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# Sveconorwegian crustal underplating in southwestern Fennoscandia: LAM-ICPMS U–Pb and Lu–Hf isotope evidence from granites and gneisses in Telemark, southern Norway

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**Abstract**

Laser ablation ICPMS U–Pb and Lu–Hf isotope data on granitic-granodioritic gneisses of the Precambrian Vråvatn complex in central Telemark, southern Norway, indicate that the magmatic protoliths crystallized at 1201±9 Ma to 1219±8 Ma, from magmas with juvenile or near-juvenile Hf isotopic composition ($^{176}Hf/^{177}Hf$=0.2823±11, epsilon-Hf>+6). These data provide supporting evidence for the depleted mantle Hf-isotope evolution curve in a time period where juvenile igneous rocks are scarce on a global scale. They also identify a hitherto unknown event of mafic underplating in the region, and provide new and important limits on the crustal evolution of the SW part of the Fennoscandian Shield. This juvenile geochemical component in the deep crust may have contributed to the 1.0–0.92 Ga anorogenic magmatism in the region, which includes both A-type granite and a large anorthosite–mangerite–charnockite–granite intrusive complex. The gneisses of the Vråvatn complex were intruded by a granitic pluton with mafic enclaves and hybrid facies (the Vrådal granite) in that period. LAM-ICPMS U–Pb data from zircons from granitic and hybrid facies of the pluton indicates an intrusive age of 966±4 Ma, and give a hint of ca. 1.46 Ga inheritance. The initial Hf isotopic composition of this granite ($^{176}Hf/^{177}Hf$=0.28219±13, epsilon-Hf=−5 to +6) overlaps with mixtures of pre-1.7 Ga crustal rocks and juvenile Sveconorwegian crust, lithospheric mantle and/or global depleted mantle. Contributions from ca. 1.2 Ga crustal underplate must be considered when modelling the petrogenesis of late Sveconorwegian anorogenic magmatism in the region.


*Keywords:* Fennoscandia (Baltic Shield); Granite; Zircon; U–Pb ages; Lu–Hf

## 1. Introduction

According to most models of continental evolution, the global volume of continental crust has increased throughout Proterozoic time (e.g., Condie, 2005, and references therein). In terms of volume, lateral crustal growth by magmatism and accretion of juvenile terranes

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along convergent plate margins may be the most important process of continental growth. However, vertical growth by crustal underplating, in which mantle-derived, mafic magmas are emplaced into the lower crust, may also contribute significantly (e.g., Frost et al., 2001). Crustal underplating may occur in different tectonic settings, including regions of intracontinental magmatism, continental rifts, and above active subduction zones (Condie, 2005).

Throughout the late Archaean and early Proterozoic, the Fennoscandian Shield grew laterally by convergent margin processes (e.g., Nironen, 1997; Brewer et al., 1998; Karlstrom et al., 2001; Andersen et al., 2004a). After ca. 1.5 Ga, crustal growth by arc magmatism and lateral accretion was only of local significance (Smalley and Field, 1985; Bingen and van Bremen, 1998; Knudsen and Andersen, 1999). During the Sveconorwegian/Grenvillian period (here broadly defined as 1.25–0.9 Ga), southwestern Fennoscandia was affected by several anorogenic igneous events of regional extent (see compilation by Andersen, 2005). Different studies based on Rb–Sr, Sm–Nd, Pb and Re–Os isotope systematics have lead to contrasting conclusions on the involvement or absence of new, mantle-derived material in the petrogenesis of these rocks (see, for example, Schiellerup et al., 2000; Andersen and Griffin, 2004 for contrasting views). Part of this ongoing controversy is due to the poor temporal resolution of the whole-rock-based isotopic methods used (e.g., Andersen, 1997).

Emplacement of mantle-derived material into old continental crust will modify the bulk radiogenic isotope signature of the lower crust, moving the isotopic compositions of Sr, Nd, Hf and Pb in the direction of mantle compositions (e.g., Frost et al., 2001). Shifts in the radiogenic isotope signature of these elements of the lower crust will, in turn, affect the corresponding isotopic compositions in magmas formed by deep crustal anatexis. Radiogenic isotope data from anatectic granites may thus be used to identify mantle-derived components introduced into the deep crust in the past, and ideally to date their emplacement and map their spatial distribution. The Hf isotope system in zircon has proven a very useful tracer of mantle-derived and crustal components in the source regions of granitic magmas (Griffin et al., 2002). Zircon has the advantage of being datable by the U–Pb method, and its chemical and mechanical robustness and low Lu/Hf ratio causes it to retain a faithful memory of the Hf isotopic signature of the melt from which it crystallized (Kinney and Maas, 2003). When U–Pb and Lu–Hf isotope data from igneous zircon are used as a combined chronometer and petrogenetic indicator, ambiguities inherent in whole-rock-based radiogenic isotope studies can largely be avoided.

In the present study, new Lu–Hf and U–Pb data from zircons in two groups of granitoids from central Telemark in southern Norway are presented. The new data identify and date a previously unknown event of crustal underplating in southwestern Fennoscandia and provide constraints on the nature of the source material in the youngest, widespread event of Sveconorwegian anorogenic magmatism in Fennoscandia.

## 2. Geologic setting

Fennoscandia (a.k.a. Baltica, the Fennoscandian Shield or the Baltic Shield) consists of an Archaean core in northwestern Russia and northern Finland, successively ringed by Paleoproterozoic and Mesoproterozoic domains toward the southwest. The Precambrian crust of southern Norway and southwestern Sweden (Fig. 1) was built by ca. 1.85 to 1.66 Ga granitoids (Transscandinavian Igneous Belt — TIB) and their reworked equivalents, 1.73–1.5 Ga (Gothian), arc-related metaigneous and metasedimentary rocks, and several 1.7–1.15 Ga intracontinenal volcanosedimentary basins (Lundqvist, 1979; Åhäll and Larson, 2000; Koistinen et al., 2001; Bingen et al., 2001; Laajoki et al., 2002; Andersen et al., 2004a). In southern Norway, there is evidence of intracontinental magmatism at 1.51–1.47 Ga (Nordgulen, 1999; Andersen et al., 2002c), 1.16–1.12 Ga (Bingen et al., 2001, Laajoki et al., 2002; Bingen et al., 2003), and at 1.0–0.92 Ga (Eliasson and Schöberg, 1991; Schärer et al., 1996; Nordgulen, 1999; Andersen et al., 2002a; Andersen and Griffin, 2004). In the final event of intracontinental anorogenic magmatism at ca. 1.0 to 0.92 Ga, a large anorthosite–mangerite–charnockite–granite (AMCG) complex (the Rogaland Intrusive Complex—RIC), and a widespread suite of potassium-rich granites were emplaced (Fig. 1; Killeen and Heier, 1975; Eliasson and Schöberg, 1991; Schärer et al., 1996; Andersen et al., 2002a; Andersen and Griffin, 2004).

The Precambrian bedrock of southwestern Fennoscandia (Fig. 1) can be divided into blocks separated by ductile Precambrian shear zones and Phanerozoic brittle faults (Andersen, 2005). The *Telemark block* is limited in the west toward the *Hardangervidda–Rogaland block* by the Mandal–Ustaoset shear zone (MANUS in Fig. 1), against the *Kongsberg–Marstrand block* to the east by a system of shear zones and brittle faults (the “Kongsberg–Telemark boundary”, KTB) and against the *Bamble–Lillesand block* to the southeast by the Kristiansand–Porsgrunn shearzone and fault (KPS/F).

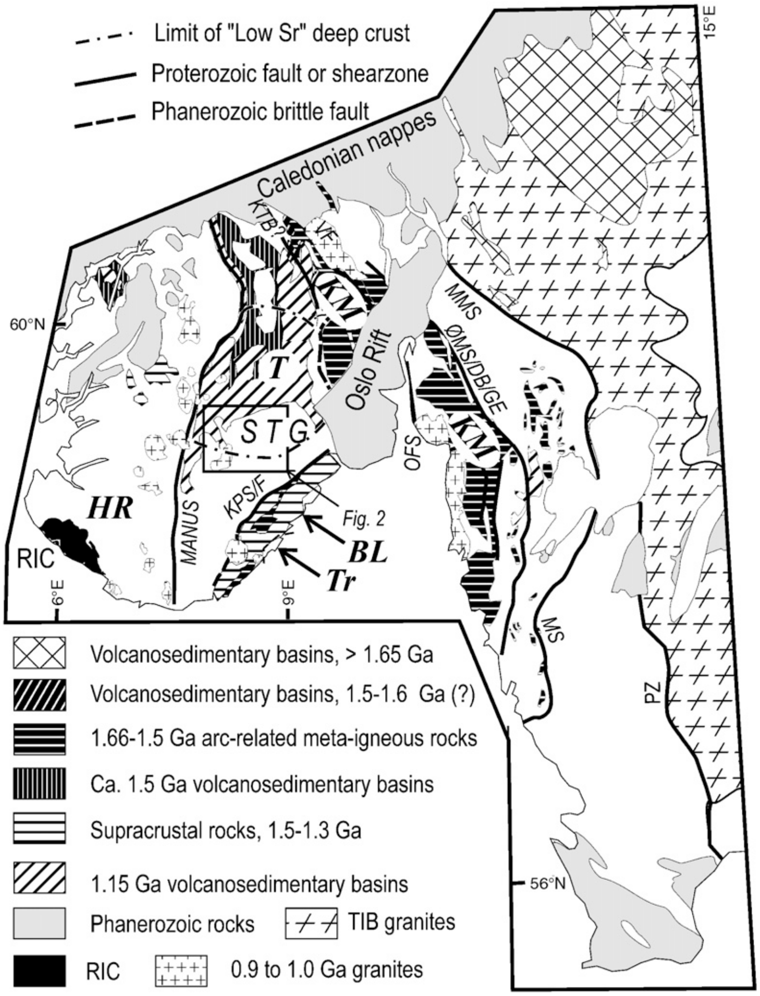



Fig. 1. Sketch map of the Precambrian geology of southern Norway and adjacent parts of Sweden, after Koistinen et al. (2001). Limits of "low-Sr" deep crust in central Telemark from Andersen et al. (2001). *Tectonic blocks* (after Andersen, 2005): HR: Hardangervidda-Rogaland, T: Telemark, BL: Bamble-Lillesand, Tr: Tromøy, KM: Kongsberg-Marstrand. *Other geological units*: STG: South Telemark Gneisses, RIC: Rogaland Intrusive complex (AMCG rocks). Late Sveconorwegian "Low Sr" granites (Andersen et al., 2001) are found only in central Telemark, north of the dash-dot limit. *Tectonic lines:* KPS/F: Kristiansand–Porsgrunn shearzone/brittle fault. ØMS/DB/GE: Ørje Mylonite Zone (Berthelsen et al., 1996)/Dalsland Boundary Thrust/Göta Elv Zone (Park et al., 1991). MS: Mylonite Zone (Gaàl and Gorbatschev, 1987), MMS: Mjøsa–Magnor Shear Zone (Nordgulen, 1999). PZ: Protogine Zone (Gaàl and Gorbatschev, 1987), MANUS: Mandal–Ustaoset line (Sigmond et al., 1997). KTB: Informal "Kongsberg–Telemark Boundary", VF: Åmot–Vardefjell shear zone (Bingen et al., 2001). OFS: Oslofjord shear zone (Berthelsen et al., 1996).

Calc-alkaline gneisses (metatonalite, metagranodiorite and associated amphibolite) are abundant in southwestern Fennoscandia, including the Kongsberg–Marstrand and Bamble–Lillesand blocks east and southeast of the Telemark block (Fig. 1), where they represent the metamorphosed and deformed products of Gothian, arc-related magmatism (Andersen et al., 2004a). There is no evidence for the presence of Gothian, calc-alkaline rocks in the Telemark block. A suite of deformed granitoids with a calc-alkaline affinity from southeastern Telemark (the Drivheia granitic gneiss), initially thought to be of Gothian age (Smalley and Field, 1985), has since been dated to ca. 1.2 Ga (Heaman and Smalley, 1994). Metavolcanic and metasedimentary rocks, known as the Telemark supracrustals, crop out in the central and northern parts of the Telemark block (Dons, 1960; Laajoki et al., 2002). This supracrustal belt consists of (1) an older sequence, comprising ca. 1.50 Ga metarhyolite (Rjukan group), which is possibly related to continental rifting (Sigmond et al., 1997), overlain by quartzite-dominated strata of poorly constrained age (the Vindeggen group, deposited between ca 1.50 and

1.15 Ga), and (2) several 1.17–1.12 Ga volcanosedimentary sequences (Oftefjell, Lifjell, Heddal, Høydalsmo, Nore groups, Eidsborg formation; Laajoki et al., 2002; Bingen et al., 2003). In the present work, these 1.17–1.12 Ga supracrustal rocks are collectively referred to as "the younger Telemark supracrustals". From geochemical data on volcanic members, a continental back-arc setting has been suggested for the Telemark block at 1.16 Ga (Brewer et al., 2002). In the north, the Rjukan group metarhyolites are penetrated by a belt of felsic to mafic intrusions, a granitic member of which (the Tinn granite) has been dated at 1476±13 Ma by SIMS U–Pb on zircons (Andersen et al., 2002c). South of the main supracrustal belt is a large area dominated by granitic gneisses, informally named the South Telemark Gneisses (STG in Fig. 1). The age and nature of the protoliths of the STG are uncertain, both reworked Rjukan group rhyolites and 1.47 Ga granitoids have been suggested (Ploquin, 1980; Andersen et al., 2002c).

The present study is concerned with rocks from the northern part of the STG, in the Vrådal–Kviteseid area (Fig. 2), which is made up of gneisses that are penetrated by a late, undeformed Sveconorwegian granitic intrusion with mafic enclaves (the *Vrådal granite*, Sylvester, 1964, 1998). The gneisses of the Vrådal–Kviteseid area belong to the *Vråvatn complex*, as defined by Laajoki et al. (2002), and are predominantly granitic with minor granodioritic components (Dons and Jorde, 1978). Deformed and metamorphosed volcanosedimentary sequences crop out north and south of the area. In the north, a metarhyolitic member has been dated at 1155±2 Ma (TIMS-ID U–Pb on zircon, Laajoki et al., 2002). The isolated outcrop area of metasupracrustals to the south (the Nissedal supracrustals; Mitchell, 1967) remains undated.

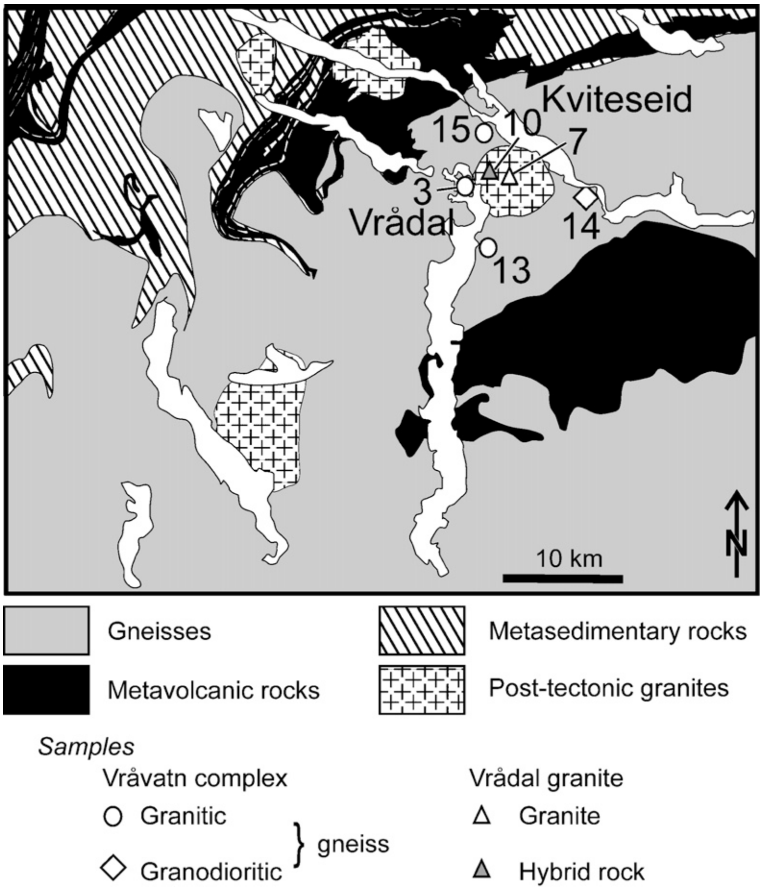



Fig. 2. The geology of the Vrådal–Kviteseid area, Telemark (Dons and Jorde, 1978; Sylvester, 1998), showing the distribution of metasupracrustal rocks, South Telemark Gneisses and late Sveconorwegian granites. Sample localities are identified by the last two digits of the sample number (i.e. *XX* in TA01-*XX*). Metasediments to the N and NW of the Vrådal area belong to the younger Telemark supracrustals (ca. 1.15 Ga), those to the south of the area remain undated. Blank areas denote lakes.

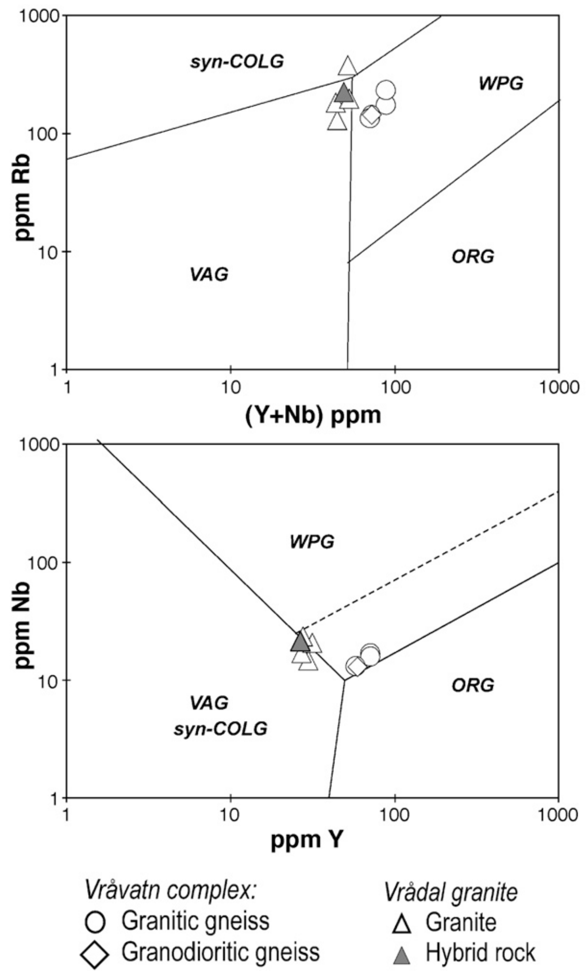



Fig. 3. Y–Nb–Rb concentrations in granites and gneisses from the Vrådal–Kviteseid area used as a tectonic indicator (Pearce et al., 1984). Symbols as in Fig. 2. Acronyms: ORG: Orogenic granites, WPG: Within plate granites, VAG: Volcanic Arc Granites, syn-COLG: Syn-collisional granites. a: Y+Nb vs. Rb diagram. b: Y vs. Nb diagram.

### 2.1. The material studied

#### 2.1.1. Vråvatn complex

Samples TA01-3, TA01-13 and TA01-15 are medium-grained, felsic two-feldspar granitic gneisses with a slightly chloritized black biotite as the only abundant mafic silicate mineral. TA01-3 was sampled at Fossøyi island in Lake Nisser, and TA01-13 in the southern parts of Vrådal. TA01-15 is a similar, granitic gneiss from Kviteseid (Fig. 2). TA01-14 is a medium-grained, greyish white, biotite-bearing granodioritic gneiss, which was sampled in a roadcut at Kviteseidvatn, southeast of the Vrådal granite.

#### 2.1.2. Vrådal granite

Two samples were selected from a larger series of samples of the Vrådal granite (Sylvester, 1964, 1998), one representing a medium-grained, unfoliated two-feldspar biotite granite (TA01-07), the other (TA01-10) a fine-grained, plagioclase-rich rock, with biotite and abundant titanite. This type of rock has been interpreted as a hybrid formed by mixing of granitic and mafic melts (Sylvester, 1998).

In Y+Nb–Rb and Y–Nb discrimination diagrams, the granitic and granodioritic gneisses of the Vråvatn complex plot in the within-plate granite (WPG) field, whereas the samples from the Vrådal granite straddle the limit between WPG and the fields of volcanic arc and syn-collisional granites (Fig. 3, data from Table 1). In contrast, calc-alkaline Mesoproterozoic metaigneous gneisses from the Bamble–Lillesand and Kongsberg–Marstrand blocks plot within the VAG field in the Y+Nb–Rb diagram (Andersen et al., 2004a).

## 3. Results

### 3.1. Analytical methods

Sample processing was done at the Laboratory of Isotope Geology, Mineralogical–Geological Museum, University of Oslo. Zircons were separated from the <250 µm size-fraction using standard heavy liquid and magnetic separation methods. Selected zircons were mounted in epoxy and polished. All imaging and analytical work was done at the GEMOC Key Centre, Department of Earth and Planetary Sciences, Macquarie University, NSW, Australia. Zircons were imaged on a Cameca SX50 electron microprobe prior to analysis,

Table 1
Y, Nb and Rb concentrations

| | | Y | Nb | Rb |
|---|---|---|---|---|
| *Vråvatn complex* | | | | |
| TA01-3 | Granitic gneiss | 73 | 17 | 171 |
| TA01-13 | Granitic gneiss | 73 | 16 | 228 |
| TA01-14 | Granodioiritic gneiss | 60 | 13 | 141 |
| TA01-15 | Granitic gneiss | 59 | 13 | 132 |
| *Vrådal granite* | | | | |
| TA01-7 | Granite | 27 | 22 | 219 |
| TA01-8 | Granite | 28 | 24 | 369 |
| TA01-9 | Granite | 32 | 21 | 194 |
| TA01-10 | Hybrid rock | 30 | 15 | 126 |
| TA01-11 | Granite | 28 | 16 | 173 |

Analysis by XRF on pressed powder pellets, Department of Geosciences, University of Oslo. T. Winje, analyst.

using a combination of backscattered electron and cathodoluminescence imaging.

U–Pb analyses were made using a Hewlett–Packard 4500 quadrupole ICPMS, coupled to Merchantec/New Wave 213 nm Nd–YAG laser microprobe. Analytical methods follow Jackson et al. (2004). Common lead corrections were made using the algorithm of Andersen (2002), assuming recent lead loss. The detection limit of the method is ca. 0.2% common $^{206}$Pb. This correction method fails for zircons which have a more complicated lead-loss history; such grains are left uncorrected, as indicated in Appendix 1. Lu–Hf analyses were made using a Nu Plasma multicollector ICPMS in the time-resolved mode, with a Merchantek/New Wave 193 nm eximer laser microprobe, by methods described by Griffin et al. (2000) and Andersen et al. (2002b). Epsilon-Hf values were calculated assuming $\lambda(^{176}\mathrm{Lu})=1.93 \cdot 10^{-11}\ \mathrm{a}^{-1}$ and the CHUR parameters of Blichert-Toft and Albarède (1997). The depleted mantle model of Griffin et al. (2000) was adopted. This curve is very

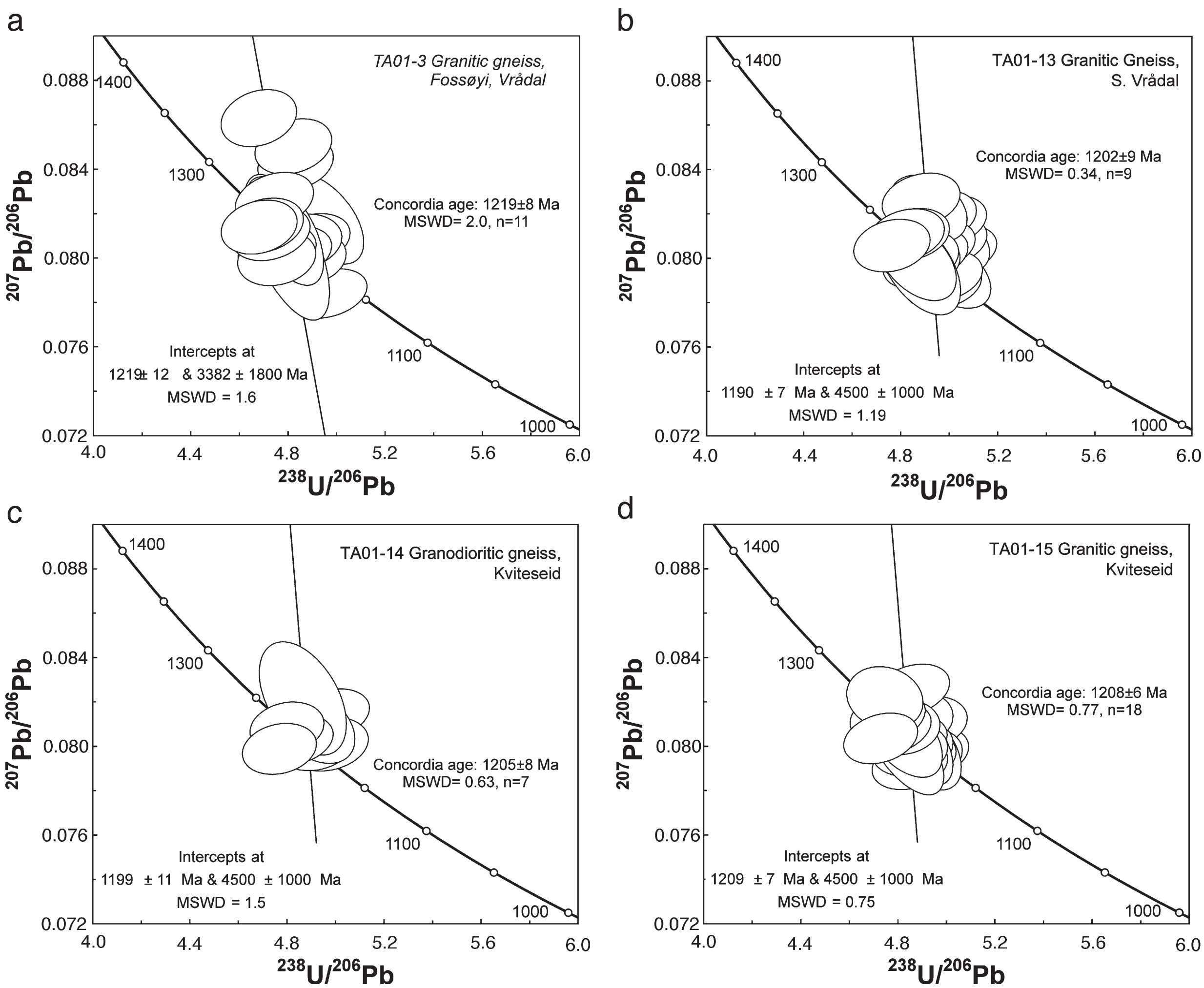



Fig. 4. Tera-Wasserburg U–Pb concordia diagrams for zircons from granitic and granodioritic gneisses, Vrådal–Kviteseid area. Data from Appendix 1, samples are plotted with $\pm 2\sigma$ error ellipses, assuming an error correlation coefficient of 0.9. Regression lines to (unconstrained) upper intercepts in the range 3.0–4.0 Ga are common-lead lines.

a: Sample TA01-3, granitic gneiss from Fossøyi, Vrådal
b: Sample TA01-13, granitic gneiss from S. Vrådal
c: Sample TA01-14, granodioritic gneiss from Kviteseid.
d: Sample TA01-15, granitic gneiss from Kviteseid.

Table 2
Summary of LAM-ICPMS U–Pb age data, Vrådal-Kviteseid

| | Sample | Rock type | Locality | Age | 2σ | MSWD | Comment |
|---|---|---|---|---|---|---|---|
| | | | | Ma | Ma | | |
| *Gneisses* | | | | | | | |
| | TA01-3 | Granitic gneiss | Fossøyi, Vrådal | 1219 | ±8 | 1.6 | Intercept of common-lead line. |
| | | | | 1219[c] | ±8 | 2.0 | Concordia age, 11 concordant zirco |
| | TA01-13 | Granitic gneiss | S. Vrådal | 1190 | +7/−8 | 1.19 | Intercept of common-lead line. |
| | | | | 1202[c] | ±9 | 0.34 | Concordia age, 9 concordant zircon |
| | TA01-14 | Granodioritic gneiss | Kviteseid | 1199 | ±11 | 1.5 | Intercept of common-lead line. |
| | | | | 1205[c] | ±8 | 0.63 | Concordia age, 7 concordant zircon |
| | TA01-15 | Granitic gneiss | Kviteseid | 1209 | ±7 | 0.75 | Intercept of common-lead line. |
| *Late Sveconorwegian intrusion* | | | | | | | |
| | | | | 1208[c] | ±6 | 0.77 | Concordia age, 18 concordant zirco |
| | TA01-7 | Granite | Vrådal pluton | 960 | +8/11 | 2.0 | Intercept of common-lead line. |
| | | | | 964[c] | ±18 | 0.53 | Concordia age, 2 concordant zircon |
| | | | | 1458 | +42/−36 | 1.6 | Upper intercept of inherited grains |
| | TA01-10 | Hybrid rock | Vrådal pluton | 966 | +5/−7 | 0.52 | Intercept of common-lead line. |
| | | | | 970[c] | ±6 | 3.5 | Concordia age, 9 concordant zircon |

[c]: Concordia ages (Ludwig, 2003).

similar to the $f_{Lu}=0.16$ curve of Vervoort and Blichert-Toft (1999). Isoplot 3.00 (Ludwig, 2003) was used for plotting of radiogenic isotope data and for U–Pb age calculation.

### 3.2. U–Pb geochronology

Zircons from the Vråvatn complex gneisses (Appendix 1) are concordant to near-concordant and have limited variation in isotopic composition, resulting in clusters in Tera–Wasserburg concordia diagrams (Fig. 4). Multi-grain groups of concordant zircons yield concordia ages (Ludwig, 1998) ranging from 1202±9 Ma (TA01-13) to 1219±8 Ma (TA01-3), with overlapping uncertainties (Table 2). The concordia age of the granodioritic gneiss from Kviteseid (TA01-14, 1205±8 Ma) is indistinguishable from the concordia age of granitic gneiss samples from Kviteseid (TA01-15, 1208±6 Ma) and Vrådal. A few normally discordant zircons, which typically contain ≤2% (uncorrected) common lead (Appendix 1) are displaced along incipient discordia lines to upper intercepts at 3 to 4 Ga, which represent the presence of minor amounts of uncorrected common lead. The lower intercepts of these common-lead lines are within error of the concordia ages of the groups of concordant zircons (Table 2, Fig. 4).

Zircons from the two samples of the Vrådal granite range in U–Pb compositions from concordant to normally discordant along lines to 3 to 4 Ga upper intercepts, again suggesting the presence of uncorrected common lead (Fig. 5a,c). The lower intercept age of the common-lead line at 960±8/−11 Ma for the granitic sample (TA01-7) is indistinguishable from a concordia age for two zircons from this sample at 964±18 Ma (Fig. 5a). Nine concordant zircons from the hybrid rock (TA01-10) gave a concordia age at 970±8 Ma, with a 966±5/−7 Ma intercept for the corresponding common-lead line (Fig. 5c). Both of these ages are identical within error to the age of the concordant zircons in sample TA01-7. The weighted average of the four ages reported in Table 2 is 967±4 Ma, which is taken as the best estimate of the intrusive age of the Vrådal granite. This age is older than the 939±20 Ma internal lead isochron age reported by Andersen et al. (2002a), and places the Vrådal granite among a group of 960–970 Ma granitic intrusions, which includes the Byklom, Sæbyggjenut, and Høvring granites (Andersen et al., 2002a).

Five grains from sample TA01-7 plot along a lead-loss line between an upper intercept at 1458±42/−36 Ma and an early Paleozoic lower intercept (Fig. 4b). The upper intercept age is marginally younger than the ages of mid-Proterozoic arc magmatism in southern Norway (Andersen et al., 2004a) and the ca. 1500 Ma early Telemark felsic volcanism (Nordgulen, 1999), but it is equal within error to the U–Pb zircon SIMS age of 1476±13 Ma of the Tinn granite in northern Telemark (Andersen et al., 2002c). This age suggests that these zircons have been inherited from ca. 1.5 Ga felsic rocks in the unexposed basement of central Telemark.

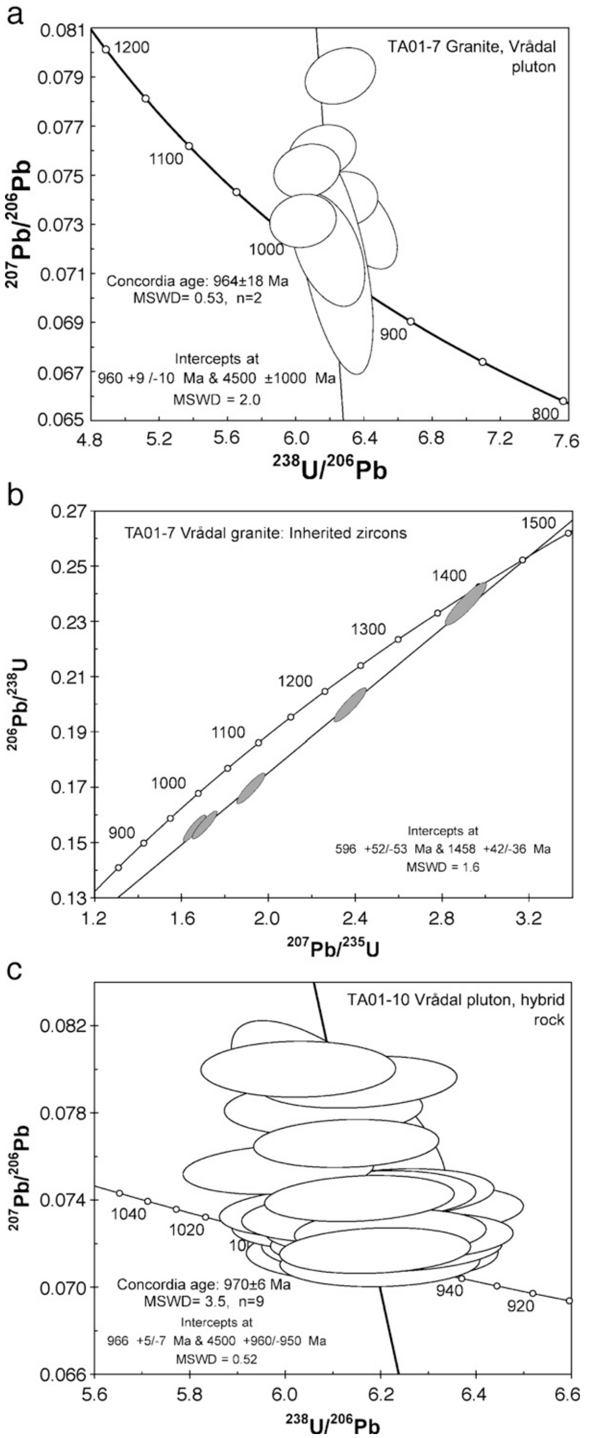



Fig. 5. U–Pb concordia diagrams for zircons from two samples from the late Sveconorwegian Vrådal intrusion. Data from Appendix 1. Plotting conventions as in Fig. 4.a: Sample TA01-7, granite, showing the main, magmatic zircon population. b: Conventional concordia diagram showing inherited zircons in sample TA01-7. c: Sample TA01-10, hybrid rock, showing the main, magmatic zircon population.

### *3.3. Hf isotope composition of zircons*

Zircons from the Vråvatn gneisses and the Vrådal granite have $^{176}$Lu/$^{177}$Hf ratios between 0.0005 and 0.003 (Appendix 2), with no systematic difference in Lu/Hf ratio between zircons from the gneisses and the granite. $^{176}$Lu/$^{177}$Hf in the granodioritic gneiss (TA01-14) is in the lower part of this range. The zircons from granitic and granodioritic gneiss have overlapping present-day Hf isotope compositions, with an average $^{176}$Hf/$^{177}$Hf=0.28232±11 (107 analyses); all but oneof the analysed zircons have $^{176}$Hf/$^{177}$Hf>0.28220 (Fig. 6a). In contrast, the two samples from the Vrådal granite have lower $^{176}$Hf/$^{177}$Hf, with an average ratio of 0.28219±13. Zircons that have suffered partial lead-loss in the Paleozoic, and therefore give young, meaningless $^{207}$Pb/$^{206}$Pb ages (<900 Ma), and zircons with uncorrected common lead (Appendix 1) overlap the undisturbed zircons in hafnium-isotope composition, illustrating the robustness of the Lu–Hf system in zircon. Zircons having suffered ancient lead-loss have been plotted at their apparent $^{207}$Pb/$^{206}$Pb ages in Fig. 6, whereas zircons containing uncorrected common lead are plotted at the corresponding lower intercept ages.

At the time of crystallization, the observed variation in Hf isotopic composition corresponds to $^{176}$Hf/$^{177}$Hf in the range 0.28246–0.28216 (epsilon-Hf=+16 to +6) for the granitic and granodioritic gneisses, and 0.28230–0.28205 (epsilon-Hf between +4 and −5) for the Vrådal granite (Fig. 6). This implies that the 1.21 Ga granitic to granodioritic protoliths of the Vråvatn complex gneisses must have been generated from a source with a distinctly more elevated average $^{176}$Hf/$^{177}$Hf than the source of the 0.97 Ga Vrådal granite magma.

The initial Hf isotope composition of the Vrådal granite overlaps fully with the range of initial $^{176}$Hf/$^{177}$Hf in late Sveconorwegian granites analysed by Andersen et al. (2002b). Initial $^{176}$Hf/$^{177}$Hf of zircons from the Rogaland Intrusive Complex (Andersen and Griffin, 2004) and baddeleyites and whole-rock samples from 0.95–0.98 Ga mafic intrusions from central and western Sweden fall within the upper $^{176}$Hf/$^{177}$Hf range of the Vrådal granite (Fig. 6b).

The crustal residence age of the Hf in a zircon can be calculated from the present-day isotopic composition of the zircon, its U–Pb age and an assumed $^{176}$Lu/$^{177}$Hf ratio for the reservoir in which the Hf resided prior to being incorporated into the zircon (e.g., Andersen et al., 2002b). At$^{176}$Lu/$^{177}$Hf=0.015, corresponding to an average crustal reservoir (Vervoort et al., 1999; Griffin et al., 2002), zircons in the gneisses have a well-defined frequency peak at ca. 1.3 Ga. 75% of the zircons have protolith ages younger than 1.43 Ga, which is younger than any other group of Precambrian metaigneous rocks from the region so far analysed (Fig. 7). In contrast, the samples from the Vrådal granite peak at ca. 1.75 Ga, with 75% of the zircons yielding ages above 1.64 Ga and 10% above 1.9 Ga, as is shown by a marked shoulder on the probability density curve. This maximum value overlaps with crustal residence ages typical of Mesoproterozoic calc-alkaline gneiss complexes from S Norway (CAG in Fig. 7, data recalculated from Andersen et al., 2002b). However, mixtures between an older, crustal Hf component (>1.7 Ga) and juvenile, Sveconorwegian Hf would give similar results. The presence of such older components is supported by the overlap between the >1.9 Ga shoulder of the Vrådal granite zircons and the range of Nd model ages from the Trans-Scandinavian Igneous Belt (e.g., Andersson, 1997).

## 4. Discussion

### 4.1. Origin and significance of the South Telemark Gneisses

The gneisses from the Vråvatn complex belong to the enigmatic South Telemark Gneisses, age and origin of which are poorly constrained. The U–Pb ages obtained in this study demonstrate that at least this part of the gneiss complex consists of early Sveconorwegian granitoids, which have gone through significant deformation and metamorphism in later Sveconorwegian time. The protoliths are thus unrelated to the ca. 1.5 Ga Rjukan group rhyolite from central Telemark (e.g., Ploquin, 1980), and to granites genetically related to those rhyolites (Andersen et al., 2002c).

The initial Hf isotopic composition of zircons from the Vråvatn complex are further illustrated in Fig. 8. There is a clear histogram peak close to 0.28230, with a corresponding probability density maximum at 0.28228 (epsilon-Hf= +10.4), with three grains making up a separate group at higher $^{176}Hf/^{177}Hf$. These initial $^{176}Hf/^{177}Hf$ values are higher than those of ca. 1.2 Ga mafic intrusions in southern Fennoscandia of an assumed lithospheric mantle origin (0.28219, epsilon-Hf<+6, Söderlund et al., 2005). They also are higher than the expected maximum $^{176}Hf/^{177}Hf$ at 1.21 Ga of 1.5–1.6 Ga juvenile crust in southern Norway and southwestern Sweden (≤0.28218, epsilon-Hf ≤+5.5, Fig. 6a). The only feasible source for hafnium of this composition is in the sub-lithospheric, global depleted mantle.

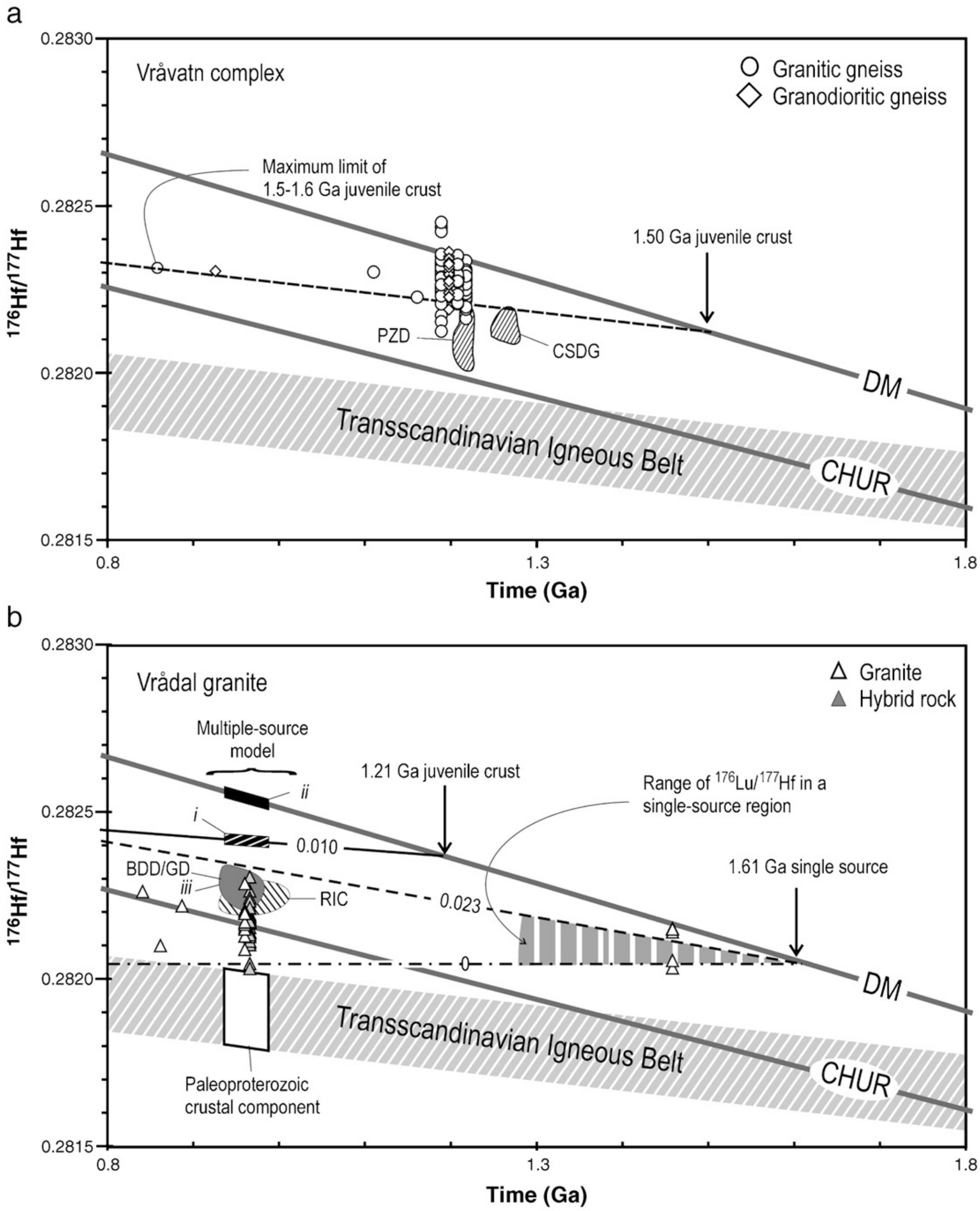
Because of the scarcity of Mesoproterozoic juvenile rocks, the Hf isotope evolution curve of the depleted mantle was initially poorly constrained by data between 1.7 and 0.7 Ga (Vervoort and Blichert-Toft, 1999). Contamination with older crustal material will reduce the $^{176}Hf/^{177}Hf$ of a mantle-derived magma; thus only the maximum observed $^{176}Hf/^{177}Hf$ can be used to constrain the composition of the mantle-derived component in the source region of the granitic magma. Using the upper cutoff value of the main histogram peak as a conservative estimate, a $^{176}Hf/^{177}Hf$ at 1.20±0.05 Ga of 0.28236 (epsilon-Hf=13±1) is indicated by the present data. This estimate agrees well with the Hf-isotope composition predicted from the depleted mantle curves of Vervoort and Blichert-Toft (1999) and Griffin et al. (2000; $^{176}Hf/^{177}Hf$=0.28235±0.00004, or epsilon Hf=12.7±0.2). The present data thus lend further support to these models of depleted mantle evolution, from a period where few juvenile rocks have been recognized worldwide (Condie et al., 2005).

Previous petrogenetic models based on Sr and Nd isotope data from whole-rock samples have invoked a mantle-derived component in the lower crust of the region, but have been unable to distinguish among components introduced from the upper mantle in Sveconorwegian and Gothian time (Andersen, 1997; Andersen et al., 2001). A mantle-derived component cannot reside in a lower crust with a Lu/Hf ratio less than that of the depleted mantle (i.e. $^{176}Lu/^{177}Hf<0.03$) for ≥300 Ma, without diverging significantly from the depleted mantle curve. The underplating event must therefore have been of early Sveconorwegian age, taking place no later than the emplacement of the 1.21 Ga granitoids. To preserve a Hf isotope composition indistinguishable from the depleted mantle curve at 1.21 Ga, underplating may have preceeded anatectic melting by ca. 50 Ma or less. Longer crustal residence times for the mafic underplate would cause a discernible deviation from the depleted mantle curve.

The 1.21 Ga zircons show a considerable range of initial $^{176}Hf/^{177}Hf$ (Fig. 6a). The lower end of the range crosses into the field of 1.6–1.5 Ga calc-alkaline rocks at 1.21 Ga, and overlaps marginally with the range of Protogine Zone dolerites from southern Sweden (Fig. 6a). This distribution range indicates that the mantle-derived component mixed with lithospheric mantle or lower crust during anatectic melting or by post-melting interaction along the magma conduit. Crustal components could include equivalents of the 1.85–1.65 Ga Trans-Scandinavian Igneous Belt, but direct evidence in the form of inherited zircons with $^{176}Hf/^{177}Hf_{1.2\ Ga}<0.2820$ is not provided by the present data.

### 4.2. Source of the late Sveconorwegian Vrådal granite

The petrogenesis of the magmas forming the late Sveconorwegian anorogenic intrusions in southwestern Fennoscandia has been debated. A lower crustal mafic granulite source has been argued for AMCG magmas (Schiellerup et al., 2000; Bolle et al., 2003). On the other hand, whole-rock radiogenic isotope data on late Sveconorwegian granites from the entire region indicate the need for a mantle-derived source component (Andersen, 1997; Andersen et al., 2001; Vander Auwera et al., 2003; Andersen and Griffin, 2004), but the resolution of the data published so far has been insufficient to distinguish clearly between mafic



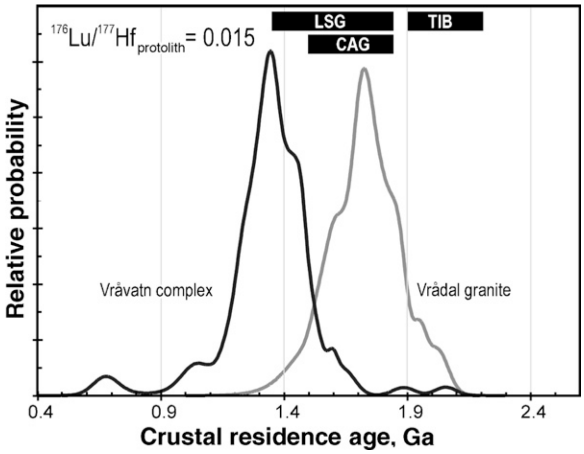



Fig. 7. Crustal residence ages of Hf in zircons from the Vrådal granite and gneisses from the Vråvatn complex, calculated from Lu–Hf data in Appendix 2 and U–Pb ages in Tables 1 and Appendix 1, assuming an average $^{176}Lu/^{176}Hf$ of 0.015 for the crustal protolith, and the depleted mantle model of Griffin et al. (2000). The black bars represent total ranges of crustal residence ages in 1.5–1.6 Ga calcaline gneisses from S. Norway (CAG) and late Sveconorwegian granites (LSG, both based on Lu–Hf data from Andersen et al. (2002b), recalculated for $^{176}Lu/^{176}Hf_{protolith}=0.015$) and TIB granitoids (based on Nd isotope data for TIB granitoids from Andersson, 1997, and additional sources compiled by Andersen et al., 2004b).

material emplaced into the crust at Sveconorwegian and Gothian time (Andersen, 1997; Andersen et al., 2004a).

The range of initial $^{176}Hf/^{177}Hf$ in the Vrådal granite resembles that of other late Sveconorwegian granites in southern Norway (Fig. 6b). This range overlaps the expected range of 1.6–1.5 Ga calc-alkaline rocks from other parts of southwestern Fennoscandia ($^{176}Hf/^{177}Hf_{1.2\ Ga}=0.2820–0.2823$), which is probably coincidental, because of the lack of evidence for such rocks in the Telemark block. Furthermore, inherited zircons with $^{176}Hf/^{177}Hf<0.2819$ (epsilon Hf<−9) have been observed in other late Sveconorwegian granites (Andersen et al., 2002b), such zircons can only have originated in rocks of age and composition similar to the Transscandinavian Igneous Belt (Fig. 6b). In the Vrådal granite, the only inherited zircons are ca. 1500 Ma, and have near-juvenile initial Hf isotope compositions. Probable source candidates for these inherited zircons would be the older Telemark supracrustal rocks (Rjukan group metarhyolites) or intrusions related to them.

The range in initial $^{176}Hf/^{177}Hf$ of the zircons from the Vrådal pluton reflects a heterogeneous magma at the time of crystallization of zircon. This range can in principle be generated by two different processes: (1): melting of source rocks in the lower crust with uniform age but variable Lu/Hf ratio and hence variable $^{176}Hf/^{177}Hf$ ("single source model") or (2): mixing of melts from two or more source rocks with different crustal residence ages, and hence different Hf isotopic composition ("multiple source model"). To evaluate the realism of a single-source model for the Vrådal granite, the age and range of $^{176}Lu/^{177}Hf$ of the hypothetical source region need to be estimated. The minimum age of a single source is given by the intersection between a horizontal line through the zircon with the lowest $^{176}Hf/^{177}Hf$, and the depleted mantle

Fig. 6. Initial Hf isotopic composition at the crystallization age of the zircons. Zircons that have been used for dating are plotted at the preferred crystallization age of their host rocks (Table 2), except for zircons showing evidence of ancient lead-loss, which have been plotted at their apparent $^{207}Pb/^{206}Pb$ ages (Appendix 1). Sample symbols as in Fig. 2. Reference lines representing meteoritic Hf evolution (CHUR, Blichert-Toft and Albarède, 1997) and the depleted mantle (DM, model of Griffin et al., 2000) are shown, as is the total range of crustal reservoirs corresponding to the Trans-Scandinavian Igneous Belt (TIB), based on initial Hf isotope ratios from zircons in TIB granitoids (Andersen et al., 2006) at an assumed $^{176}Lu/^{177}Hf$ of 0.015. Growth curves of model systems are labeled with their $^{176}Lu/^{177}Hf$ ratios.

a: Hf isotope compositions of zircons in gneisses from the Vråvatn complex. The maximum limit of 1.5 Ga juvenile crust is defined by a growth curve with $^{176}Lu/^{177}Hf=0.015$ run through the most primitive initial $^{176}Hf/^{177}Hf$ observed in 1.5–1.6 Ga calc-alkaline gneisses from the Bamble–Lillesand and Kongsberg–Marstrand blocks (Andersen et al., 2002b, 2004a). The fields PZD and CSDG represent initial Hf isotope compositions of Protogine Zone dolerites and the Central Scandinavian Dolerite group, respectively (Söderlund et al., 2005).

b: Hf in zircons from the Vrådal granite, showing possible single-and multiple-source models for the origin of the granite. The ruled field marked RIC shows the variation of igneous zircons from the Rogaland Intrusive Complex (Andersen and Griffin, 2004), and the shaded field (BDD/GD) the variation of initial $^{176}Hf/^{177}Hf$ in mafic igneous rocks belonging to the Blekinge–Dalarne (BDD) and Göteborg dolerites (GD; data from Söderlund et al., 2005). The minimum age of 1.61 Ga for a single, lower-crustal source for the granite is defined by the intersection with the depleted mantle curve of the dash-dot horizontal line ($^{176}Lu/^{177}Hf=0$) through the minimum observed initial $^{176}Hf/^{177}Hf$. To account for the total variation in $^{176}Lu/^{177}Hf$ at 0.97 Ga by radiogenic accumulation in a heterogeneous lower crust of uniform age, a source region must have a variation in $^{176}Lu/^{177}Hf$ from zero to (at least) 0.023 (broken line). In a multi-source model, Paleoproterozoic crust corresponding to the TIB is a possible low-$^{176}Hf/^{177}Hf$ endmember (white box). Possible high-$^{176}Hf/^{177}Hf$ endmembers include 1.21 Ga juvenile crust (i), global depleted mantle (ii) and sub-Fennoscandian lithospheric mantle (iii) corresponding to the BDD/GI source (Söderlund et al., 2005).

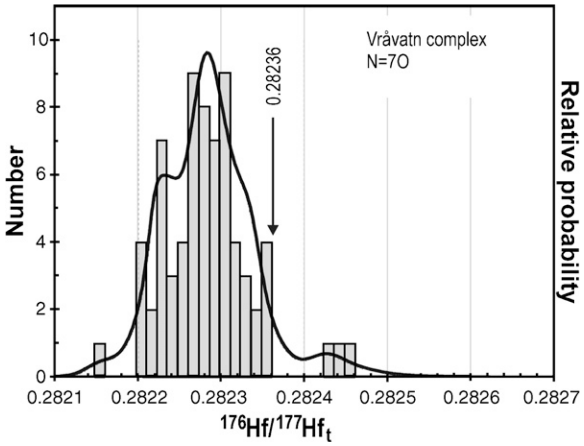



Fig. 8. Initial $^{176}Hf/^{177}Hf$ in zircons from Vråvatn complex gneisses used to constrain the position of the depleted mantle growth curve at 1.21 Ga. The arrow indicates the best estimate of the Hf isotopic composition of the depleted mantle (DM) that can be derived from these data.

curve (Fig. 6b). The horizontal line is a growth curve for a system with $^{176}Lu/^{177}Hf=0$, i.e. for a highly unrealistic, Lu-free rock. The resulting minimum age of crustal residence is 1.61 Ga, which is compatible with Gothian magmatism (Andersen et al., 2004a). To obtain the total range of $^{176}Hf/^{177}Hf$ observed in the Vrådal granite, a variation of $^{176}Lu/^{177}Hf$ in the source region from zero to 0.023 is needed (Fig. 6b). The maximum value is well above average crustal values, but is possible for garnet-bearing, depleted mafic lower crustal granulites (Vervoort and Patchett, 1996). Younger single sources are impossible, because this would require a reduction of $^{176}Hf/^{177}Hf$ in the source rock with time, in order to produce the lowest $^{176}Hf/^{177}Hf$ observed in zircons. For sources older than 1.61 Ga, minimum $^{176}Lu/^{177}Hf > 0$ are indicated, but at the same time, maximum values of $^{176}Lu/^{177}Hf > 0.023$ are needed, increasing with increasing age of the source.

If, on the other hand, the source region of granitic magma is heterogeneous in age, it is possible to generate the observed range of $^{176}Hf/^{177}Hf_{0.97\ Ga}$ without invoking lower crustal rocks with extreme Lu/Hf ratios (Fig. 6b). A multiple source model for the Vrådal granite magma would require source rocks with $^{176}Hf/^{177}Hf <$ 0.2820 and $>$0.2823, respectively. At 0.97 Ga. 1.65–1.85 Ga granitoids of the Trans-Scandinavian Igneous Belt have $^{176}Hf/^{177}Hf$ in the range 0.2818 to 0.2820 at 0.97 Ga (Andersen et al., 2006); TIB equivalents are therefore possible crustal end members in a multiple source model for the Vrådal granite. On the other hand, three different sources are possible for the more radiogenic Hf component (Fig. 6b). (i): 1.21 Ga juvenile crust, with expected $^{176}Hf/^{177}Hf_{0.97\ Ga}=$ 0.2824; (ii): Global depleted mantle, with $^{176}Hf/^{177}Hf_{0.97\ Ga}=0.2825–0.2826$; and (iii): Lithospheric mantle of composition comparable to the source of the Blekinge–Dalarne and Göteborg dolerites of Sweden ($^{176}Hf/^{177}Hf_{0.97\ Ga}=0.28232–0.28221$, Söderlund et al., 2005).

### *4.3. Tectonic implications, and suggestions for future studies*

In Neoproterozoic time, Fennoscandia was part of a supercontinent known as Rodinia (e.g., Condie, 2005). In southwestern Fennoscandia, tectonometamorphic events reflecting collisions with other continental blocks have been recorded at ca. 1.1 Ga and 0.97 Ga (Kullerud and Dahlgren, 1993; Möller and Söderlund, 1997; Knudsen and Andersen, 1999). One or both of these events may be related to the incorporation of Fennoscandia into the Rodinan supercontinent. The tectonic setting of Fennoscandia prior to 1.1 Ga is debated. It may have been continuous with Laurentia in a pre-Rodinia continental assembly at 1.16 Ga (e.g., Brewer et al., 2002), or it may have rifted away from Laurentia, to drift independently since 1.27–1.25 Ga (Elming and Mattson, 2001; Pesonen et al., 2003).

Both the mafic underplating event recorded by zircons from the Vråvatn complex, and the granitic magmatism that produced the protoliths of the granitic gneisses took place prior to assembly of Rodinia. Magmas from the sub-lithospheric mantle would be able to penetrate the lower crust in zones of crustal extension, which might be related to an event of continental rifting, or to continental back-arc spreading. The presence of ca. 1.2 Ga granitoids with calc-alkaline affinity in nearby southeastern Telemark (Heaman and Smalley, 1994) suggests that the Telemark block was situated behind a subduction margin in the early Sveconorwegian; from trace element data on 1.16 Ga volcanic rocks from Telemark, Brewer et al. (2002) also concluded that the younger Telemark supracrustals (and the Dal group of Sweden) were deposited in basins formed by back-arc extension. Magmatic underplating in a continental back-arc setting at 1.21 Ga or slightly earlier is compatible with the present Hf isotope data. Alternatively, mafic underplating may have been related to a ca. 1.25–1.27 Ga rifting event separating Fennoscandia from Laurentia, as indicated by some paleomagnetic data (Pesonen et al., 2003). The present data do not allow a choice between the two alternatives. However, Lu–Hf data from granitoids formed in the period 1.27–1.21 Ga in the Fennoscandian Shield and its potential neighbours in Rodinia would provide decisive constraints on the timing and nature of the mafic underplating event.

When the late Sveconorwegian intrusions (Vrådal granite and its coeval granites studied by Andersen et al., 2001, 2002a; Rogaland Intrusive Complex) were emplaced, Fennoscandia formed part of Rodinia. In most reconstructions of Rodinia, western Fennoscandia faces eastern Greenland, although other configurations are possible (Meert and Powell, 2001; Hartz and Torsvik, 2002; Pesonen et al., 2003; Meert and Torsvik, 2003). The presence of 0.94–0.92 Ga granitic intrusions in East Greenland has been interpreted as evidence of a Rodinian fit between southwestern Fennoscandia and the eastern Greenland margin (Kalsbeek et al., 2000; Watt and Thrane, 2001). However, this correlation is based only on a similarity of U–Pb ages, without supporting evidence from other radiogenic isotope data. Together with the data on late Sveconorwegian granites from southern Norway of Andersen et al. (2002b), the present data from the Vrådal granite provide a clear Hf isotopic fingerprint of the source region of late Sveconorwegian granitic magmas. In order to evaluate a correlation between these granites and potential counterparts in other fragments of Rodinia, it is necessary to compare such fingerprints as well as U–Pb ages. Until Hf isotope data become available from 0.92–1.0 Ga granites in Greenland and other possible neighbours in Rodinia, correlations based only on the emplacement ages of granitic intrusions remain speculative.

## 5. Conclusions

The findings reported in this study demonstrate the power of combined U–Pb and Lu–Hf isotope data on zircons from (meta)igneous rocks as indicators of petrogenetic processes in the deep crust. Unlike whole-rock based isotopic data (e.g., Sm–Nd), which only yield univariant lines in time-isotope ratio space, if not accompanied by independent age data, the present approach allows the simultaneous determination of the timing of a process and the isotopic characteristics of the material involved.

LAM U–Pb and Lu–Hf isotope data on single zircons separated from granitic and granodioritic gneisses of the Vråvatn complex in the Telemark block of southwestern Fennoscandia demonstrate that the northern part of the South Telemark Gneiss area formed from early Sveconorwegian (1.22–1.20 Ga) granitic intrusions, which have a distinct, depleted mantle-like initial Hf isotope signature (epsilon-Hf$>+6$). The magmatic protoliths of these gneisses must have formed shortly (i.e.$<50$ Ma) after, and possibly in response to, an event of crustal underplating affecting the Fennoscandian continent. This event was related to either breakup of a pre-Rodinian continental assemblage involving Fennoscandia, and probably Laurentia, or to back-arc extension behind a Fennoscandian subduction margin.

Zircons from the 967$\pm$4 Vrådal granite provide constraints for the Hf isotopic composition of the source material of late Sveconorwegian magmatism in southwestern Fennoscandia. Mixing of two or more components is needed to account for its range of initial $^{176}Hf/^{177}Hf$: A crustal component, corresponding to rocks of the 1.85–1.65 Ga Transscandinavian Igneous Belt, and one or more components, derived from 0.97 Ga lithospheric mantle, 1.21 Ga early Sveconorwegian juvenile crust, or possibly from 0.97 Ga global depleted mantle.

By constraining the age and source characteristics of Sveconorwegian granitoids from southwestern Fennoscandia, we are starting to establish a base of information of importance for the interpretation of the crustal evolution and tectonic history of Fennoscandia. Unfortunately, combined U–Pb and Lu–Hf data are still missing from strategically important rocks in both Fennoscandia and other late Mesoproterozoic – early Neoproterozoic continental terranes, so that more accurate timing of mafic underplating, and correlation with processes in potential neighbours in Rodinia, is not yet possible. The results of this study suggest that such data would be most important for the interpretation of both the status of Fennoscandia prior to assembly of Rodinia, and its position within the supercontinent.

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## Appendix A. Supplementary materials

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.lithos.2006.03.068.

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