Derivation of the 1.0–0.9 Ga ferro-potassic A-type granitoids of southern Norway by extreme differentiation from basic magmas

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Received 10 October 2001; received in revised form 28 September 2002

Abstract

Major and trace elements, Sr and Nd isotopic data as well as mineral compositions are presented for a selection of the 1.0–0.9 Ga ferro-potassic A-type granitoids (Bessefjellet, Rustfjellet, Verhuskjerringi, Vale, Holum, Svøfjell, Handeland-Tveit, Åseral, Lynghdal gabbro-norites) that occur close to the Mandal-Ustaoset Line (MUL) of southern Norway. These hornblende biotite granitoids (HBG) define an extensive differentiation trend ranging from gabbro-norites (50 wt.% SiO2) to granites (77 wt.% SiO2). This trend is interpreted as resulting from extreme fractional crystallization of several basaltic magma batches with similar major and trace elements compositions. At 930 Ma, the HBG suite displays a narrower range in I Sr (0.7027–0.7056) than in εNd(t) (+1.97 down to −4.90) suggesting some assimilation of a Rb-depleted lower crust (AFC process) or/and source variability. An age of 929 ± 47 Ma is given by a Rb-Sr isochron on the Holum granite (Sr i = 0.7046 ± 0.0006, MSWD = 1.7). Geothermobarometers indicate a low pressure of emplacement (1.3–2.7 kbar) and an oxygen fugacity close to NNO. High liquidus temperatures are given by the apatite saturation thermometer (1005–1054°C) and are in agreement with results from other studies. The basaltic parent magmas of the HBG suite are partial melts of an hydrous mafic, potassic source lying either in the lithospheric upper mantle or in the mafic lower crust derived from it. This contrasts with the 930 Ma anorthosite–mangerite–charnockite suite (AMC suite) of the Rogaland Province for which a depleted lower crustal anhydrous gabbro-noritic source has been indicated. The present data imply the penecontemporaneous melting of two contrasting sources in southern Norway. The source duality could result from an increasing degree of metamorphism (amphibolite to granulite) from East to West, an horizontal stratification of the lower crust or from the stratification of the lithosphere (melting of the lower crust or upper mantle). It may also indicate that the AMC and HBG suites formed in two distinct crustal segments. The linear alignment of the HBG suite along the Mandal-Ustaoset shear zone suggests that a linear uprise of the asthenosphere, following a lithospheric delamination under this structure, could be the vector of the mantle heat.

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Keywords: A-type; Granitoids; Sveconorwegian; Proterozoic; Southern Norway

1. Introduction

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1. Introduction

At about 1.0–0.9 Ga, the southwestern Baltic shield (southern Norway and Sweden) was intruded by a
suite of late granitoids related to the main Sveconorwegian deformation structures (Andersson et al., 1996). Considerable volume of magma was emplaced in a short time (Fig. 1) creating a large scale phenomenon affecting the Proterozoic continental crust. These late granitoids belong to the distinctive rock type, usually referred to as A-type, occurring in most Proterozoic belts. The origin of this granitoid suite is...
currently under considerable debate. Geochronological and isotopic data already exist for some of the Sveconorwegian granitoids (Andersen et al., 2001). They indicate similar ages between 1.0 and 0.9 Ga (see Andersson et al., 1996 for a summary of available geochronological data) and the Nd, Sr and Pb systematics suggest mixing between a depleted mantle component and two or more components having an extended crustal history (Andersen et al., 2001). The purpose of this study is to examine the geochemistry (major and trace elements as well as Nd and Sr isotopes) and petrology of a selection of these granitoids outcropping in the Vest Agder and Telemark sectors of southern Norway, in order to define possible differentiation trends, the parent magma compositions as well as possible sources. Indeed, it has already been recognized that A-type Proterozoic granites encompass different granite compositions (metaluminous, peraluminous, peralkaline) which may have different sources (Poitrasson et al., 1995). On the other hand, in southern Norway as in many other Proterozoic belts, these granitoids exist in a space and time with an anorthosite-mangerite-charnockite suite (AMC suite), the Rogaland anorthosite complex, for which a complete liquid line of descent has been defined (Demaiffe and Hertogen, 1981; Duchesne et al., 1974; Vander Auwera et al., 1998a; Wilmsen et al., 1989) and a lower crustal gabbronoritic (mafic anhydrous) source has been indicated (Longhi et al., 1999). It is thus also important to constrain the source of the penecontemporaneous granitoid suite which is studied here in order to better characterize its relationship with the AMC suite and the overall evolution of this segment of the Proterozoic continental crust.

2. Geological outline

The Precambrian basement of southern Norway belongs to the southwest Scandinavian domain of the Baltic shield. It has been subdivided in five sectors (Rogaland-Vest Agder, Telemark, Bamble, Kongsberg, Osfjord-Akerhus) separated by major crustal lineaments (Fig. 1). The first two, in which occur the intrusions discussed in this paper, are separated by the Mandal-Ustaoset Line (MUL) (Simmond, 1985) that crops out as a brittle fault zone in its northern and central parts and as an elongated augen gneiss in its southern part. Similarly, the Feda augen gneiss which is also N-S elongated (Fig. 1) could delineate a major crustal structure (shear zone) separating the Rogaland anorthosites from the West and the Vest Agder migmatitic province to the East (Duchesne et al., 1999). The Telemark sector is bounded to the East by the Kristiansand-Porsgrunn shear zone (KPS in Fig. 1).

The Telemark and Rogaland-Vest Agder sectors are made up of high grade gneisses (migmatites with a supracrustal protolith, granitic gneisses and anorthosite) and low grade supracrustal formations (the Telemark Supergroup), intruded by pre-, syn- and postcollisional Sveconorwegian intrusions. Geochronological data obtained in these two sectors range from 1600 to 800 Ma and reveal a complex metamorphic and tectonic evolution (i.e. Jacobsen and Heier, 1978; Menage, 1982; Pasteels and Michot, 1975; Priem and Verschure, 1982; Tobi et al., 1985). Classically, ages between 1.75 and 1.5 Ga have been related to the Gothian orogeny and those ranging from 1.2 to 0.9 Ga belong to the Sveconorwegian orogeny.

In the Rogaland-Vest Agder and Telemark sectors, magmatism occurred at various stages of the tectonic evolution. The Hidderskog (Zhou et al., 1995), Botnavatnet and Gloppurdi massifs (Wielens et al., 1981) were emplaced at about 1160 Ma, prior to the main Sveconorwegian event, and display an A-type signature (Zhou et al., 1995). The high-K calc-alkaline augen gneisses were emplaced at the climax of the orogeny around 1.05 Ga (Bingen et al., 1993). The postcollisional intrusions can be separated in two groups. One group corresponds to the Rogaland AMC suite, emplaced in a short period of time between 930 and 920 Ma in the western Rogaland-Vest Agder sector (Pasteels et al., 1979; Schärer et al., 1996). The second group comprises a series of biotite-hornblende-bearing granitoids (HBG suite) forming a ≃300 km long belt stretching along the MUL. It includes the Lyngdal granodiorite (Bogaerts et al., 2001; Bogaerts et al., in press; Falkum and Petersen, 1974; Falkum et al., 1972; Falkum et al., 1979; Petersen, 1980a,b). This HBG suite belongs to the voluminous 1.00–0.90 Ga postcollisional plutonism occurring in southern Norway and eastwards and represented, among others, by the Grimstad pluton in the Bamble sector, the Herefoss granite crossing the Kristiansand-Porsgrunn shear zone (Andersen, 1997).
Table 1
Sample location (X-Y coordinates in the EURF89 kilometric UTM grid) and description

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926 ± 8 Ma, Sr* = 0.7046; εNd = −3.2 to −0.8) and the Bohus-Flå granite belt stretching for at least 300 km from northern Telemark down to the west coast of southern Norway, result from the mixing between a depleted mantle-derived component and two or more major components having a long crustal history.

We present here a petrological study of eight intrusions (90 samples) selected among the granitoids related to the MUL. These intrusions are from North to South Verhuskjerringi, Bessefjellet, Valle, Rustfjellet, Aseral, Gneisset, and Skjeiestad.

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Andersen et al., 2001; Andersson et al., 1996 (Fig. 1). Recently, Andersen et al. (2001) have suggested that most of the granites occurring all over the Bohus-Flå granite belt and Telemark down to the west coast of Sweden along several large scale tectonic lineaments.
Svøfjell, Åseral (not shown), Handeland-Tveit (not shown) and Holum (Fig. 1). The following discussion will also include the undeformed Lyngdal hyperites (gabbroenorites to monzonorites) cropping out as two small intrusions north of the Lyngdal granodiorite and studied by Domaïff et al. (1990). The intrusions are shown in Fig. 1 and the X-Y coordinates of samples are given in Table 1. A detailed geochemical and experimental study of the Lyngdal granodiorite is in progress (Bogaerts et al., 2001; Bogaerts et al., in press).

3. Analytical methods

Microprobe analyses of feldspars, amphiboles, biotites, Fe-Ti oxides and clinopyroxenes have been performed with the Cameca SX50 of the CAMST (“Centre d’analyses pour les Sciences de la Terre”, Louvain-La-Neuve, Belgium) on selected samples of the granitoids. Accelerating voltage was set at 15 kV and elements were counted for 20 s (Ca, K, Ti, Cl), 30 s (Si, Al, Fe, Mg, Mn, Cr) or 45 s (Na, F) at a beam current of 20 nA. A combination of synthetic and natural standards were used and X-ray intensities were reduced using the Cameca PAP correction program. Results are shown in Tables 2–5.

Ninety whole-rock samples were selected for a geochemical study. Major and some trace elements were analyzed by X-ray fluorescence (ARL 9400 XP for Holum samples and CGR ALPHA 2020 for all other samples) following the method described in Bologne and Duchesne (1991) for samples analyzed with the CGR ALPHA 2020. Other trace elements including rare earths were analyzed with the ICP-MS (VG Plasma Quad PQ2) following the method described in Vander Auwera et al. (1998b) or/and INAA (Pierre Sue Laboratory: CEA, Saclay, France). FeO was measured by titration. Results are shown in Table 6.

Sr and Nd isotopes were measured on a selection of samples. Results are shown in Table 7. After acid dissolution of the sample and Sr and Nd separation on ion-exchange resin, Sr isotopic compositions have been measured on Ta simple filament and Nd isotopic compositions on triple Ta-Re-Ta filament on a Micromass Sector 54. Repeated measurements of Sr and Nd standards have shown that between-run error is better than ±0.000015. Within-run errors are generally lower. The NBS987 standard has given a value for 87Sr/86Sr of 0.710274 ± 0.000011 (2σ on the mean), four measurements, normalized to 86Sr = 0.1194 and the Rennes Nd standard (Chauvel and Blichert-Toft, 2001), a value for 143Nd/144Nd of

Table 2

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<td>101.04</td>
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Normalized to eight oxygens

Si 2.995 2.850 2.729 2.996 2.702 2.980 2.798 2.705 2.989 2.770 3.001 2.858
Al 0.997 1.141 1.260 0.998 1.287 1.000 1.190 1.282 1.004 1.221 0.988 1.135
Fe 0.000 0.000 0.004 0.000 0.004 0.008 0.008 0.008 0.011 0.005 0.002 0.000 0.006
Ca 0.002 0.007 0.295 0.002 0.316 0.008 0.220 0.311 0.000 0.246 0.000 0.150
Na 0.071 0.851 0.691 0.127 0.675 0.220 0.768 0.669 0.079 0.751 0.049 0.837
K 0.953 0.013 0.016 0.888 0.015 0.800 0.013 0.018 0.946 0.008 0.976 0.017
Ab 6.94 82.21 68.99 12.52 67.06 20.65 76.66 67.04 7.69 74.67 4.82 83.39
An 0.21 16.47 29.44 0.20 31.43 0.75 22.00 31.18 0.00 24.52 0.00 14.90
Or 92.86 1.32 1.57 87.29 1.51 78.60 1.34 1.79 92.31 0.81 95.18 1.71
Table 3

Selected electron microprobe analyses of amphiboles

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<td>101.15</td>
<td>101.13</td>
<td>101.70</td>
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Normalized to 13 cations

| Si     | 6.556    | 6.534    | 6.664    | 6.358    |
| Ti     | 0.107    | 0.139    | 0.190    | 0.048    |
| Al     | 1.607    | 1.656    | 1.394    | 1.819    |
| Fe³⁺   | 0.620    | 0.524    | 0.361    | 0.725    |
| Fe²⁺   | 1.940    | 1.855    | 1.465    | 1.977    |
| Mn     | 0.086    | 0.073    | 0.075    | 0.093    |
| Mg     | 2.085    | 2.218    | 2.852    | 1.952    |
| Ca     | 1.880    | 1.885    | 1.961    | 1.886    |
| Na     | 0.416    | 0.423    | 0.534    | 0.510    |
| K      | 0.272    | 0.281    | 0.282    | 0.332    |
| Cr     | 0.000    | 0.000    | 0.000    | 0.000    |
| F      | 0.318    | 0.307    | 0.466    | 0.497    |
| Cl     | 0.038    | 0.030    | 0.022    | 0.054    |
| OH     | 1.644    | 1.662    | 1.512    | 1.449    |
| Mg(Mg + Fe³⁺) | 0.518    | 0.545    | 0.661    | 0.497    |

4. Field relationships and petrography

The Besselfjellet intrusion (~25 km²) is a medium-grained (~2 mm) pink granite, grossly circular in shape and intrusive into the supracrustal formations (metabasalts, metasandstones and metarhyolites) of the Telemark Province (Dons, 1960; Killeen and Heier, 1975) (Fig. 1). It is homogeneous and generally displays a porphyritic texture with centimetric phenocrysts of perthitic microcline dispersed in a matrix essentially made of plagioclase, smaller grains of perthitic microcline, rounded quartz as well as minor biotite, apatite and accessory zircon, titanite and fluorite. Leucogranitic facies are abundant and...
contain secondary muscovite associated with fluorite. Locally, it is totally devoid of biotite. Aplitic and pegmatitic dykes crosscut the intrusion. It has been dated at 923 ± 16 Ma (Rb-Sr isochron) by Kileen and Heier (1975).

The Verhuskjerringi massif (~50 km$^2$) is also almost circular and intrusive into the Telemark supracrustals (Fig. 1). The most common facies is a coarse-grained heterogranular (1 mm to 1 cm) granite containing perthitic microcline, plagioclase, quartz, biotite and locally amphibole as major phases. Zircon, apatite, titanite and opaques are accessory and titanite usually surrounds the opaques. The mafics (biotite, amphibole, opaque, apatite, zircon and titanite) commonly form aggregates of millimetric size. Some mesocratic and leucocratic facies are also present. The former is richer in biotite and amphibole containing relic cores of clinoxyroxene, whereas secondary muscovite associated with fluorite have been observed in the latter. Aplitic dykes crosscut the intrusion.

The Rustfjellet pluton (~15 km$^2$) straddles the MUL and is intrusive into the supracrustal formations of Telemark (metasediments of the Bandak group) along its eastern margin and in the banded and granitic gneisses of the Rogaland-Vest Agder sector along its western margin. It is mainly leucogranitic but granitic faces also occur. Its texture is essentially equigranular with a variable grain size ranging from 1 to 5 mm. The major phases are plagioclase, quartz and perthitic microcline (containing inclusions of plagioclase and quartz) associated with minor biotite and secondary muscovite. Accessory phases are zircon,

Table 4

Selected electron microprobe analyses of biotites

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Table 5
Selected electron microprobe analyses of clinopyroxenes

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Fe$^{3+}$ has been calculated using the method of Droop (1987).

apatite, opaque and fluorite. The latter mineral is usually interstitial and associated with muscovite. In some samples, the texture clearly suggests that muscovite was formed at the expense of biotite. Large enclaves of banded gneisses have been observed as well as some aplitic and pegmatitic dykes.

The Svöfjell massif is much larger than the other intrusions and covers around 350 km$^2$ (Fig. 1). It intrudes the various gneisses (banded, granitic and augen) of the Rogaland-Vest Agder sector. It has a coarse-grained equigranular to heterogranular texture with a grain size ranging from 1 to 10 mm. The composition is granodioritic to granitic and contains angular enclaves of the surrounding gneisses as well as scarce lobate microgranular mafic enclaves. The major phases are plagioclase, K-feldspar (perthitic microcline or orthoclase), quartz, brown to brown-green biotite and bluish green amphibole. Accessory phases are apatite, zircon and opaques commonly surrounded by titanite. The latter are usually present in aggregates together with biotite and amphibole. The Svöfjell massif is crosscut by aplitic and pegmatitic dykes. At several places, the granite is intrusive into migmatised gneisses which locally display agmatitic textures. In these agmatites, slightly tilted blocks (m to dm) of gneisses are embedded in a leucocratic matrix. This leucocratic matrix (samples 84-48, 84-49, 84-50, SV90-6, SV90-8—called Svöfjell dykes hereafter) is very poor in mafics (biotite, orthopyroxene in samples 84-48, 84-59, SV90-6; apatite, zircon and opaques with locally a small amount of amphibole in samples 84-59, SV90-8). Some of these samples (84-48, 84-50, SV90-8) contain mesoperthites locally associated with plagioclase (84-50, SV90-8); the other ones contain antiperthitic plagioclase and perthitic K-feldspar (SV90-6, 84-59). Quartz is always abundant (>20%). The presence of orthopyroxene and mesoperthite in this leucocratic material suggests that it belongs to the charnockitic suite and is distinct from the main body of Svöfjell. Samples 84-52...
Table 6
Major (wt. %; XRF data) and trace elements (ppm) compositions of a selection of samples

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| [Eu/Eu+] | 0.63  | 0.48 | 0.71 | 0.46 | 1.33| 0.23| 0.18| 0.25 | 0.79 | 0.33| 0.19 | 0.31 | 0.30 | 0.23 |
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Ni – – – – – – – – – – – – – – 90.0
Zn 171 99 173 51 112 31 85 98
Pb 6 14 7 18 3 3 10 –
Sr 32 40 38 51 13 33 24 20
Ba 76 88 104 108 29 73 56 46

Pr 10 10 13 12 3.81 8.47 7.20 –
Nd 45 44 53 49 17 33 30 27
Sm 9.9 10 13 9.4 4.0 7.0 8.7 6.0
Eu 2.2 2.0 2.3 1.6 1.4 1.4 1.3 2.0
Gd 10 11 12 8.37 4.44 5.98 7.09 –
Dy 9.41 11 14 7.34 4.35 4.56 7.79 –
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Er 5.16 5.78 8.08 3.82 2.61 2.40 4.99 –
Tm 0.78 0.85 1.26 0.54 0.41 0.37 0.79 –
Yb 5.06 5.62 8.44 3.38 2.60 2.47 5.39 2.36
Lu 0.71 0.82 1.23 0.46 0.38 0.37 0.78 0.34
Y 5.4 56 87 46 28 28 57 29
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[La/Nb] 5 5 3 11 4 10 3 5.99

(Ex/Eu*): 0.66 0.59 0.56 0.55 0.98 0.65 0.57 1.14

Rh, Sr, Ba, Zn by XRF; V, Sc, Cr, Co, Ni by INAA; U, Th, Ta, Cs, INAA except for 84-43, 84-47, 84-48, 84-52, 84-53, 84-64; REE by ICP-MS except for SV1, SV3 to SV5, SV7 to SV9, SV11, SV12 (INAA); Y by ICP-MS; Zr by XRF except for 84-43, 84-47, 84-48, 84-52, 84-53, 84-64, SV5 to SV7, SV9, SV11, SV12; Nb by XRF except for BE4, BE5, VA1, VA2, 84-43, 84-47, 84-48, 84-52, 84-53, 84-64, SV5, SV10, S1, R3, R5, R7; F determined by PIGE (Roelandts et al., 1987). For samples of Holmen (sample number have a 98BN prefix): Rh, Sr average of XRF and ICP-MS values, other trace elements by ICP-MS.

* Demaiffe et al. (1990).
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and 84-53 have been sampled at the northern contact of the Svoefjell massif and according to the 1:250,000 geological map, could belong to a separate intrusion. The granodioritic to granitic intrusion of Valle (∼150 km²) displays a porphyritic texture with phenocrysts (several cm) of perthitic microcline embedded in a matrix of plagioclase, microcline, quartz, green biotite and bluish-green amphibole ranging in size from 0.25 to 2 mm. Biotite is much more abundant than amphibole. The microcline phenocrysts contain inclusions of quartz, biotite and plagioclase displaying an albite rim at the contact with the K-feldspar. Titanite, zircon, apatite and opaque are accessory phases. This intrusion contains lobate mafic microgranular enclaves which together with the microcline phenocrysts are locally slightly oriented, probably due to magmatic flow.

The Holum granite outcrops close to the southernmost tip of the MUL (Fig. 1). It forms a north-south elongated pluton (∼83 km²) intruded into banded, granitic and augen gneisses of the Rogaland-West Agder sector, metamorphosed under amphibolite facies. As already noticed by Wilson et al. (1977), no metamorphic aureole can be detected around the granite. Xenoliths from the country rocks are found throughout the pluton, the most abundant being several centimeters to hundred of meters size enclaves of faintly-foliated biotite–granite, presumably coming from the granitic gneisses. Mafic microgranular enclaves have not been found. Pegmatite veins occur, especially along the margins of the pluton. The main rock-forming minerals are K-feldspar (usually microperthitic microcline), unzoned plagioclase, quartz, reddish-brown biotite, and green-brown hornblende. Accessory phases include opaque(s), usually surrounded by titanite, and the mafic minerals constitute about 5–25% of the rock. Mineral modal proportions indicate almost exclusively monzogranitic compositions. The rock is usually coarse-grained (average grain size about 2–3 mm) and displays an heterogranular hypidiomorphic texture, with frequent K-feldspar and/or plagioclase phenocrysts up to 1.5–2 cm long. The mafic minerals tend to form more or less elongated aggregates, up to several millimetres long. The Holum granite is considered to be the oldest Sveconorwegian posttectonic granite of southernmost Norway and its emplacement age (980 Ma, based on a Rb-Sr whole-rock isochron: Wilson et al., 1977) is usually taken as a lower limit for the last Sveconorwegian regional folding phase in the area (Falkum, 1998). A recent structural study of the Holum granite, based on a survey of its anisotropy of low-field magnetic susceptibility, has shown, however, that this pluton is not posttectonic, but that it was emplaced in a tectonic strain field (Bolle et al., in press). A model of emplacement and deformation synchronous with the last Sveconorwegian regional folding phase evidenced in southernmost Norway, has been proposed (Bolle et al., in press).

The small intrusions of Handeland (∼2 km²) and Åseral (∼5 km²) were emplaced at a short distance south of Svoefjell (Fig. 1) and their mineralogy is similar to that of the main granites. The Handeland body is essentially quartz dioritic with biotite and hornblende as the main mafic phases as well as plagioclase and quartz. Relics of clinopyroxene partly reacted to hornblende are often present and can be abundant. Apatite, zircon and opaque(s), usually surrounded by titanite, are accessory phases. In the Åseral intrusion, composition ranges from quartz monzonite to granite. The major phases are plagioclase, K-feldspar (often microcline), quartz, biotite and hornblende. Accessory phases include opaque(s), usually surrounded by a rim of titanite, apatite, zircon and common fluorite.

The Lyngdal hyperites include the most mafic samples of the granitoid series and have been described by Demaiffe et al. (1990). Hyperite is an old Swedish name for a rock composed of hypersthene, plagioclase and augite and its application has been restricted to Scandinavia (Demaiffe et al., 1990). In the rocks discussed here, pyroxenes (orthopyroxene, clinopyroxene) are the dominant mafic minerals associated with rare amphibole, biotite is a minor phase. Ilmenite, magnetite, apatite, K-feldspar, quartz and pyrite are accessory minerals (Demaiffe et al., 1990). Using the exact IUGS terminology (Streckeisen, 1976), these samples will be called gabbronorites. Similar rocks have been described in the Laramie Anorthosite Complex (Fuhrman et al., 1988; Kolker and Lindsley, 1989) and named biotite gabbros or high-Al gabbros (Mitchell et al., 1995). The belonging of these gabbronorites to the granitoid series is discussed in detail in Section 8.1.
5. Mineral chemistry

Plagioclase composition (Table 2) ranges from An15 (Bessefjell) up to An31 (Svöfjell, Verhuskjerringi), An50 (Handeland: optical determination) and An58 (Lyngdal gabbronorites: Demaiffe et al., 1990). The amphibole (Table 3) present in Valle, Verhuskjerringi and Svöjell is an edenitic to magnesian hastingsitic hornblende (Leake, 1978) with a fluorine content ranging from 0.50 wt.% (Svöfjell) up to 1.06 wt.% (Valle). In the Lyngdal gabbronorites, the secondary amphibole is a ferroan pargasitic hornblende (Demaiffe et al., 1990). Biotite is richer in fluorine than amphibole with contents ranging from 0.63 wt.% (Svöfjell) up to 2.55 wt.% (Valle) (Table 4). Relic cores of clinopyroxene are salite with an average composition of En34Fs17Wo49 (Table 5). In the Lyngdal gabbronorites, the pyroxenes have the following average compositions: En39Fs13Wo48 (salite), En58Fs41Wo1 (hypersthene) (Demaiffe et al., 1990).

6. Geochemistry

6.1. Nomenclature

In the cationic classification of Debon and Le Fort (1983) (Fig. 2A), samples define a whole series, intermediate between the CALK (calc-alkaline) and SALKD (dark subalkaline potassic) trends, and stretching from the gabbro to the granite fields. Samples of the dykes crosscutting the agmatites in the surrounding gneisses of Svöfjell fall in the granite (84-48, 84-50), adammellite (SV90-6, SV90-8) and tonalite (84-59) fields. Most samples are metaluminous (A/CNK < 1; see Table 6) whereas a group of samples from Svöfjell, Verhuskjerringi (not shown in Table 6), Rustfjellet (not shown in Table 6) and Bessefjell are slightly peraluminous.

The granitoids plot in the subalkaline field in a TAS (Na2O + K2O versus SiO2) diagram (not shown) and their aqapatic index (Fig. 2B, Table 6) is below 0.87 except for a few samples of Bessefjell and Rustfjellet which lie between 0.87 and 1.00. This granitoid series is thus not alkaline (Légeois, 1988). In the Peacock diagram (Fig. 2C), the overall trend is calc-alkaline and in the K2O versus SiO2 diagram of Peccei and Taylor (1976) (Fig. 2D), they plot in the high-K calc-alkaline and shoshonitic fields. Nevertheless, these granitoids are also characterized by high FeOt/MgO, they plot in the theoleitic field in the AFM diagram (Fig. 2E).

Besides their characteristic ferro-potassic geochemical signature, the granitoids have also A-type affinities. Indeed, they display high contents of Ga (Ga/Al × 10,000 > 2.6: Whalen et al., 1987) and of incompatible elements (Zr + Nb + Ce + Y > 350 ppm: Whalen et al., 1987), relatively high alkali (Na2O + K2O ~ 8 wt.% at 70 wt.% SiO2: Eby, 1990) and F contents (1126–6532 ppm). Originally defined by Loiselle and Wones (1979) (with the prefix A—standing for anorogenic, anhydrous and alkaline), the A-type definition has been revised later (e.g. Eby, 1990). The general consensus (e.g. Frost et al., 1999) is now to consider that the A-type granitoids emplace into non-compressive environments at the end of an orogenic cycle (postorogenic or postcollisional granitoids), in continental rift zones or in oceanic basins. Geochemically, they are characterized by low CaO and Al2O3, high FeOt/MgO, high K2O/Na2O and high incompatible elements contents. Moreover, A-type granites are commonly considered as being reduced (Loiselle and Wones, 1979) as clearly shown in the rapakivi-type subgroup (Emslie, 1991; Emslie and Stirling, 1993; Frost et al., 1999), but relatively oxidized A-type granites have also been recognized (i.e. Bogaerts et al., 2001; Dall’Agnol et al., 1999).

The H2O content of A-type granites is also a matter of discussion. These granites were originally thought as nearly anhydrous (Loiselle and Wones, 1979), but experimental data have clearly shown that they may contain several wt.% of H2O (Bogaerts et al., 2001; Clemens et al., 1986; Dall’Agnol et al., 1999). Finally, A-type granites are not necessarily alkaline. Consequently, the A-type granitoids encompass a rather large group of rocks to which belongs the HBG suite.

6.2. The granitoids trend

6.2.1. Major elements

In Harker diagrams (Fig. 3), the nine intrusions define a single general trend with a small gap between 55
Fig. 2. Selected geochemical characteristics of the HBG suite. (A) $Q = \text{SiO}_2 - ((\text{K} + \text{Na} + 2\text{Ca})/3)$ (at.%) vs. $P = \text{K} - (\text{Na} + \text{Ca})$ (at.%) of Debon and Le Fort (1983); (B) agpaitic index $(\text{Na} + \text{K})/\text{Al}$ (at.%) vs. SiO$_2$ (wt.%). The limit at AI = 0.87 is from Liégeois and Black (1987); (C) peacock index $(\text{CaO}/(\text{Na}_2\text{O} + \text{K}_2\text{O}))$ (wt.%) vs. SiO$_2$ (wt.%). The limits are from Brown (1981); (D) K$_2$O (wt.%) vs. SiO$_2$ (wt.%); the limits are from Rickwood (1989); (E) AFM diagram (wt.%); the limit between tholeiitic and calc-alkaline fields are from Irvine and Baragar (1971).

- Gabbronartes
- Åseral
- Handeland
- Svitjell
- Svitjell dykes
- Holman
- Holman inclusions
- Rustfjell
- Valle
- Verhaukjerringi
- Besselfjell
Fig. 3. Major elements content (wt.%) vs. wt.% SiO₂. Same symbols as in Fig. 2.
and 59 wt.% SiO$_2$. The latter could result from a lack of samples as mafic samples represent a very small proportion of the outcrops. Indeed, the three samples of the small Åseral intrusion plot on each side of the gap suggesting a possible continuity in the trend. There is some scatter in the data points (i.e. TiO$_2$ and P$_2$O$_5$ for Svöfjell) which probably results from samples not being representative of pure liquids but of crystal (ilmenite and/or apatite) laden liquids. With increasing SiO$_2$, there is a regular decrease in CaO, MgO, FeO$_T$ and TiO$_2$ from 55 wt.% SiO$_2$ whereas P$_2$O$_5$ first increases from the gabbronorites to sample 98BN41C of differentiation processes. The REE content first increases and then decreases. A slight increase of FeO$_T$ and TiO$_2$ at low SiO$_2$ content is also observed. K$_2$O (see Fig. 2D) increases up to 72 wt.% SiO$_2$ and then decreases. Na$_2$O remains relatively constant with increasing SiO$_2$ whereas Al$_2$O$_3$ (not shown) slightly increases and then decreases. A slight increase of Sr and TiO$_2$ at low SiO$_2$ content is also observed. The N-MORB-normalized spidergrams of the average composition of the different massifs are shown in Fig. 6. These patterns are characterized by negative anomalies in Ba, Nb, Ta, Sr, P and Ti. The negative anomalies in Ba, Sr, P and Ti are more or less pronounced, consistent with different degrees of differentiation in the selected samples.

6.2.2. Trace elements
Sr, Zr, Ba, Ce (not shown), Y, Nb and Sc display a bell-shape trend with increasing SiO$_2$, whereas Co and V (not shown) regularly decrease and Rb increases (Fig. 4). The high Zr contents (900 up to 1100 ppm) observed in several samples of Holum likely results from some accumulation of zircon. Nb behaves similarly to Zr, as commonly observed (e.g. Duchesne and Wilmut, 1997), except in Bessefjellet where the amplitude of the variation is an order of magnitude larger than in the general trend.

REE patterns of selected samples from the different massifs and corresponding to various SiO$_2$ contents are shown in Fig. 5. The various patterns differ by the (La/Sm)$_N$ ratios and the magnitude of the Eu anomaly, and might correspond to different degrees of differentiation (see below for a discussion about the possible differentiation processes). The REE content first increases from the gabbronorites to sample 98BN41C of Holum and the least differentiates sample of the Svoifjell massif. Then the REE content decreases down to sample BE6 where a strong increase in the (La/Sm)$_N$ is observed. The Eu anomaly is first absent (Sk4, A10), then slightly positive (98BN39A) or, more frequently, slightly negative (SV90-13, 98BN41C, R7, H14). Finally, a strong negative Eu anomaly is displayed in samples BE4, BE5 and BE6. The average (La/Yb)$_N$ ratio (see Table 6) is higher in Rustfjellet (36.2) and Valle (40.2) than in the main granite of Svöfjell (7.6), Vehuskjerringi (7.8) and Bessefjellet (7.6). On the other hand, the gabbronorite is characterized by small (La/Yb)$_N$ values (average of 7.39: Demaiffe et al., 1990).

6.2.3. Isotopic data
Sr and Nd isotopic data are presented in Table 7 together with data from various typical gneisses. All calculations have been performed following Ludwig (2001), implying that errors were multiplied by $\sqrt{\text{MSWD}}$ when the latter was >1.2. The ages and thus initial isotopic compositions are known for the gabbronorites and some of the selected granitoids: gabbronorites (910 ± 82 Ma (Rb-Sr), $^{147}$Sm$^{143}$Nd = 0.7052-0.7054, $^\delta$Nd$^{143}$NdSM = +0.4 to +1.97, $^{206}$Pb/$^{206}$Pb = 17.45, $^{207}$Pb/$^{205}$Pb = 15.51: Demaiffe et al., 1990), Vehuskjerringi (932 Ma: U/Pb, Dahlgren, personal communication in Sylvester, 1998) and Bessefjellet (923 ± 16 Ma (Rb/Sr): Killeen and Heier, 1975). Enough samples have been measured on the Holum pluton for geochronological purposes and an age indication of 929 ± 47 Ma is obtained ($\text{Sr}_{i}^\delta$ 0.7046 ± 0.0006, MSWD = 1.7 for 7 WR; Fig. 7). An older age of 980 ± 34 Ma was proposed by Wilson et al. (1977) for this intrusion but if errors are considered, both ages overlap. When all measured HBG samples are considered, including samples from the Lyngdal massif (Bogaerts et al., in press), an errorchron is obtained but with a reasonable MSWD for such a large number of samples collected in a series of plutons extending along a distance of more than 100 km: 965 ± 19 Ma, $\text{Sr}_{i}^\delta$ 0.70433 ± 0.00056, MSWD = 14 for 39 WR (Fig. 7). When samples of the Bessefjellet pluton are excluded from the errorchron as their very high Rb/Sr could impose the slope, results are similar: 957 ± 25 Ma, $\text{Sr}_{i}^\delta$ 0.70447 ± 0.00064, MSWD = 15 for 37 WR (Fig. 7). These ages are identical within error limits, the weighted average being: 961 ± 14 Ma with a MSWD = 0.97. This latter value gives more weight to the individual pluton ages than the former ones and initial ratios are also nearly identical. Based
Fig. 4. Trace elements content (ppm) vs. wt.% SiO$_2$. Same symbols as in Fig. 2.
mainly on errorchrons, these ages are not geologically entirely meaningful but are in agreement with the U-Pb zircon age of 950 ± 5 Ma on the Lyngdal massif (Pasteels et al., 1979). However, the actual time span during which the HBG suite emplaced is not precisely constrained. For instance, the Holum massif (929 ± 47 Ma) and the small Verhuskjertring massif could be younger than 961 Ma if the cf. 932 Ma U-Pb zircon age for this latter massif is confirmed (Dahlgren, personal communication in Sylvester, 1998). Moreover, at 961 Ma, the $I_{Sr}$ of the Rustfjellet pluton is unrealistically low (0.7011–0.7015; Table 7). Consequently, $I_{Sr}$ and $\varepsilon_{Nd}$ have been calculated at 930 Ma, the emplacement age of the Rogaland AMC suite (Scharer et al., 1996). Note that recalculating the initial ratios at 961 Ma would not change the results much except for samples with very high Rb/Sr ratios but for these samples, the error on their initial ratios (Table 7) is so high that they are not very useful. At 930 Ma (Fig. 8), the granitoids display a characteristic narrow range in $I_{Sr}$ from 0.7027 (R7) up to 0.7056 (SV6) and a larger range in $\varepsilon_{Nd}$ ($+2.0$: Sk10: Demaiffe et al., 1990 down to $-4.9$ (VA2)). In Bessefjellet, the $^{87}Rb/^{86}Sr$ ratio is relatively high (62–66, Table 7) implying that the calculated $I_{Sr}$ is rather imprecise and strongly age dependent (at
Fig. 6. N-MORB normalized spidergrams of average compositions. N-MORB values from Sun and McDonough (1989). Same symbols as in Fig. 2.

$930 \text{ Ma}, I_s = 0.7329-0.7569$ and $\varepsilon_{\text{Nd}}(t) = -6.5$ and $-3.9$. Data for the Herefoss granite (Andersen, 1997) have also been plotted in Fig. 8 and fall in the range of the granitoids studied here. Our data are also in agreement with results of Andersen et al. (2001). The intrusions of Svojfell, Rustfjellet, Valle, Verhukskjerri and Holum have Sr contents higher than 150 ppm, $^{87}\text{Rb}/^{86}\text{Sr}$ ratio $<5$, $I_s < 0.710$ and $\varepsilon_{\text{Nd}}(t) < 0$, they thus belong to the ‘normal Sr concentration granite’ of Andersen et al. (2001). Bessefjellet has a Sr content lower than 150 ppm, a $^{87}\text{Rb}/^{86}\text{Sr}$ ratio $>5$, $I_s > 0.710$ and $\varepsilon_{\text{Nd}}(t) > 0$ and thus fall in the ‘low Sr granite’ group of the same authors, the gabbroenorites have an $I_s < 0.705$ and $\varepsilon_{\text{Nd}}(t) > 0$.

Model ages have been calculated according to the depleted mantle model of Nelson and DePaolo (1984) and are shown in Table 7. They range from 1.4 to 1.7 Ga (plus sample BE5 at 2.2 Ga) and are in agreement with model ages presented by Andersen et al. (2001). The two samples collected close to Svojfell (84-48: charnockitic dyke and 84-53: granite) have clearly distinct $I_s$, 0.7168 and 0.7075, respectively; they are quite different from Svojfell ($I_s = 0.7039$ to 0.7049) and are thus not comagmatic with this body.

7. Geothermobarometry

Several granitoids contain the appropriate assemblage (plagioclase, K-feldspar, quartz, ilmenite or magnetite, titanite, hornblende and biotite) to use the Al-in-hornblende geobarometer. We have used the two experimental calibrations of this geobarometer (fluid-saturated with varying proportion of H$_2$O and CO$_2$: Johnson and Rutherford, 1989; H$_2$O-saturated fluid: Schmidt, 1992) as well as the Anderson and Smith (1995) calibration which incorporates temperature as well as the above-mentioned experimental data. For this latter calibration, temperature was...
estimated with the hornblende-plagioclase thermometer of Holland and Blundy (1994), in which amphibole composition was recalculated according to the formula embedded in this thermometer. Results (Table 8) indicate a pressure range from 1.3 up to 5.6 kb. When temperature is taken into account, the pressure range is reduced to 1.3 up to 2.7 kbar (Anderson, 1996). These granitoids appear thus to be rather low pressure plutons; it is important to keep in mind however that a pressure of 2 kbar is the lower limit of the calibration range of the geobarometer.

Nevertheless, these pressure estimates are in agreement with recent experimental data obtained on the Lyngdal granodiorite which belongs to the HBG suite (Bogaerts et al., 2001). The calculated pressure range for the HBG suite overlaps the pressure of emplacement of the Rogaland AMC suite (≤5 kbar; Vander Auwera and Longhi, 1994; Vander Auwera et al., 1998a) suggesting that both suites were emplaced approximately at the same level of the upper crust and maybe at the same age (~930 Ma) (Schärer et al., 1996).
Several geothermometers have been used to estimate temperatures of crystallization. Temperatures estimated with the hornblende-plagioclase geothermometer of Holland and Blundy (1994) range between 798 and 825 °C and the hornblende-clinopyroxene equilibrium (Perchuk et al., 1985) gives a temperature of 770 °C for sample SV90-13.

The equations derived from the experimental data of Harrison and Watson (1984) and Watson and Harrison (1983) on the solubility of zircon and apatite in subaluminous melts can be used as geothermometers. Results are shown in Table 8: apatite saturation temperatures range from 1005 up to 1054 °C whereas zircon saturation temperatures are much lower (774

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**Table 8**

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to 876 °C. Temperatures obtained with these geothermometers agree with petrographic and chemical data. For example, in sample SV90-13 (Svöfjell) apatite occurs as small needles dispersed in the other phases, suggesting its early crystallization and thus the presence of apatite at or near the liquidus of the magma. From the P2O5–SiO2 diagram, it can be deduced that apatite saturation occurred around 53 wt.% SiO2. On the other hand, the low temperatures of zircon saturation are similar to those obtained for amphibole crystallization and the occurrence of rims of hornblende surrounding clinopyroxene supports their late crystallization. Moreover, this hypothesis is corroborated by the bell-shape trend observed in the Zr versus SiO2 diagram (see Fig. 4) pointing to zircon saturation at 63 wt.% SiO2. These data thus show that several phases occurring in a single sample did not necessarily crystallize simultaneously and further indicate that the mineral composition may register intervals of equilibrium crystallization.

The assemblage magnetite + quartz + titanite in the granitoids suite suggests an oxygen fugacity above NNO (Wones, 1989).

8. Discussion

8.1. Possible differentiation processes

In Harker diagrams, the small intrusions of Åseral, Handeland-Tveit as well as the Lyngdal and Skoland gabbro-norites plot on the same trend as the larger granitoids (Figs. 3 and 4) suggesting that they all belong to the same suite of rocks. The Lyngdal and Skoland gabbro-norites were previously thought to belong to the AMC suite (Demaiffe et al., 1990) as they display similar mineralogical and geochemical compositions to the jotunites (hypersthene bearing monzodiorites) of this latter suite. The gabbro-norites and jotunites show comparable enrichments in TiO2, P2O5, K2O as well as similar REE patterns and isotopic data suggest that they derive from the same isotopic reservoir (Demaiffe et al., 1990). Nevertheless, significant mineralogical and geochemical differences also occur. Biotite is a late stage phase in both gabbro-norites and jotunites but it is much more abundant in the former rocks suggesting that these contain a few wt.% of H2O whereas the jotunites are practically anhydrous. The gabbro-norites are distinctly lower in (Na + K)/Al (aluminian index) and FeOt and higher in CaO than jotunites, and in most variation diagrams, the AMC and HBG (including the gabbro-norites) suites define two distinct trends (Vander Auwera et al., 2001; Vander Auwera et al., in preparation). It is also worth emphasizing that the Lyngdal and Skoland gabbro-norites outcrop at the northern margin of the Lyngdal granodiorite, which is the southernmost HBG pluton of the Rogaland-Vest-Agder sector (Bogaerts et al., in press) (Fig. 1), and, in this massif, lobate enclaves of gabbro-norites are mingled with the granodiorite clearly indicating that both rock types are representative of penecontemporaneous liquids and are thus genetically linked. The relic cores of clinopyroxenes observed in some amphiboles of Svöfjell (see Table 5) have a Mg# of 0.66. Using the Fe-Mg exchange distribution coefficient of 0.23 (Grove and Bryan, 1983), the calculated Mg# of the liquid in equilibrium with these clinopyroxenes is 0.31. This value is lower than the Mg# of the gabbro-norites (≈0.5), but higher than the Mg# of the least differentiated sample of Svöfjell (Mg# = 0.25). Moreover, the relic cores of clinopyroxene have certainly not retained their liquidus composition as they reacted to amphiboles: their Mg# is thus a minimum value. As discussed above, the gap occurring between 55 and 59 wt.% SiO2 probably results from the small proportion of mafic samples as the higher density of these magmas traps them in the lower crust and/or from the fact that they have differentiated to produce more evolved magmas plus cumulates. Moreover, as already mentioned, the three samples of the Åseral intrusion plotting on each side of the gap give further support to a continuous trend. The trend ranging from 50 up to 78 wt.% SiO2 defined by the HBG suite could result from mixing processes between two (basaltic and granitic) or three (basaltic, intermediate and granitic) components, from partial melting processes or from a fractional crystallization process with (AFC process) or without assimilation. Mixing between two components can be precluded here. It should result in linear evolution in all variation diagrams, which is obviously not the case here. Indeed, several elements (P2O5, Zr, Ba, Sr and Sc: Figs. 3 and 4) show a bell-shape evolution indicating that with increasing content of an incompatible element, the liquid becomes saturated in a phase containing the element (which therefore
The presence of these bell-shape curves thus supports a fractional crystallization process. However, these trends could also result from two mixing stages between three components: a first mixing stage between the gabbronorites and an intermediate component similar in composition to the least differentiated sample of Svøffjell (e.g. SV90-13; see Table 6) and a second mixing stage between the intermediate and the granitic components. This hypothesis is however unlikely for four reasons: (1) when mixing occurs between more than two components, representative samples are usually much more dispersed in Harker diagrams and particularly occur inside the concavity of the bell-shape curves (see the example of the Tismana pluton; Duchesne et al., 1998); (2) a Rb-Sr isochron has been obtained for Holum which makes mixing between the intermediate and granitic components unlikely; (3) the first mixing stage should have produced linear arrays in all variation diagrams which is not the case here as TiO$_2$, P$_2$O$_5$, and FeOt display a slight increase and then a steady decrease with increasing SiO$_2$ (Fig. 3); (4) the isotopic composition of the gabbronorites ($\varepsilon$Nd(t) = 0.7052–0.7054, $\varepsilon$Sr = 0 to +1.97; Demaiffe et al., 1990) is close to that of the less contaminated granitoid (Verhuijskenning: $\varepsilon$Nd = 0.70385; $\varepsilon$Sr = −0.68; Table 7 and Fig. 8). One could further argue that the lack of samples inside the concavity of the bell-shape curves in Harker diagrams is not a definitive argument as the two stages of the mixing process could have occurred at two different crustal levels in successive magma chambers at decreasing depths. Nevertheless, representative samples of this supposed first mixing stage (that took place at a lower crustal level), were finally emplaced in the upper crust where they should have been mixed with the granitic end-member, hence producing samples plotting inside the concavity of the bell-shape curves. Moreover, such a two-stage mixing process does not encounter the other arguments: isochron obtained for Holum; non-linear arrays in the TiO$_2$, P$_2$O$_5$, and FeOt variation diagrams. In conclusion, even if rather complex mixing processes occurring in separate systems cannot be completely ruled out here, we suspect that they would have produced much more scatter in the variation diagrams and we thus not retain a mixing process to explain the observed trend. The lobate inclusions of gabbronorites observed in the Lyngdal granodiorite are thus interpreted as mafic injections in a more differentiated, still partially liquid magma chamber.

Two melting processes of two different protoliths could produce a broken linear array. Nevertheless, it is shown by Bogaerts et al. (in press) that a melting process does not fit the representative data points of the Lyngdal granodiorite (very similar to Svøffjell) which corresponds to the second part of the HBG trend. On the contrary, a fractional crystallization process is successfully modelled for major and trace elements using the least squares regression method and experimental data (Bogaerts et al., 2001). The fractionating cumulate is made of clinopyroxene, hornblende, plagioclase, oxides, biotite, apatite, zircon and allanite (Bogaerts et al., in press). A melting process could indeed explain the first part of the HBG trend but this hypothesis implies that the second part of the trend would correspond to the fractional crystallization of the low degree melts of this first melting process. A question thus remains open: why the other liquids of this melting process did not evolve through fractional crystallization? Other observations do not favor this melting process. The three data points of the Åseral intrusion straddle the two parts of the HBG trend (see Co versus SiO$_2$ in Fig. 4) pointing to an identical differentiation process, fractional crystallization, for both parts of this trend. The Rb-Sr isochron (910 ± 82 Ma, MSWD = 0.74) obtained by Demaiffe et al. (1990) on the gabbronorites is also better explained by a fractional crystallization process. Finally, in a bilogarithmic Co (compatible element) versus Rb (strongly incompatible element) plot (Fig. 9) (Allègre et al., 1977; Hanson, 1978), the HBG trend can be approximated by a broken line supporting the hypothesis of a fractional crystallization process. The first segment of this line includes the Handeland-Tveit intrusion and two samples of Åseral and the second, the rest of the samples. These two segments could correspond to the subtraction of two distinct cumulates (Allègre et al., 1977), in agreement with observations from variation diagrams.
This hypothesis of extensive fractional crystallization of basalts to produce A-type granites was already pointed out by Frost and Frost (1997), Loiselle and Wones (1979) and Turner et al. (1992). Cumulates formed during this fractional crystallization process were probably trapped lower in the crust as no layered intrusion belonging to the HBG suite has ever been observed at the present level of exposure. It is worth noting that in the penecontemporaneous AMC suite the whole series of corresponding cumulates has been described in the Bjerkreim-Sokndal layered intrusion (Wilson et al., 1996).

The narrow range of $I_{Co}$, the existence of reliable Rb-Sr age indications together with a quite large variation in $\varepsilon_{Nd}$ (+1.9 to −6.51) in the members of the HBG suite can result either from sources with variable $\varepsilon_{Nd}$ or from assimilation during fractional crystallization (AFC process), the two processes may have played a role together. The two gabbro-norites have similar $SiO_2$ content but very different $\varepsilon_{Nd}$ (Sk10: +1.9 for 51.38 wt.% $SiO_2$; Ly11a: +0.4 for 51.62 wt.% $SiO_2$; Demaiffe et al., 1990) indeed suggesting source variability. Andersen et al. (2001) noted that for most granites, the depleted mantle Nd model ages ($T_{DM}$: 1.38–1.67 Ga) decrease slightly westwards from the

![Fig. 9. Co and Rb (ppm) contents in bilogarithmic scale.](image)
Østfold-Akershus sector to the Rogaland-Vest Agder sector. They attributed this slight decrease to a progressive increase of the proportion of juvenile component in the source of the granitic magmas. In this latter model, the TDM model ages are intermediate between the ages of the two possible components (juvenile and crustal) of the source, indicating an old crustal contaminant at least older than 1.67 Ga, possibly of Svecofennian age (cf. 2 Ga). In the hypothesis of an AFC process, the banded gneisses sampled in the vicinity of the plutons (Tables 6 and 7) and at some distance from the Rogaland anorthositic complex as well as the Bamble metasediments (Andersen et al., 1995) have too high \( I_{\text{Sr}} \) and \( \epsilon_{\text{Nd}} \) to be plausible contaminants. Isotopic data favor contamination by a Rb-depleted material characterized by strongly negative \( \epsilon_{\text{Nd}} \) and intermediate \( I_{\text{Sr}} \) (<0.710), i.e., an old granulitic lower crust.

8.2. Non-CHARAC or/and tetrad effects

The Besserjet intrusion displays geochemical characteristics significantly different from those of the other intrusions: high Rb/Sr ratio (Table 6), low K/Rb ratio, almost vertical trends in the Rb versus SiO\(_2\), Nb versus SiO\(_2\) (Fig. 4) and Ta versus SiO\(_2\) (not shown) plots. Moreover, some samples have a peraluminous character. The Rb and Nb vertical trends are difficult to ascribe to magmatic differentiation; fluid/rock interaction (non-CHARAC or tetrade effects) at the magmatic stage can be suspected. In a geochemical system characterized by charge-and-radius-controlled (CHARAC) trace elements behaviour, elements having close charges and radii are expected to show coherent behavior (chondritic ratios) and normalized patterns of the trivalent REE should be smooth.

![Tetrad effect](image1)

![Tetrad effect](image2)

![F vs. K/Rb and Zr/Hf](image3)

![F vs. K/Rb and Zr/Hf](image4)

Fig. 10. Tetrad effect (t1 parameter from Iler, 1999: \( T_{\text{E}_1} = (t1 \times t_3)^{0.5} \)), \( t_1 = (\text{Ce/Ce}_{\text{t}} \times \text{Pr/Pr}_{\text{t}})^{0.5} \) with \( \text{Ce/Ce}_{\text{t}} = \text{Ce}_{\text{N}}/(\text{La}_{\text{2/3}} \times \text{Nd}_{\text{1/3}}) \) and \( \text{Pr/Pr}_{\text{t}} = \text{Pr}_{\text{N}}/(\text{La}_{\text{1/3}} \times \text{Nd}_{\text{2/3}}) \). The \( t_3 = (\text{Tb/Tb}_{\text{t}} \times \text{Dy/Dy}_{\text{t}})^{0.5} \) with \( \text{Tb/Tb}_{\text{t}} = \text{Tb}_{\text{N}}/(\text{Gd}_{\text{2/3}} \times \text{Ho}_{\text{1/3}}) \) and \( \text{Dy/Dy}_{\text{t}} = \text{Dy}_{\text{N}}/(\text{Gd}_{\text{1/3}} \times \text{Ho}_{\text{2/3}}) \) vs. K/Rb and Zr/Hf and F content (ppm) versus K/Rb and Zr/Hf. Chondritic values are from Anders and Grevesse (1989). See text for explanation.
functions of atomic number and radius (Bau, 1996; Jahn et al., 2001). Highly evolved magmas, like Bessefjellet, are potentially very enriched in H2O, F, P, Cl and appear therefore transitional between pure silicate melts and hydrothermal fluids. In such systems, the behavior of trace elements can also be governed by complexation processes resulting in non-chondritic Y/Ho, Zr/Hf ratios or even in the lanthanide tetrad effect (Bau, 1996; Jahn et al., 2001). The tetrad effect refers to the subdivision of the REE into four groups called tetrads (La-Nd, Pm-Gd, Gd-Ho, Er-Lu) displaying upwards or downwards concavities with minima at La, between Nd and Pm, at Gd, between Ho and Er, and at Lu (Bau, 1996; Iber, 1999). In order to quantify the tetrad effect, Iber (1999) has proposed to use the $T_{E1,3} = (t_1 \times t_3)^{0.5}$ parameter (see Fig. 10) which determines the deviation of a REE pattern with a tetrad effect from a hypothetical effect-free REE pattern. We have used here only the $t_1$ parameter which evaluates this deviation for the first tetrad (La to Nd) as the third tetrad (Gd to Ho) does not show any concavity ($t_3 = 1$) (Fig. 5). In Fig. 10, the tetrad effect $t_1$ parameter of Iber (1999) and the F content are shown as functions of the K/Rb and Zr/Hf ratios and compared with the chondritic values of these parameters. Fig. 10 shows that in Bessefjellet, a very slight tetrad effect is significant (above 1.1; Iber, 1999) only in three samples. Nevertheless, the K/Rb, Zr/Hf and Nb/Ta ratios (not shown: average of 11 in Bessefjellet compared to 17 for the chondritic value) are indeed much lower than chondritic values and are correlated with high F contents in this intrusion.

9. Possible sources and geodynamic implications

In order to better constrain the geodynamical setting of the HBG suite, the possible sources of the parent gabbroic magmas have to be discussed. We have mentioned above that the magmas that gave rise to the gabbronorites of the HBG suite probably contained several percent H2O. This moderate H2O content must be ultimately derived from the source which contained hydrated phases (e.g. mica, amphibole). Moreover, the gabbronorites are characterized by low (La/Yb)N values (average of 7.39; Demaiffe et al., 1990) suggesting that garnet was absent in the residue. It is thus plausible that these gabbronorites were derived by partial melting of a garnet-free, hydrated, undepleted to slightly depleted ($\varepsilon_{Nd} > 0$) and potassic mafic source, lying either in the lithospheric upper mantle or in the mafic lower crust derived from it. It is worth mentioning here that experimental data obtained by Rapp and Watson (1995) on dehydration melting of metabasalts are in agreement with this hypothesis. Experimental liquids obtained at 8 kbar and 1075, 1050 and 1000 °C from an amphibolite are very similar in composition to the gabbronorites and the Handeland small intrusion of the HBG suite (Auwera et al., in preparation).

As already mentioned above, the HBG suite is close in age and space with the AMC suite of Rogaland (≈930 Ma). The parent magmas of this suite are the least differentiated jotunites (Demaiffe and Hertogen, 1981; Duchesne et al., 1974; Vander Auwera et al., 1998a), called primitive jotunites by Vander Auwera et al. (1998a), and phase equilibria based on experimental data further indicate that these primitive jotunites are products of the partial melting of an anhydrous gabbronorite source (mafic granulite or gabbronoritic cumulates) in the lower crust (11 kbar) (Longhi et al., 1999). Consequently, partial melting of two distinct sources (a gabbronoritic one and an hydrated potassic mafic one) probably occurred penecontemporaneously beneath southern Norway (Vander Auwera et al., 2001). These two distinct sources may reflect increasing degree of metamorphism (amphibolite to granulite) from East to West (Bingen and van Breemen, 1998a) if the gabbronoritic source is a granulite. It could also point to an horizontal stratification of the lower crust (Rudnick and Fountain, 1995), a stratification of the lithosphere (melting of the lower crust or upper mantle) or may indicate that the AMC and granitoid suites belong to two distinct crustal segments as proposed by Duchesne et al. (1999).

Demaiffe et al. (1990) and Demaiffe et al. (1986) indicated that isotopic data on jotunites (AMC suite) and gabbronorites (HBG suite) suggest slightly depleted upper-mantle origin or an origin in the lower crust by melting of depleted mafite-derived basic rocks, shortly (<200 Ma) after their formation. Isotopic data corroborate phase equilibria and bring the additional constraint that, in the case of a partial melting process, emplacement of the basic protolith in
the lower crust must have occurred at an age younger than about 1130 Ma. It is worth emphasizing that a significant magmatic event occurred at 1050 Ma in southern Norway with the emplacement of the calc-alkaline augen gneisses series (Bingen et al., 1993; Bingen and van Breemen, 1998b). The geochemical evolution of these augen gneisses can be accounted for by fractional crystallization of a parent magma resulting from mixing between ultrapotassic, mantle-derived mafic magma (20–25%) with a granodioritic magma generated in the lower crust presumably in a subduction-related geodynamic regime (Bingen et al., 1993). Emplacement of ultrapotassic basalts in the lower crust may thus have occurred around 1050 Ma yielding a plausible protolith for the gabbroic parent magmas of the HBG suite.

The production of large volumes of magmas during a small period of time implies that a major thermal pulse occurred at that time in southern Norway and was likely linked to an important geodynamic feature of the Sveconorwegian evolution. A similar suggestion was proposed by Bingen and van Breemen (1998a) who pointed out that the M2 low-pressure thermal metamorphism dated at 930–925 Ma had a “much broader significance than a local phase of contact metamorphism associated with the intrusion of anorthosite plutons” (p. 351). Similarly, in southwestern USA, the 1.4 Ga A-type magmatism is associated with a broad thermal anomaly (Frost et al., 1999). Regional extension and these authors suggested that the necessary heat was probably supplied by the mantle. The vector of this mantle heat was more likely a linear uprise of the asthenosphere following a lithospheric delamination rather than a mantle plume in an intracontinental setting (Albarède, 1998; Frost et al., 2001). Indeed, the HBG suite is roughly linear along the Mandal-Ustaoset shear zone and deep seismic profiles have shown that this shear zone corresponds to a significant Moho offset demonstrating its lithospheric scale (Andersson et al., 1996; Duchesne et al., 1999). Moreover, a plume geodynamic setting has not been favored for AMC suites (Ashwal, 1993). This asthenospheric uprise could be at the origin of the melting of both the hydrated potassic mafic source for HBG generation and the lower crustal gabbroinoritic source for AMC generation. Such a geodynamical environment is typical during the postcollisional period (Laget et al., 1999).

10. Conclusions

A suite of granitoids, the HBG suite, related to the MUL belongs to the distinctive group of Proterozoic ferro-potassic A-type granites, also recognized in many cratonic areas and usually defined as “anorogenic”. This suite displays an extensive differentiation trend ranging from gabbronorites (50 wt.% SiO₂) to granites (77 wt.% SiO₂) which most probably results from fractional crystallization of several batches of parent basaltic magmas with similar major and trace elements compositions. Moreover, contrary to what was currently admitted for A-type granites (Loselle and Wones, 1979), the HBG suite is characterized by relatively high water contents and oxygen fugacity (NNO).

The HBG suite is probably penecontemporaneous with the AMC suite of Rogaland and the parent magmas of these two suites resulted from the partial melting of two different sources: a lower crustal anhydrous, gabbroinoritic source for AMC and an hydrous, undepleted to slightly depleted potassic mafic source for the HBG. The penecontemporaneous melting of these two contrasting sources, anhydrous mafic lower crust versus hydrous mafic–ultramafic potassic crust or mantle could reflect increasing degree of metamorphism from East to West, stratification of the lithosphere (mantle versus crust) or of the continental crust itself (Rudnick and Fountain, 1995) or may indicate that the two suites belong to two distinct lithospheric segments as formerly proposed by Duchesne et al. (1999). Linear lithospheric delamination along a major shear zone with consequent asthenospheric uprise could explain the HBG alignment within the MUL.

Acknowledgements

G. Bologne and G. Delhaze are greatly thanked for their analytical and samples preparation work, respectively. I. Roelandts provided F analyses for a selection of samples. Part of the trace element analyses were performed by INAA at Pierre Sue Laboratory, CEN, Saclay, under the supervision of J.-L. Joron, by E.W. who has benefited from an EC doctoral grant at the University of Paris VI. XRF analyses and ICP-MS analyses were performed at the “Collectif

Interinstitutionnel de Géochimie Instrumentale” (University of Liège). Isotopic analyses were performed at the “Centre Belge de Géochronologie” (University of Brussels and Africa Museum, Tervuren). This work was funded by the Belgian Fund for Joint Research. R.F. Emslie and S. Fourcade are greatly acknowledged for their constructive reviews.

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