dating of south-vergent folding at 976 ± 7 Ma. At a second locality, similar recrystallized zircon rims in leucosome date pre-Sveconorwegian migmatization at 1425 ± 7 Ma; an upper age bracket of 1394 ± 12 Ma for two overprinting phases of deformation (upright folding along gently SSW-plunging axes and stretching in ESE) was set by zircon in a folded metagranitic dyke. Lower age brackets for these events were set at 952 ± 7 and 946 ± 8 Ma by zircon in two crosscutting and undeformed granite–pegmatite dykes. Together with previously published data the present results demonstrate: (i) Tectonometamorphic reworking during the Hallandian orogenesis at 1.44–1.42 Ga, resulting in migmatization and formation of a coarse gneissic layering. (ii) Sveconorwegian continent–continent collision at 0.98–0.96 Ga, involving (a) emplacement of an eclogite unit, (b) regional high-pressure granulite facies metamorphism, (c) southvergent folding, subhorizontal, east–west stretching and migmatization, all of which caused overprint or transposition of older Mesoproterozoic and Sveconorwegian structures. The Sveconorwegian migmatization and folding took place during or shortly after the emplacement of Sveconorwegian eclogite and is interpreted as a result of north–south shortening, synchronous with east–west extension and unroofing during late stages of the continent–continent collision.

Key words: dating deformation; dating migmatization; Mesoproterozoic orogenesis; Sveconorwegian orogenesis; zircon geochronology.

INTRODUCTION

High-grade metamorphic terranes are commonly characterized by a multiphase metamorphic and tectonic history. Metamorphic and deformation phases are not homogeneously distributed throughout these terranes but generally partitioned into specific lithologies, specific areas or specific shear zones. Successful dating of individual phases of deformation or metamorphism requires not only appropriate samples from localities where these phases are well expressed but also reliable linking between the geochronological data from specific minerals and the geological, petrological and structural data. U–Pb dating of zircon is com-

monly the method of choice for high-grade terranes, but linking individual zircon generations to large-scale processes may prove difficult.

In this work, we address the tectonometamorphic evolution of the Eastern Segment of the Sveconorwegian Province in south Scandinavia, a tectonic counterpart to the Grenville Province of North America. Available data indicate the existence of two high-grade metamorphic events in the Eastern Segment, one pre-Sveconorwegian at *c.* 1.43 Ga and one Sveconorwegian at *c.* 0.97 Ga. The relative importance and geochronology of tectonometamorphic events in the segment are controversial and have been discussed at length in scientific publications. The primary aim of

the present study was to use zircon formation during partial melting and systematic sampling to date defined phases of migmatization and high-temperature deformation at two representative key localities.

Several classical studies have demonstrated growth of zircon in migmatites (e.g. Oliver *et al.*, 1999; Vavra *et al.*, 1999; Rubatto *et al.*, 2001) and other high-grade metamorphic rocks (e.g. Vavra *et al.*, 1996; Williams *et al.*, 1996; Bingen *et al.*, 2001; Möller *et al.*, 2002). In the present study, we have attempted to apply this knowledge in investigation of polymetamorphic migmatitic orthogneisses. Our results demonstrate different zircon geochronological records from two different localities in the same terrane, and show that the difference is connected to the local variations in partial melting history. In most reports from migmatitic rocks, the morphology and textures of zircon suggest formation by dissolution and reprecipitation; here we present examples where recrystallization was the dominant process for zircon formation during migmatization.

Different parts of the migmatitic orthogneisses were sampled with the aim of finding a connection between zircon textures and partial melting. Also, at one of the two localities in our study, the aim is to date a fold phase that was synchronous with migmatization. Complex zircon populations from two different generations of granitic dykes were investigated with the aim of obtaining age brackets for another fold phase. Microtextures of individual zircon grains were documented by backscatter electron (BSE) and cathodoluminescence (CL) images; this work was followed by U–Pb–Th ion probe analysis. The results are integrated with results from new geophysical and geological mapping by the Geological Survey of Sweden. Together, the new data illustrate the imprints of the two major tectonometamorphic events and provide constraints for the complex tectonic evolution in this part of the Sveconorwegian Province.

ZIRCON DATING OF HIGH-GRADE METAMORPHISM AND DEFORMATION

Zircon U–Pb dating is a reliable and widely used method to investigate the geochronology of orogenic belts. Few other isotope systems have closure temperatures high enough (>650 °C) to escape diffusion during cooling after high-grade metamorphism or during reheating connected with later metamorphism. U, Pb and Th diffusion in crystalline zircon is known to be extremely slow at temperatures below 900 °C (Lee *et al.*, 1997; Möller *et al.*, 2002; Cherniak & Watson, 2003) and the system therefore has the qualification for remaining undisturbed even at elevated temperatures. Age determination of metamorphism with the U–Pb zircon system is, however, restricted to the type of rocks and external conditions in which either neocrystallization or recrystallization of zircon takes place.

During metamorphic recrystallization, neocrystallization of zircon can take place by the release of Zr and Si from minerals other than zircon (Fraser *et al.*, 1997). This process appears common in basic and silica-undersaturated rocks in which Zr has been contained in igneous minerals other than zircon (cf. Bingen *et al.*, 2001).

New zircon growth may also take place during partial melting by partial dissolution of zircon by an aggressive fluid or melt, followed by reprecipitation as the melt crystallizes (cf. Watson, 1996). Investigations of migmatitic rocks (mainly formed from metasedimentary protoliths) have confirmed that the amount of zircon growth is dependent on several factors, notably temperature, bulk rock composition and the volume of leucosome (Oliver *et al.*, 1999; Vavra *et al.*, 1999; Rubatto *et al.*, 2001). In these rocks zircon formed by dissolution and reprecipitation and occurs as unzoned, weakly zoned or oscillatory zoned overgrowths with distinct boundaries to older, remnant zircon cores. Also in migmatitic orthogneisses this process appears responsible for the formation of new zircon (e.g. Andersson *et al.*, 2002).

In situ recrystallization has been documented in some high-grade metamorphic rocks and migmatites. The recrystallized zircon domains typically have diffuse transitional, generally concordant boundaries to the protolith zircon and commonly also exhibits a relict igneous zoning ('ghost zoning', Hoskin & Black, 2000; Söderlund *et al.*, 2002; Hoskin & Schaltegger, 2003).

As secondary zircon can form by different processes, dating of a specific metamorphic event requires not only identification of secondary zircon as such but also a petrological clue to the process behind its formation. Discrimination between different textural types and generations of secondary zircon is essential for this interpretation and thereby for the connection between a zircon age and a geological event.

GEOLOGICAL SETTING

Sveconorwegian Province

The Sveconorwegian Province forms a 500 km wide mobile belt in the present-day south-west Baltic Shield and is a tectonic counterpart to the Grenville Province in the Canadian Shield (Bingen *et al.*, 2005). It is composed of several Proterozoic gneiss belts that are separated by roughly N–S-trending deformation zones (Fig. 1). The province is delimited to the east by the Protogine Zone and the Sveconorwegian Frontal Deformation Zone (Wahlgren *et al.*, 1994 and references therein). East of the Sveconorwegian Province are *c.* 1.92–1.75 Ga rocks belonging to the Svecofennian orogen, non-penetratively deformed rocks of the 1.81–1.66 Ga Transscandinavian Igneous Belt, and bodies of younger, *c.* 1.45 Ga granite. Across the Sveconorwegian Province metamorphic ages point to a

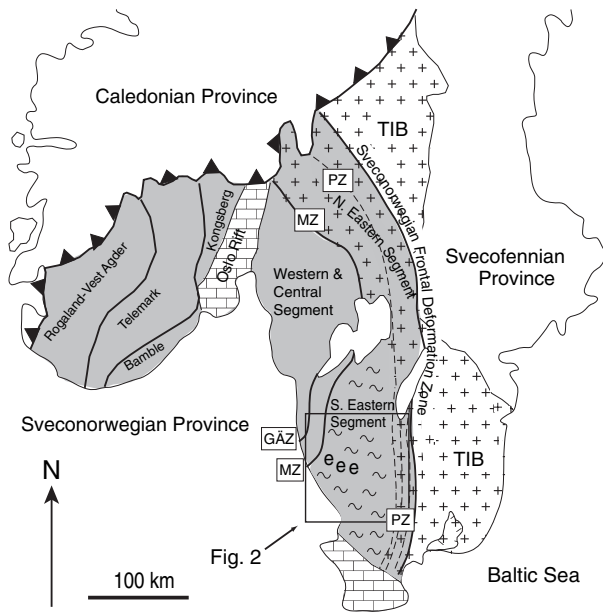


Fig. 1. Sketch map of southern Sweden and Norway. The Sveconorwegian Province is shown in shaded grey. Sveconorwegian deformation zones: PZ, Protogine Zone; MZ, Mylonite Zone; GÄZ, Götå Älv Zone. TIB = Transscandinavian Igneous Belt. Brick pattern = Phanerozoic rocks. Location of the Sveconorwegian Frontal Deformation Zone north of Lake Vänern follows Stephens *et al.* (1996). Box marks the location of Fig. 2.

sequence of orogenic events that took place over a time period of more than 150 Myr, between *c.* 1.13 and 0.96 Ga (Bingen *et al.*, 2005, and references therein). The final event at 0.97 Ga involved tectonic emplacement of eclogites followed by regional-scale metamorphism in the high-pressure granulite and upper amphibolite facies in the south-easternmost part of the orogen (Fig. 1), and is interpreted as a continent–continent collision event (Möller, 1998, 1999; Johansson *et al.*, 2001).

The Eastern Segment forms the easternmost, parautochthonous basement of the Sveconorwegian Province and is made up of orthogneisses and different generations of mafic rocks. In the southern part (south of Lake Vänern, Fig. 1) the metamorphic rocks are high-grade and largely migmatitic. The bulk of the gneisses have protolith ages of 1.73–1.66 Ga and show geochemical similarities with largely undeformed coeval rocks of the Transscandinavian Igneous Belt east of the Sveconorwegian Province (Connelly *et al.*, 1996; Söderlund *et al.*, 1999, 2002; Andersson *et al.*, 2002). Younger rocks include an igneous suite of 1.40–1.38 Ga granites and monzonites (Andersson *et al.*, 1999 and references therein) as well as metabasic rocks of different generations, all of which are overprinted by ductile deformation, migmatization and high-grade metamorphism.

Two major shear zones bound the Eastern Segment, both of which have accommodated significant dis-

placements during late stages of the Sveconorwegian orogeny. The Protogine Zone, a late Sveconorwegian, ~25 km wide deformation zone characterized by sub-vertical and discrete, N–S-trending shear, forms the transition to largely non-penetratively deformed and unmigmatized rocks of the Transscandinavian Igneous Belt in the east. The western boundary is a shallowly west-dipping shear zone named the Mylonite Zone. This zone is overlain by allochthonous gneiss belts and represents a major lithological, metamorphic and structural terrane boundary (Andersson *et al.*, 2002). In its northern parts, the Mylonite Zone is characterized by left-lateral transpressive deformation (Stephens *et al.*, 1996), but the southern parts have been strongly overprinted by later down-to-the-west extension (Berglund, 1997).

Sveconorwegian metamorphism in the southern Eastern Segment reached high-pressure granulite and upper amphibolite facies conditions. The *P–T* estimates obtained from metabasic rocks in the region yield temperatures between 680 and 770 °C and corresponding pressures of 9–12 kbar (Johansson *et al.*, 1991; Wang & Lindh, 1996; Möller, 1998, 1999). Secondary zircon dates the high-grade event at 0.98–0.96 Ga (Söderlund *et al.*, 2002 and references therein). Next to the Protogine Zone in the east, metamorphism took place at similar pressures but temperatures were lower, at ~600 °C (Söderlund *et al.*, 2004).

A dismembered unit of 972 ± 14 Ma eclogite occurs interfolded with the Eastern Segment gneisses (Möller, 1998, 1999; Johansson *et al.*, 2001), and has been overprinted by high-pressure granulite and amphibolite facies metamorphism and deformation. This distinct tectonometamorphic unit is evidence of *c.* 0.97 Ga thrust-related displacements within the deep-seated basement in this part of the Sveconorwegian orogen. High-angle discordant pegmatitic and granitic dykes dated at 956 ± 7 Ma (Möller & Söderlund, 1997; Andersson *et al.*, 1999), set a lower age bracket for the eclogite emplacement and the high-grade metamorphic interfolding.

Locally metabasic dykes, metamorphosed in the high-pressure granulite facies, are found discordant to veined gneisses. Although their igneous emplacement ages are unknown the high-grade metamorphic overprint sets a minimum age at *c.* 0.95 Ga. They may be equivalents to 0.98–0.94 Ga dolerites occurring in the Protogine Zone and east of the Sveconorwegian Province (Söderlund *et al.*, 2005), in which case they formed in response to orogenic collapse. In the southern parts of the Protogine and Mylonite Zones lower amphibolite facies and greenschist facies metamorphism and associated deformation has been dated at *c.* 920 Ma (Johansson & Johansson, 1993; Andréasson & Dallmeyer, 1995; Page *et al.*, 1996). At this time, movements in these large deformation zones accommodated E–W extension and uplift of the Eastern Segment.

Polymetamorphism in the Eastern Segment

Numerous studies have interpreted the regional deformation structures with concordant migmatite veins as older than 1.55 Ga (Connelly *et al.*, 1996; Christoffel *et al.*, 1999 and references therein). The recent application of single-grain and ion-microprobe dating of secondary zircon has, however, provided growing evidence for Sveconorwegian migmatization and deformation (Andersson *et al.*, 1999, 2002; Söderlund *et al.*, 2002 and references therein), results which are consistent with remnant Sveconorwegian eclogites and high-pressure granulites. Pre-Sveconorwegian, *c.* 1.46–1.42 Ga, monazite and zircon have also been dated at some different localities (Christoffel *et al.*, 1999; Söderlund *et al.*, 2002; Hegardt *et al.*, 2005). This, hitherto comparatively vaguely defined, metamorphic event was succeeded at 1.40–1.38 Ga by intrusion of granite to monzonite, and contemporaneous charnockitization of country rock gneisses (Hubbard, 1975; Åhäll *et al.*, 1997; Andersson *et al.*, 1999; Rimsa *et al.*, 2007).

Recent and ongoing regional geological mapping (1:250 000) by the Geological Survey of Sweden, supported by a new high-resolution airborne magnetic survey, provide a new framework to interpret regional geological and structural data in the Eastern Segment. The magnetic map reveals a banded and folded pattern (Fig. 2), reflecting a tectonic banding of alternating gneisses and metabasites and gneissic and/or migmatitic structure. The dismembered Sveconorwegian eclogite unit is concordant with this banding. The regional foliation has been folded in a complex pattern (Fig. 2) of which certain elements have been possible to define:

1. Roughly E–W to WNW–ESE-trending folds with gently plunging fold axes.
2. In the south-western parts of the segment, two phases of upright to moderately overturned folding intersect each other, with axes trending roughly WNW–ESE and NNE–SSW, respectively. Some of the limbs of the WNW–ESE folds are high strain zones.
3. Along the southern part of the Mylonite Zone and in areas within 25 km to the east of it, the structural trend is NNW or N–S, and appears to have resulted from tectonic activity along the zone overprinting and rotating older structures.

ANALYTICAL METHODS

Zircon mineral separates were obtained from density separation of ~1 kg of crushed rock sample using a Wilfley water table. The magnetic minerals were removed by a hand magnet. About 100–200 hand-picked crystals from each sample were cast in epoxy. The zircon mounts were polished and examined by BSE and CL imaging, using standard electron microscopy at the Department of Geology, Lund University, and the Museum of Natural History in Stockholm, Sweden, respectively.

High-spatial resolution secondary ion mass spectrometer analysis was made using a Cameca IMS 1270 at the Nordsim facility at the Swedish Museum of Natural History in Stockholm. Detailed descriptions of the analytical procedures for these analyses are given in Whitehouse *et al.* (1999) and Whitehouse & Kamber (2005). Analytical data are given in Tables S1 and S2. Diagrams and age calculations of isotopic data were made using software Isoplot 3.00 (Ludwig, 2003). The age calculations were based on concordant data only and statistical errors are given in a 95% confidence interval unless explicitly stated otherwise.

LOCALITY 1: SYNCHRONOUS DEFORMATION AND MIGMATIZATION

Structural and metamorphic relations

The specific objective for this locality was to date the E–W-trending fold structures that form conspicuous patterns in the magnetic anomaly map and geological map in area 1 (Figs 2 & 3). The large-scale folds are upright to moderately overturned, have wavelengths of 4–15 km and subhorizontal, undulating fold axes; folds are generally south vergent. In these structures, migmatitic gneisses have been tightly folded and garnet amphibolite layers have been stretched and boudinaged along the same direction. Older, small-scale folds are commonly tight to isoclinal and intrafolial and in places rootless. Stretching lineations are generally fold-axis-parallel and defined by strung-out mineral aggregates, leucosome veins and amphibolite bands. In general, the linear fabric is more strongly developed than the planar fabric and locally a planar fabric is lacking.

Samples were taken in a quarry (Oxanäset, Fig. 3) situated in the southern limb of a synformal fold. Outcrop-scale folds are tight, have gently north-dipping axial planes and gently west-plunging fold axes. The main lithology is granitic gneiss with a migmatitic layering that is generally stromatic (Fig. 4a). Remnants of a coarser-grained, relict igneous protolith fabric are locally partly preserved. The gneiss is intercalated with cm- to dm-wide bands and lenses of amphibolite (Fig. 4b) and with thick, up to 10 m wide, boudins of compositionally layered garnet amphibolite (Fig. 4c). Leucosome occurs in the boudin necks, and is also present as veins perpendicular to the stretching direction within garnet amphibolite (Fig. 4c). In amphibolite, garnet is resorbed and has been partly replaced by coronas of plagioclase and hornblende, a texture that suggests decompression and/or cooling.

Three structural generations of leucosomes are recognized. The oldest one is penetrative, mm- to cm-wide, fine- to medium-grained granitic vein that define a stromatic layering concordant with the gneissic layering and which is folded at outcrop scale (Fig. 4a,d). A structurally younger generation of leucosomes

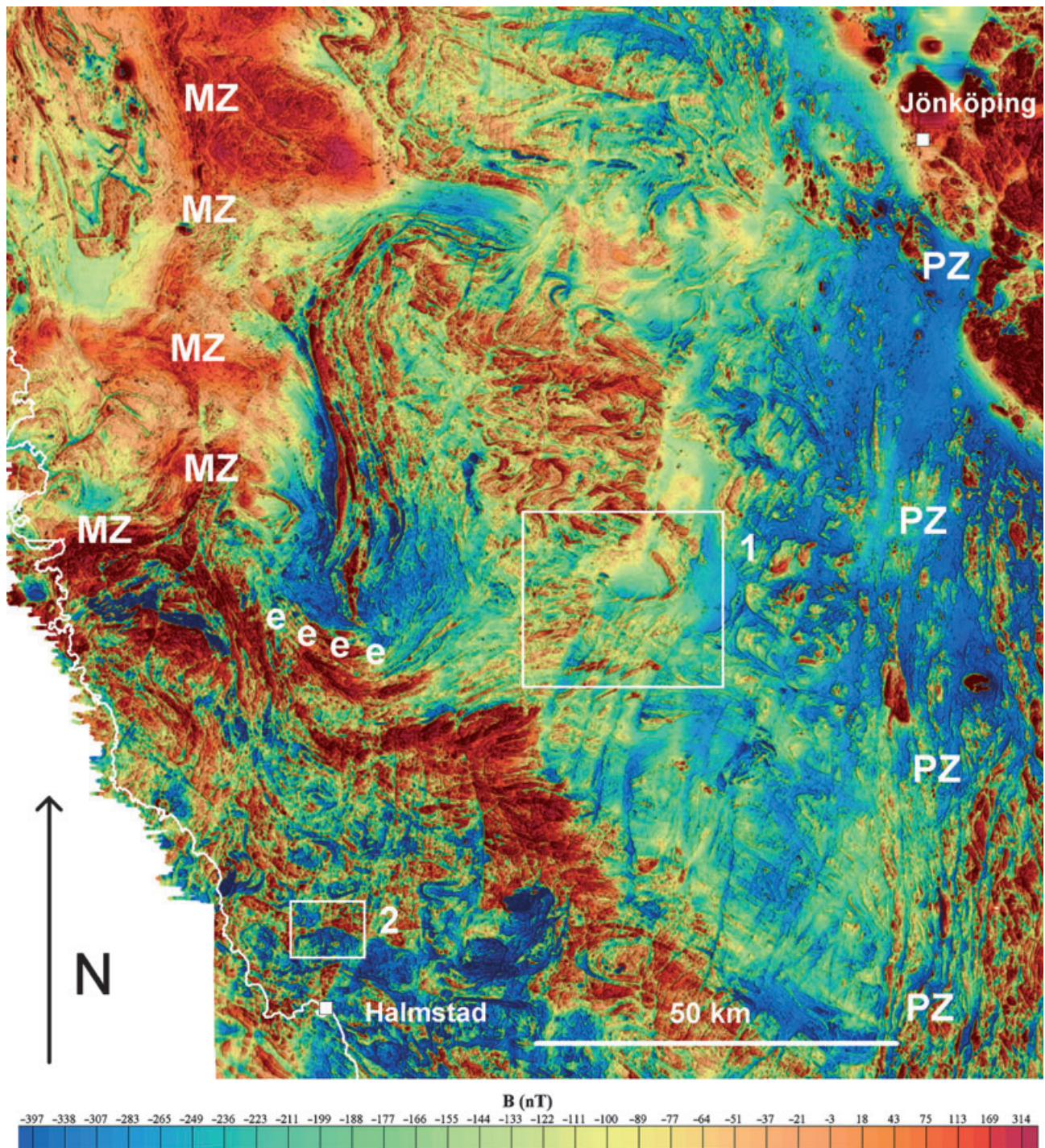


Fig. 2. Magnetic anomaly map illustrating the tectonic banding and fold patterns in the south-east Sveconorwegian Province. PZ, Protogine Zone; MZ, Mylonite Zone (over large areas MZ has a gentle dip making it not to appear as a sharp lineament); e, Sveconorwegian eclogite. The white line in the south-west outlines the coast. Boxes mark the locations of key localities 1 and 2. Part of the airborne magnetic map (total field) over SW Sweden. Data source: Geological Survey of Sweden.

occurs as more than cm-wide, medium- to coarse-grained veins that commonly are concordant to the gneissic layering and folded, but that in places crosscut the fold structures (Fig. 4e,f). These veins locally host porphyroblasts of magnetite or hornblende. Both types

of veins are bordered by a biotite-rich melanosome and are interpreted to have formed from *in situ* partial melting of the granite gneiss protolith. A third type of leucosome crosscuts or just blurs the folded veined gneissic structures. Thus, migmatization took place

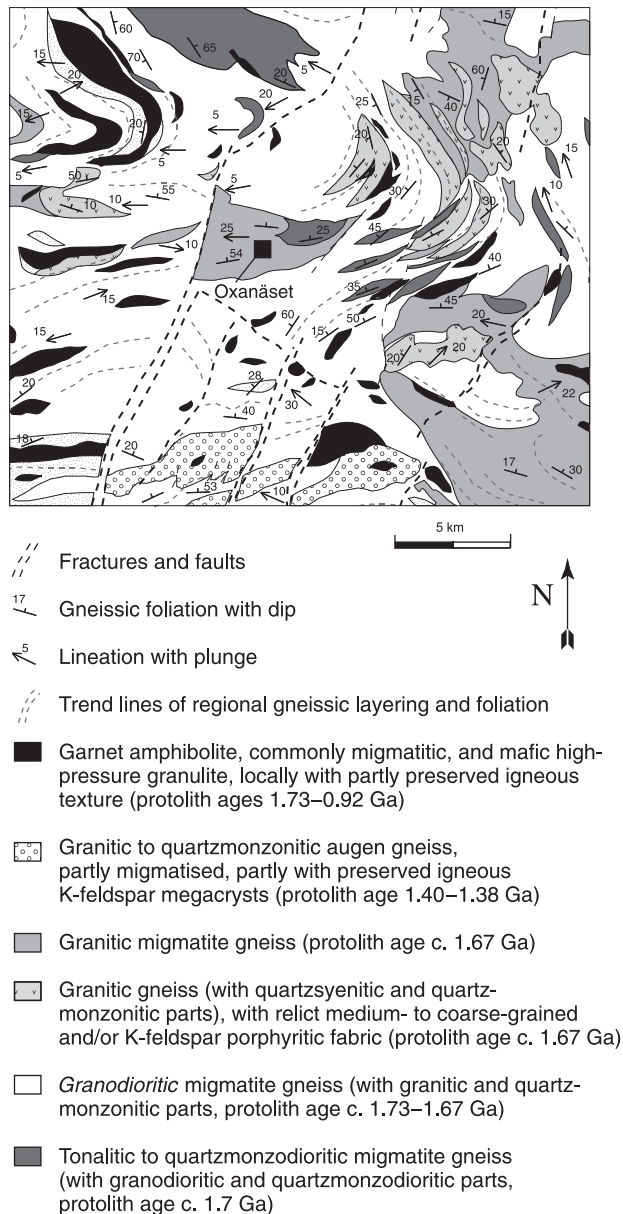


Fig. 3. Geological map of the area of locality 1. Black square marks the sample locality (Oxanåset). Data source: Geological Survey of Sweden.

simultaneously with the south-vergent folding along E–W axes, as demonstrated by the type of leucosome that occurs in both folded and semi-discordant relations to these folds (referred to as syn-tectonic leucosome, Fig. 4e,f). Although the meso-scale structures show strong stretching (Fig. 4f) the mineral fabric is isotropic in both the folded and the syn-tectonic veins.

U–Pb zircon ion probe dating

Three samples were chosen in order to characterize zircon in different varieties of granitic gneiss: (i) reddish grey, fine- to medium-grained, strongly foliated gneiss without distinct leucosome (gneiss mesosome, OX-1), (ii) gneiss with fine-grained, greyish mesosome and distinct, light reddish, 3–15 mm wide, folded leucosome of granitic composition (gneiss with folded leucosome, OX-2, cf. Fig. 4a,d), (iii) coarse-grained leucosome that occurs in both folded and discordant relations to the fold structures (syn-tectonic leucosome OX-3, cf. Fig. 4e,f). Analytical data are given in Table S1 and ages summarized in Table 1. The three samples are described separately below, followed by a summary and interpretation of the results.

Gneiss mesosome

The mesosome (sample OX-1) is made up of quartz, microcline and plagioclase (locally myrmekite) with granoblastic or smoothly sutured grain boundaries. Oriented grains of olive green biotite and titanite define the gneissic foliation.

Zircon is typically 100–200 μm long, short prismatic (length to width ratios: 1.5–2), and anhedral with rounded terminations. Euhedral morphologies are rare. Most crystals are colourless but contain inclusions and cracks, commonly with hematite staining. In polished section the crystals are CL-bright and BSE-dark with oscillatory zoning and thin, 5–15 μm wide, CL-dark and BSE-bright rims (Fig. 5a,b). Crystals with heterogeneous, patchy BSE-bright interiors occur, but in subordinate amounts.

Eight concordant analyses in oscillatory zoned, BSE-dark domains gave a concordia age of 1668 ± 11 Ma (MSWD = 0.01; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$

Table 1. Summary of samples and age results (recommended ages in bold).

Sample	Lithology	Mineral assemblage	Intrusion	Migmatization
Locality 1, Oxanåset. Swedish National Grid: 634196/135970				
OX-1	Gneiss mesosome	Qtz, Kfs Pl, Bt, Ttn, Opq, Grt, Ap, Zrn, Fl (Hem)	1668 $\pm$ 11 Ma	
OX-2	Gneiss with pre-kinematic leucosome	Qtz, Kfs, Pl, Bt, Ttn, Opq, Grt, Zrn, Ap, Aln (Ms, Chl, Cal)	1688 $\pm$ 26 Ma	970 $\pm$ 7 Ma
OX-3	Syn- to late-kinematic leucosome in gneiss	Qtz, Kfs, Pl, Opq, Ttn, Bt, Ap, Zrn, Aln	1.68–1.62 Ga	976 $\pm$ 7 Ma
Locality 2, Högabjär. Swedish National Grid: 629648/132281				
HB-1	Gneiss mesosome	Kfs, Pl, Qtz, Hbl, Grt, Bt, Opq, Ap, Zrn, Aln	1686 $\pm$ 12 Ma	
HB-2	Gneiss leucosome	Kfs, Pl, Qtz, Hbl, Bt, Opq, Ap, Grt, Zrn, Aln	1693 $\pm$ 22 Ma	1425 $\pm$ 7 Ma
HB-3	Folded metagranitic dyke	Kfs, Qtz, Pl, Opq, Bt, Grt	1394 $\pm$ 12 Ma	
HB-4	Undeformed cross-cutting granitic dyke	Kfs, Pl, Qtz, Hbl, Bt, Ttn, Zr	945 $\pm$ 7 Ma	
HB-6	Undeformed cross-cutting granitic dyke	Kfs, Pl, Qtz, Mag	952 $\pm$ 7 Ma	

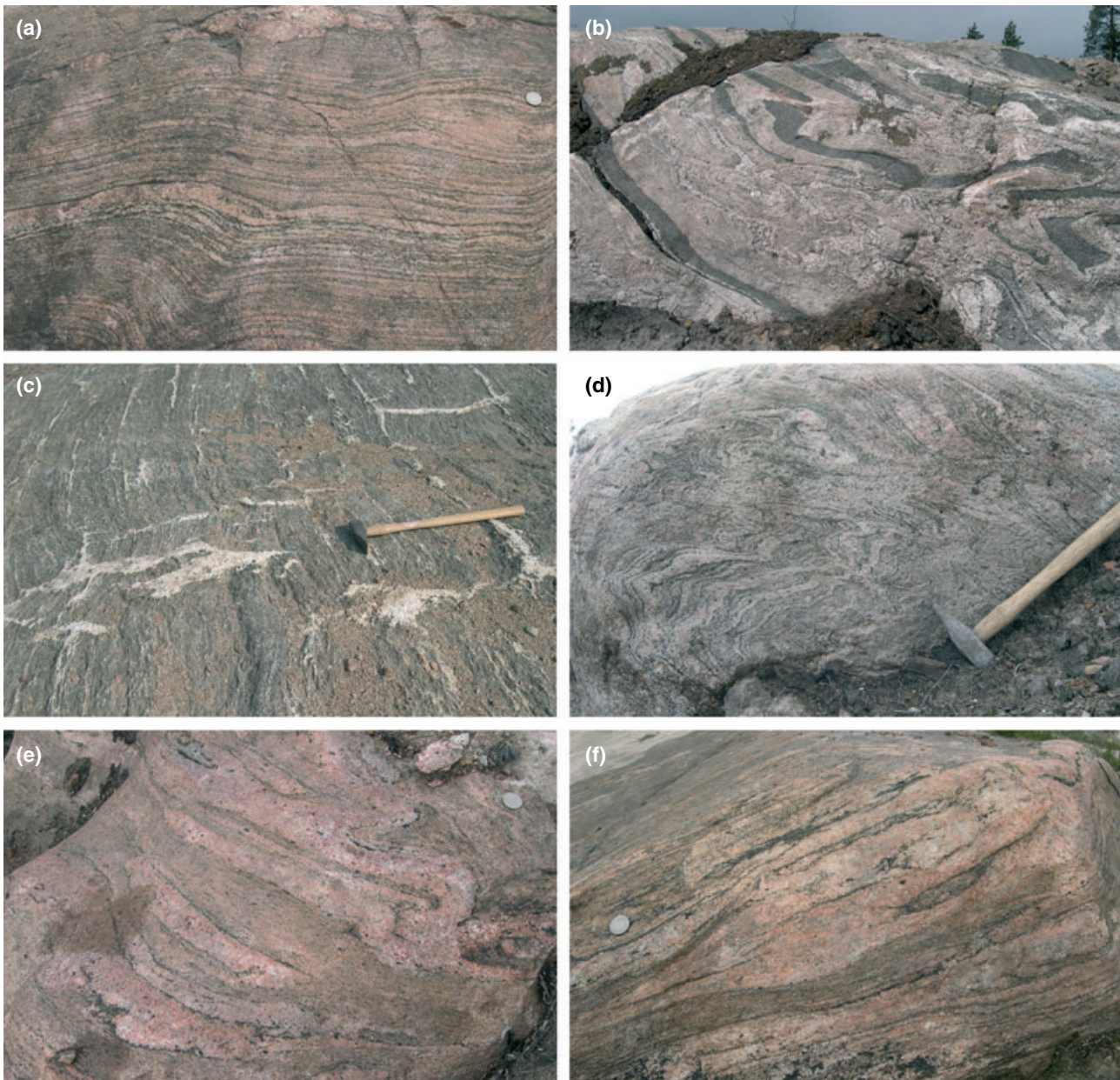


Fig. 4. Structural relations at Oxnäset (locality 1). Coin (25 mm) and hammer (0.55 m) for scales. (a) Stromatic migmatite layering in granitic gneiss, representative of sample OX-2. (b) Overview, showing disrupted and folded amphibolite bands, up to 0.2 m wide, in migmatite gneiss. (c) Concordant garnet amphibolite layer, boudinaged in E–W direction and carrying leucosome concordant and perpendicular (orientation of hammer) to the stretching direction. (d) Migmatite gneiss, folded along gently west-plunging fold axis and gently north-dipping axial plane. Small, late leucosome obscures the folded fabric. (e) Syn-tectonic, folded leucosome (representative of sample OX-3) partly cross-cutting a stromatic layering. Section oblique to the stretching direction. (f) Syn-tectonic, concordant and coarse-grained leucosome (representative of sample OX-3) in migmatite gneiss in a section parallel with the stretching direction.

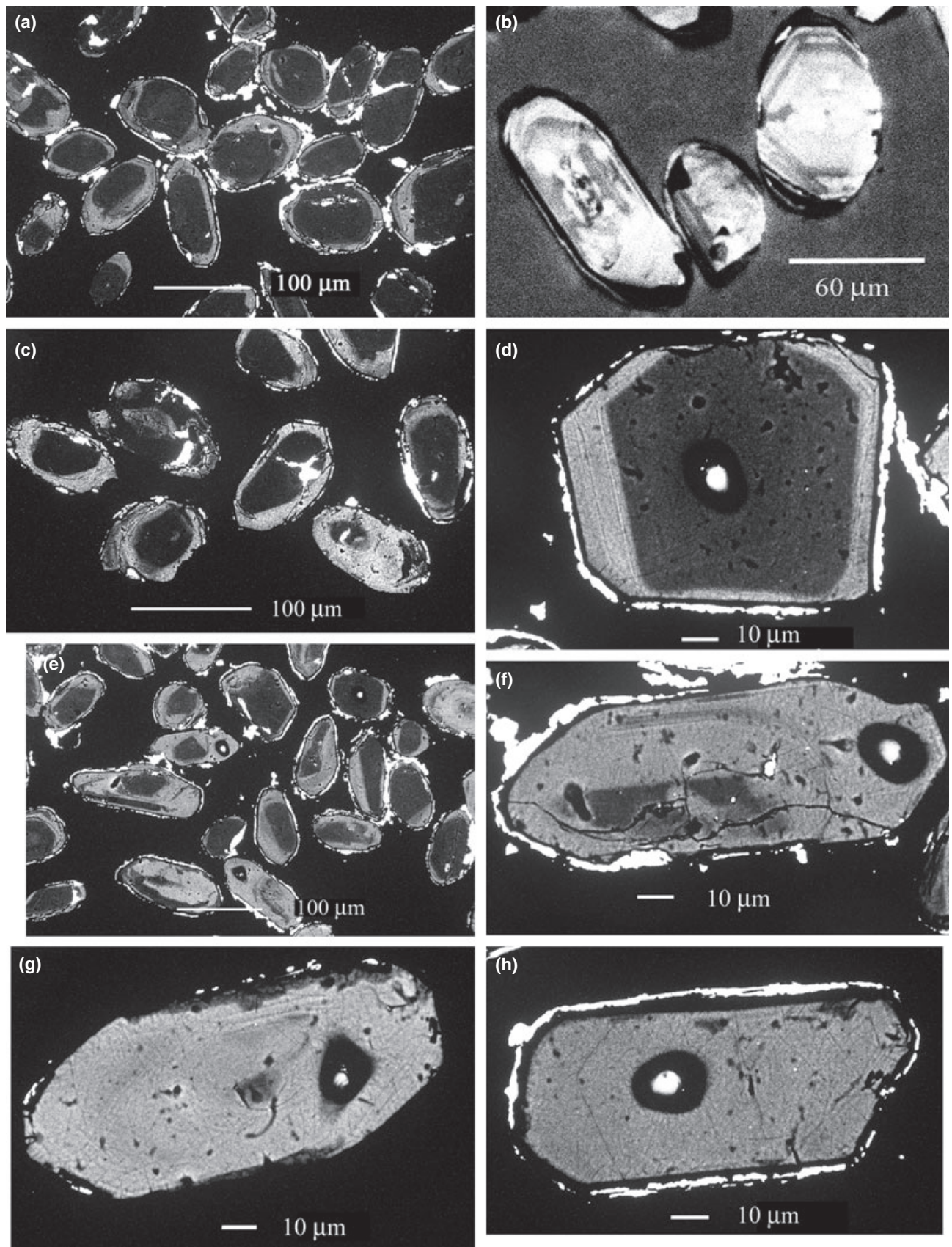
age = 1667 ± 15 Ma, MSWD = 0.89). Th/U-ratios are high (0.76–0.94) and U-concentrations low (61–105 ppm). BSE-bright rims were too thin to analyse (spot size = 15–20 μm).

Gneiss with folded leucosome

The gneiss with folded leucosome (sample OX-2) is composed of quartz, microcline, plagioclase and small

amounts of muscovite. Dark minerals are biotite, garnet, titanite and opaque minerals. Muscovite occurs in small amounts in the leucosomes where quartz may form up to 2–3 mm large grains.

The morphology of zircon is similar to that in sample OX-1, with the exception of the occurrence of larger, slightly turbid and clouded greyish grains. CL-bright and BSE-dark zircon with igneous oscillatory zoning dominates, typically surrounded by up to



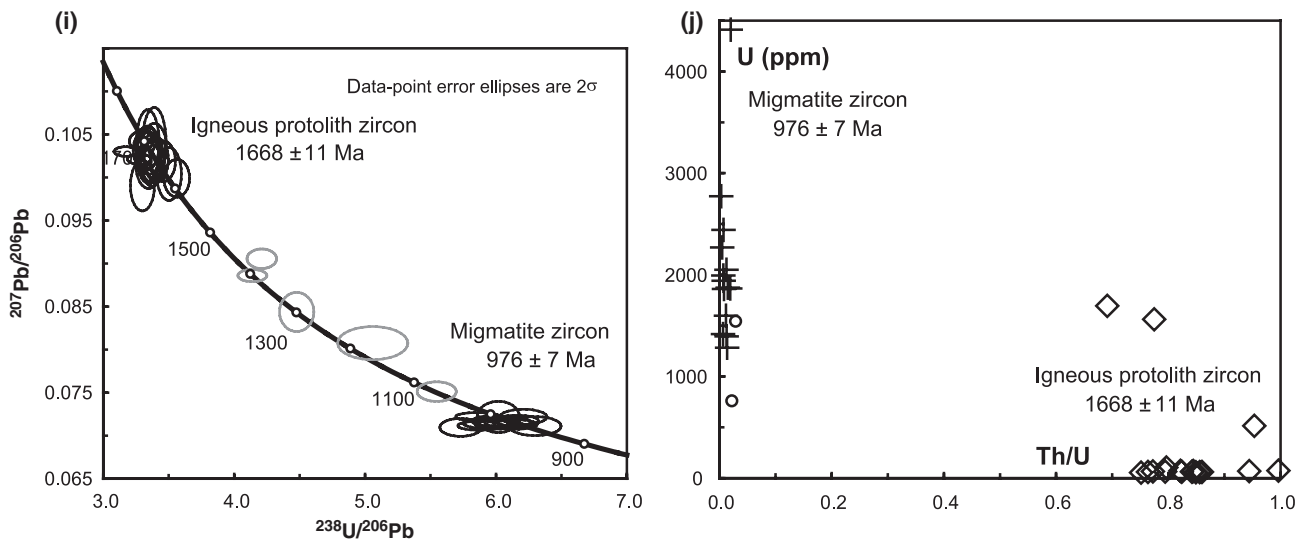


Fig. 5. *Cont'd.*

30 μm wide, CL-dark and BSE-bright rims (Fig. 5c). Most rims are anhedral but some grains have subhedral to euhedral crystal faces (Fig. 5d). BSE-bright zircon is also found in the interior parts of some grains, in places with a faint, but blurred oscillatory zoning inferred to be of igneous origin (analysis n1753-08a). BSE-bright rims and interiors were the main targets for analysis.

Five analyses of BSE-dark and -bright core domains yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1700 ± 8 and 1659 ± 17 Ma. A calculation of concordia age based on three concordant analyses, from one oscillatory-zoned BSE-dark and two BSE-bright domains, yielded 1688 ± 26 Ma (MSWD = 1.08; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1692 ± 45 Ma, MSWD = 2.4). The age is within error identical with that obtained from the gneiss mesosome (sample OX-1), but based on too few analyses to be statistically significant. The analysed core domains all have high Th/U-ratios (> 0.69) similar to the analysed zircon in sample OX-1. In the BSE-bright domains the U-content is, however, higher (> 500 ppm).

One near-concordant analysis (0.7% discordance, n1753-02a), of an unzoned, BSE-bright zircon gave a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 970 ± 7 Ma. This zircon yielded

the highest U (~ 2800 ppm U) and lowest Th-content (Th/U < 0.01) among the zircon in the sample.

Two analyses in BSE-bright core domains and one analysis in a BSE-bright rim (not a distinct overgrowth) yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages at 1437 ± 10 , 1396 ± 6 and 1215 ± 19 Ma. This zircon has 760–1543 ppm U and Th/U ratios of 0.03–0.02, values between those of the 1688 ± 26 and the 970 ± 7 Ma zircon. The youngest of these three ages (n1753-04a) is from a heterogeneously zoned domain and most likely a mix between domains of different generations.

Syn-tectonic leucosome

The syn-tectonic leucosome (sample OX-3) is unequigranular with coarse-grained quartz, microcline, plagioclase and opaque minerals and fine-grained titanite and biotite. Aggregates of titanite contain rounded zircon grains, $\sim 5 \mu\text{m}$ wide.

The external zircon morphology is similar to that of OX-1 and -2, but some grains also have euhedral crystal faces. The internal textures are, however, distinctly different. BSE-dark core domains, in places with igneous oscillatory zoning, occur in some grains, but CL-dark and BSE-bright domains are significantly

Fig. 5. Electron microscope images and U–Pb–Th ion microprobe analytical data of zircon from Oxnäset. In BSE-images the analysed spot is seen as a black area with a white centre; very bright spots/areas are remnants of gold coating. (a) BSE-image of zircon crystals from gneiss mesosome (sample OX-1). The crystals are dark igneous zircon with thin light rims. (b) CL-image of zircon in gneiss mesosome. The crystals are dominantly brightly luminescent with dark rims. (c) BSE-image of zircon crystals from gneiss with folded leucosome (sample OX-2). Most crystals are dark igneous with light rims but some have irregularly shaped light interior domains. (d) Near-euhedral section of igneous zircon crystal with light, probably igneous, rim (sample OX-2, analysis n1753-03a). (e) BSE-image of zircon from syn-tectonic leucosome (sample OX-3). Crystals are composite with dark igneous cores and bright domains and rims of secondary zircon. (f) Composite zircon crystal with BSE-bright, U-rich rim, up to 40 μm wide, and a remnant core of protolith zircon (n1754-05a). (g) Composite zircon crystal with BSE-bright, U-rich rim with oscillatory 'ghost zoning', and a barely visible core of protolith zircon (n1754-19a). (h) Simple BSE-bright secondary zircon (n1754-09a). (i) Inverse concordia (Tera-Wasserburg) diagram of zircon analyses from Oxnäset (all samples). Grey ellipses mark five analyses (including three mixed analyses; see text for details) that fall outside the two clusters at 1.67 Ga (protolith zircon) and 0.98 Ga (migmatite zircon). (j) U versus Th/U-diagram. Diamonds = 1.67 Ga igneous zircon, crosses = 0.98 Ga migmatite zircon. Circles = two analyses yielding *c.* 1.4 Ga. Three analyses of mixed domains have been excluded (see text for details).

more voluminous and more homogenous than in samples OX-1 and OX-2 (Fig. 5e). The majority of the BSE-bright domains occur as wide rims enclosing rounded remnants of BSE-dark zircon (Fig. 5f). Many of the secondary BSE-bright domains have a faint 'ghost structure' that resembles oscillatory zoning (e.g. spots 06, 04 & 19; Fig. 5g). Commonly, the outermost parts of these domains are structureless (Fig. 5f). Structureless, BSE-bright zircon with euhedral crystal faces is also present (Fig. 5h). One single crystal was found to be dark in both CL and BSE.

Four analyses in BSE-dark core domains with zonation bands yielded $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1682 ± 17 and 1622 ± 17 Ma, within error (2σ) identical to the ages of texturally similar zircon in samples OX-1 and -2. These domains also have low U-concentrations (< 75 ppm U) and high Th/U ratios (> 0.79).

Eleven concordant analyses were obtained from CL-dark domains, all but one bright in BSE. These analyses included both unzoned homogenous rims and domains with oscillatory 'ghost zoning', and yielded a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 976 ± 7 Ma (MSWD = 1.7). The domains are high in U (> 1250 ppm) and have low Th/U ratios (< 0.02). The highest U-concentration (4412 ppm) was found in the CL-dark and BSE-dark grain (n1754-13a). Two analyses in BSE-bright domains were excluded from age calculation as they crosscut into cores (analyses n1754-03a & 19a, Table S1).

Summary and interpretation

CL-bright and BSE-dark, commonly oscillatory zoned zircon is present in all three samples (Fig. 5i) and is interpreted as remnants of igneous protolith zircon. It typically has high Th/U ratios (> 0.69) and relatively low U-contents (< 105 ppm; Fig. 5j). Four analyses taken in BSE-brighter domains demonstrate, however, compositions richer in U (500–1700 ppm). Eight concordant analyses of igneous zircon from the gneiss mesosome (OX-1) define a protolith age at 1668 ± 11 Ma (concordia age, MSWD = 0.01). The external zircon morphology is similar in all three samples, but the proportion of younger, BSE-bright and CL-dark secondary zircon varies widely between the three samples. Secondary zircon is abundant and voluminous in syn-tectonic leucosome (OX-3; Fig. 5e), and is also present in gneiss with early leucosome (OX-2). The secondary zircon has low Th/U ratios (< 0.02) and high U contents (1300–4400 ppm) and is chemically distinct from the protolith zircon (Fig. 5j). In the mesosome (OX-1), 5–15 μm BSE-bright wide rims are too thin for analysis, but are likely of the same origin. The correlation between the size of leucosome and amount of secondary zircon, and the corresponding decrease in size of igneous zircon remnants, demonstrate that the secondary zircon rims formed during migmatization, of one and

the same protolith. The timing for this event is set at 976 ± 7 Ma (weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of concordant analyses in the syn-tectonic leucosome; Fig. 5i).

The secondary, BSE-bright zircon show textures characteristic of recrystallization, with oscillatory-like 'ghost zoning' similar in appearance to domains formed by diffusion after igneous crystallization (*in situ* recrystallization, cf. Hoskin & Black, 2000). In the syn-tectonic leucosome at Oxnäset these domains must be the result of partial to complete recrystallization during migmatization. Secondary unzoned domains, locally with euhedral crystal faces, may have formed by accompanying neocrystallization or simply be the result of efficient homogenization during recrystallization. The positive correlation between leucosome volume and amount of new zircon suggests that also for recrystallization processes the interacting melt volume is important for the formation of new zircon.

Five of the zircon analyses from the migmatitic samples give ages that spread along the concordia from 1437 ± 10 to 1072 ± 12 Ma, between the age of the igneous protolith and the age of the partial melting event (Fig. 5i). The three youngest are artefacts produced by mixed volumes of 1668 ± 11 Ma protolith zircon and domains completely or partly reset at 976 ± 7 Ma, but the origin of the two older at 1396 ± 6 and 1437 ± 10 Ma (n1753-01a, BSE-bright interior, and n1753-09a, BSE-bright rim, discordant) is less certain. These domains may possibly represent pre-Sveconorwegian secondary zircon.

Geological interpretation

The migmatitic gneiss at Oxnäset belong to the 1.73–1.66 Ga generation of granitic and related intrusive rocks in the Eastern Segment, as demonstrated by the ages of igneous zircon cores in three variably migmatized samples. The age of the granite protolith is 1668 ± 11 Ma.

The timing of migmatization and formation of secondary zircon in the syntectonic leucosome is set at 976 ± 7 Ma. As this type of leucosome occurs in semi-discordant folded positions in relation to the E–W trending folds, it directly dates the folding and the associated amphibolite facies stretching and boudinage. By implication it sets the age for one of the regional deformation phases in the southern Eastern Segment, the phase that resulted in upright to southwards overturned folds with E–W to WNW–ESE trending, subhorizontal axes, and associated stretching along the same direction.

It is feasible that the Oxnäset gneiss experienced also earlier metamorphism at *c.* 1.4 Ga, but a still larger data set is needed in order to verify this. If a 1.4 Ga zircon generation is present in the Oxnäset gneiss it is clearly in subordinate amounts in relation to the Sveconorwegian generation.

LOCALITY 2: MIGMATIZATION AND YOUNGER DEFORMATION

Structural and metamorphic relations

Polyphase ductile structures are well-exposed in south-western, near-coastal parts of the Eastern Segment. They are the result of superposed phases of deformation connected with folding along WNW–ESE and NNE–SSW trending axes, respectively. The youngest structures are brittle or ductile-brittle fault zones, the most prominent oriented along N–S (Fig. 2), that offset the ductile structures. There are, however, also ductile high strain zones, apparently located in tight limbs of regional WNW–ESE folds that disrupt the NNE–SSW trending gneissic layering (Figs 2 & 6). A locality at Högabjär (Fig. 6) was chosen for investigation, with the objectives to date the migmatization and to set age brackets for the deformation phases that overprint the migmatization, particularly the regional NNE–SSW-trending fold phase.

Reddish grey migmatite gneiss with flame-like, folded leucosome structures is one of the most common rocks of the area. Other rocks are red granitic gneiss with less conspicuous migmatization, leucocratic granitic gneiss and granitic augen gneiss with subordinate migmatite veining only. Mafic layers and boudins of garnet amphibolite and high-pressure granulite occur intercalated and interfolded with the felsic gneisses (cf. Fig. 6). Due to differences in competence the rocks differ in strain state, but share a common metamorphic and deformational history. Exceptions are structurally young, high-grade metadolerite dykes and granitic to pegmatitic dykes that locally occur discordant to the migmatitic layering in the host rock.

A quarry at Högabjär (10 km north of Halmstad, Figs 2 & 6) exposes the structural relations in migmatite gneiss in three dimensions. The gneiss has medium-grained, greyish red, granitic leucosome that makes up ~25–40% of the rock volume (Fig. 7a–d). The leucosome veins are generally up to 5 cm thick but decimetre-wide leucosome occurs locally. Melanosome is lacking. The veins are oriented subparallel to one another and define a layering with an overall strike along NNE and a steep dip towards WNW. Red, medium- to coarse-grained metagranite dykes, up to 0.5 m wide, occur with low- to high-angle discordant relations to the migmatitic layering (Fig. 7b,c). The metagranitic dykes and their host gneiss have been folded (Fig. 7b,c) and the original geometry at the time of dyke intrusion has thus been severely modified.

On horizontal surfaces the structures appear complex with highly irregular leucosome pods (Fig. 7a,c). The structural relations are best illustrated on south- and north-facing vertical surfaces, roughly perpendicular to the axial surface of outcrop-scale folds (Fig. 7d). The folds are tight, upright to slightly overturned, with axial surfaces striking NNE and dipping ~60° to the ESE. In places leucosome material is located along the axial-planar fold limbs (Fig. 7d), possibly developed during the folding. Fold axes plunge 20–50° to the SSW. A linear deformation fabric, oriented parallel with the axial-plane of the folds, has developed in medium and coarse-grained rock domains, i.e. in leucosome (Fig. 7d,e) and folded granitic dykes. The fabric is a stretching lineation defined by elongated and recrystallized mineral aggregates. It is, however, not parallel to the fold axis, but plunges ~70° towards ESE.

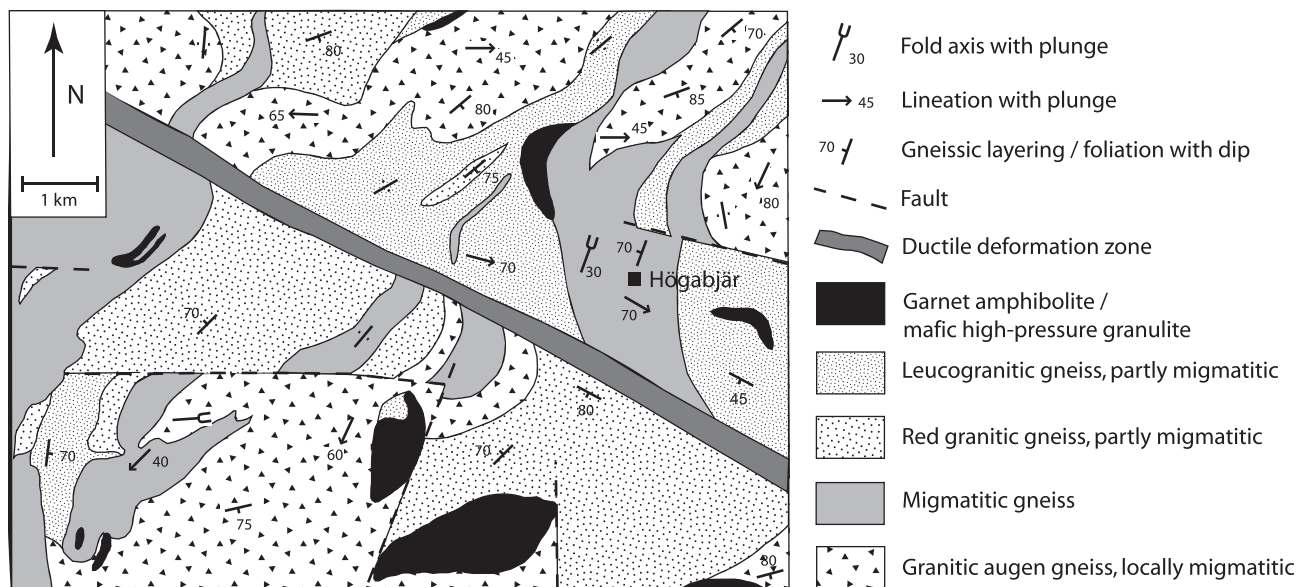


Fig. 6. Geological map of the area of locality 2 (north of Halmstad). Black square mark the sample locality (Högabjär). Data source: Geological Survey of Sweden.

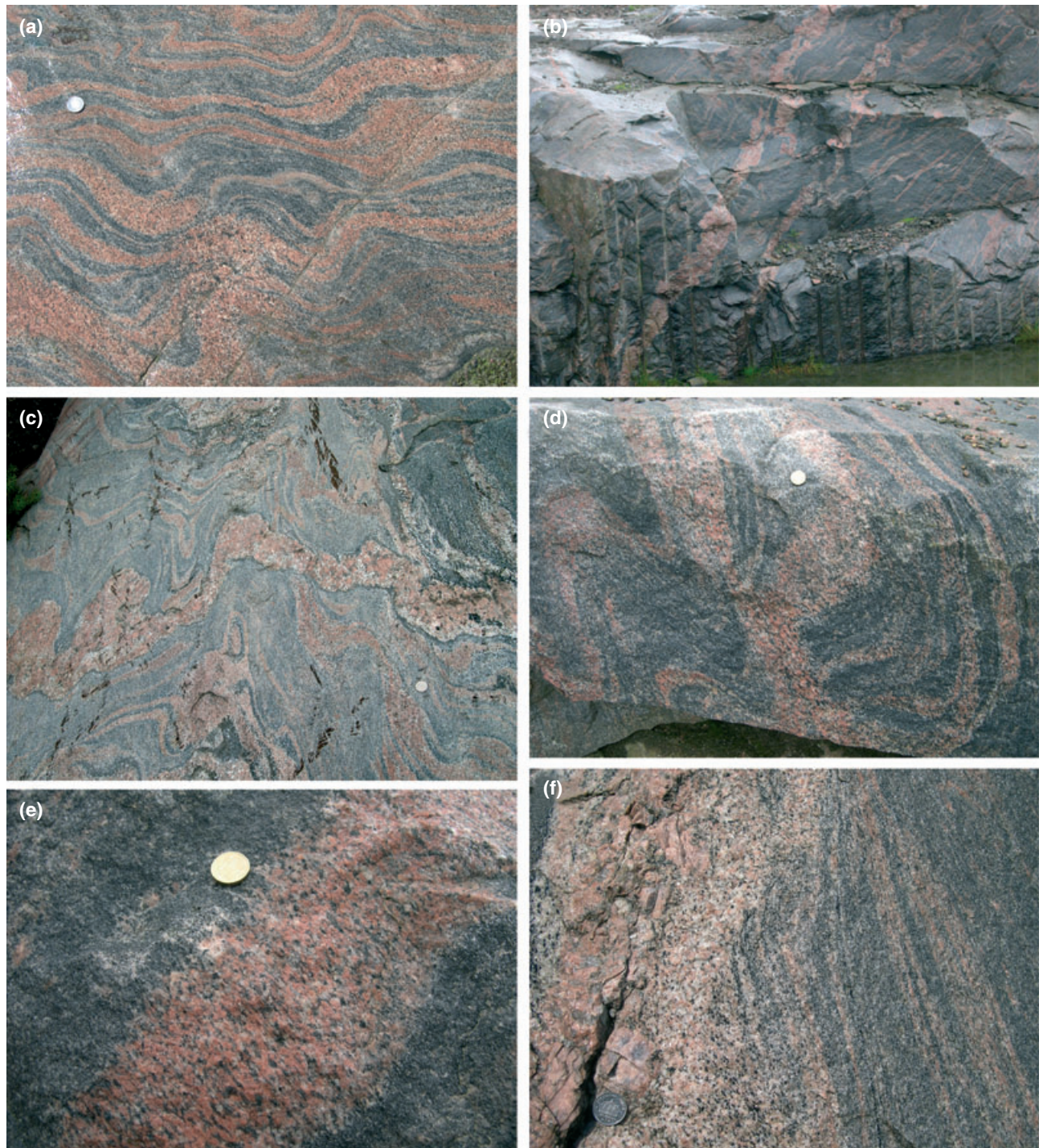


Fig. 7. Structural relations at quarry Högabjär (locality 2). Coin (25 mm) for scale. (a) Mesoscale banding and flame-like leucosome (the latter representative of sample HB-2) in migmatitic gneiss; outcrop on sub-horizontal surface. (b) Vertical, east-facing surface (~4 m high) of the quarry, showing the coarse gneissic layering in migmatitic gneiss and two discordant metagranite dykes (folded). (c) Discordant relationships between migmatitic gneiss and a metagranite dyke (representative of sample HB-3), overprinted by folding (subhorizontal surface). (d) Geometry of upright to moderately overturned folds with gently SSW-plunging axis. Vertical, south-facing surface, perpendicular to the fold axis. (e) Steeply ESE-plunging stretching lineation, best developed in leucosome, close up of the fold in (d). (f) Undeformed granite-pegmatite dyke (left), crosscutting the structures in the migmatite gneiss (sample HB-6; vertical E–W trending surface).

Late, crosscutting dykes occur at a few places. One granulite facies metadolerite is located in the southern part of the quarry. Undeformed granite dykes are up to 20 cm wide, fine- to coarse-grained, pinkish and have an isotropic mineral fabric. They are oriented around NNW, dips steeply towards WSW, and cross-cut the folded structures of the host rock (Fig. 7f).

U–Pb zircon ion probe dating

Five samples were selected at Högabjär. Two samples of migmatite gneiss, mesosome HB-1 and leucosome HB-2, were investigated to connect a specific zircon generation to the leucosome formation and thereby obtain a direct age of the migmatization. One sample, HB-3, of deformed metagranitic dyke, discordant to the leucosome in the host gneiss but folded together with the gneiss, was investigated with the aim of obtaining an upper age bracket for the upright folding along SSW-plunging axes and for the ESE-stretching. Two samples of undeformed and crosscutting granitic dykes, HB-4 and HB-6, were dated in order to obtain lower age bracket for the same events.

Analytical data from the two samples of migmatitic gneiss are given in Table S1; data from the granitic dykes are found in Table S2. The samples are described separately below, followed by a summary and interpretation of the results. Ages are summarized in Table 1.

Migmatitic gneiss: mesosome

The mesosome (sample HB-1) is grey, fine-grained, devoid of distinct leucosome but has diffuse, red and 1–3 mm wide, granitic micro-domains. Besides micro-perthitic K-feldspar, antiperthitic plagioclase and quartz, it contains small amounts of hornblende, garnet and biotite.

Zircon typically occurs as anhedral, short and stubby grains (length to width ratios: 1–2) with rounded terminations, and vary in size from 50 to 200 μm . Some grains are long prismatic and subhedral with length to width ratios up to 4. Most crystals are colourless and clear with subordinate inclusions and cracks. In CL and BSE they show igneous oscillatory zoning, and commonly have a 5–25 μm wide, CL-dark and BSE-bright outermost rim (Fig. 8a,b).

Six concordant analyses in oscillatory zoned and BSE-dark domains gave a concordia age of 1686 ± 12 Ma (MSWD = 2.5; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1691 ± 21 , MSWD = 2.5). U-concentrations are 90–300 ppm and Th/U-ratios 0.5–1.5. Four analyses mixing rim and core domains (n1759-02a, 04a, 06a & 08a) and two discordant analyses (01a & 03a) were discarded from age calculation. The mixed age (1484 ± 28 Ma) from one of the displaced analyses (02a) suggests that the thin BSE-bright rims represent a distinctly younger generation.

Migmatitic gneiss: leucosome

The leucosome (sample HB-2) is greyish red, medium-grained, granitic in composition and has a linear fabric. It is dominated by microperthitic K-feldspar, plagioclase (partly antiperthitic) and quartz (Table 1).

The external morphology of zircon is essentially similar to that of zircon in the mesosome, except for the occurrence of reddish hematite staining, mainly along micro-cracks and along core–rim boundaries. Most crystals have, however, voluminous CL-dark and BSE-bright domains, up to 100 μm wide, surrounding a remnant core (Fig. 8c). The boundary between rim and core appears transitional and the rims typically show an internal texture that resembles a relict oscillatory zoning (Fig. 8d). In some grains, the BSE-bright rims are homogeneous and lack remnants of igneous zoning (Fig. 8e). A few sections of grains lack remnant cores (Fig. 8f). Approximately two-fifths of the zircon domains are texturally young and BSE-bright.

Four analyses (n1757-06a, 12a, 15a & 24a) from core domains with apparently intact igneous oscillatory zoning gave a concordant age of 1693 ± 22 Ma (MSWD = 0.20; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1697 ± 27 Ma, MSWD = 1.9). U-contents are 40–150 ppm and Th/U ratios are 0.5–1.2. Three analyses in cores with slightly blurred oscillatory zoning yielded younger ages (n1757-03a, 09a & 10a), interpreted as affected by post-crystallization Pb-loss (excluded from age calculation).

Eighteen concordant analyses of BSE-bright zircon domains gave a range of $^{207}\text{Pb}/^{206}\text{Pb}$ ages between 1482 ± 18 and 1414 ± 14 Ma (Table S1). This type of zircon is rich in U (800–2800 ppm) and has low Th/U ratios (0.02–0.17). The analytical dataset for the secondary zircon domains show a too wide spread for calculation of a concordia age. A weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of all concordant analyses of the secondary domains give an age of 1444 ± 10 Ma (MSWD = 4.6). In the Tera-Wasserburg plot (Fig. 8g), however, two subgroups may be recognized in the analytical data set. A statistical cluster analysis of the same data confirmed these two subgroups, where the first ($n = 11$) has higher $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ ratios (Table S2) and consistently very low Th/U ratios (< 0.07). Comparison of the two data groups with the images reveal that the first is composed of both ghost zoned ($n = 6$) and unzoned ($n = 5$) secondary zircon domains. The second group ($n = 7$) is with the exception of one analysis composed of unzoned zircon. The groups give weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 1462 ± 6 (MSWD = 0.78) and 1425 ± 7 (MSWD = 0.55), respectively.

One analysis of a BSE-bright, ~ 35 μm wide rim (n1757-23a) with a high U-content (1361 ppm) and a low Th/U ratio (0.09), gave a $^{207}\text{Pb}/^{206}\text{Pb}$ age that is within error identical to the age of BSE-dark, oscillatory zoned cores. This analysis was excluded from age calculations.

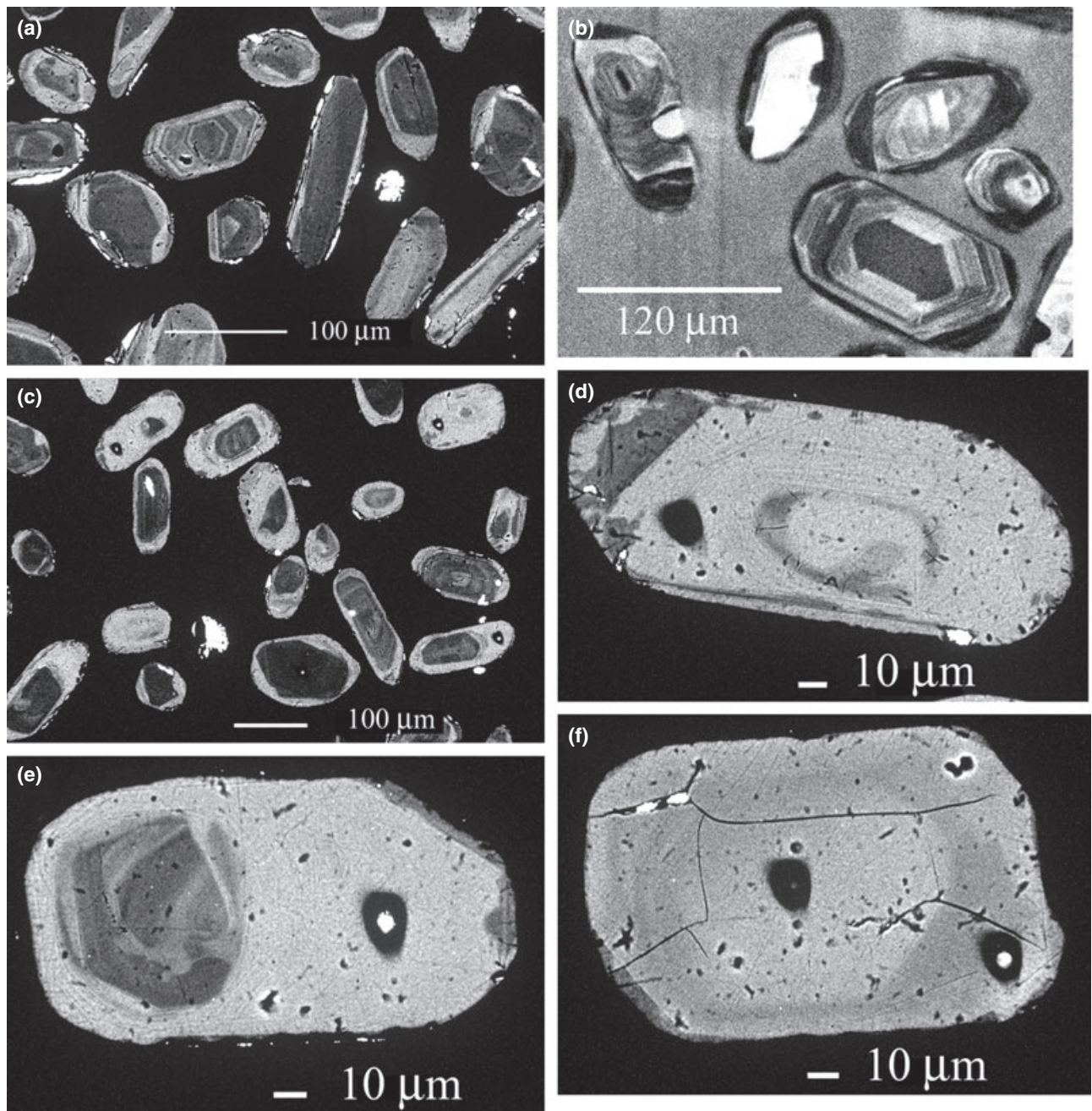
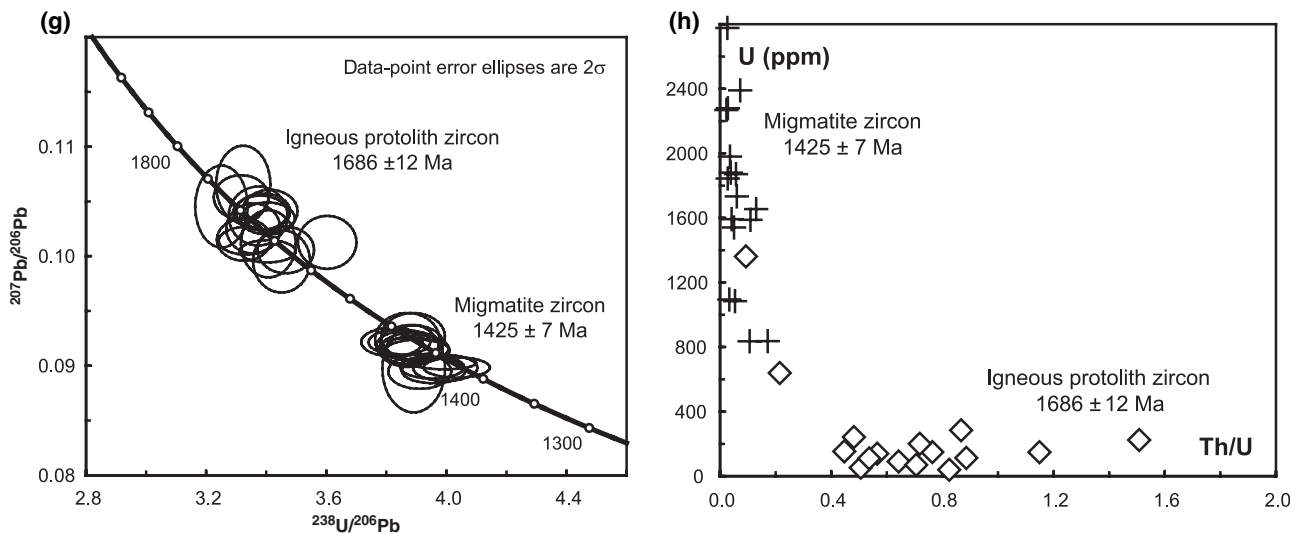


Fig. 8. Electron microscope images and U–Pb–Th ion microprobe analytical data of zircon in migmatite gneiss from Högbjör (samples HB-1 and HB-2). In the BSE-images, the analysed spots occur as black areas with or without a white centre. Very bright spots are remnants of gold coating. (a) BSE-image of zircon in mesosome (HB-1), showing dark and oscillatory zoned igneous crystals with thin, light rims. (b) CL-image of zircon in mesosome. The crystals are relatively brightly luminescent and have up to 15 μm wide dark rims. (c) BSE-image of zircon in leucosome (HB-2). Secondary, BSE-bright rims surround igneous zircon cores. (d) Composite crystal from leucosome, made up of BSE-bright secondary zircon with oscillatory 'ghost zoning' and a resorbed core of igneous zircon (analysis n1757-22a). (e) Composite crystal from leucosome with a comparatively homogeneous, BSE-bright rim and a core of protolith zircon (n1757-05a). (f) Simple secondary zircon crystal (n1757-13a and b). (g) Inverse concordia (Tera Wasserburg) diagram showing zircon analyses from gneiss mesosome and leucosome. Analyses cluster at 1.72–1.64 Ga (igneous protolith zircon, both samples) and 1.48–1.41 Ga (secondary BSE-bright zircon in leucosome). Three analyses, representing mixed domains, have been excluded for clarity. See text for details. (h) U v. Th/U diagram. Diamonds = 1.69 Ga protolith zircon and crosses = 1.42 Ga migmatite zircon. See text for details.

Fig. 8. *Cont'd.*

Folded metagranitic dyke

The folded, metagranitic dyke (HB-3, cf. Fig. 7c) has an originally medium- to coarse-grained fabric that has been strongly stretched and recrystallized.

Zircon crystals in the dyke are subprismatic to oval with rounded terminations. Igneous zircon domains with oscillatory zoning are present (Fig. 9a) as well as BSE-bright domains (Fig. 9b) of similar appearance to the rims in leucosome sample HB-2. Most grains are complex and few domains are homogeneous and large enough for analysis, which is why the data set is limited. Seven analyses of igneous zircon domains (from four crystals) and eight analyses of CL-dark and BSE-bright zircon (from eight crystals) were acquired (Table S2).

Eight analyses of BSE-bright domains, similar in appearance to the BSE-bright zircon in the leucosome sample HB-2, spread along the concordia between 1457 ± 9 and 1411 ± 6 Ma (Fig. 9f, weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1434 ± 12 Ma, MSWD = 3.3). This zircon has low to moderate Th/U ratios (0.03–0.45) and moderate to high contents of U (650–2800 ppm, Fig. 9e), similar to zircon rims in the gneiss leucosome (cf. sample HB-2, Fig. 8d). Thin, BSE-bright domains, up to $25 \mu\text{m}$ wide, that occur along the rims of the oldest igneous zircon grains (below) are probably of the same generation.

Seven analyses from four igneous zircon crystals range from 1491 ± 12 to 1377 ± 17 Ma. All are low in U (150–370 ppm), with Th/U ratios between 0.2 and 0.6 (Fig. 9e). The two older grains (with thin BSE-bright rims) yield a concordia age at 1469 ± 14 Ma ($n = 4$, MSWD = 2.4, Fig. 9d; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1477 ± 26 Ma, MSWD = 1.7). The two younger grains yield a concordia age at 1394 ± 12 Ma (2σ , $n = 3$, MSWD = 0.35, Fig. 9c; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1399 ± 47 Ma, MSWD = 1.5).

Undeformed and cross-cutting granitic dykes

Two crosscutting and undeformed granitic dykes were dated. One is 1–2 dm wide, pink, fine- to coarse-grained and has an isotropic fabric (HB-4, Fig. 7f) with microcline, plagioclase, quartz and lesser amounts of hornblende, biotite, euhedral titanite, and zircon. The second dyke (HB-6) is less than 10 cm wide but can be followed for ~ 100 m. It is medium- to coarse-grained, isotropic and consists of grey quartz, red K-feldspar and white plagioclase, and some biotite and magnetite.

Zircon crystals in sample HB-4 are up to $200 \mu\text{m}$ long and euhedral. BSE-images show that the crystals are composite. Distinct, rounded cores with variable textural characteristics (BSE-dark, BSE-bright, irregularly and oscillatory zoned) are surrounded by euhedral, unzoned rims, up to $50 \mu\text{m}$ wide and medium to light grey in BSE-image (Fig. 10a). Seven analyses in euhedral rims gave a concordia age of 945 ± 7 Ma (2σ , MSWD = 1.6, Fig. 10b; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 939 ± 10 Ma, MSWD = 1.2). The zircon rims are characterized by moderate U-contents (300–400 ppm) and Th/U-ratios between 0.10 and 0.28 (Fig. 10c, Table S2).

More than 60% of the zircon grains in sample HB-6 are similar to the composite grains in HB-4. The remaining crystals are up to 0.5 mm large, sharp-edged fragments of large grains and smaller, euhedral crystals, both of which are homogeneous and medium light in BSE (Fig. 10d). Five spots in euhedral, texturally homogenous grains yielded a concordia age of 952 ± 7 Ma (2σ , MSWD = 2.6, Fig. 10e; weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age = 946 ± 8 Ma, MSWD = 0.8). U-contents (530–610 ppm) and Th/U-ratios (0.17–0.21) are similar to the values in sample HB-4 (Fig. 10c).

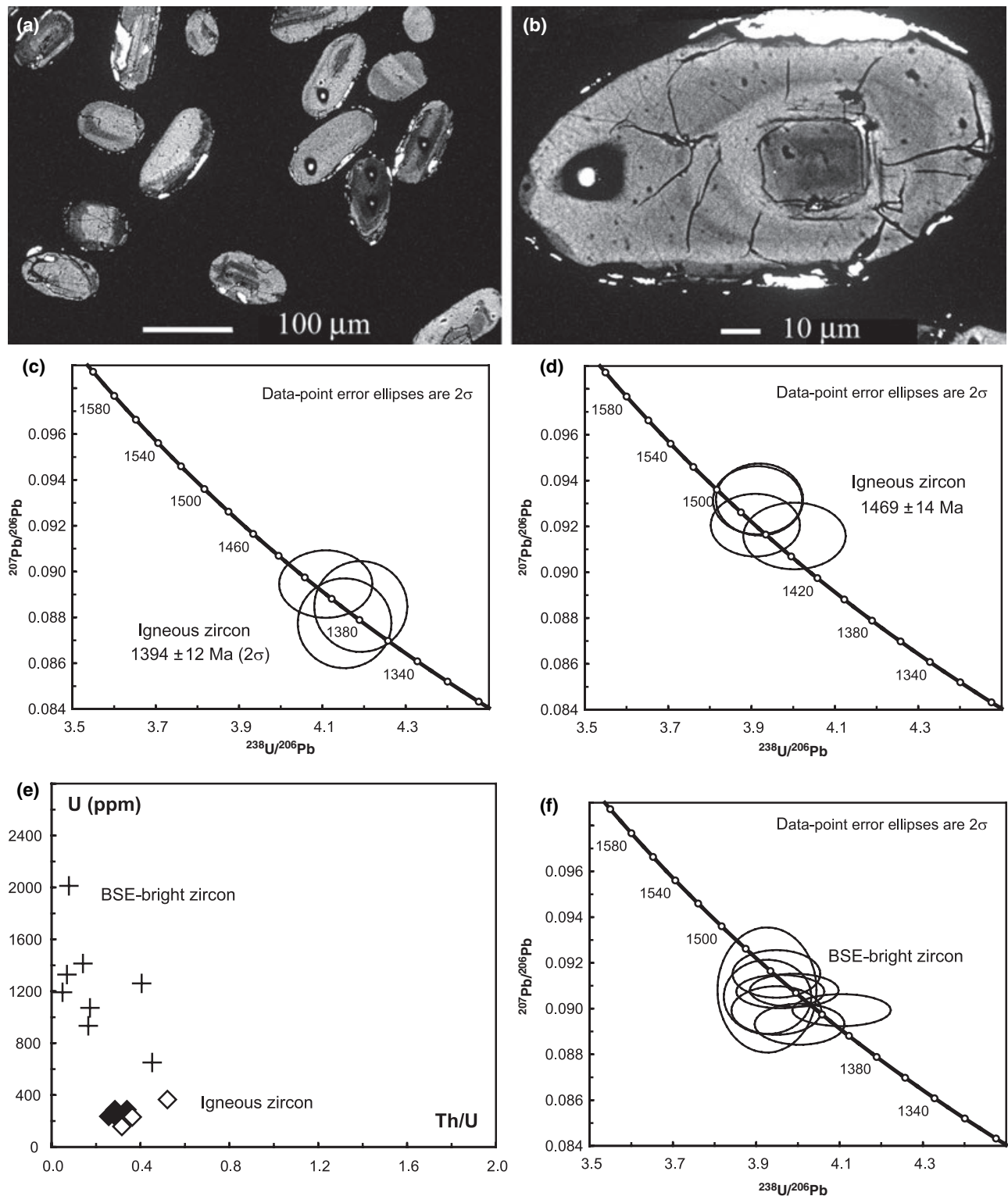


Fig. 9. BSE-images and U–Pb–Th ion microprobe analytical data of zircon in the folded metagranitic dyke (sample HB-3) at Högbjör. Black spots with white centres are analysis craters. Very bright spots are remnants of gold coating. (a) Overview, illustrating the variety and complexity of the zircon crystals with igneous domains and BSE-bright secondary domains. (b) Dominantly BSE-light zircon (analysis n1756-03a, $^{207}\text{Pb}/^{206}\text{Pb}$ age = 1424 ± 9 Ma) with dark core remnant. (c) Inverse (Tera Wasserburg) concordia diagram with analyses of igneous 1.40 Ga zircon. (d) Inverse concordia diagram with analyses of igneous 1.47 Ga zircon. (e) U v. Th/U diagram. Filled diamonds = 1.47 Ga igneous domains, open diamonds = 1.40 Ga igneous zircon, crosses = BSE-bright secondary zircon. (f) Inverse concordia diagram with analyses of BSE-bright secondary zircon (1457–1411 Ma, no concordia age).

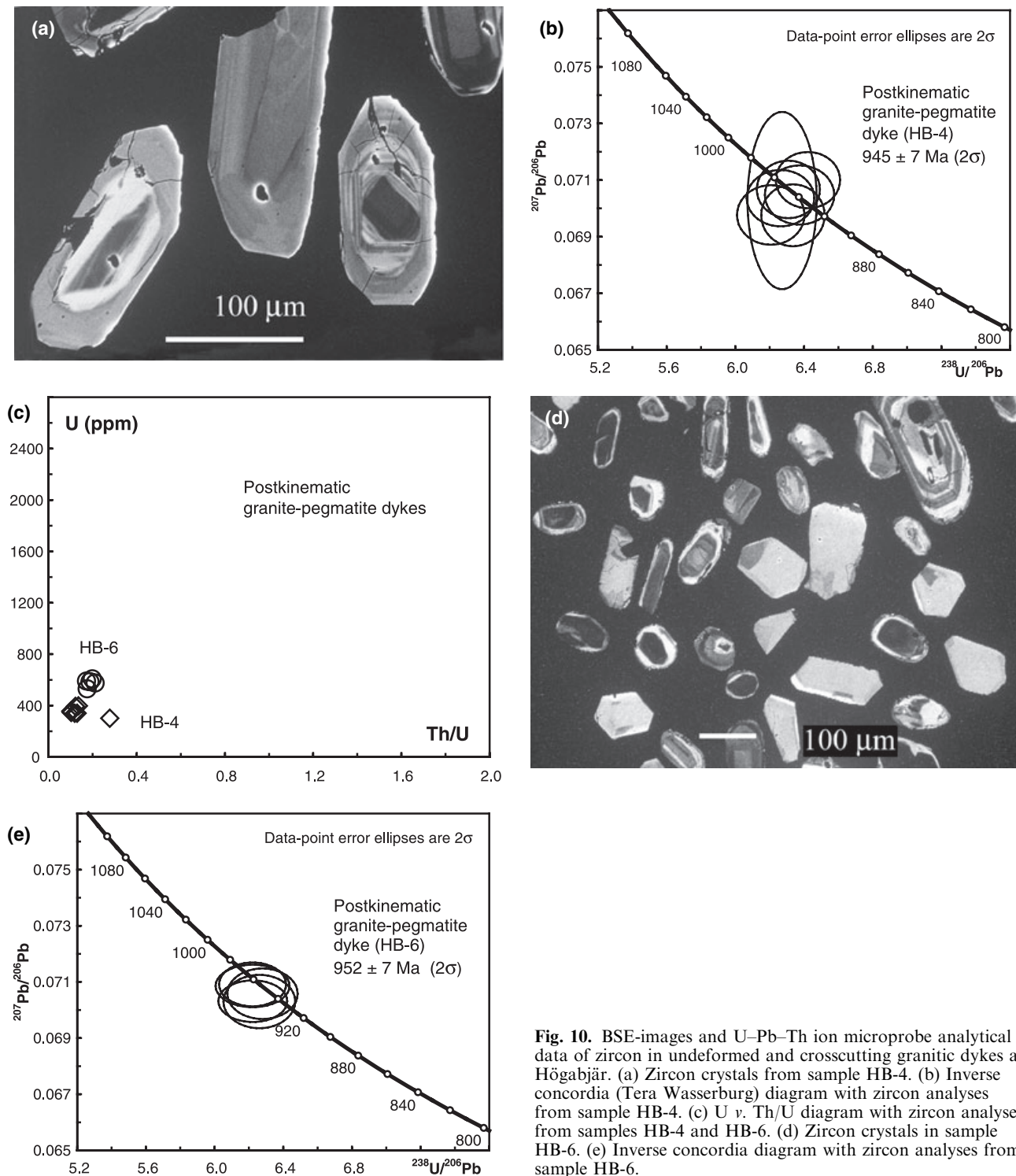


Fig. 10. BSE-images and U–Pb–Th ion microprobe analytical data of zircon in undeformed and crosscutting granitic dykes at Högabjär. (a) Zircon crystals from sample HB-4. (b) Inverse concordia (Tera Wasserburg) diagram with zircon analyses from sample HB-4. (c) U v. Th/U diagram with zircon analyses from samples HB-4 and HB-6. (d) Zircon crystals in sample HB-6. (e) Inverse concordia diagram with zircon analyses from sample HB-6.

Summary and interpretation

CL-luminescent and BSE-dark and/or oscillatory zoned zircon in the migmatite gneiss is interpreted as primary igneous zircon. It is present in both mesosome and leucosome; zircon from the mesosome yielded an

age of $1686 \pm 12 \text{ Ma}$ for igneous crystallization of the protolith (Fig. 8g). This zircon (one analysis excepted) is characterized by high Th/U-ratios (0.2–1.6) and U-contents lower than 650 ppm (Fig. 8h), distinctly different from the BSE-bright rims, which are U-enriched (800–2800 ppm) and have low Th/U ratios (0.02–0.17).

The rounded edges and tips of zircon, in both mesosome and leucosome, indicate resorption (by melt). The igneous zircon in the mesosome has, however, retained its primary character in terms of U–Pb–Th isotopic signature and largely undisturbed oscillatory zoning despite the high-grade metamorphism. In the leucosome, by contrast, secondary zircon makes up 40% or more of the total zircon volume and igneous zircon is present as remnant cores surrounded by secondary rims up to 100 μm wide. The consumption of igneous zircon and formation of secondary zircon is thus intimately connected with the migmatization process.

The BSE-bright secondary domains in the leucosome are either ‘ghost zoned’ or unzoned; in grains that host both types the ‘ghost-zoned’ domains usually form an inner zone. This textural relation and the apparent transitional boundaries between secondary and igneous zircon suggest that the secondary domains formed by *in situ* recrystallization rather than by precipitation from a melt.

Concordant analyses yielded a weighted average $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1444 ± 10 Ma but two subgroups are identified at 1462 ± 6 and 1425 ± 7 Ma. The older group includes both types of textural domains whereas six out of seven analysed domains are unzoned in the younger group. Therefore, we suggest that the older ages reflect incomplete isotopic resetting during recrystallization, a feature observed in zircon from other localities (Rimsa *et al.*, 2007). If so, the younger age of 1425 ± 7 Ma ($n = 7$) is the ‘best’ age for migmatization. Within this group there is no scatter towards lower ages indicating post-anatectic Pb-loss. However, we cannot rule out that part of the analytical spread along the concordia for the igneous and migmatite zircon generations (Fig. 8g) was caused by later Pb-diffusion, if so probably during Sveconorwegian high-pressure granulite facies metamorphism at *c.* 970 Ma. In their study of amphibolite and granulite facies zircon from the Ivrea Zone, Vavra *et al.* (1999) considered the weighted mean of the highest apparent ages as the real age of zircon growth during anatexis. Based on the combination of textural and isotopic data, the younger ages were interpreted as modified by post-anatectic Pb-loss during later events. In the investigated rocks at Högbjörn it is extremely difficult to evaluate possible Pb-loss effects on the recrystallized zircon domains as they do not offer the textural control as, for example, a primary zoning would. This is a problem that should be generally acknowledged in dating of early metamorphic zircon generations in polymetamorphic rocks. It is also noteworthy that no Sveconorwegian secondary zircon has been found in the gneiss and metagranite samples, despite the proven overprint of Sveconorwegian high-pressure granulite facies metamorphism in the region; this issue is treated below.

Zircon from the folded metagranite dyke yielded an age range between 1491 ± 12 and 1377 ± 17 Ma and

more data are required to reach satisfactory conclusions. The oldest (1469 ± 14 Ma), BSE-dark, oscillatory zoned zircon grains are interpreted as igneous in origin. These and the grains that are dominantly U-rich (1457 ± 9 to 1411 ± 6 Ma) are interpreted as xenocrysts. The youngest, 1394 ± 12 Ma, BSE-dark, oscillatory zoned zircon is interpreted as igneous and its significance is open for two different interpretations. It may represent zircon formed in the melt that fed the granitic dyke, in which case the 1394 ± 12 Ma age is an approximate age of the dyke intrusion. If, on the other hand, these zircon grains are also xenocrysts the age represents an upper bracket only for the intrusion of the dyke.

In the two crosscutting granitic dykes, the majority of zircon is variably resorbed, xenocrystic grains, distinctly overgrown by euhedral, comparatively homogeneous rims. Based on their euhedral habit and textural position as outermost rims and single crystals, these domains are interpreted as crystallized directly from the granitic melt. In both dykes, this zircon has moderate U-contents and Th/U-ratios, and yielded ages of 952 ± 7 and 946 ± 8 Ma, respectively.

Geological interpretation

Igneous zircon in migmatitic gneiss (mesosome and leucosome) at Högbjörn dates the crystallization of the protolith at *c.* 1686 ± 12 Ma. The rock is thus yet an example of a strongly migmatized and deformed variety of the 1.73–1.66 Ga felsic intrusions that dominate the Eastern Segment.

The migmatization is dated at 1425 ± 7 Ma by the secondary zircon generation formed in the leucosome. The orientation of this leucosome defines a layering, but field relations clearly show that multiphase deformation has modified the original character and orientation of this migmatite structure. It is emphasized that, at this locality, the migmatite age is not a direct age of deformation but a bracket for the various deformation structures. Provided that the youngest generation of igneous zircon in one of these dykes, dated at 1394 ± 12 Ma, is not made up of xenocrysts, it is clear that the origin of the layering at Högbjörn is pre-Sveconorwegian. It is probable that the layered structure originated in connection with the 1425 ± 7 Ma migmatization and was later modified by post- 1394 ± 12 Ma deformation.

The age of the metagranitic dyke sets an upper bracket at 1394 ± 12 Ma for the tight folding along SSW-plunging axes (Fig. 7; and for the steep ESE-stretching). The absolute age of the fold phase is, however, not known and it may be as young as Sveconorwegian in age. The geological map (Fig. 6) shows that this fold phase has oriented the gneissic and migmatitic layering into a regionally relatively consistent NNE–SSW strike. Additional 1.40 Ga granite–pegmatite dykes are known from three other localities

NW of Halmstad (Söderlund, 1996; Christoffel *et al.*, 1999; Rimsa *et al.*, 2007) and at one of these localities the dyke is tightly folded (the locality investigated by Söderlund, 1996). ESE-stretching fabrics may be coeval with this fold phase or, considered more likely, developed in connection with a different phase of late Sveconorwegian folding, along WNW–ESE axes, at 976 ± 7 Ma (cf. locality 1 above). The ages of unmetamorphosed, undeformed and crosscutting granite–pegmatite dykes set lower brackets for these two ductile deformation phases at 952 ± 7 and 946 ± 8 Ma. Both generations of granitic dykes (1.40 & 0.95 Ga) are leucogranitic in composition and represent late- to post-orogenic melts.

DISCUSSION

Linking secondary zircon to migmatization and deformation

The results presented in this paper demonstrate polymetamorphic migmatization of the crust in the south-east Sveconorwegian Province, developed during two different orogenic episodes at *c.* 1425 ± 7 and 976 ± 7 Ma. The younger age can be directly linked to deformation, as the dated leucosome has a syndeformational structural relation to south-vergent folding (locality 1), whereas the older age obtained from deformed leucosome (locality 2) is not a direct age of deformation but a bracket for different deformation phases.

Zircon is a mineral that cannot be connected to a specific metamorphic stage or a deformation phase unless a petrogenetic link is provided. Whatever the target is for the dating, the choice of samples is therefore critical. The potential link between zircon and deformation is evident where partial melting can be linked to a phase of synchronous deformation. Requirements for successful dating are clear, syntectonic structural relations and samples of comparatively large leucosome (large enough to provide secondary zircon growth, cf. Oliver *et al.*, 1999); in our study 5–10 cm wide pockets were sufficient. Investigation of zircon in mesosome and leucosome from the same rock will provide the correlation between melt and zircon type.

Zircon characteristics and age determination

The character of migmatite zircon varies with bulk rock geochemistry and external factors. For example in metasedimentary rocks of the Reynolds Range in Australia, zircon overgrowths in mesosome and leucosome consistently have weak CL-zoning and Th/U-ratios < 0.1 (Rubatto *et al.*, 2001). By contrast, most secondary zircon in metasedimentary migmatites from the Ivrea zone is distinctly luminescent and zoned; some granulite facies secondary zircon has oscillatory growth zoning (Vavra *et al.*, 1999). Typical, however, for migmatite zircon formed by dissolution and re-

precipitation are euhedral crystal faces and formation as rims with distinct boundaries to resorbed cores of protolith zircon (and, to a lesser extent, as single grains; e.g. Vavra *et al.*, 1999; Rubatto *et al.*, 2001; Andersson *et al.*, 2002).

The data from migmatites in this study demonstrate that, in addition to possible minor neocrystallization, secondary zircon principally formed by recrystallization in leucosomes. This recrystallized zircon is different from migmatite zircon that has formed by dissolution and reprecipitation. It is remarkably U-rich (800–4400 ppm), commonly shows faint (relict) oscillatory zoning, and has transitional boundaries to the remnants of protolith zircon. The external morphologies of the zircon grains (50–150 μm long) show evidence of light resorption but have not been significantly modified. The processes behind recrystallization and U-enrichment in zircon, such as found in the presently studied migmatites, have not been fully explained (cf. Hoskin & Schaltegger, 2003, and references therein). Present knowledge is not sufficient to define the external conditions and driving forces for this recrystallization. Clues may possibly be found in the (near-) absence of dissolution and reprecipitation. Zircon solubility is commonly low in granitic melts, which is why large zircon grains commonly survive without complete dissolution (e.g. Watson, 1996). The limited amount of zircon resorption in the migmatites of this study demonstrates that zircon solubilities in the partial melts were very low. Furthermore, the lack of neocrystallization indicates that saturation conditions were close to equilibrium between zircon and melt. It is possible that under these conditions zircon instead reacts by recrystallization.

Micro-scale analyses of U-rich, recrystallized zircon were made both in domains with a faint oscillatory ghost structure and in homogeneous domains. The analyses from locality 2 (Högabjär) yielded a considerable age-spread, which suggests that some domains were either incompletely reset or subject to late Pb-loss, or both. The ‘ghost zoned’ secondary zircon domains are commonly located closest to the remnant core of individual zircon grains and, on average, yielded older ages than the homogeneous domains, and the older group was for this reason excluded from age calculation. It is probable that age determination of recrystallized zircon as far as possible should be restricted to homogeneous and structureless domains.

On the absence of secondary zircon

In rocks that host Zr-bearing phases other than zircon (i.e. basic and pelitic rocks) secondary zircon is likely to form during a first metamorphic event but less likely to form during later events. The reason for this is that once Zr, mobilized from reactant primary minerals, has been consumed and formed secondary zircon it is generally not available for further reaction. In silica-rich igneous rocks most Zr is from the start bound in

protolith zircon, which renders even the formation of a first generation of metamorphic zircon difficult. Only through a process such as partial melting that readily dissolves and re-precipitates zircon, or significantly enhances recrystallization of zircon, can secondary zircon form.

At locality 2 (Högbjör), Sveconorwegian secondary zircon was not detected despite the fact that Sveconorwegian temperatures were high enough for partial melting ($\sim 700^\circ\text{C}$). Christoffel *et al.* (1999) and Söderlund *et al.* (2002) have previously discussed an apparent absence of Sveconorwegian secondary zircon in rocks older than 1.4 Ga in the Eastern Segment. The present data, however, show that the majority of the polymetamorphic orthogneisses contain one generation of secondary zircon, either 1.43 or 0.97 Ga. Only a minority of rocks contain both secondary zircon generations [i.e. Mårdaklev gneiss of Söderlund *et al.*, 2002; Söndrum gneiss of Rimsa *et al.*, 2007; possibly also Oxanäset (locality 1), this study]. Significant for these rocks is that they are granitic and comparatively leucocratic in composition, and will relatively easily produce melt, and that one secondary zircon generation strongly dominates over the other.

A few different explanations are feasible for the polymetamorphic rocks that lack two generations of zircon. Partial melting may not have taken place due to lack of hydrous fluids, or partial melting may have produced very small volumes of zircon only due to small volumes of leucosome. A third explanation is that partial melting did take place but that melts escaped by the aid of simultaneous deformation. At Högbjör (locality 2), a certain amount of late, syn-deformational partial melting is evidenced by the presence of leucosome veinlets along axial surfaces of meso-scale, post-1394 $\pm$ 12 Ma folds (cf. Fig. 7d), structures common to large parts of the southern Eastern Segment. If the deformation promoted disequilibrium melting (cf. Sawyer, 1991) the melts would have been undersaturated in Zr and caused dissolution of pre-existing zircon, an explanation that is attractive for the resorbed zircon habit in the gneiss mesosome, gneiss leucosome and metagranitic dyke at Högbjör (locality 2). Regardless of the cause for the lack of Sveconorwegian zircon at this particular locality, it is emphasized that the absence of secondary zircon in a rock does not prove the absence of metamorphism. At Oxanäset (locality 1), the relations are the opposite: Sveconorwegian zircon dominates and the existence of a 1.4 Ga zircon generation is only weakly indicated. These data demonstrate a high degree of partial melting in Sveconorwegian time but, for the reasons discussed above, do not rule out early metamorphism at 1.43 Ga.

Late-orogenic granitic dykes as structural age markers

Our study exemplifies that discordant or semi-discordant, granitic or pegmatitic dykes can be used as

structural age brackets, but some requirements must be met. First, in the cases of deformed discordant dykes, pre-intrusive and post-intrusive features must be critically distinguished. If deformation and metamorphism have affected the rocks after dyke intrusion, the present geometry, metamorphic assemblage, texture, and deformation fabrics in the host rock will be different from those at the time of dyke intrusion. Second, the abundance of zircon xenocrysts and complex zircon grains in this type of dykes requires micro-scale dating techniques. Even if analyses are obtained from several types of zircon domains in the dykes it may still be difficult to verify whether the youngest zircon generation precipitated from the melt itself or represents xenocrysts.

The melts from which the late-orogenic dykes crystallized were most likely extracted from high-temperature metamorphic sources beneath the investigated level, and apparently efficiently extracted zircon crystals from their sources. The assembly and escape of melt indicate that the lower, high-temperature levels still were undergoing ductile deformation at the time of melt extraction (cf. e.g. Sawyer, 1991; Brown *et al.*, 1995, and references therein).

Tectonic implications pertaining to the Sveconorwegian orogeny

Analyses from locality 1 (Oxanäset) define late Sveconorwegian migmatization and synchronous folding in the Eastern Segment at 976 ± 7 Ma. The age is within error identical with 0.98–0.96 Ma ages of migmatization and metamorphism in the upper amphibolite and high-pressure granulite facies metamorphism in other parts of the southern Eastern Segment (Andersson *et al.*, 1999, 2002; Söderlund *et al.*, 2002, and references therein).

The age of migmatization and folding is also within error identical to the age of 972 ± 14 Ma for prograde eclogite metamorphism (Johansson *et al.*, 2001) recorded from eclogite remnants situated 30–50 km west of the Oxanäset locality. The retrogressed eclogites occur as a dismembered and amphibolitized, up to kilometre-thick unit (Möller, 1998, 1999). It is overlain and underlain by lower-pressure rocks and thereby clearly a tectonically emplaced unit. It is also strongly affected by E–W stretching and upright to roughly south-vergent folding (Fig. 11). The folding and migmatization exemplified by locality 1 (Oxanäset) in this paper must therefore have taken place in close connection to the emplacement of the eclogites. Amphibolite facies assemblages and garnet breakdown textures in the metabasites at locality 1 suggest that the E–W folding, migmatization and associated stretching formed during decompression and/or cooling. It is proposed that the folding took place in response to N–S shortening and E–W extension, shortly after the emplacement of the eclogites, in early stages of uplift of the high-grade, southern parts of the Eastern Segment.

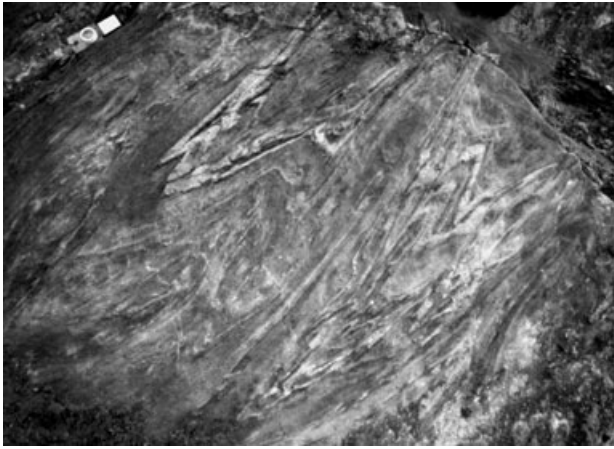


Fig. 11. Retrogressed and tightly folded Sveconorwegian eclogite. Felsic veins are folded and occur oriented along axial-planar surfaces. Compass, 18 cm long, for scale.

The same structural components (ductile foliation, E–W stretching lineation, folding along E–W-trending axes, migmatization) have been documented in Mesoproterozoic, 1.40–1.38 Ga and *c.* 1.20 Ga, granitic and quartz-monzonitic augen gneisses (Berglund, 1997; Andersson *et al.*, 1999; M. Ekdahl, unpublished data). Similarly, however lower-grade (amphibolite to greenschist facies), Sveconorwegian structures characterize a shear zone immediately north of Lake Vänern (Fig. 1, Berglund *et al.*, 1997). Collectively, these data suggest that the E–W folding expressed in the aeromagnetic map pattern (Fig. 2) is Sveconorwegian and took place during amphibolitization and initial exhumation of the southern parts of the Eastern Segment.

The regular pattern of upright to overturned E–W folds suggests deformation in response to N–S compressional forces simultaneous with extension in E–W direction. North–south thrusting and extrusion of eclogite (and possibly other units) is suggested, followed by tight to open folding and high-grade metamorphism in a similar stress field. Unroofing was likely aided by E–W extension and vertical stripping. North–south shortening at 976 ± 7 Ma is a different Sveconorwegian tectonic scenario than commonly envisaged (e.g. Bingen *et al.*, 2005 and references therein) and needs to be considered in reconstruction of the Sveconorwegian orogen and in correlation with tectonic counterparts in the Grenville Belt.

East of the Protogine Zone Sveconorwegian shortening has not affected the bedrock. If a scenario with Sveconorwegian N–S compression is correct, this force must have been taken up as strike-slip strain along the Protogine Zone. Modern analogous, large-scale examples are the strike-slip zones bounding present India to the east and west.

Mesoproterozoic Hallandian orogenesis

Identification of pre-Sveconorwegian structures and metamorphism in the Sveconorwegian Province

Early migmatization (locality 2, Högbjär) was dated at 1425 ± 7 Ma. It is interpreted to have taken place roughly simultaneously with deformation and produced a coarse layered structure with concordant leucosome. This structure was severely modified during two phases of deformation after 1394 ± 12 Ma. Similar structural relations characterize the south-westernmost portions of the Eastern Segment (Figs 2 & 6; cf. Hubbard, 1975; Söderlund, 1996; Christoffel *et al.*, 1999). Also from other parts of the Eastern Segment there is chronological evidence of Mesoproterozoic thermal reworking (Söderlund *et al.*, 2002; Hegardt *et al.*, 2005). In previous studies, links between 1.46 and 1.42 Ga secondary zircon formation and the metamorphic process were not uniquely established. The conformity of these age data suggests, however, that migmatite-associated structures developed on a regional scale during a tectonometamorphic event at *c.* 1.43 Ga.

An orogenic model has long been maintained in which the migmatitic and regional deformation structures in the Eastern Segment have been interpreted as older than 1.55 Ga (called Gothian, M1/M2 and D1/D2 events). The interpretation has largely relied on field relationships with semi-discordant, deformed and metamorphosed granitic and mafic dykes, similar to that of the 1394 ± 12 Ma dyke described in the present paper (Larson *et al.*, 1986; Åhäll, 1995; Connelly *et al.*, 1996; Christoffel *et al.*, 1999; and references in these papers). The identification of Sveconorwegian high-grade metamorphism and deformation challenged this model and the lack of direct age determinations supporting the Gothian model has been emphasized (cf. Möller, 1998, 1999; Andersson *et al.*, 1999, 2002; Söderlund *et al.*, 2002; and references in these papers). The available data, including the ages of migmatization in the present study, suggest that both Sveconorwegian and pre-Sveconorwegian migmatitic and deformation structures are present on a regional scale, but that the pre-Sveconorwegian migmatitic structures developed during orogenesis at 1.43 Ga. Primary *c.* 1.69 Ga igneous structures, such as compositional layering, mingling structures, synmagmatic mafic dykes and structures formed by syn-emplacement deformation, must be recognized as distinct from the 1.43 Ga structures. Evidence of tectonometamorphic reworking prior to 1.48 Ga has not been detected in our study or, to our knowledge, in other studies of the Eastern Segment.

It is important to recognize that Sveconorwegian high-grade metamorphism and ductile deformation have everywhere overprinted the pre-Sveconorwegian structures to some extent. Pre-Sveconorwegian mineral assemblages have undergone complete or partial

recrystallization at high temperatures and mineral equilibria have been changed. Where the Sveconorwegian deformational overprint is of moderate intensity the Mesoproterozoic structures can be identified (locality 2), however reoriented, partly transposed and with overprinting fabrics. In other places the Sveconorwegian deformational overprint has been intense and/or accompanied by significant degrees of partial melting (locality 1). As exemplified by the present study, distinction between 1.44–1.42 and 0.98–0.96 Ma migmatite structures must be aided by age determination.

The Hallandian orogeny: definition and regional context

Compared to the younger Sveconorwegian orogenic events, relatively little is known of the effects of Mesoproterozoic tectonometamorphism on different parts of the southern Baltic Shield. However, data presented in this study and previous publications demonstrate an orogenic event east of the Mylonite Zone (Fig. 1) involving regional metamorphism, migmatization and deformation at *c.* 1.43 Ga.

Following the pioneering work of Hubbard (1975), metamorphism dated at 1.46–1.42 Ga was referred to as 'Hallandian' by Christoffel *et al.* (1999) and Söderlund *et al.* (2002). Hubbard (1975) originally coined the term for isolated pyroxene-bearing granite–monzonite magmatism (1.40–1.38 Ga) and associated and overprinting high-grade metamorphism, which was later dated as Sveconorwegian (Johansson *et al.*, 1991). We concur with previous suggestions that the term Hallandian be used for the *c.* 1.43 Ga regional metamorphism as well as for the younger 1.40–1.38 Ga intrusions, for two main reasons. First, the term Hallandian has been used for high-grade metamorphism and granite–monzonite intrusions in the region for more than 30 years. Second, structural relations, petrology and geochronology for these events are characterized, linked and well exposed in the region where the term was coined.

Due to the Sveconorwegian overprint, further characterization in the Eastern Segment of map-scale structural patterns and *P–T* conditions related to the Hallandian orogeny is difficult. For this purpose it is of interest to move outside the Sveconorwegian Province, east of the Protogine Zone. Recent studies have proposed that 1.45 Ga granite magmatism in the southeasternmost Baltic Shield took place in response to convergent tectonism (Bogdanova *et al.*, 2006 and references therein), and thin overgrowths on zircon and resetting of the U–Pb system in titanite at 1.45–1.40 Ga have been documented in older rocks (Johansson *et al.*, 2006). Direct connections between 1.45 Ga magmatism and regional metamorphism and deformation structures are, however, still to be confirmed. If *c.* 1.43 Ga tectonometamorphic features can be identified in the south-eastern Baltic Shield, they will most likely be genetically connected with the

Hallandian orogeny in southwest Sweden. Any correlations across the Protogine Zone must, however, consider significant post-1.4 Ga tectonic displacements of the crust within the Sveconorwegian Province, displacements that did not significantly affect the crust east of the zone.

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SUPPLEMENTARY MATERIAL

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Table S1. SIMS U–Pb–Th data of zircon from migmatite gneisses at localities 1 and 2, Oxnäset and Högabjär.

Table S2. SIMS U–Pb–Th data of zircon from granitic dykes at locality 2, Högabjär.