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Mid-Proterozoic magmatic arc evolution at the southwest margin of the Baltic Shield☆

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Abstract

Mid-Proterozoic calc-alkaline granitoids from southern Norway, and their extrusive equivalents have been dated by LAM-ICPMS U–Pb on zircons to ages ranging from 1.61 to 1.52 Ga; there are no systematic age differences across potential Precambrian terrane boundaries in the region. U–Pb and Lu–Hf data on detrital zircons from metasedimentary gneisses belonging to the arc association show that these were mainly derived from ca. 1.6 Ga arc-related rocks. They also contain a minor but significant fraction of material derived from (at least) two distinct older (1.7–1.8 Ga) sources; one has a clear continental signature, and the other represents juvenile, depleted mantle-derived material. The former component resided in granitoids of the Transscandinavian Igneous Belt, the other in mafic rocks related to these granites or to the earliest, subduction-related magmatism in the region. Together with published data from south Norway and southwest Sweden, these findings suggest that the western margin of the Baltic Shield was the site of continuous magmatic arc evolution from at least ca 1.66 to 1.50 Ga. Most of the calc-alkaline metaigneous rocks formed in this period show major- and trace-element characteristics of rocks formed in a normal continental margin magmatic arc. The exceptions are the Stora Le-Marstrand belt in Sweden and the Kongsberg complex of Norway, which have an arc-tholeiitic chemical affinity. The new data from south Norway do not justify a suggestion that the crust on the west side of the Oslo Rift had an early to mid-Proterozoic history different from the crust to the east. Instead, they indicate that the different parts of south Norway and southwest Sweden were situated at the margin of the Baltic Shield throughout the mid-Proterozoic. Changes from arc tholeitic to calc-alkaline magmatism reflect changes with time in the subduction zone system, or lateral differences in subduction zone geometry. The NW American Cordillera may be a useful present-day analogue for the tectonomagmatic evolution of the mid-Proterozoic Baltic margin.

Keywords: Baltic Shield; Magmatic arc; Mesoproterozoic; U–Pb; Lu–Hf; Laser ablation-ICPMS

1. Introduction

The Baltic Shield (Fig. 1, inset) is composed of an Archaean core in the northeast, and progressively younger Proterozoic crustal domains towards the southwest (e.g. Gaàl and Gorbatschev, 1987). A substantial part of the regional protolith of the south-

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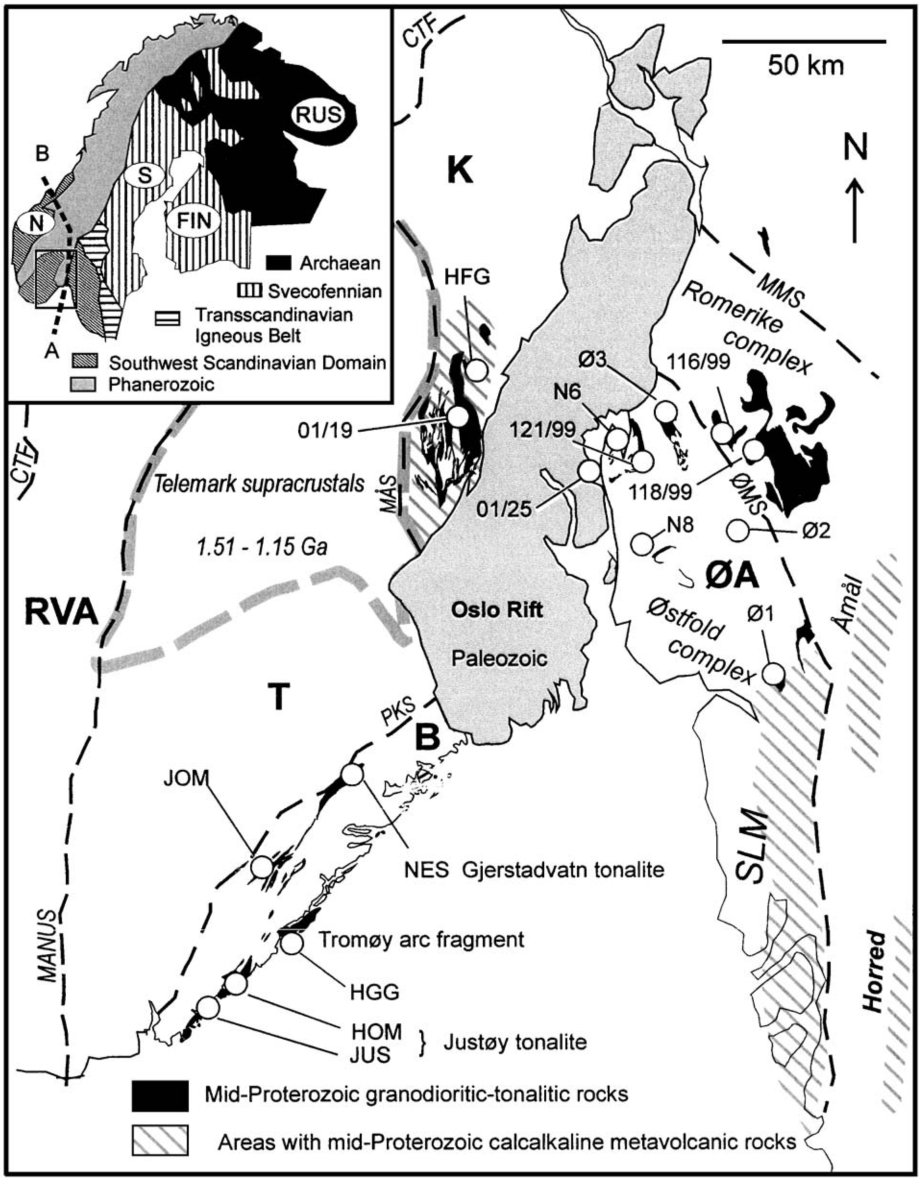


Fig. 1. Simplified geological map of southeast Norway and adjacent parts of Sweden, showing the distribution of mid-Proterozoic (Gothian) calc-alkaline metaigneous rocks, and the position of the samples of the present study. Major crustal domains: T: Telemark sector, K: Kongsberg sector, B: Bamble Sector, RVA: Rogaland-Vest Agder sector, ØA: Østfold–Akershus sector. Precambrian shear zones: PKS: Porsgrunn-Kristiansand shear zone (and Paleozoic brittle fault), KTB: Kongsberg-Telemark boundary, MANUS: Mandal-Ustaoset shear zone, ØMS: Ørje Mylonite Zone, MMS: Mjøsa-Magnor shear zone (continues as the "Mylonite zone" in Sweden). Areas with preserved calc-alkaline metavolcanic rocks in SW Sweden: H: Horred, Å: Åmål, SLM: Stora Le-Marstrand belt. The inset shows the main features of the Baltic Shield, the broken line AB marks a possible mid-Proterozoic suture between the Baltic Shield and a hypothetical "SW Norwegian Craton" (Åhäll and Gower, 1997; Åhäll et al., 1998). For samples, see Table 1.

westernmost part of the shield, the Southwest Scandinavian Domain (abbreviated SSD) may have formed during the controversial, mid-Proterozoic event known as the Gothian orogeny, in the interval ca. 1.75 to 1.5 Ga (e.g. Gaàl and Gorbatschev, 1987). According to a much-debated tectonic model for the mid-Proterozoic SW margin of the Baltic Shield, fragments of magmatic arcs of contrasting geochemical character have been accreted onto the Baltic shield in distinct events in the period 1.66 to 1.59 Ga (Åhäll and Gower, 1997; Åhäll et al., 1995, 1998, 2000; Brewer et al., 1998). This period of accretionary orogeny was terminated by merging of a continental fragment, making up present-day southern Norway west of the Oslo Rift (the "SW Norwegian craton"), with the Baltic Shield (Berthelsen, 1980). This final accretion event has been tentatively dated to ca. 1.58 Ga (Åhäll et al., 1998), 1.55 Ga (Åhäll and Larson, 2000) or 1.55 to 1.50 Ga (Åhäll et al., 2000). A central feature of this model is that the area west of the "possible suture line" AB in Fig. 1 (inset) is an exotic continental terrane, whose internal history and compositional character should be different from that of the Baltic Shield sensu stricto.

This model appears to be in conflict with the finding that clastic metasediments in south Norway west of the Oslo Rift show an important contribution from 1.70 to 1.90 Ga source rocks, i.e. similar to rocks found today in easternmost south Norway and in central Sweden (Knudsen et al., 1997; Haas et al., 1999; Bingen et al., 2001). However, this cannot be seen as conclusive evidence against an exotic origin for south Norway, as the sedimentary sequences studied were deposited after the proposed docking event, and transport of clastic material over great distances cannot be ruled out. Furthermore, few of the provenance-ages have been supported by radiogenic isotopes or other data on the rocks studied by SIMS, which could help to identify the source terranes as belonging to the Baltic Shield.

Perhaps more significantly, Andersen et al. (2001a), using late Proterozoic granitic intrusions as indicators of the composition of the unexposed, continental crust at ca. 930 Ma, found no evidence of lateral compositional differences in the deep crust that could be attributed to the existence of an exotic "SW Norwegian Craton". Recently published hafnium isotope data confirm the presence of early Proterozoic rocks in central and western south Norway, whose Hf isotope signatures are indistinguishable from those of pre-Gothian rocks of the Baltic Shield. Such rocks occur both in the deep crust, and in the once exposed source terranes of Proterozoic clastic metasediments (Andersen et al., 2002a).

The present knowledge of the mid-Proterozoic evolution of south Norway is to a large extent based on indirect evidence, mainly from rocks formed well after the event in question. However, the products of mid-Proterozoic arc-related magmatism are quite abundant in southern Norway, both east and west of the Oslo Rift (e.g. Jacobsen and Heier, 1978; Skjernaa and Pedersen, 1982; Graversen, 1984; Graversen and Pedersen, 1999; Berthelsen et al., 1996; Nordgulen, 1999). In the east, metasedimentary rocks deposited in arc environments are also abundant (e.g. Graversen, 1984; Mansfeld and Andersen, 1999). Unfortunately, few reliable geochronological data have been published for these rocks, and little is known about their geochemistry and petrogenesis, or about the nature of the arc(s) in which they formed.

The present paper reports aims to test whether or not Mesoproterozoic subduction-related metaigneous rocks in south Norway show patterns of timing or geochemical character which can be attributed to the existence of a "SW Norwegian craton" with an internal history distinct from that of the Baltic Shield sensu stricto. The new major and trace element analyses, U–Pb ages and Lu–Hf isotope data are found to be in conflict with the presence of an exotic continental fragment in southern Norway, and to support alternative tectonic models for the region.

2. Geological setting

In the central and southwest parts of the Baltic Shield, orogenic events have been identified in the periods 1.9–1.75 Ga (Svecofennian), 1.75–1.55 Ga (Gothian), and 1.2–0.9 Ga (Sveconorwegian, i.e. Grenvillian; Gaàl and Gorbatschev, 1987). The southwestern part of the Baltic Shield acquired its final structural and metamorphic habit in the Sveconorwegian orogeny (e.g. Starmer, 1993), and the overall crustal architecture may be largely controlled by Sveconorwegian tectonics (e.g. Torske, 1985; Haas et al., 1999; Bingen et al., 2001). Large volumes of granitoids were emplaced in a belt extending from southernmost Sweden to central Norway (Transscandinavian Igneous Belt, TIB) in the period 1.85–1.65 Ga (e.g. Andersson, 1997; Andersson and Wikström, 2001; Åhäll and Larson, 2000). In the central and SE part of the shield, two suites of anorogenic mafic to felsic intrusions were emplaced in the periods 1.65–1.58 and 1.54–1.45 Ga, respectively; the younger of these may be a response to collision between the Baltic Shield and a continent to its present south, possibly Amazonia (the Dano-Polonian orogeny; Bogdanova, 2001).

The Precambrian continental crust of Norway south of the Caledonian thrust front and of SW Sweden (Fig. 1) is divided into domains separated by major Sveconorwegian shear zones, some of which have been reactivated as brittle faults in the Phanerozoic. In the present paper, the regional terminology for the Norwegian part of the SSD of Andersen and Knudsen (2000) is used. This is a geographic and nongenetic nomenclature; it does not assign tectonostratigraphic terrane status to any of its units, which are referred to as *sectors*. In traditional Swedish usage, the term *segment* is used in a similar sense (e.g. Larson et al., 1990; Söderlund et al., 2002). The area between the Oslo Rift and the Mjøsa-Magnor shear zone (MMS in Fig. 1, which continues as the Mylonite Zone in Sweden, e.g. Söderlund et al., 2002) has been called the "Idefjord terrane" (Åhäll and Gower, 1997); we reject this term, as this crustal block should not be elevated to tectonic terrane status (see discussion in Section 7.6, below).

From east to west, the area considered here consists of the *Østfold–Akershus Sector* (ØA in Fig. 1), the *Kongsberg Sector* (K in Fig. 1) and the *Bamble Sector* (B in Fig. 1). The Østfold–Akershus sector is a northward continuation of the Western Segment and parts of the Eastern Segment of SW Sweden, and is separated from the other sectors by the late Paleozoic Oslo Rift (Fig. 1). The Bamble Sector (B in Fig. 1) is separated from the *Telemark Sector* (T in Fig. 1) by a major Sveconorwegian shear zone (the *Porsgrunn-Kristiansand shear zone*, PKS), along which upper amphibolite-facies to granulite facies gneisses and metasediments of the Bamble Sector have been thrust over amphibolite-facies granitic gneisses of the Telemark sector (Starmer, 1993). Although outside of the area covered by this study, it should be noted that the central part of the Telemark sector was the site of voluminous rhyolitic volcanism at 1.51–1.50 Ga (Sigmond, 1998). This volcanism and subsequent intrusive activity have been related to crustal extension (Sigmond et al., 1997; Andersen et al., 2002b).

Remnants of mid-Proterozoic magmatic arcs occur in the Østfold–Akershus, Kongsberg and Bamble Sectors. The *Østfold complex* in the Østfold–Akershus sector consists of metasupracrustal gneisses, associated with several generations of granitic to tonalitic orthogneisses. Amphibolites, metarhyolites and metasedimentary gneisses in the Østfold complex can be correlated with the ca. 1.59–1.60 Ga Stora Le-Marstrand supracrustals in SW Sweden (Graversen, 1984; Åhäll et al., 1998). SIMS U–Pb ages on detrital zircons from a metasedimentary unit in the Østfold complex suggest the dominance of young, arc-related sources (Mansfeld and Andersen, 1999). The granitic to tonalitic orthogneisses associated with the metasupracrustals are interpreted as deformed intrusions (Berthelsen et al., 1996). Further north, the *Romerike complex* consists of mid-Proterozoic migmatitic gneisses of possible supracrustal origin, intruded by calc-alkaline granitoids (Skjernaa and Pedersen, 1982; Berthelsen et al., 1996). Granodioritic to tonalitic gneisses associated with metasupracrustal rocks also occur in the *Kongsberg complex* of the Kongsberg sector (Jacobsen and Heier, 1978; Dons and Jorde, 1978); these may be correlative with similar rocks in the Østfold complex (Nordgulen, 1999; Bingen et al., 2001).

The Bamble Sector consists of metasedimentary gneisses and quartzites, associated with metaigneous quartzofeldspathic gneisses and metagabbroic rocks (Starmer, 1993; Padget and Brekke, 1996). Metamorphism to the upper amphibolite to granulite facies (Nijland, 1993; Knudsen, 1996) has been dated to 0.99–1.13 Ga (Kullerud and Dahlgren, 1993; Knudsen and Andersen, 1999; Haas et al., 2002). Tonalitic to granodioritic gneisses occur as lenses, elongated bodies and large sheets, whose contacts with their country-rocks are in general concordant with the (Sveconorwegian) regional foliation (Fig. 1, see also Padget and Brekke, 1996). On a regional scale, they define a SW–NE striking array, whose endpoints are defined by the *Justøy tonalite* in the SW and the *Gjerstadvatn tonalite* in the NE. Tromøy island, at the coast of the Bamble sector, is a Sveconorwegian

(1198 ± 13 Ma, SIMS U–Pb age of zircons) island arc fragment, known as the *Tromøy complex* (Knudsen and Andersen, 1999).

2.1. The material studied

Summary data on the rocks studied are given in Table 1. Late- and post-magmatic alteration processes commonly affect fine-grained, felsic extrusive rocks (e.g. Brewer and Menuge, 1998). Most samples collected for dating and geochemistry therefore come from medium- to coarse-grained metaintrusive gneiss complexes, which better preserve their initial major and trace element signatures. To obtain information on the magmatic history and the nature of the basement of the western part of the Østfold–Akershus sector and the Kongsberg complex, two additional samples of metavolcanic rocks were included (01/25: an ignimbritic metarhyolite from westernmost part of the Østfold–Akershus sector and 01/19: a metadacitic gneiss of assumed supracrustal origin from the Kongsberg complex). Lu–Hf isotope data on most of the metaigneous rocks considered here were given by Andersen et al. (2002a); data on samples not included in that study are published here. One sample of granulite-facies tonalitic gneiss from Hisøy island, situated along the regional strike from Tromøy island, was included in the dating study, in order to check the areal extent of Sveconorwegian metatonalitic rocks in the Bamble sector; no such rocks have previously been identified outside of Tromøy island itself.

Previous SIMS U–Pb studies on detrital zircons in metasedimentary rocks have indicated the importance of pre-1.7 Ga sources of detritus in central south Norway (Knudsen et al., 1997; Haas et al., 1999; Bingen et al., 2001). Preliminary SIMS U–Pb data

Table 1
Summary data on rocks studied

Regional unit	Locality	Rock type	Sample	UTM-coordinate		Previous dating	Reference
				E	N		
Østfold–Akershus sector							
Østfold complex	Tistedal	Granodiorite	Ø1	6440	65578		Berthelsen et al. (1996)
	Feiring	Quartz diorite	Ø3	6105	66423	1.56 ± 0.04 Ga (Rb–Sr)	Graversen and Pedersen (1999)
	Norstrand-Sørmarka	Granodiorite	TA121	6034	66273	1.58 ± 0.13 Ga (Rb–Sr)	Graversen and Pedersen (1999)
	Mysen	Metagreywacke	Ø2	6355	66056		Berthelsen et al. (1996)
	Nesodden	Metapelite	N6	5940	66363		Graversen (1984)
	Våler	Metapelite	N8	6020	65960		Berthelsen et al. (1996)
	Slemmestad	Metarhyolite	01/25	5837	66268		Berthelsen et al. (1996)
Romerike complex	Børkelangen	Granodiorite	TA118	6417	66343	1.63 ± 0.05 Ga (Rb–Sr)	Skjernaa and Pedersen (1982)
	Midtskog	Tonalite	TA116	6278	66391	1.63 ± 0.05 Ga (Rb–Sr)	Skjernaa and Pedersen (1982)
Kongsberg sector							
Kongsberg complex	Snarum	Granodiorite	HFG	5452	66584		Nordgulen (1999)
	Bingen	Calc-alkaline gneiss (metadacite?)	01/19	5385	66364		Dons and Jorde (1978)
Bamble sector							
Justøy tonalite	Justøy	Tonalite	JUS	4624	64517		Padget and Brekke (1996)
Justøy tonalite	Homborsund	Tonalite	HOM	4707	64595		Padget and Brekke (1996)
Gjerstadvatn		Tonalite	NES	5029	65247		Padget (1993)
Jomås		Granodiorite	JOM	4761	64968		Padget and Brekke (1996)
Tromøy complex	Hisøy	Tonalite	HGG	4843	64764	1.20 Ga (U–Pb)[a]	Knudsen and Andersen (1999)

[a] SIMS U–Pb age on zircons in tonalitic gneisses from Tromøy island. Metamorphic overprint at 1.13 Ga.

seem to exclude this type of source for metasedimentary rocks in the Østfold complex, although the metasedimentary unit studied by Mansfeld and Andersen (1999) is situated much closer to present-day outcrops of TIB and other potential Paleoproterozoic source rocks than any other metasediment so far studied by SIMS U–Pb. To further characterize the source(s) of detritus to the arc system, and if possible to confirm or exclude an influence from the Baltic Shield, three samples of metasedimentary rocks from the Østfold–Akershus sector were selected for combined LAM-ICPMS U–Pb and Lu–Hf analysis. One of these (N8) duplicates the sample of Mansfeld and Andersen (1999), another (N6) gives Nd model age of 1.97 Ga (Haas et al., 1999; depleted mantle model of DePaolo, 1981).

3. Analytical methods

Samples (5–10 kg) collected from fresh outcrops were crushed to a grain size <250 μm, using a steel jaw crusher and a percussion mill. Zircons were separated by a combination of Wilfley-table washing, heavy liquid separation (1,1,2,2-tetrabromoethane and diiodomethane) and magnetic separation. Selected zircon grains from the least magnetic fractions were cast in epoxy disks for LAM-ICPMS analysis. A single mount consisted of several tens of zircons from each of several samples. The mounts were ground and polished to expose the zircons, and thoroughly cleaned prior to analysis. Zircons were examined in transmitted and reflected light, and by electron backscatter imaging. Standards were mounted in separate epoxy disks. Imaging, LAM-ICPMS U–Pb and Lu–Hf analyses were made at the GEMOC National Key Centre, Macquarie University.

3.1. LAM U–Pb dating

U–Pb dating was performed using a HP 4500 series 300 inductively coupled plasma quadrupole mass spectrometer (ICP-MS), attached to a custom built laser ablation microsampling system (Norman et al., 1996). This incorporates a petrographic microscope that provides the high quality visual image of the sample necessary for U–Pb work. A very fast scanning data acquisition protocol was employed to minimise plasma and other contributions to signal noise. Data acquisition for each analysis took 3 min. Instrumental-operating conditions are listed in Table 2. Helium was used to transport ablated material out of the sample cell and was mixed with Ar via a T-junction prior to entering the ICP torch. Relative to ablation in Ar, ablation in He minimises deposition of ablation products around ablation sites, improves sample transport efficiency, provides more stable signals and gives more reproducible U–Pb fractionation. Provided that constant ablation conditions are maintained, accurate correction for U–Pb fraction-

Table 2
LAM-ICP-MS operating conditions and data acquisition parameters

ICP-MS	
Model	HP 4500 Series 300
Shield torch	Used
Forward power	1450 kW
Gas flows	
Plasma	16 l/min
Auxiliary	0.8 l/min
Carrier	ca. 1 l/min He, mixed with ca. 09 l/min
LAM	
Model	Custom built (Norman et al., 1996)
Wavelength	266 nm
Apertured beam diameter	ca. 5 mm
Laser repetition rate	10 Hz
Degree of defocusing	150 μm (above sample)
Incident pulse energy	0.6 mJ/pulse
Ablation spot diameter	30–40 μm
Data acquisition parameters	
Data acquisition protocol	Time Resolved Analysis
Scanning mode	Peak hopping, 1 point per peak, pulse counting
Isotopes determined	^{206}Pb, ^{207}Pb, ^{208}Pb, ^{232}Th, ^{238}U
Dwell time per isotope	15 ms, except ^{207}Pb (20 ms)
Time/mass scan	90 ms
No. of mass scans	ca. 2000
Analysis time	180–60 s background, 120 s (max) ablation
Samples and standards	
Samples	Hand picked, best quality grains
Mounts	5 mm diameter polished grain mounts
Standards	02123, 295 ± 1 Ma, (Ketchum et al., 2001) JG-1, 609 Ma (Corfu, personal communication, 2000)

ation can then be achieved using an isotopically homogeneous zircon standard.

Samples were analysed in ''runs'' of ca. 20 analyses which included 12 unknown points, bracketed, beginning and end, by four analyses of a concordant zircon standard (02123, which is concordant at 295 Ma (Ketchum et al., 2001) or GJ-1, nearly concordant at 608.5 Ma (Corfu, personal communication, 2000). Three concordant standard zircons were run frequently as unknowns, to provide an independent control on reproducibility and instrument stability (Fig. 2, Table 3a). It should be noted that shifts in the elemental U/Th/Pb ratios caused by uncorrected drift in degree of elemental fractionation during a ''run'' of 20 analyses can result in displacement of points along straight lines indistinguishable from recent lead-loss lines (Knudsen et al., 2001). The standardization procedure used in this study has been designed to correct for this effect. Furthermore, the effects of minor, short-term variation of elemental fractionation during ablation of individual points, or between ablations, are accounted for by the error-propagation procedure used (see below). Any residual effect of short-term fluctuations in elemental fractionation is therefore incorporated into the errors assigned to the $^{206}Pb/^{238}U$, $^{207}Pb/^{235}U$ and $^{208}Pb/^{232}Th$ ratios.

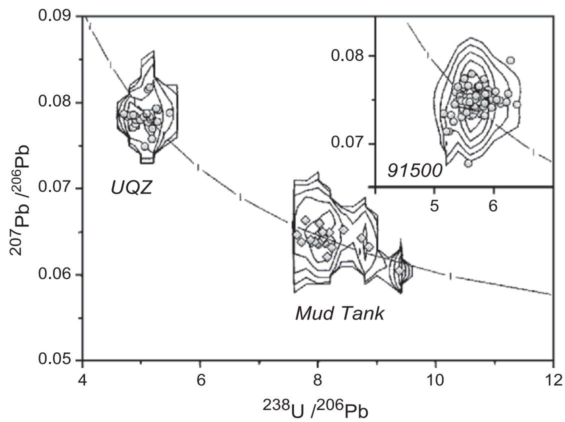


Fig. 2. Standards run as unknowns during the period of analysis, plotted in a Tera-Wasserburg concordia diagram. The diagram contains 277 runs of 91500 (age 1065 Ma), 46 runs of Mud Tank (TIMS age 734 Ma; bimodal age distribution 734 and 720 Ma) and 24 runs of UQZ (age 1143 Ma). Each analysis is represented by a single spot. The contours represent an accumulated relative probability surface, which takes into account the errors in both isotopic ratios, and are equally spaced at an arbitrary scale. See also Table 3.

Individual U–Pb ages were calculated from the raw signal data using the off-line software package GLITTER (Van Achterbergh et al., 2000). GLITTER calculates the relevant isotopic ratios for each mass sweep and displays them as a coloured pixel map and time-resolved intensity traces. This allows isotopically homogeneous segments of the signal to be selected readily for integration, which is particularly important when the laser beam has ablated through successive zones of different ages or compositions. GLITTER then corrects the integrated ratios for ablation-related fractionation and instrumental mass bias by calibration, for each selected, homogeneous time-segment, against the identical integrated ablation time segments of the standard zircon analyses.

3.1.1. Error propagation

Isotope ratios are built from background-subtracted signals for the corresponding isotopes. Uncertainties in these combine the uncertainties of signal and background, arising from counting statistics, added in quadrature. The same propagation is used for unknowns and standard analyses. The standard ratios are interpolated between standard measurements to estimate the standard ratios at the time of the measurement of the unknowns. Uncertainties in the standard ratio measurements are propagated through this procedure to estimate the standard ratio uncertainties relevant to each unknown ratio measurement. Relative uncertainties estimated for the standard ratios are combined with the unknown ratio uncertainties in quadrature. A further 1% uncertainty (1σ) is assigned to the published values of the isotope ratios for the standard and propagated through the error analysis.

3.1.2. Common-lead correction

As ^{204}Pb could not be measured due to low peak-to-background ratio and isobaric interference from ^{204}Hg, the conventional correction for common-lead used in thermal and secondary ionization mass spectrometry could not be used. Commonly used correction methods for U–Pb analyses that do not report ^{204}Pb assume that the only source of discordance in a zircon is the presence of common lead (e.g. Ludwig,

Table 3

(a) The precision and accuracy obtained by LAM-ICPMS compared with TIMS data on some well-characterised zircons

	TIMS 207/206	206/238	mswd	207/235	mswd	207/206	mswd
UQ-Z5	1143 ± 1	Machado and Gauthier (1996) (n = 9)					
n = 14		1138 ± 17	5.5	1137 ± 11	5.7	1134 ± 11	0.32
n = 33		1136 ± 14	4.7	1134 ± 11	5	1135 ± 9	6.2
Temora	417	Black (personal communication, 2000)					
n = 11		416.7 ± 2.9*	0.69	414.7 ± 3.7	0.35	406 ± 23	0.31
91500	1065.4	Wiedenbeck et al. (1995) (n = 11)					
n = 75		1051 ± 7	4.4	1056 ± 3.3	3.1	1066 ± 5	0.72
Mud Tank	734 ± 32	Black and Gulson (1978) (n = 5)					
n = 67		736 ± 3	2.9	737 ± 3	2.6	740 ± 7	0.46

207/206 ages preferred for older samples

(b) Data on Hf isotope standards

Sample	No. analyses	Yb (ppm)	Lu (ppm)	Hf (ppm)	$^{176}Lu/^{177}Hf$	$^{176}Yb/^{177}Hf$	$^{176}Hf/^{177}Hf$	± 2SD	Hf signal (volts)
JMC475 Hf	208			0.1–1.0			0.282161	0.000021	5–18[a]
JMC475 Hf (+Yb)	27	0.04		0.105		0.260	0.282159	0.000060	0.8–4.0
	18	0.01		0.1		0.070	0.282161	0.000040	3.0–4.5
Zircon 91	60	62.5	13.9 12.4[b]		0.00030		0.282297	0.000044	
TIMS	7		12	5895	0.00029		0.282290	0.000014	
Zircon 61308	45				0.00169	0.04996	0.282999	0.000083	
	12				0.00022	0.00654	0.282966	0.000113	
	22	352	78		0.00138	0.04080	0.282992	0.000093	
TIMS	9		83	5658	0.00207	na	0.282975	0.000050	

Yb and Lu concentrations of zircons by LAM-ICPMS; TIMS data from Wiedenbeck et al. (1995).

[a] 5–10 V for 0.1-ppm solution using MCN6000; 9–18 V for 1-ppm solution using Meinhard nebuliser.

[b] Lu calculated from $^{176}Lu/^{177}Hf$ and given Hf content.

* LAM-ICPMS 206/238 ages preferred for samples with age less than 1 Ga.

2003). Using such corrections on lead analyses that contain a discordance component caused by lead loss will invariably lead to overcorrection, and hence to meaningless, young ages. The contents of common lead has therefore been estimated from the 3D discordance pattern in $^{206}Pb/^{238}U-^{207}Pb/^{235}U-^{208}Pb/^{232}Th$ space, and corrected accordingly (Andersen, 2002). The corrections have been made assuming recent lead-loss, which will not induce a systematic bias greater than analytical uncertainty in zircons which have suffered moderate lead loss during the Phanerozoic (Andersen, 2002).

3.2. Hf isotope analysis

The Hf-isotope analyses reported here were carried out in situ using a New Wave Research LUV213 laser-ablation microprobe, attached to a Nu Plasma multi-collector ICPMS. The analytical methods, including extensive data on the analysis of standard solutions and zircons, are discussed in detail by Griffin et al. (2000, 2002).

A value for the decay constant of ^{176}Lu of 1.93×10^{-11} $year^{-1}$ has been used in all calculations. For the calculation of ε_{Hf} values, we use chondritic $^{176}Hf/^{177}Hf = 0.279742$ and $^{176}Lu/^{177}Hf = 0.0332$ (Blichert-Toft and Albarède, 1997). These values were reported relative to $^{176}Hf/^{177}Hf = 0.282163$ for the JMC475 standard, well within error of our reported value (Table 3). We have adopted the depleted mantle model of Griffin et al. (2000), which produces a value of $^{176}Hf/^{177}Hf$ (0.28325) similar to that of average MORB over 4.56 Ga, from $(^{176}Hf/^{177}Hf)_i = 0.279718$ at $^{176}Lu/^{177}Hf = 0.0384$. This mantle evolution curve is indistinguishable from the $f_{Lu/Hf} = 0.16$ curve of Vervoort and Blichert-Toft (1999).

Table 4
Major and trace element data for calc-alkaline granitoids, South Norway

	Ø1	Ø3	121/99	116/99	118/99	NES	HOM	JUS	JOM	HGG	Standards	
											GSP-1	JG-1
Weight percent oxides												
SiO_2	63.17	60.61	68.37	58.22	67.25	66.19	65.22	64.93	63.93	71.93	67.11	72.15
TiO_2	0.72	0.94	0.49	0.71	0.44	0.51	0.44	0.74	0.61	0.42	0.66	0.26
Al_2O_3	15.32	14.66	15.08	17.08	15.71	14.77	16.43	16.08	15.69	13.07	15.06	14.19
$Fe_2O_3^t$	6.02	8.33	3.62	7.41	3.58	3.96	4.50	4.83	5.89	4.88	4.29	2.09
MnO	0.09	0.11	0.06	0.12	0.05	0.06	0.07	0.06	0.06	0.09	0.04	0.06
MgO	2.75	4.26	1.21	3.46	1.71	2.04	1.80	1.87	2.72	0.88	0.99	0.75
CaO	4.54	7.76	3.23	6.68	3.60	3.77	4.39	4.37	3.24	3.21	1.98	2.17
Na_2O	3.45	2.92	3.89	3.55	4.39	4.13	4.32	4.44	3.71	3.59	2.84	3.36
K_2O	2.19	0.50	3.06	1.79	2.33	1.23	1.52	1.58	2.24	1.20	5.43	3.91
P_2O_5	0.16	0.17	0.15	0.20	0.15	0.09	0.10	0.20	0.12	0.06	0.29	0.09
LOI	0.59		0.36	0.54	0.63	0.43	0.66	0.45	0.78	0.34	0.6	0.46
Sum	99.00	100.26	99.49	99.76	99.81	97.17	99.44	99.56	99.00	99.68	99.29	99.49
CIPW norm												
qz	19.44	18.42	24.54	10.09	21.83	25.15	20.75	20.04	20.88	36.48		
co	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.00	1.56	0.18		
or	12.94	2.95	18.08	10.58	13.74	7.27	8.98	9.33	13.23	7.09		
ab	29.18	24.69	32.86	30.02	37.08	34.93	36.53	37.55	31.38	30.36		
an	19.85	25.42	14.67	25.39	16.33	18.11	20.96	19.29	15.29	15.53		
di	1.34	9.89	0.32	5.34	0.44	0.00	0.14	0.90	0.00	0.00		
hs	11.49	13.61	5.94	13.19	7.16	8.52	8.51	8.19	12.05	6.67		
ilm	1.37	1.79	0.92	1.35	0.83	0.97	0.84	1.41	1.16	0.80		
mt	2.31	2.89	1.39	2.57	1.37	1.52	1.73	1.86	2.26	1.88		
ap	0.38	0.40	0.34	0.47	0.36	0.21	0.24	0.47	0.28	0.14		
											SY3	MAG1
Parts per million												
Rb	136	5.4	96	43	75	26	55	39	85	39	215	148
Y	28	26	14	26	10	8.5	27	13	22	28	720	27
Zr	174	132	148	164	147	116	130	178	159	121	365	127
Nb	9.6	8.3	8.8	6.9	7.5	5.3	4.3	7.3	7.9	3.0	149	11
Hf	5.3	4.0	4.0	4.4	4.0	3.6	4.0	4.8	5.2	3.8	11	3.6
Ta	0.63	0.48	0.61	0.31	0.43	0.22	0.33	0.51	0.61	0.16	24	1.22

Ranges	Feiring			Midtskog			Nordstrand-Sørmarka		
	Min	Mean	Max	Min	Mean	Max	Min	Mean	Max
	$n=11$			$n=4$			$n=26$		
Weight percent oxides									
SiO_2	54.01	57.54	60.61	52.71	61.59	74.93	61.3	69.45	77.2
TiO_2	0.85	0.98	1.27	0.22	0.56	0.72	0.1	0.58	1.1
Al_2O_3	14.66	17.04	18.69	12.46	16.43	18.80	11.3	14.28	19.6
$Fe_2O_3^t$	4.99	5.99	9.65	2.39	6.21	8.76	1.2	4.24	7.8
MnO	0.10	0.12	0.15	0.03	0.10	0.14	0.03	0.76	2.78
MgO	3.07	3.70	4.38	0.57	2.74	4.21	0.42	1.82	5.09
CaO	6.26	7.01	8.07	2.06	5.62	8.02	1.42	2.90	4.03
Na_2O	2.92	3.55	3.87	2.73	3.42	3.90	1.83	3.61	5.29
K_2O	0.46	0.86	1.38	1.79	2.24	3.52	0.04	1.27	5.11
P_2O_5	0.17	0.27	0.33	0.03	0.14	0.20	0.08	0.47	2.78

(continued on next page)

Table 4 (*continued*)

Ranges	Feiring			Midtskog			Nordstrand-Sørmarka		
	Min	Mean	Max	Min	Mean	Max	Min	Mean	Max
	$n=11$			$n=4$			$n=26$		
CIPW norm									
qz	4.8	11.5	18.4	0.4	15.8	39.6	20.9	29.0	40.7
co	0.0	0.0	0.0	0.0	0.1	0.5	0.0	1.4	11.2
or	2.7	5.1	8.2	10.6	13.2	20.8	8.6	20.3	31.3
ab	24.7	30.0	32.7	23.1	28.9	33.0	12.3	27.4	34.1
an	25.4	28.0	33.1	10.0	22.5	30.2	2.9	11.3	22.7
di	1.6	4.1	9.9	0.0	3.7	7.0	0.0	0.1	1.3
hs	13.6	15.1	18.4	3.3	11.0	15.7	1.6	6.6	13.0
ilm	1.6	1.9	2.4	0.4	1.1	1.4	0.2	1.1	2.1
mt	2.5	2.8	3.0	1.1	2.2	3.0	0.5	1.7	3.0
ap	0.4	0.6	0.8	0.1	0.3	0.5	0.1	0.4	3.6
Parts per million									
Rb				43.31	60.11	90.51	87.54	118.71	167.04
Y				5.07	20.86	27.31	8.23	20.51	44.79
Zr				108.33	156.64	215.86	46.07	164.14	293.06
Nb				4.43	5.67	6.89	4.98	9.27	16.64
Hf				2.99	4.20	5.81	1.76	4.62	7.67
Ta				0.22	0.30	0.46	0.49	0.79	1.31

3.3. Major and trace elements

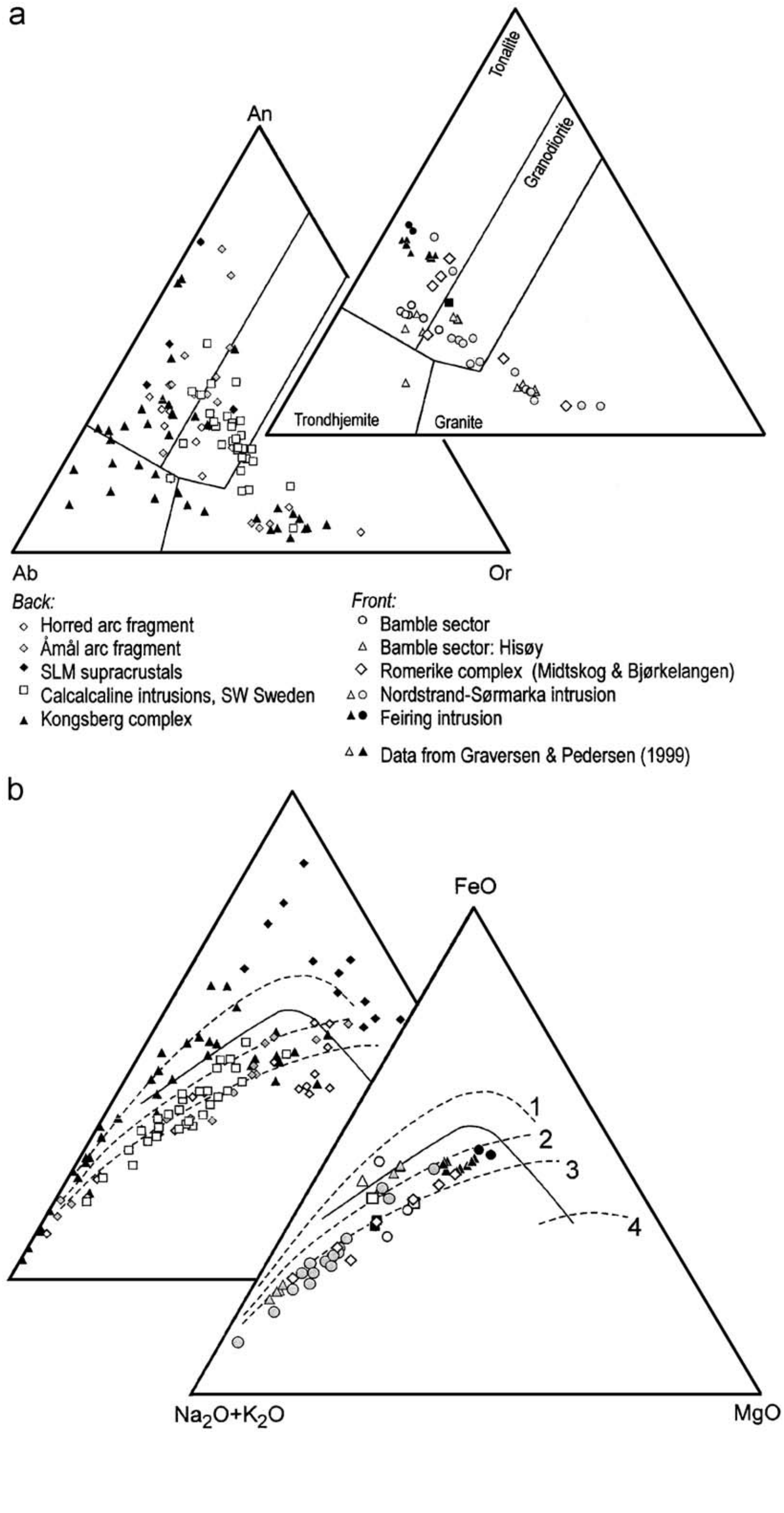
A ca. 20 g fraction of material from the jaw crusher was reduced to powder in a steel disk mill for whole-rock geochemistry. Major elements were analysed by XRF on lithium borate glass pellets, using a Phillips X-ray fluorescence spectrometer at the Department of Geology, University of Oslo, Norway. Trace elements were analysed by fusion ICPMS by ACTLAB Canada. Data for international standards analysed as unknowns are included in Table 4.

4. Major and trace element geochemistry of calc-alkaline metaigneous rocks

Major and trace element data for the calc-alkaline metaintrusive rocks dated in the present study are given in Table 4. Additional analyses from the authors' database and from Graversen and Pedersen (1999) are included in Figs. 3 and 4. Full listings of data are available on request to the corresponding author.

The samples analysed for major and trace elements span an overall range from gabbro to granite, with a clear majority along a trend from tonalite through granodiorite to granite. SiO_2 ranges from 52% in a leucogabbroic sample from Midtskog, to 75% in a granite from the Nordstrand-Sørmarka intrusion. The intrusions range from metaluminous to weakly peraluminous; normative corundum amounts to less than 0.5% or is absent in the Justøy, Gjerstadvatn and Feiring intrusions. One sample from the Midtskog intrusion has 0.49 % *co*, the others are metaluminous. The Nordstrand-Sørmarka intrusion stands out as systematically peraluminous,

Fig. 3. Major element variation in mid-Proterozoic calc-alkaline rocks in S Norway and SW Sweden. The front diagrams show data from the present study, the background diagrams published data. (a) Normative feldspar components, with classification fields of Barker (1979). (b) Weight % AFM diagram. The solid line is the calc-alkaline/tholeitic division line of Irvine and Baragar (1971), the broken lines 1–4 are contours of increasing arc maturity, from Brown (1982) and Knudsen and Andersen (1999). Sources of data: Åmål, Horred and Stora Le-Marstrand metavolcanic rocks: Brewer et al. (1998). Calc-alkaline intrusions, SW Sweden: Samuelsson (1978, 1982). Kongsberg complex: Jacobsen (1975). The data on the Feiring and Nordstrand-Sørmarka intrusions include analyses from Graversen and Pedersen (1999), shown as triangles.


a
An
Tonalite
Granodiorite
Trondhjemite
Granite
Ab
Or
Back:
Horred arc fragment
Åmål arc fragment
SLM supracrustals
Calcalcaline intrusions, SW Sweden
Kongsberg complex
Front:
Bamble sector
Bamble sector: Hisøy
Romerike complex (Midtskog & Bjørkelangen)
Nordstrand-Sørmarka intrusion
Feiring intrusion
Data from Graversen & Pedersen (1999)
b
FeO
1
2
3
4
Na2O+K2O
MgO

with *co* reaching 11% in one sample. There is no clear correlation between normative *co* and SiO_2 or normative *qz*.

In terms of normative feldspar components, the calc-alkaline metaintrusive rocks from southern Norway define a range from gabbro and quartz diorite through granodiorite to granite, with one only sample plotting in the trondhjemite field (Fig. 3a). The Feiring intrusion defines the low-*or* end of the trend, ranging from quartz diorite to tonalite, whereas some of the samples from the Nordstrand-Sørmarka intrusion mark its granitic end. In an AFM diagram, most samples plot within the calc-alkaline field, defining a trend of moderate

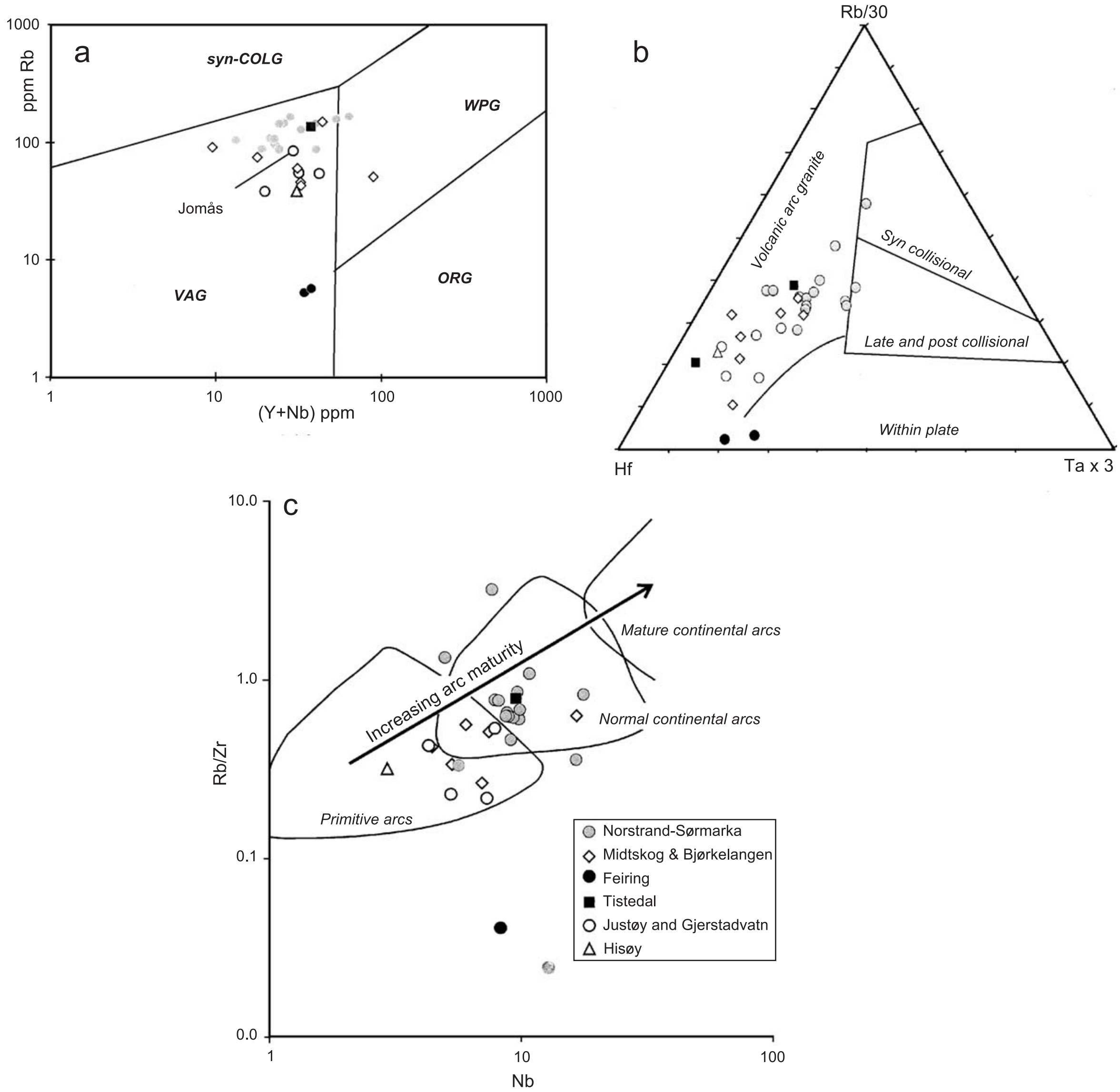


Fig. 4. Trace element indicators of tectonic setting and arc maturity. Key to symbols in c applies to all three diagrams. (a) Rb vs. Y + Nb discrimination diagram for granitic rocks from Pearce et al. (1984). (b) Hf–Ta–Rb discrimination diagram for granitic rocks (Harris et al., 1986). (c) Trace element arc maturity indicators, after Brown et al. (1984).

arc maturity (Fig. 3b). Again, the Feiring and Nordstrand-Sørmarka intrusions make up the extremes of the trend. The Hisøy tonalite falls within the tholeitic field.

In trace element discrimination diagrams for granitic rocks (Rb vs. Y + Nb, Fig. 4a; Hf–Ta–Rb, Fig. 4b), the gneiss complexes classify as "volcanic arc granites", with only a few of the more silicic samples straddling the boundary to "within-plate granites" (Fig. 4a) or to "syn-, late- and post-collisional granite" (Fig. 4b), which probably reflect differentiation rather than differences in tectonic setting. For volcanic arc granites, a plot of Rb/Zr vs. Nb may be used as a qualitative indicator of arc maturity (Brown et al., 1984). The samples span a range from the "primitive arc" into the "normal continental arc" field; none of the rocks analysed fall within the field of mature continental arcs (Fig. 4c). The samples from the Feiring intrusion plot significantly below the main array, because of their low Rb concentrations, which may be due to the presence of accumulated plagioclase. The moderate variation in Rb for the rest of the samples suggests that metamorphic remobilization of Rb was not a major problem in any of these metaintrusive rocks, probably because they have been holocrystalline from the start; however, none of the conclusions drawn from Fig. 4 are critically dependent on the Rb concentration.

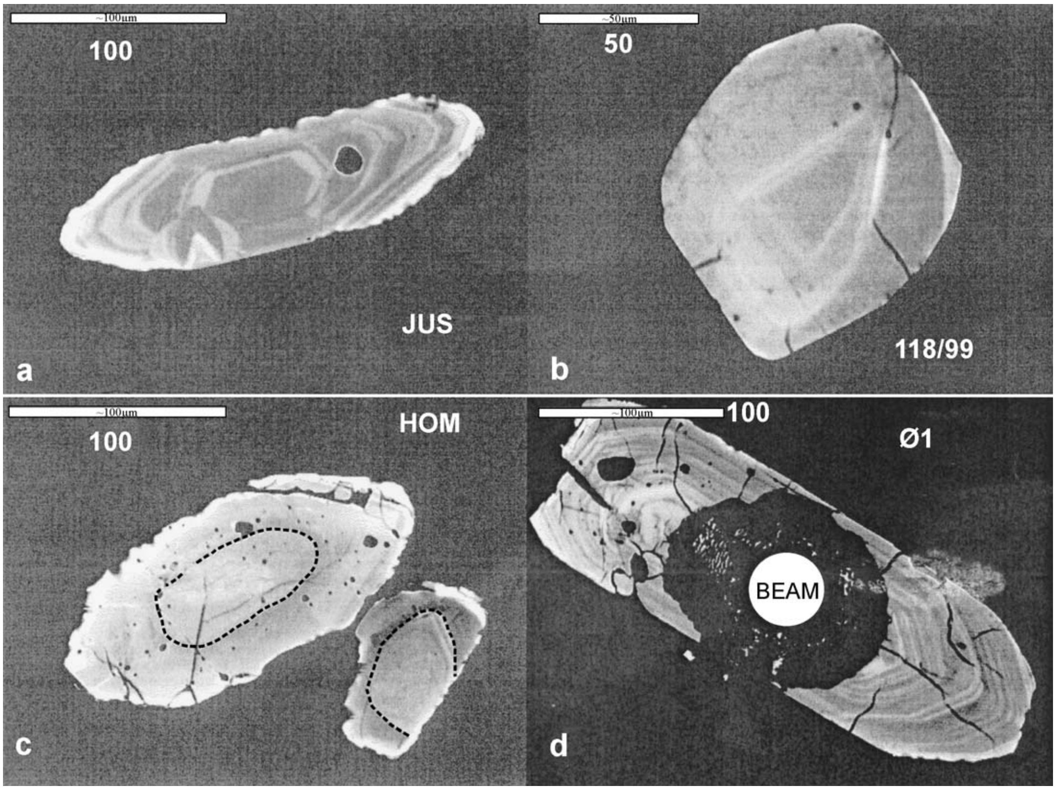


Fig. 5. Backscatter electron images of zircons from calc-alkaline gneisses in southern Norway. Scale bars in μm. (a) Oscillatory zoned grain from the Justøy tonalite, sample JUS. Irregularities in the oscillatory zonation pattern (left) are due to twinning or intergrowth. The dark inclusions consist of low atomic number, BSE-dark material (feldspar?). The crystal shows a discontinuous BSE bright outer rim due to Sveconorwegian (?) metamorphism. (b) Cross-section through an oscillatory zoned zircon grain from the Bjørkelangen granodiorite, sample TA118. The outer bright zone to the right may be a discontinuous metamorphic overgrowth. (c) Two grains with xenocrystic cores from the Justøy tonalite, sample HOM. The cores (outlined by broken lines) have faint oscillatory zoning; the zoning of the enclosing zircon is discordant to the zoning and the core-rim boundary. (d) An oscillatory zoned grain without visible metamorphic overgrowth (Tistedal tonalite, sample Ø1) has been drilled by the laser beam (highlighted by white circle). The dark halo surrounding the ablation pit consists of ablated material.

5. Timing of calc-alkaline magmatism

5.1. Zircon morphology, internal structure and U/Th ratio

Zircons show moderately elongated prismatic habits, with terminal pyramid faces (Fig. 5). Most grains have thin BSE-bright overgrowths (Fig. 5a,b,c) which are generally too thin for analysis (< 10 μm). The central parts of the grains show regular oscillatory "magmatic" zoning (Fig. 5a and c). Infrequent xenocrystic cores show oscillatory zoning crosscut by the enclosing grain (Fig. 5c). The diameter of the laser beam and the deposition of an ablation halo are illustrated in Fig. 5d. Some zircon crystals contain BSE-dark inclusions (Fig. 5a and c). Grains or parts of grains containing visible inclusions were avoided during analysis; the presence of sub-surface inclusions that were penetrated by the laser beam could easily be identified by irregularities in the time-resolved intensity and isotope ratio traces, and excluded from integration. In

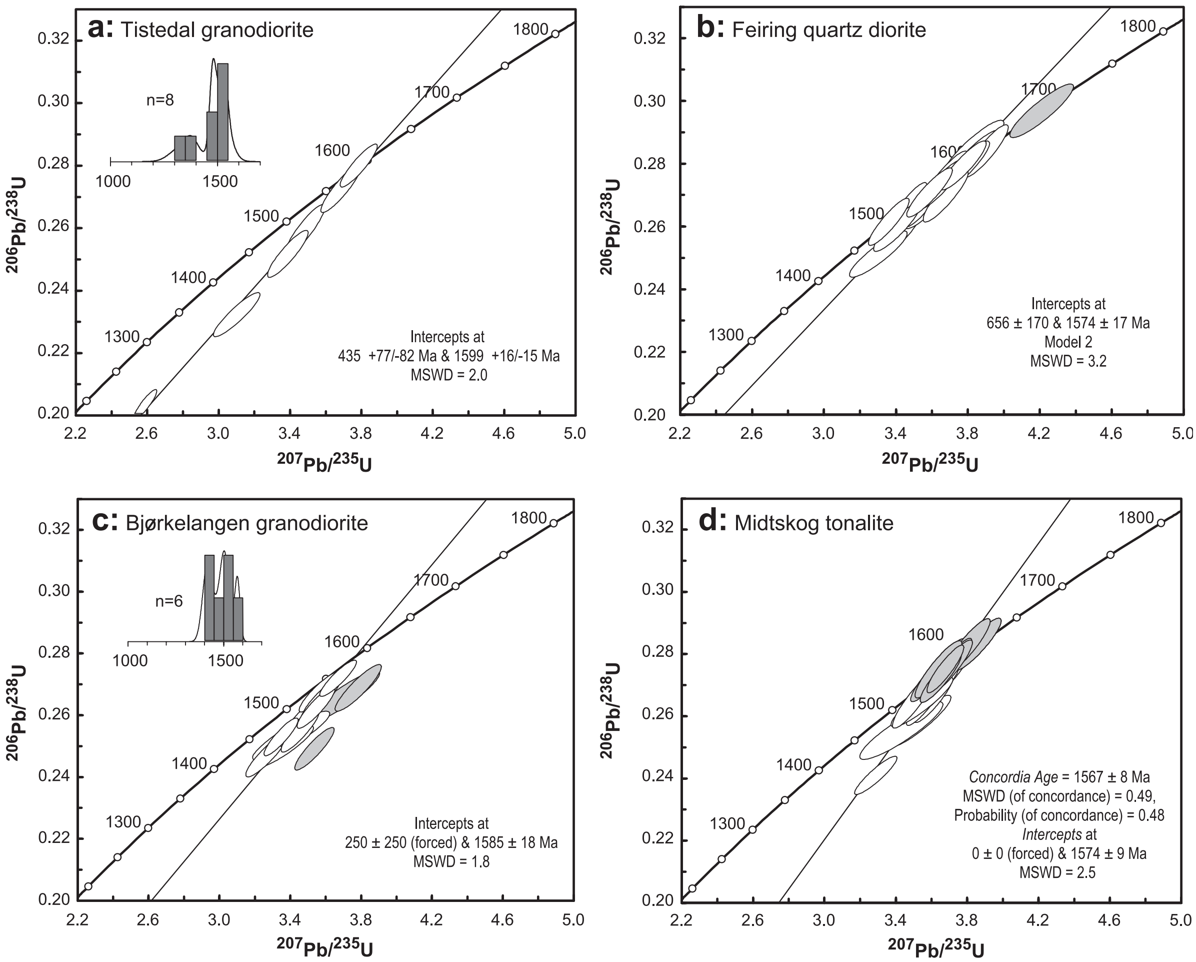


Fig. 6. (a–m) Conventional U–Pb concordia diagrams of the samples dated by LAM-ICPMS U–Pb. The compositions of single zircons are shown as $\pm 2\sigma$ error ellipses, with an assumed error correlation of 0.9 (data from Table 5). Geochronological regressions have been made using free or forced lower intercepts as indicated in Table 6. Only zircons used in regressions (black and white), and inherited zircons (grey), are plotted. Zircons, which have been partly or fully reset by Sveconorwegian metamorphism, are represented by accumulated probability plots of $^{207}Pb/^{206}Pb$ ages (number of zircons as indicated).

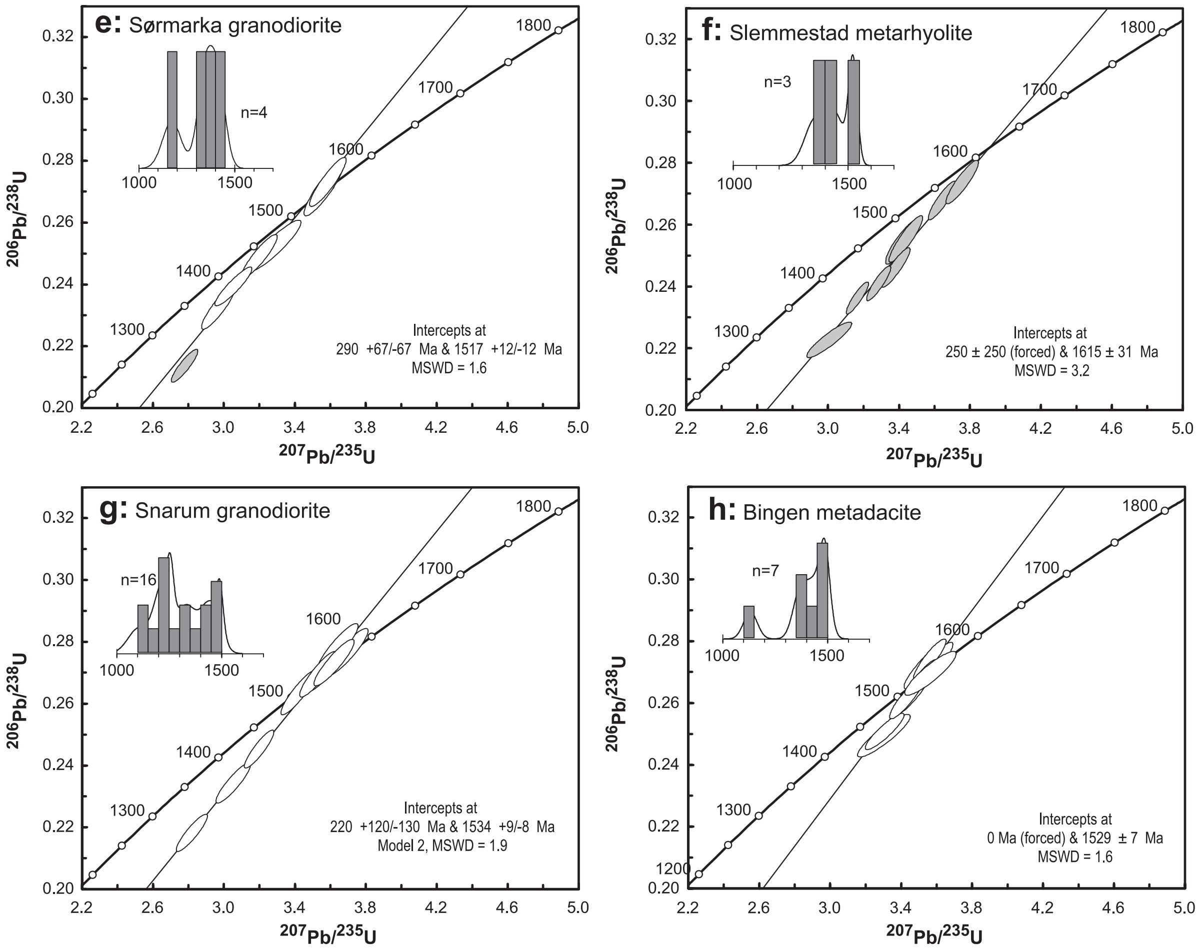


Fig. 6 (*continued*).

general, the grains analysed have $^{238}U/^{232}Th < 5$ (Table 5, available as an electronic supplement). Elevated ratios ($^{238}U/^{232}Th > 10$) are restricted to a few grains from the granulite-facies Hisøy tonalite (Table 5).

5.2. Dating

Several TIMS-ID U–Pb studies in the SSD have reported Phanerozoic or near-recent lower intercept ages (e.g. Pasteels and Michot, 1975; Bingen and van Bremen, 1998; Connelly and Åhäll, 1996; Åhäll et al., 2000; Haas et al., 2002), which may overlap with the age of tectonothermal events during the Caledonian orogeny, or the regional rifting event which affected northern Europe in the late Paleozoic (e.g. Oslo Rift). In addition, the rocks studied have suffered Sveconorwegian amphibolite- to granulite facies metamorphism, at ages ranging from ca. 1100 Ma in the Bamble Sector (Kullerud and Dahlgren, 1993) and 1090 Ma in the northern part of the Østfold–Akershus sector/Western Segment (Ali, 2000) to ca. 970 Ma in SW Sweden (Söderlund et al., 2002). Zircon images (Fig. 5) suggest that little new zircon crystallized in secondary events, but whole zircon crystals or of domains within single crystals may have suffered partial lead loss. In most cases, the precision of individual LAM-ICPMS points is too poor to allow clear separation between concordant zircons of intermediate age (i.e. 1200–

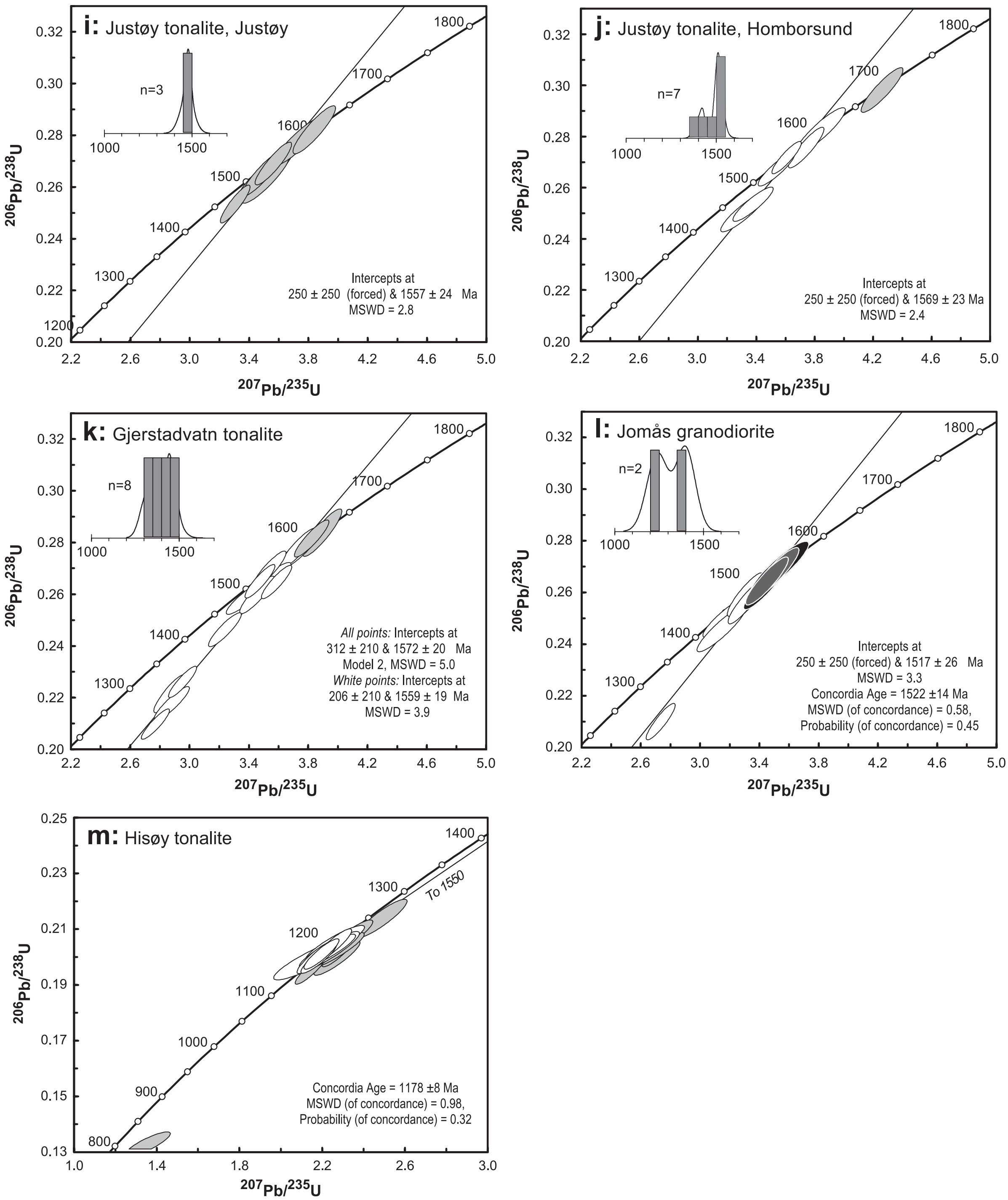


Fig. 6 (*continued*).

1400 Ma) and points on a discordia line from a Mesoproterozoic upper intercept to a Sveconorwegian lower intercept; similar problems affect SIMS U–Pb data on zircon populations with an identical resetting history (e.g. Knudsen et al., 1997). The $^{207}Pb/^{206}Pb$ age of such an anomalously young, apparently "concordant" zircon merely reflects the amount of lead lost during metamorphic resetting, or the proportions of preserved and newly crystallized or reset zircon penetrated during the analysis, and is thus meaningless.

The approach of the present study has been to identify multi-grain populations of zircons which are unaffected by post-crystallization lead loss, or which share a simple and consistent post-crystallization history, which involves only Phanerozoic resetting; in practice, this amounts to identifying and excluding from further consideration those zircons that have suffered significant Sveconorwegian lead loss. Zircons, which have been rejected from regression because of significant Sveconorwegian lead loss, are represented by accumulated probability plots of $^{207}Pb/^{206}Pb$ ages shown as insets in Fig. 6. The concordia age (i.e. a weighted average incorporating information from the $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ratios in all grains of a population of concordant and equivalent zircons; Ludwig, 2003), or the upper intercept defined by a population of zircons having suffered late lead loss gives the best estimate of the emplacement age of the host rock. The lower intercept of such lines is commonly too poorly constrained to allow positive identification of the secondary tectonothermal event involved, and the geological significance of lower intercept ages are therefore not further considered here. Some zircon populations giving imprecise, Paleozoic lower intercepts have been recalculated with a forced lower intercept of 250 ± 250 Ma, and two populations which suggested recent lower intercepts with large uncertainties were forced through zero.

U–Pb data for individual grains are given in Table 5. Single zircons are plotted with 2σ errors and an assumed error correlation of 0.90 (Fig. 6), which is a conservative estimate of error correlations in raw data on standards analysed during the period this study was made. Regressions on multigrain populations have been calculated using the geochronology package Isoplot 2.49 (Ludwig, 2003); ages are given with 2σ errors (Fig. 6 and Table 6). Some of the regression lines (Fig. 6) show scatter in

Table 6

Complex/locality	Age	2σ		Lower intercept, Ma		MSWD	N
Østfold–Akershus sector							
Tistedal granodiorite	1599	+15/−16		435+77/−82		2.0	6
Feiring quartz diorite	1574	±17	M2	656±170		3.2	20
Bjørkelangen granodiorite	1585	±18		250±250	F	1.8	11
Midtskog tonalite	1567	±8	Conc.			0.49	9
	1574	±9		0	F	2.5	20
Sørmarka granodiorite	1517	±12		290±67		1.6	7
Slemmestad metarhyolite	1615	±31		250±250	F	3.2	9
Kongsberg sector							
Snarum granodiorite	1534	+9/−8		220±130		1.9	9
Bingen metadacite	1529	±7		0	F	1.6	9
Bamble sector							
Justøy tonalite, Justøy	1557	±24		250±250	F	2.8	7
Justøy tonalite, Homborsund	1569	±23		250±250	F	2.4	6
Gjerstadvatn tonalite	1559	±19		206±210		3.9	14[a]
	1572	±20	M2	312±210		5.0	16
Jomås granodiorite	1517	±26		250±250	F	3.3	9
	1522	±14	Conc.			0.58	4
Hisøy tonalite (Tromøy complex)	1178	±9	Conc.			0.98	9

F: forced lower intercept, M2: model 2 age (Ludwig, 2003), Conc.: Concordia age (Ludwig, 2003). N: number of zircons in regression.

[a] Tow possibly inherited zircons omitted.

excess of what is expected if analytical error was the only source of scatter; this is most likely due to minor Sveconorwegian disturbance. Regression model 2 of Ludwig (2003) has been used for populations with elevated MSWD where indicated in Fig. 6 and Table 6.

Of the five intrusions dated from the Østfold–Akershus sector, four (Tistedal, Feiring, Bjørkelangen, Midtskog) give relatively precise ages in the range 1.60–1.57 Ga. At an age of 1615 ± 31 Ma, the metarhyolite from Slemmestad also belongs in this age group, or is slightly older. The Sørmarka granodiorite is significantly younger, at 1517 ± 12 Ma. The two samples from the Kongsberg sector dated in this study give ages of 1535 + 9/ − 8 (HFG) and 1529 ± 7 Ma (01/19); these ages are indistinguishable from each other and from the age of the Sørmarka granodiorite. Nordgulen (1999) reported ages of 1555 ± 3 and 1500 + 5/ − 3 Ma for two dioritic to granodioritic intrusions, suggesting that calc-alkaline magmatism in the Kongsberg sector lasted for (at least) 50 Ma. In the Bamble sector, the Gjerstadvatn and Justøy tonalites are coeval at 1550–1570 Ma, and are thus comparable in age to the older intrusions reported from the Kongsberg sector; they have overlapping uncertainties with several of the older intrusions in the Østfold–Akershus sector. The Jomås granodiorite is significantly younger, at 1522 ± 14 Ma, which is indistinguishable from the younger calc-alkaline rocks in the Kongsberg and Østfold–Akershus sectors.

The Hisøy tonalite gives a concordia age of 1178 ± 9 Ma, indistinguishable from the SIMS U–Pb age of 1198 ± 13 Ma reported for mafic and tonalitic gneisses at the neighbouring Tromøy island (Knudsen and Andersen, 1999). The present data thus indicate the existence of Sveconorwegian, metaigneous rocks belonging to the Tromøy arc fragment along the coast of the Bamble sector, outside of Tromøy island itself (Fig. 1).

Clearly inherited zircons with ages in the range 1.7–1.9 Ga were observed in samples Ø1, Ø3, HFG and HOM (Fig. 6). The presence of inherited zircons of this age range in intrusions from the Østfold–Akershus, Kongsberg and Bamble sector suggests the presence of such rocks at depth throughout the region at the time of arc magmatism. Younger, but still possibly inherited zircons (1.6–1.7 Ga) were observed in TA118 and 01/19; zircons of this age-range may come from older, arc-related sources (Åhäll and Larson, 2000 and references therein). In the Hisøy tonalite, some discordant zircons cluster along a discordia line towards a poorly constrained mid-Proterozoic upper intercept (Fig. 6m), which is compatible with a source within the older arc-related magmatic rocks in the region. Sr, Nd and Pb isotope data suggest the presence of a component with a crustal history in the rocks of the Tromøy complex (Knudsen and Andersen, 1999).

6. U–Pb and Hf isotope systematics of detrital and inherited zircons in supracrustal rocks

Accumulated probability plots of detrital zircon ages from metasedimentary rocks from the Østfold–Akershus sector, and of inherited zircons in the Slemmestad metarhyolite are shown in Fig. 7 (data from Table 8, available as an electronic supplement). Fig. 7 also shows provenance age spectra obtained from quartzites, conglomerates and metapelites in the Kongsberg, Telemark and Bamble sector, covering a range of depositional ages from mid-Proterozoic to Sveconorwegian (Knudsen et al., 1997; Haas et al., 1999; Bingen et al., 2001) and from a ca. 1.59 Ga Stora Le-Marstrand fm. metapsammite (Åhäll et al., 1998).

The two metapelitic samples (N6 and N8) show dominant age peaks close to 1.6 Ga, confirming the preliminary results of Mansfeld and Andersen (1999). These age distribution profiles closely resemble that of the Stora Le-Marstrand Formation metapsammite. The dominant source of detritus for these sediments must have been situated within the magmatic arc itself. However, N8 and, in particular, N6 show tails towards older ages, suggesting a minor input from a 1.7- to 1.8-Ga source. The metagreywacke (sample Ø2) differs clearly from the two other Østfold metasediments, showing a wide range of detrital zircon ages with a prominent peak above 1.8 Ga. Although an arc-related source has contributed to this sediment, older sources were also of major importance.

The Slemmestad metarhyolite contains inherited zircons ranging in age from 1.63 to 1.86 Ga. Although a majority of the dated, inherited zircons may be

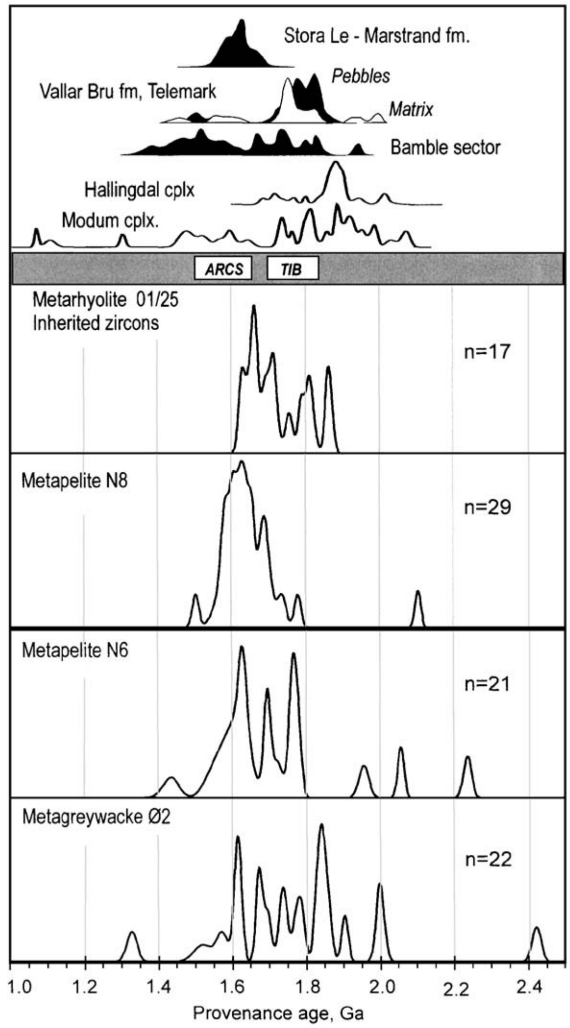


Fig. 7. Accumulated probability diagrams for U–Pb ages of inherited zircons from sample 01/25 and detrital zircons in arc metasediments (N6, N8, Ø2). The upper part shows similar curves (at compressed vertical scale) for continental metasediments from southern Norway (Knudsen et al., 1997; Haas et al., 1999; Bingen et al., 2001) and from the Stora Le-Marstrand fm, SW Sweden (Åhäll et al., 1998). Approximate age ranges for the TIB granitic magmatism and arc magmatism at the Baltic margin (Åhäll et al., 2000 and references therein; this work) are indicated. n = number of grains analysed.

derived from arc-related sources, the rhyolite must have interacted with older material, either during ascent or at the surface during eruption.

Further insight into the nature of the sources of detrital or inherited material in the supracrustal rocks can be obtained by considering the initial Hf isotopic composition of the zircons (Fig. 8). Hf isotope data for granites from the Svecofennian domain of the Baltic Shield were published by Patchett et al. (1981) and Vervoort and Patchett (1996), and some supplementary data from TIB in southern Norway by Andersen et al. (2002a). The limit for Baltic Crust shown in Fig. 8 is defined by the most elevated published initial $^{176}Hf/^{177}Hf$ from the TIB and Svecofennian granites, assuming $^{176}Lu/^{177}Hf = 0.005$ (Andersen et al., 2002a). Data from Andersen et al. (2002a) for tonalitic to granodioritic intrusions dated in this paper have been plotted in Fig. 8 at ages given in Table 6, Lu–Hf data on samples not included in that study are given in Table 7. Compared to the similar diagram given in the original publication, which relied on preliminary age estimates from Andersen et al. (2001b), there is a small shift towards younger ages for the Gjerstadvatn and Justøy tonalites, which with the new ages plot within error of the depleted mantle curve, as does sample 01/19.

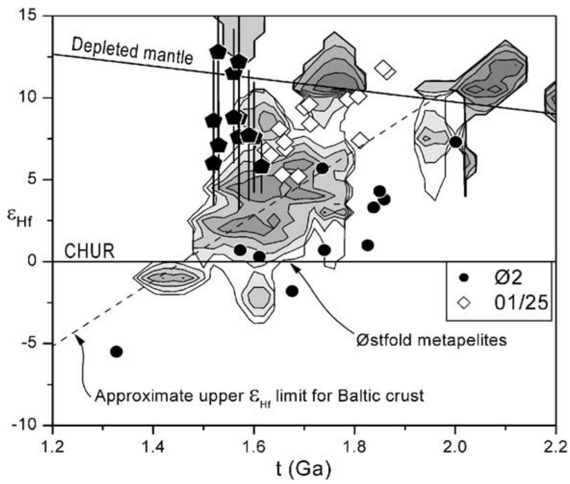


Fig. 8. Initial Hf isotope systematics of zircons in supracrustal, arc related rocks and arc granitioids. Contoured field: Accumulated probability surface of detrital zircons in metapelites (samples N6 and N8), with equally spaced contours at an arbitrary scale. Black dots: Zircons from metagreywacke Ø2. White rhombs: Zircons (both primary crystallized and inherited) from metarhyolite 01/25. Black pentagons with error bars: calc-alkaline granitoids (including gneiss sample 01/19 from the Kongsberg sector). Hf data from Table 7 and Andersen et al. (2002a) have been replotted at ages given in Table 6. The limit of the Baltic shield gives the upper ε_{Hf} values as a function of time expected for pre 1.7 Ga granitoids and other rocks of the Baltic Shield, based on data from Patchett et al. (1981), Vervoort and Patchett (1996) and Andersen et al. (2002a).

Table 7
Lu–Hf data on zircons used for dating

Point no.	$^{176}Hf/^{177}Hf$	1σ	$^{176}Lu/^{177}Hf$	1σ	$^{176}Yb/^{177}Hf$	1σ	$\varepsilon_{Hf}(t)$	2σ
Sample 01/19 (1527 ± 8 Ma)								
ta0119-9	0.282119	0.000018	0.001107	0.000011	0.029537	0.000190	10.9	1.3
ta0119-12	0.282100	0.000025	0.001118	0.000025	0.029950	0.000750	10.3	1.8
ta0119-13	0.282175	0.000032	0.000709	0.000010	0.020777	0.000340	13.3	2.3
ta0119-17	0.282358	0.000043	0.001009	0.000032	0.032848	0.000880	19.5	3.1
ta0119-18	0.282114	0.000017	0.000863	0.000035	0.026699	0.000800	11.0	1.2
ta0119-20	0.282108	0.000029	0.001337	0.000006	0.035290	0.000210	10.3	2.1
ta0119-21	0.282200	0.000020	0.001211	0.000016	0.032924	0.000670	13.7	1.4
ta0119-22	0.282146	0.000030	0.001257	0.000010	0.036091	0.000820	11.7	2.1
ta0119-23	0.282221	0.000027	0.001468	0.000027	0.031201	0.000730	14.2	1.9
Average							12.8	5.9
Sample 01/25 (1585 ± 7 Ma)								
TA0125-3	0.281958	0.000016	0.001780	0.000021	0.056732	0.001100	4.6	1.1
TA0125-5	0.282036	0.000021	0.003333	0.000130	0.102722	0.002500	5.7	1.5
TA0125-14	0.281994	0.000012	0.001913	0.000059	0.058220	0.002100	5.7	0.9
TA0125-15	0.282031	0.000016	0.002644	0.000042	0.083665	0.002600	6.3	1.1
TA0125-17	0.282006	0.000010	0.001580	0.000071	0.048718	0.002000	6.5	0.7
TA0125-19	0.281981	0.000023	0.001791	0.000008	0.058379	0.000550	5.4	1.6
TA0125-22	0.282026	0.000009	0.001594	0.000073	0.047233	0.001500	7.2	0.6
ta0125-26	0.282013	0.000016	0.002946	0.000110	0.084558	0.003500	5.3	1.1
Average							5.8	1.6

For U–Pb data on the individual zircons, see Table 5.

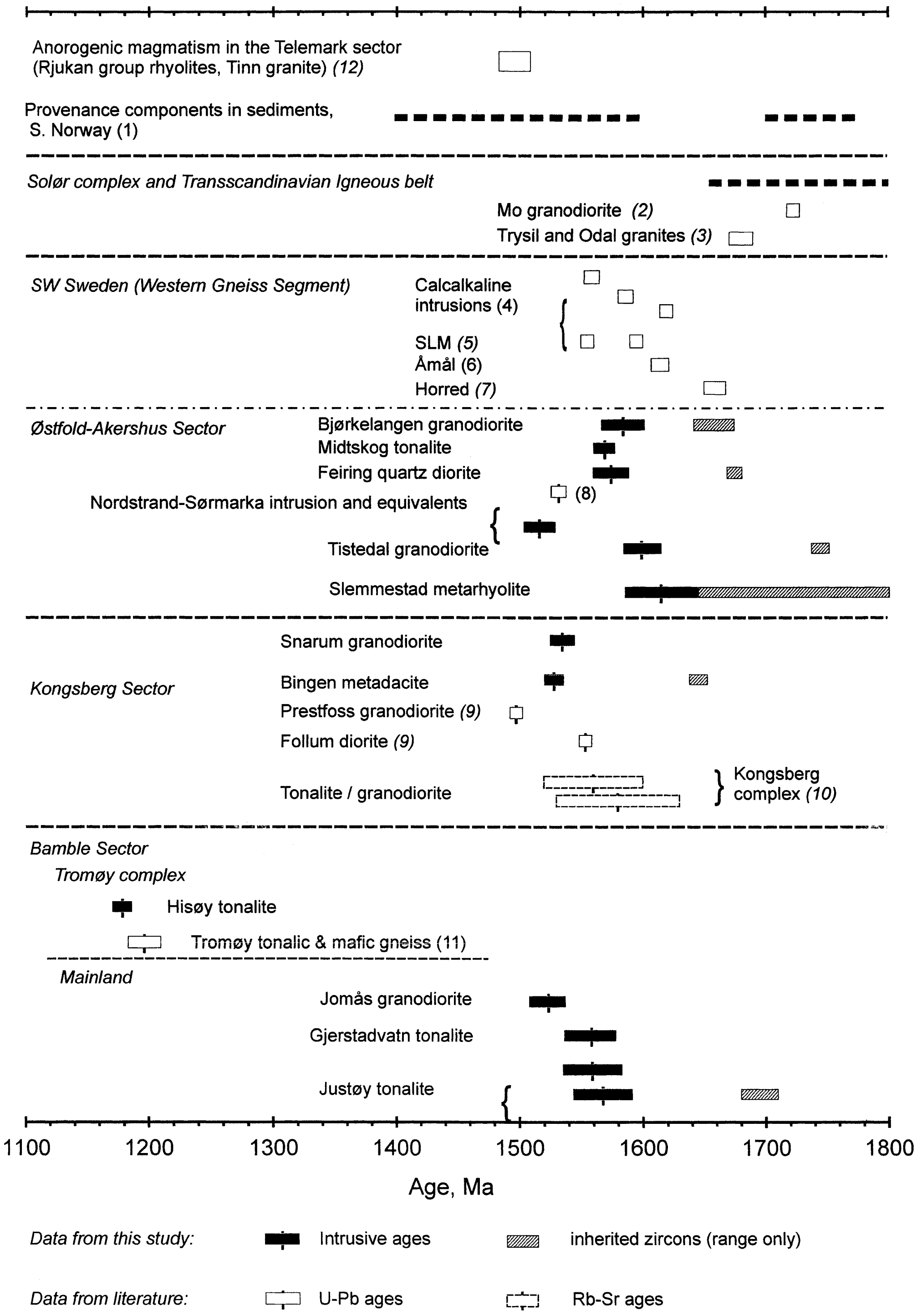
The Østfold metapelites show a considerable range of initial $^{176}Hf/^{177}Hf$ (given as $\varepsilon_{Hf}(t)$ in Fig. 8). A large proportion of the ca. 1.6 Ga grains range from $\varepsilon_{Hf}=+10$ to the upper limit for Baltic crust. However, the total range of variation extends into the field of Baltic crust, and there are subordinate maxima well within this field for 1.7–1.8 Ga and older zircons. Furthermore, there is another, subordinate, but still prominent frequency maximum close to the depleted mantle curve at 1.7–1.8 Ga, suggesting the presence of pre-arc juvenile rocks in the source terrane. Inherited zircons of this age and composition are also present in the Slemmestad metarhyolite, although most zircons in this sample show ε_{Hf} values quite similar to those of the tonalitic to granodioritic gneisses. 1.8-Ga juvenile zircons are absent from the Mysen metagreywacke (Ø2), which yielded zircons plotting well within the field of Baltic crust.

7. Discussion

7.1. Timing and duration of calc-alkaline magmatism in South Norway

Fig. 9 summarizes the timing of mid-Proterozoic calc-alkaline magmatism in the SSD, constrained by the results of the present study and by published data. This compilation shows that the present-day southwestern margin of the Baltic Shield was the site of arc-related magmatism from ca. 1.66 Ga (or possibly even earlier, cf. Mansfeld, 1998) to ca. 1.50 Ga. Of the individual sub-areas, the Østfold–Akershus sector shows the earliest preserved record of magmatic activity (at ca. 1.6 Ga), in the Bamble and Kongsberg sectors, magmatism may have started later, at ca. 1.55–1.57 Ga, but in all three areas, calc-alkaline magmatism has persisted until shortly before

Fig. 9. Compilation of geochronological data from mid-Proterozoic rocks from south Norway and the Western Gneiss Segment of SW Sweden. Sources of data: (1) Knudsen et al. (1997), Haas et al. (1999) and Bingen et al. (2001). (2) Mansfeld (1998). (3) Heim et al. (1996), Sundvoll (personal communication). (4) Åhäll et al. (2000) and references therein. (5) Age bracket from Åhäll et al. (1998). (6) Lundquist and Skiöld (1992). (7) Åhäll et al. (1995) and Connelly and Åhäll (1996). (8) Ali (2000). (9) Nordgulen et al. (1997). (10) Jacobsen (1975) and Jacobsen and Heier (1978). (11) Knudsen and Andersen (1999). (12) Dahlgren et al. (1990), Sigmond (1998) and Andersen et al. (2002b).



Anorogenic magmatism in the Telemark sector
(Rjukan group rhyolites, Tinn granite) (12)
Provenance components in sediments,
S. Norway (1)
Solør complex and Transscandinavian Igneous belt
Mo granodiorite (2)
Trysil and Odal granites (3)
SW Sweden (Western Gneiss Segment)
Calcalkaline
intrusions (4)
SLM (5)
Åmål (6)
Horred (7)
Østfold-Akershus Sector
Bjørkelangen granodiorite
Midtskog tonalite
Feiring quartz diorite
Nordstrand-Sørmarka intrusion and equivalents
(8)
Tistedal granodiorite
Slemmestad metarhyolite
Snarum granodiorite
Kongsberg Sector
Bingen metadacite
Prestfoss granodiorite (9)
Follum diorite (9)
Tonalite / granodiorite
Kongsberg
complex (10)
Bamble Sector
Tromøy complex
Hisøy tonalite
Tromøy tonalic & mafic gneiss (11)
Mainland
Jomås granodiorite
Gjerstadvatn tonalite
Justøy tonalite
1100
1200
1300
1400
1500
1600
1700
1800
Age, Ma
Data from this study:
Intrusive ages
inherited zircons (range only)
Data from literature:
U-Pb ages
Rb-Sr ages

1.5 Ga. An earlier start of magmatism in the Kongsberg sector may be indicated by Rb–Sr isochron ages of 1.58 ± 0.05 and 1.56 ± 0.04 Ga reported for enderbitic and granitic gneisses by Jacobsen and Heier (1978), although the relatively large uncertainties of these ages leave their significance open to debate.

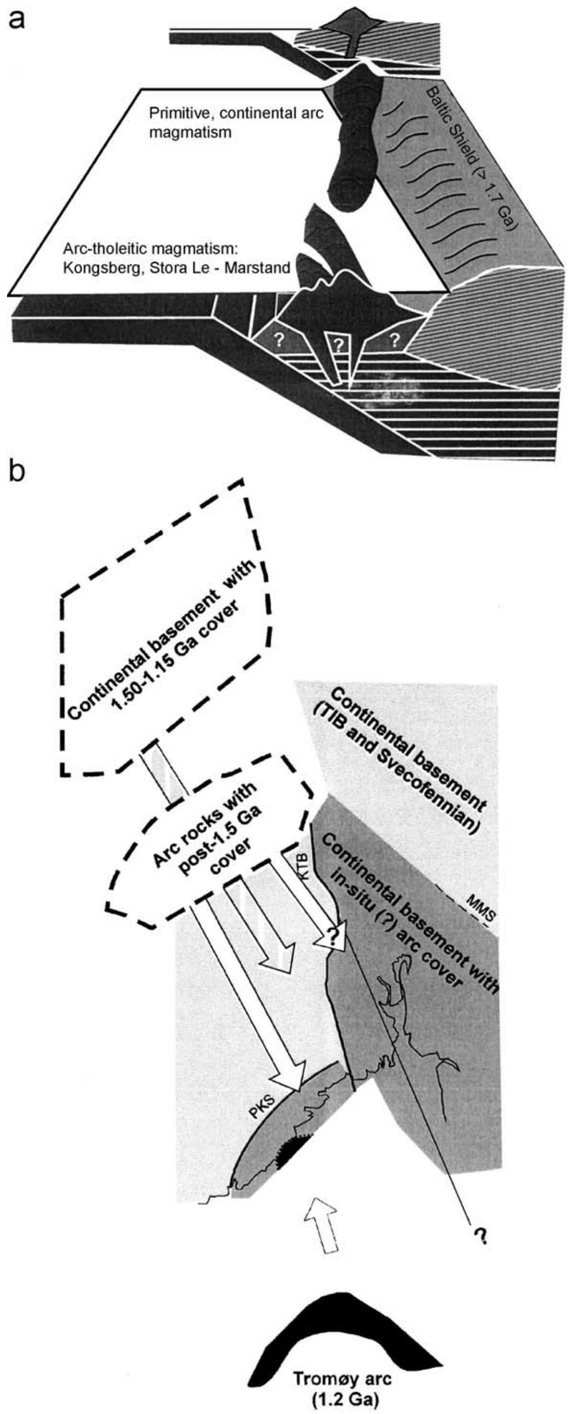
Rocks formed during the final 60–100 Ma of Mesoproterozoic arc magmatism are found on both sides of the Oslo Rift, as well as across other tectonic zones identified in S Norway. Although the data available at present are insufficient for definition of local or region-wide magmatic events within this period, they show no evidence of different age patterns in the east and west, which would be expected if two different subduction zone systems were involved, as required by the microcontinent model. In addition, arc-related magmatism must have persisted in the region until, or possibly even after, the onset of extension related magmatism in central Telemark at 1.51–1.50 Ga (Sigmond, 1998). The present geochronological data thus do not support the existence of a "SW Norwegian craton", distinct from the Baltic Shield, nor do they agree with tectonic models involving an end of active subduction significantly earlier than 1.50 Ga.

7.2. Geochemistry and tectonic setting

The calc-alkaline intrusions analysed in the present study show clear geochemical evidence of having formed in an immature to moderately mature volcanic arc setting. There is no evidence of a change of tectonic regime from east to west, nor from the older to the younger mid-Proterozoic magmatic intrusions.

Åhäll and Persson (1992) and Åhäll et al. (1995) introduced the concept of the Åmål–Horred belt, for the discontinuous areas of calc-alkaline rocks situated to the east of the Stora Le-Marstrand belt in SW Sweden (Fig. 1). These rocks were assumed to have formed in response to continuous subduction in the period 1.66 to 1.60 Ga. From trace element data, Brewer et al. (1998) suggested different tectonic settings for the older Horred (oceanic) and younger Åmål (continental) supracrustals within this belt, and these two sequences were reinterpreted as fragments of two separate magmatic arcs, accreted onto the continental margin in distinct tectonic events. However, metavolcanic rocks from the two arc fragments overlap in most major and trace element parameters. For example, they fall on similar trends of moderate arc maturity in an AFM diagram (Fig. 3b). The only geochemical parameter of potential tectonic importance to show a difference between the two supracrustal rock suites is the Y/Zr ratio of their basaltic members (Fig. 7a of Brewer et al., 1998). However, if basaltic andesites from Åmål and Horred are also included in the comparison, there is considerable overlap between the two arc fragments, which leaves the oceanic arc nature of the Horred event open to debate.

In contrast, metavolcanic rocks of the Stora Le-Marstrand (SLM) belt and the gneisses of the Kongsberg Complex show a consistent arc-tholeiitic character in major and trace elements, suggesting that they formed in a more primitive arc setting (Brewer et al., 1998). However, the presence of pre-1.60 Ga zircons in rocks of the Stora Le-Marstrand metasediments from SW Sweden (Åhäll et al., 1998; see Fig. 7) and in metasedimentary gneisses from Østfold (Fig. 7) suggest that even the primitive magmatic arc must have been situated close to the continental margin. Calc-alkaline intrusions postdating the SLM supracrustals in SW Sweden overlap in major element characteristics with the calc-alkaline intrusions from south Norway. In Cenozoic magmatic arcs, changes from tholeitic to calc-alkaline (and even high-K) magma types with time (e.g. central America, Andes, Cascades, Antilles; Brown, 1982; Brown et al., 1984; Maury et al., 1990) or along the arc (e.g. Tonga-Kermadec-New Zealand, Antilles; Smith et al., 1997; Maury et al., 1990) are well documented, reflecting progressive petrological evolution within the arc system, or changes from an oceanic to a more evolved plate margin along a single subduction zone. The differences in composition observed among the arc-related rocks in S Norway and SW Sweden can therefore neither be interpreted as evidence for the existence of several, compositionally distinct mid-Proterozoic arc systems, nor for several docking events in the region. As low-K magma types are not unique to oceanic arc settings, and no primary basement-cover contacts appear to have been preserved, the type of basement on which the Kongsberg and SLM rocks were deposited remains unconstrained (Fig. 10a).



7.3. Southern Norway and the Baltic Shield

The microcontinent model of Åhäll et al. (2000) (and references therein) requires that the pre-1.5 Ga basement of southern Norway is made up by rocks formed outside of the Baltic Shield. It should therefore have its own internal history and age distribution pattern. If this is true, a discontinuity in geochemical parameters relevant to crustal evolution should be expected across the suture. Such parameters include the Sr, Nd, Pb and Hf isotopic signatures of the deep crust, which show no discontinuities across the Oslo Rift and other potential suture lines in the region (Andersen et al., 2001a, 2002a). The microcontinent model furthermore requires that arc-related rocks formed east and west of the suture line must have formed above separate subduction zones. It may be regarded as highly improbable that two separate, convergent continental margins should show identical trends of magmatic evolution and indistinguishable age patterns and Lu–Hf characteristics, all of which is the case for the Mesoproterozoic arc-related rocks of SW Sweden and S Norway.

Collision (or “docking”) between the Baltic Shield and this hypothetical exotic terrane would terminate active subduction along the margins involved in the event, and in most cases generate a structural and

Fig. 10. (a) Schematic representation of the situation along the (present-day) SW margin of the Baltic Shield in the mid-Proterozoic. A subduction zone (or system of subduction zones) along which oceanic lithosphere was consumed, and whose position with respect the continental margin changed along strike, can adequately explain the more or less simultaneous production of arc-tholeitic and more evolved calc-alkaline rocks in the region between 1.60 and 1.50 Ga. The question marks on the basement of the Kongsberg complex and SLM indicates that its oceanic or continental nature remains unconstrained by data. The present crustal configuration is due to lateral terrane displacement in Sveconorwegian time (see Haas et al., 1999; Bingen et al., 2001; Andersen et al., 2002a), see (b) for visualization of a possible scenario. (b) Possible Sveconorwegian terrane displacements in SE Norway, based on data from Haas et al. (1999), Knudsen and Andersen (1999), Bingen et al. (2001) and the present work. The Telemark sector consists of continental rocks with a 1.50 Ga and younger volcanic and sedimentary cover, the Bamble sector of mid-Proterozoic arc rocks with a post-1.5 Ga sedimentary cover (Haas et al., 1999, 2002). Both of these originate further north than their present position, during the Sveconorwegian orogeny, both are displaced southwards relative to the continental margin, and the Bamble sector is thrust above the continental rocks of Telemark.

metamorphic overprint in the volumes of crust involved in the process, as well as a change in the type of magmatism. The age data summarized in Fig. 9 indicate that active subduction continued until 1.50 Ga, when it was succeeded by magmatism related to crustal extension (Sigmond et al., 1997; Andersen et al., 2002b). No evidence of Gothian post-collisional magmatism, or of high-pressure metamorphism attributable to collision has been reported from S Norway or SW Sweden.

The present data thus add to the previously published evidence against a microcontinent model for S Norway (e.g. Andersen et al., 2001a, 2002a). Southern Norway must therefore be regarded as an integral part of the Baltic Shield, whose present position may be controlled by Sveconorwegian strike-slip displacement, as suggested by Haas et al. (1999) and Bingen et al. (2001). The sources of detrital and inherited zircons with a continental affinity should therefore be sought within the Baltic Shield. Source rocks with ages in the range 1.7–1.9 Ga include TIB-related granitoids or sediments derived from such granites (e.g. Haas et al., 1999). The presence of 1.7–1.8-Ga zircons with initial Hf isotopic compositions near the depleted mantle curve is a minor, but significant feature of the populations of detrital and inherited zircons studied from the Østfold–Akershus sector (Fig. 8). Such zircons must have been derived from a juvenile source within the Baltic Shield or at its margin. Possible source rocks are mantle-derived igneous rocks within the TIB (e.g. Andersson, 1997), and possible pre-1.7 Ga arc-fragments. One such arc fragment, occurring as an enclave of calc-alkaline granitoids in 1.67 Ga TIB granite in the northern part of the Østfold–Akershus sector (Fig. 1), was dated to 1.73 Ga by Mansfeld (1998). Unfortunately, no Hf isotope data have been published from any of these rocks.

7.4. A cordillera model for the Baltic margin

As an alternative to the tectonic models for the SSD suggested by Åhäll et al. (1998) and Brewer et al. (1998), a scenario for the evolution of the Baltic margin is sketched in Fig. 10. In the period 1.66 to 1.50 Ga (Fig. 10a), a system of magmatic arcs built up at or beyond the margin of the Baltic continent, on continental and, possibly, oceanic crust in close proximity to the continent, in response to the subduction of oceanic lithosphere, which may have started around 1.66 Ga, or even significantly earlier (Mansfeld, 1998). This scenario represents a cordillera-type tectonic setting, the best modern analogues of which can be found along the western margin of the American continental plates, which has been situated above an active subduction zone since the Mesozoic; magmatic rocks have formed in response to different stages of evolution of the subduction zone (Lipman, 1992 and references therein). The present data show no indication of a continent–continent collision event in the mid-Proterozoic, but individual arc fragments may of course have been thrust onto the continental margin, as part of arc–continent collision events at an evolving, convergent plate margin.

Åhäll et al. (1998) suggested that final ‘‘cratonization’’ of the Åmål, Horred and Stora Le-Marstrand arc terranes occurred prior to 1.53 Ga. The emplacement of calc-alkaline intrusions in south Norway as late as 1.50 Ga suggests that active subduction must have continued along parts of the Baltic margin after 1.53 Ga (Fig. 9). However, there is also abundant evidence of anorogenic intracontinental magmatism, both in S Norway and the Western Segment of SW Sweden at 1.51–1.50 Ga and later (Sigmond et al., 1997; Åhäll and Connelly, 1998; Nijland et al., 2000; Andersen et al., 2002b). This may indicate that termination of active subduction was not synchronous along the length of the margin, and that some areas have seen back-arc type extension towards the end of the period of active subduction. The example of western North America shows that changes in the geometry of the subducting oceanic plate(s) can cause changes from compressional to extensional tectonomagmatic regimes within the continental plate, as a function of time, or of position along the trend of the continental margin. Thus, active subduction in an area can end without the need of a continental collision event or of ‘‘docking’’ of an exotic, continental fragment.

The life time of a single volcanic centre within an arc system will in general be much shorter than the total life time of an arc, on the order of a few millions or tens of millions of years. Therefore, when reconstructing the plate-tectonic setting of Precambrian metaigneous rocks, it is important to keep the fragmentary nature of the preserved geological record in mind. Dating rocks from discontinuous arc fragments to ages differing by more than the precision of the analytical method is not sufficient evidence to conclude that the fragments represent different magmatic arcs, that they formed in response to distinct subduction events, or that they were accreted onto the continent in separate events.

At some time during the Sveconorwegian orogenic period, certainly before deposition of supracrustal rocks in the Telemark Sector at ca. 1.15 Ga (Laajoki et al., 2002), the present-day Telemark and Bamble Sectors were displaced southwards from their original position along the Baltic margin (Torske, 1985; Haas et al., 1999), along a shear zone that may be partly obscured by the Oslo Rift (Bingen et al., 2001). Similar, large-scale lateral displacement of portions of the American margin is ongoing today. At ca. 1.18 Ga, a primitive island arc was developing offshore, to be brought in tectonic contact with the Baltic Shield in time for the arc-related rocks to be metamorphosed in the same high-grade event as the rocks of the mainland Bamble Sector at ca. 1.10–1.13 Ga (Kullerud and Dahlgren, 1993; Knudsen and Andersen, 1999).

7.5. The relationship to Laurentia at 1.6 Ga

Baltica and Laurentia were parts of the Proterozoic supercontinent Rodinia (e.g. Condie, 1997). In classical reconstructions, the present-day western margin of Baltica is placed adjacent to Laurentia, in which case the Mesoproterozoic arc-related magmatism in SW Sweden and S Norway can be traced to Laurentia (e.g. Karlstrom et al., 2002). However, neither geological correlation nor paleomagnetic evidence is conclusively in favour of such reconstructions, and an alternative reconstruction has been proposed, in which the present-day western margin of Baltica faces away from Laurentia (Hartz and Torsvik, 2002; Torsvik, 2003). In eastern Canada, there is abundant evidence of arc-related magmatism prior to 1650 Ma (the Labradorian orogeny), but a conspicuous absence of calc-alkaline rocks of 1620–1500 Ma age (Åhäll and Gower, 1997 and references therein). The lack of temporal overlap with the calc-alkaline magmatism in S Norway and SW Sweden suggests that the SSD and the present-day eastern margin of the Canadian Shield have formed in response to different subduction events, as in the reconstruction of Hartz and Torsvik (2002). Thus, a correlation between the two crustal domains in the mid-Proterozoic based on arc-related magmatic events may be questionable.

7.6. Towards a consistent terrane analysis for the Mesoproterozoic SSD?

Based on results from SW Sweden, Åhäll and Gower (1997) divided the Southwest Scandinavian Domain into three tectonostratigraphic terranes (from east to west: “Klaraälven”, “Ätran” and “Idefjord”). The status of the two eastern terranes as separate tectonic units has been questioned by more recent data from SW Sweden (e.g. Andersson and Wikström, 2001; Söderlund et al., 2002), a criticism which has been implicitly accepted by Åhäll and Larson (2000) and Åhäll et al. (2000), who substituted the term “Klaraälven-Ätran segment”. The “Idefjord terrane” as defined by Åhäll and Gower (1997) referred to the area between the Mylonite Zone (i.e. the southern continuation of the MMS in Fig.1) and the Oslo Rift, which was supposed to consist of juvenile Mesoproterozoic rocks that were accreted onto the Baltic Shield by “Gothian” tectonism. However, in the sense of Åhäll and Gower (1997) and Brewer et al. (1998), the “Idefjord terrane” is built up of several separate arc fragments (Åmål, Horred, Stora Le-Marstrand), which formed at different times and in different tectonic environments, and which were accreted onto the Baltic Shield in separate events. By normal definitions (e.g. Jones et al., 1983; Howell, 1995), each of these should be given status as a separate accreted terrane. As originally defined, the “Idefjord terrane” therefore can be neither a single *accreted terrane* nor a *composite terrane* amalgamated in an offboard position and later accreted onto the continent. If so, its status is reduced to that of a regional descriptive unit defined by geographical boundaries, not unlike the “sectors” or “segments” used in the present paper, and the “Klaraälven-Ätran segment” in the sense of Åhäll et al. (2000).

In view of the results of Andersen et al. (2002a) and the present paper, it seems probable that at least some of the calc-alkaline igneous rocks of SW Sweden and S Norway make up a younger igneous cover above an older continental margin basement, which is most likely indistinguishable from the "Klaraälven-Ätran segment" further east. If so, these rocks should not be assigned to a separate terrane (cf. Jones et al., 1983).

The Mesoproterozoic accretion of the "Idefjord terrane" was questioned by Bingen et al. (2001), who argued that the present-day crustal architecture of the SSD is largely due to Sveconorwegian tectonic processes. Bingen et al. (2001) introduced a hybrid nomenclature, with tectonic ('terrane') and geographic-descriptive ('sector') components. Furthermore, much of their terrane analysis is based on provenance ages from post-1.51 Ga sedimentary cover sequences, which makes their analysis irrelevant for the Mesoprotoerozoic situation. Whereas some of their terms probably refer to valid Sveconorwegian (as opposed to Mesoproterozoic) terranes (e.g. the Rogaland-Hardangervidda terrane), others do not. An example of an invalid terrane would be Telemark-Bamble (Bingen et al., 2001)—the two units were brought in tectonic contact only in late Sveconorwegian time (e.g. Starmer, 1993, Andresen and Bergundhaugen, 2001; Andersen et al., 2002a), and should not be assigned to a single Sveconorwegian terrane (Fig. 10b). The Bamble sector itself may be composed of two lithotectonic terranes: Mainland Bamble with a pre-Sveconorwegian crustal history, and a Sveconorwegian arc fragment along the central part of the coast (Tromøy and adjacent islands).

It is our opinion that a consistent terrane analysis of the Mesoproterozoic SSD can only be carried out when the nature of Sveconorwegian tectonic displacements has been further clarified, and the age and nature of the basement beneath the cover sequences studied by Bingen et al. (2001) have been identified. In the meantime, we see no reason to maintain the concept of an "Idefjord terrane"; at the present state of knowledge, an appropriate geographic-descriptive term with no tectonic connotations should be substituted. The traditional use of different regional terms by Swedish and Norwegian authors complicates the choice of descriptive name; the terminology in the present paper is far from ideal, but it is preferred for the time being, as it acknowledges well-established usages in both Swedish and Norwegian literature, as well as the problems inherent in the terrane analyses by Åhäll and Gower (1997) and Bingen et al. (2001).

8. Conclusions

In the early Mesoproterozoic, south Norway and adjacent parts of southwest Sweden were part of a cordillera-type continental margin. Voluminous intermediate to felsic calc-alkaline igneous rocks formed as part of this magmatic arc system, fragments of which have been preserved as calc-alkaline gneisses. Together with recently published data (Nordgulen, 1999; Åhäll et al., 2000; Andersen et al., 2002a and references therein), the results of the present study indicate that the margin of the Baltic Shield was the site of more or less continuous calc-alkaline magmatism, at least from 1.66 to 1.50 Ga. Rocks formed during the final 100 Ma of this period are recognized across southern Norway. There are no systematic differences in the timing of arc-related magmatic activity across the Oslo rift or any of the major Precambrian shear zones in the area, which calls into question several tectonic models in which parts of south Norway belong to an exotic continental fragment brought in contact with the Baltic Shield some time between 1.58 and 1.50 Ga.

The calc-alkaline intrusions in S Norway show the major and trace element characteristics of intrusive rocks formed in a moderately evolved volcanic arc setting, and represent a more evolved stage of a magmatic arc evolution that may have started with eruption of the low-K mafic volcanic rocks of the Stora-Le Marstrand belt in SW Sweden and their possible equivalents in the Kongsberg complex of south Norway. There are no systematic changes in geochemical character from east to west, or from the older to the younger intrusions, and there is thus no indication that merging of the Baltic margin with a separate, "SW Norwegian craton" took place during the period considered here.

U–Pb and Hf isotope data on detrital and inherited zircons indicate that young, arc-related crust was an important source of clastic input to sedimentary basins along the Baltic margin, and show that older sources within the shield itself also contributed. The most important of these can be tentatively identified as 1.7–1.8 Ga granites and mafic rocks of the Transscandinavian Igneous Belt.

The present findings support the evidence of Knudsen et al. (1997), Haas et al. (1999), Andersen and Knudsen (2000) and Andersen et al. (2001a) in establishing a continuity of crustal composition and history across the Oslo Rift and the large Precambrian shear zones of southeast Norway. It may now be concluded with confidence that the central parts of present-day south Norway (i.e. the Telemark, Bamble and Kongsberg Sectors) were part of the Baltic continental margin prior to 1.6 Ga (and probably well before); probably this is also true for western southern Norway (Andersen et al., 2001a).

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