

Geothermobarometry of igneous and metamorphic rocks around the Åna-Sira anorthosite massif: Implications for the depth of emplacement of the South Norwegian anorthosites

EDITH WILMART & JEAN-CLAIR DUCHESNE

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Orthopyroxenes and clinopyroxenes in metabasites in contact with the massif-type Åna-Sira anorthosite yield the following equilibrium temperatures using various geothermometers: 840°C (Wells 1977); 800°C (Wood & Banno 1973), 650°C (Lindsley & Andersen 1983). Garnet–cordierite–plagioclase–sillimanite–quartz assemblages in neighbouring metapelites give values around 600°C and 3 ± 0.5 kb by means of several geothermobarometers. Armoured relics of symplectitic quartz + spinel in garnet suggest that equilibrium was attained during decreasing temperature. Fayalite + quartz and Ca-poor + Ca-rich clinopyroxene assemblages in charnockitic rocks of late tectonic intrusions yield 800–850°C (Lindsley & Andersen 1983) and 6–7.5 kb for the upper part of the Bjerkreim–Sokndal lopolith and a minimum pressure of 5–5.5 kb for the Hidra massif (Bohlen & Boettcher 1981). When interpreted in terms of the regional geological evolution, the conditions measured reflect two successive stages of a PT path characterizing the post-magmatic evolution. It is inferred that the emplacement of the Rogaland massif-type anorthosites took place at greater depths than these massifs, i.e. not less than 22–28 km near the Bjerkreim–Sokndal massif and not less than 18–20 km near the Hidra body.

E. Wilmart & J. C. Duchesne, Lab. assoc. de Géologie, Pétrologie et Géochimie, Université de Liège, B-4000 SART TILMAN (Belgium)

Anorthositic massifs south of the Grenville Front are associated with medium- to high-pressure granulite-facies rocks, thus suggesting an emplacement at deep levels in the crust (e.g. Morse 1982; Duchesne 1984). However, isotopic data (Valley & O'Neil 1982) seem to indicate that in the Adirondacks, south of the Grenville Front, anorthosites were emplaced at relatively shallow depths (<10 km), prior to burial and metamorphism at high pressure.

The anorthosites of the Rogaland igneous province (South Norway), belonging to the extension of the Grenville Province, have long been considered syntectonic (Michot 1969). Consequently the PT conditions of emplacement seem to be those of the regional granulite-facies metamorphism. The minimum temperature of 660°C, resulting from the occurrence of mesoperthite, suggested to Michot depths of about 20 km (5 kb), assuming a normal geothermal gradient (30°C per km). This view was later supported by the occurrence of Fe-rich orthopyroxene (Duchesne 1972b), and the garnet–cordierite association indicating 6–7 kb (Henry 1974; Demaiffe 1977).

Subsequently, however, geochronological studies (Demaiffe & Michot 1985) and new petrological (Tobi et al. 1985; Jansen et al. 1985) and structural data (Hermans et al. 1975) have led to the suggestion that the emplacement of anorthosites was not coeval with the regional deformation. Moreover, according to Maquil & Duchesne (1984), the anorthosites are late- to post-tectonic intrusions, which were capable of deforming their envelope. These new views differ from the earlier approach which emphasized the syntectonic character of anorthosites.

The purpose of the present paper is to supply data about the PT conditions recorded in rocks from the vicinity of the Åna-Sira anorthosite massif and the upper part of the neighbouring Bjerkreim–Sokndal lopolith (Fig. 1).

Regional geological evolution

In the general model of evolution of the Rogaland igneous province it is widely accepted that the Åna-Sira massif, by analogy with the Egersund–

Ogna massif, had completed its emplacement by the end of the Sveconorwegian orogeny (ca. 1000 Ma) in an envelope of gneisses previously deformed and metamorphosed on a regional scale (Falkum & Petersen 1980; Demaiffe & Michot 1985; Duchesne et al. 1985).

The eastern envelope of the Åna-Sira massif (Fig. 1) comprises a series of banded gneisses. The latter is 1–2 km thick and made up of a variety of rocks, banded on a dm–m scale, intensely deformed, affected by several episodes of migmatization, and locally blastomylonitic. The banded series occurs at the contact with the magmatic rocks and is conformable on a large scale with the Åna-Sira massif.

Several igneous bodies have intruded the contact zone between the Åna-Sira massif and the banded series. They are: The Farsund charnockite; the Apophysis (a thin mangero-monzonoric sheet intrusion); the Hydra leuconorite body and the Breimyrknutan charnockite (Duchesne et al. 1987). The last two show igneous structures, are free of any post-emplacment deformation, and, because of their post-tectonic character, are of particular interest.

According to Demaiffe & Hertogen (1981), the *Hydra massif* results from the fractional crystallization of a monzonoritic liquid which yielded charnockitic residual liquids forming a stockwerk of sharp-walled dykes in the leuconorite of the central part of the massif. The *Breimyrknutan charnockitic intrusion* (Fig. 1A) has been emplaced into the banded series and is probably a product of anatexis triggered by the intrusion of the Hydra massif (Duchesne et al. 1987). These two intrusions show petrographic and geochemical similarities with the upper part of the *Bjerkreim-Sokndal lopolith* (BKSK) (Fig. 1) (Michot 1965). Structurally BKSK is also situated between the metamorphic envelope and the anorthositic massifs; its final crystallization took place at about the same time as the other two intrusions (ca. 930 Ma) (Wielens et al. 1980).

The metamorphic and igneous rocks around the

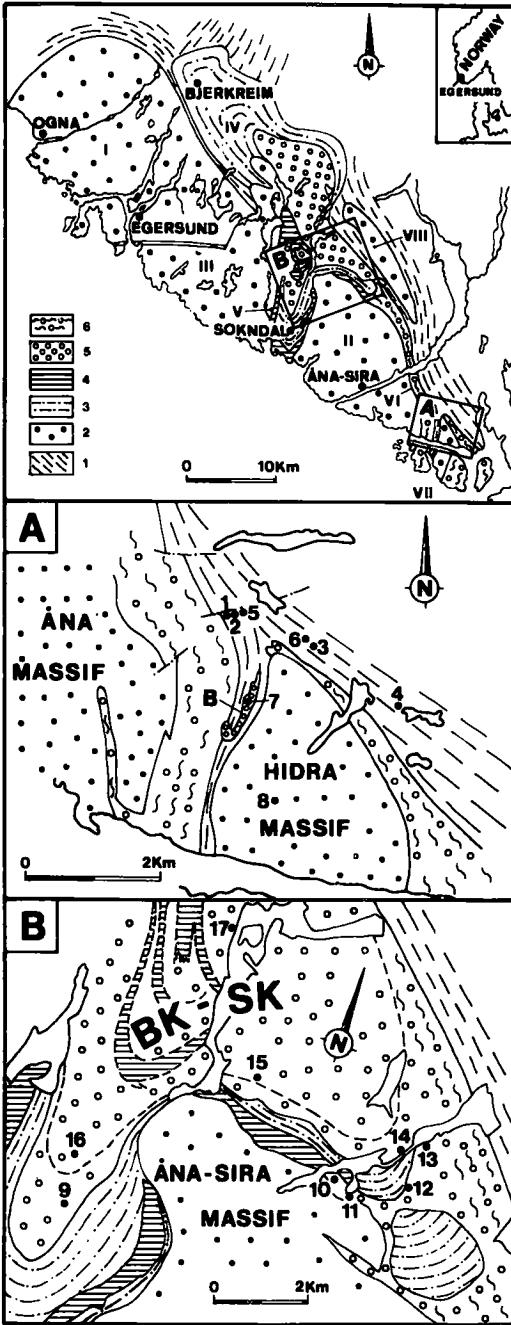


Fig. 1. Schematic geological maps of the Rogaland igneous province (S.W. Norway) (after Michot & Michot 1969; Rietmeijer 1979; Duchesne et al. 1987). Legend: 1 – high-grade metamorphic rocks; 2 – anorthosites; 3 – leuconorites and norites; 4 – monzonorites; 5 – intermediate to acidic rocks (upper part of the Bjerkreim-Sokndal lopolith); 6 – slightly oriented to foliated intermediate to acidic rocks (Apophysis and Farsund charnockite). I, II and III: massif-type anorthosites of Egersund-Ogna, Åna-Sira and Håland-Mølleren, respectively; IV: Bjerkreim-Sokndal layered lopolith; V: Eia-Rekefjord intrusion; VI: Apophysis; VII: Hydra massif; VIII: Garsaknatt massif. A. Detailed map of the southeastern contact zone of the Åna-Sira massif. B. Breimyrknutan charnockitic intrusion. Sample locations indicated by numbers 1 to 8. B. Detailed map of the northern contact zone of the Åna-Sira massif with the Bjerkreim-Sokndal lopolith. Sample locations indicated by numbers 9 to 17.

Åna-Sira anorthosite exhibit mineral associations typical of granulite-facies conditions, which have lasted for a long time in the tectonometamorphic evolution: During the regional metamorphism dated to 1350 ± 200 Ma (Weis & Demaiffe 1983), then during the emplacement of the Åna-Sira intrusion and the concomitant blastomylonitic and migmatitic reworking, finally during the ensuing intrusion and consolidation of the latest charnockitic bodies. The conformable structure of the banded series at its contact with the Åna-Sira massif suggests that the rocks underwent a deformation and a recrystallization connected with the emplacement of the anorthosite. Their mineral association results therefore from contact metamorphism superimposed on an earlier regional metamorphism. It is not excluded, however, that they were also affected by the intrusion of the charnockitic masses or by cooling during isostatic uplift.

A model for the PT evolution of the high-grade metamorphic terrain has been proposed by Jansen et al. (1985) (Fig. 2). Three stages of metamorphism have been distinguished. The last one

(M3) is of retrograde type varying in temperature from 700°C to 600°C at about 3 kb total pressure. The second stage M2, related to the intrusion of the igneous bodies and particularly of the BKSK lopolith, yields temperatures of $800\text{--}1050^\circ\text{C}$ at pressures of 3–4 kb. The first stage (M1) is estimated at $750\text{--}900^\circ\text{C}$ and 6–8 kb. Swanenberg (1980) indicated fluid inclusions with homogenization temperatures T_h lower than -50°C , and surmised the occurrence of trapped fluids of ultra-high equivalent CO_2 densities, quite incompatible with the M1 to M3 evolution. A recent reinvestigation of Swanenberg's samples by Touret and Van den Kerkhof (pers. comm.), however, shows that the trapped fluids studied by Swanenberg are not pure CO_2 , but $\text{CH}_4\text{--N}_2$ mixtures. These authors, moreover, suggest the range of CO_2 isochores (T_h between -28°C and $+17^\circ\text{C}$) represented in Fig. 2. Only the isochores of the lowest density are compatible with M3 conditions. Moreover, the late stage fluids compositions are dominated by CH_4 and N_2 , not by CO_2 , the latter being restricted to the early stages of the metamorphic evolution.

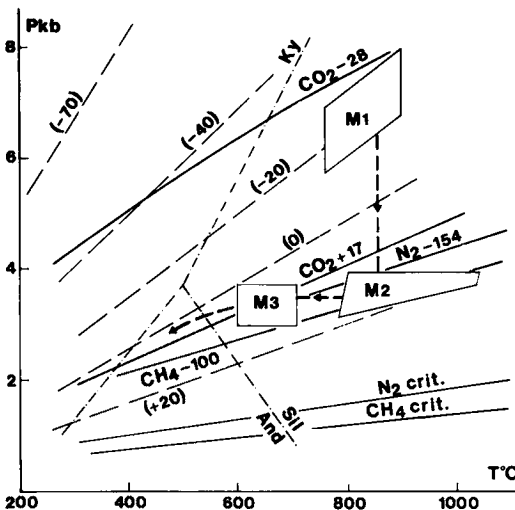


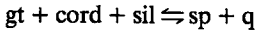
Fig. 2. Pressure-temperature diagram (after Touret 1985) showing the three stages of metamorphism (M1 to M3) of Jansen et al. (1985). The stippled lines are the isochores for the CO_2 inclusions of Swanenberg (1980). The numbers in parentheses are the homogenization temperatures (T_h) of the liquid phase corresponding to the CO_2 -equivalent density of the isochore. The continuous lines are isochores corresponding to the CO_2 , CH_4 and N_2 fluids analysed by Touret & Van den Kerkhof (pers. comm.). The nature of the fluid and the T_h are indicated on each isochore. Al_2SiO_5 stability relations from Holdaway (1971).

Petrography

In an attempt to determine the P–T conditions of equilibrium, a number of rock types have been studied: (1) metabasites (specimen nos. 1–4) and kinzigitic gneisses of the banded series (nos. 5–6), (2) the Breimyrknutan charnockite (no. 7), a charnockitic dyke of the Hidra massif (no. 8), and intermediate and acidic rocks from the upper part of BKSK lopolith (nos. 9–17). Location of the specimens is shown in Fig. 1.

The *metabasites* of the banded series (nos. 1–4) essentially comprise plagioclase (pl), orthopyroxene (opx) and clinopyroxene (cpx), with amphibole and/or biotite. Apatite, quartz and opaque minerals are accessories. Their texture is granoblastic, generally equigranular. The *kinzigitic gneisses* of the banded series (nos. 5–6) contain garnet (gt), cordierite (cord), sillimanite (sil), potassium feldspar, plagioclase (pl), aluminous spinel (sp), Fe–Ti oxides and quartz (q); biotite, always minor, is of secondary origin. The rocks have a fine-grained granoblastic matrix ($400\ \mu\text{m}$) in which garnet porphyroblasts (up to 10 mm) occur. These have irregular and lobate shapes but sometimes show idiomorphic outlines. They contain numerous rounded inclusions of

quartz, and locally vermicular aggregates of quartz and aluminous spinel. These aggregates are never found outside garnet. On the other hand, a very fine-grained garnet-spinel symplectite is common in the matrix. The minerals present seem to obey the following univariant relation (Hensen & Green 1973):



The presence of vermicular spinel-quartz aggregates points to high temperatures resulting in a shift of the equilibrium to the right, while the preservation of this association in the garnet (armoured relic) suggests a subsequent reduction in temperature.

The *Breimyrknutan charnockite* (no. 7) is a

massive, medium-grained rock containing microperthite, plagioclase, quartz, and an inverted pigeonite ($\text{Fs}_{69}\text{En}_{24}\text{Rh}_2\text{Wo}_5$) (Table 1) with wide cpx exsolutions. The *charnockite dyke of Hydra* considered here (no. 8: specimen 283 2/2 of Demaiffe (1977) is a medium-grained biotite quartz mangerite in sharp contact with the surrounding leuconorite. Orthopyroxene is $\text{Fs}_{80}\text{En}_{18}\text{Wo}_2$ (Demaiffe 1977:75) in which the Fe ratio (= $\text{Fe}/(\text{Fe}+\text{Mg})$) approximates the value reached in BKSK (0.81) (Duchesne 1972b). The upper part of the *Bjerkreim-Sokndal lopolith* (BKSK) (Duchesne 1972b; Rietmeijer 1979; Wiebe 1984) comprises (1) mangerites with mesoperthite, clinopyroxene, inverted pigeonite and olivine (no. 9); (2) (quartz) monzonites with

Table 1. Representative opx-cpx pairs in metabasites and Hydra charnockite.

| Specimen No. | 1 | | 2 | | 3 | | 4 | | 7 | |
|--------------------------------|-------|-------|--------|-------|-------|-------|-------|-------|-------|-------|
| | opx | cpx | opx | cpx | opx | cpx | opx | cpx | opx | cpx |
| SiO ₂ | 49.95 | 50.51 | 50.13 | 50.55 | 50.94 | 50.73 | 51.67 | 51.22 | 48.02 | 49.66 |
| TiO ₂ | 0.11 | 0.20 | 0.11 | 0.25 | 0.20 | 0.22 | 0.13 | 0.32 | 0.14 | 0.19 |
| Al ₂ O ₃ | 0.97 | 1.71 | 0.90 | 1.63 | 1.03 | 1.60 | 1.25 | 2.21 | 0.43 | 0.97 |
| FeO _{tot} | 31.92 | 13.33 | 34.90 | 14.88 | 28.40 | 11.14 | 24.99 | 9.21 | 39.83 | 20.16 |
| MnO | 0.59 | 0.24 | 0.57 | 0.30 | 1.10 | 0.42 | 0.61 | 0.33 | 1.08 | 0.50 |
| MgO | 14.90 | 10.71 | 13.23 | 9.98 | 17.31 | 11.99 | 20.26 | 13.24 | 7.88 | 6.74 |
| CaO | 0.69 | 21.65 | 0.92 | 21.91 | 0.69 | 21.30 | 0.67 | 22.31 | 2.25 | 20.65 |
| Na ₂ O | 0.00 | 0.27 | 0.04 | 0.27 | 0.00 | 0.35 | 0.00 | 0.01 | 0.00 | 0.42 |
| K ₂ O | 0.01 | 0.02 | 0.04 | 0.00 | 0.01 | 0.00 | 0.01 | 0.00 | 0.01 | 0.00 |
| Total | 99.14 | 98.64 | 100.84 | 99.77 | 99.68 | 97.75 | 99.59 | 98.85 | 99.64 | 99.29 |
| Si | 1.972 | 1.948 | 1.970 | 1.940 | 1.968 | 1.954 | 1.957 | 1.938 | 1.980 | 1.963 |
| Al ^{IV} | 0.028 | 0.052 | 0.030 | 0.060 | 0.032 | 0.046 | 0.043 | 0.062 | 0.020 | 0.037 |
| Al ^{VI} | 0.017 | 0.025 | 0.012 | 0.014 | 0.015 | 0.027 | 0.013 | 0.037 | 0.001 | 0.008 |
| Ti | 0.003 | 0.006 | 0.003 | 0.007 | 0.006 | 0.006 | 0.004 | 0.009 | 0.004 | 0.006 |
| Fe ³⁺ | 0.006 | 0.037 | 0.017 | 0.052 | 0.006 | 0.032 | 0.023 | 0.008 | 0.012 | 0.051 |
| Fe ²⁺ | 1.048 | 0.393 | 1.130 | 0.425 | 0.911 | 0.327 | 0.769 | 0.284 | 1.362 | 0.616 |
| Mn | 0.020 | 0.008 | 0.019 | 0.010 | 0.036 | 0.014 | 0.020 | 0.011 | 0.038 | 0.017 |
| Mg | 0.877 | 0.616 | 0.775 | 0.571 | 0.997 | 0.689 | 1.144 | 0.747 | 0.484 | 0.397 |
| Ca | 0.029 | 0.894 | 0.039 | 0.901 | 0.029 | 0.879 | 0.027 | 0.905 | 0.099 | 0.874 |
| Na | 0.000 | 0.020 | 0.003 | 0.020 | 0.000 | 0.026 | 0.000 | 0.001 | 0.000 | 0.032 |
| K | 0.001 | 0.001 | 0.002 | 0.000 | 0.000 | 0.000 | 0.000 | 0.000 | 0.001 | 0.000 |
| Total | | | | | | | | | | |
| Mole% | | | | | | | | | | |
| Ca | 1.47 | 46.78 | 1.99 | 47.25 | 1.47 | 46.04 | 1.39 | 46.48 | 4.99 | 45.90 |
| Mg | 44.33 | 32.23 | 39.48 | 29.34 | 50.53 | 36.09 | 58.57 | 38.37 | 24.41 | 20.85 |
| Fe | 53.09 | 20.56 | 57.56 | 22.29 | 46.17 | 17.13 | 39.14 | 14.59 | 68.68 | 32.35 |
| Mn | 1.01 | 0.42 | 0.97 | 0.52 | 1.82 | 0.73 | 1.02 | 0.56 | 1.92 | 0.89 |
| Temperature (°C) calculated by | | | | | | | | | | |
| Wood & Banno | 792 | | 775 | | 820 | | 847 | | 768 | |
| Wells | 831 | | 812 | | 859 | | 875 | | 803 | |
| Lindsley | 620 | 620 | 680 | 600 | 650 | 650 | 650 | 700 | ≥800 | 520 |

Specimen nos. 1-4: metabasites Specimen no. 7: Breimyrknutan charnockite.

orthopyroxene, clinopyroxene, and inverted pigeonite (nos. 10–12); (3) quartz mangerites with inverted pigeonite or olivine and clinopyroxene (nos. 13–17). These rocks contain interesting associations that are used for geothermobarometry: (1) fayalitic olivine + quartz + Ca-rich cpx, in which olivine ranges from Fo₆ to Fo₄ (Table 5); (2) Fe-rich opx either as inverted pigeonite (the most Fe-rich pigeonite is in sample no. 9: Fe ratio = 0.81) or as primary opx (Fe ratio ca. 0.80) (Table 5). A Ca-rich cpx coexists with opx or olivine. Fayalite + quartz have never been found with Fe-rich opx. Pyroxene generally shows exsolution lamellae as well as frequent externally-exsolved granules. Magmatic (solidus) bulk compositions can be reconstituted – or at least approached – through direct measurement with a defocused microprobe beam (Duchesne 1972b), or by recalculation from the composition and proportion of exsolved phases (Rietmeijer 1979).

Geothermometry and geobarometry

A. Metabasites

Twenty-nine pairs of opx and cpx grains in contact from four samples were analysed and used to calculate temperatures according to the Wood & Banno (1973) and Wells (1977) geothermometers. Four representative pairs are shown in Table 1 and plotted in the pyroxene quadrilateral (Fig. 3). Table 2 summarizes the results and their reproducibility.

The temperatures measured for each rock are remarkably homogeneous (standard deviation

Table 2. Temperatures determined from opx–cpx pairs in metabasites.

| Specimen no. | 1 | 2 | 3 | 4 |
|-------------------|-----|-----|-----|-----|
| Number of pairs | 8 | 11 | 3 | 7 |
| Wood & Banno (°C) | | | | |
| mean | 795 | 777 | 820 | 837 |
| S.D. (s) | 11 | 18 | 5 | 9 |
| Wells (°C) | | | | |
| mean | 836 | 814 | 859 | 867 |
| S.D. (s) | 15 | 27 | 7 | 25 |

<3.5%). There is a 30 to 40°C difference between the values given by the two thermometers, those of Wells being systematically higher, as commonly noted (Maquil & Duchesne 1984; Rietmeijer 1984). The mean of all samples yields a value of 804 ± 28°C (2s) for the Wood & Banno geothermometer. This value is identical to the average temperature obtained by Jacques de Dixmude (1978) on metabasites from the envelope. The Lindsley & Andersen (1983) graphic thermometer yields reasonably coherent results for temperatures measured from orthopyroxene and clinopyroxene (Table 1), but the average temperature (650°C) is far lower than those calculated by the other methods. Rietmeijer (1984) has presented convincing arguments that the temperatures obtained with the Wood & Banno thermometer are ca. 150°C too high. This suggests that the Lindsley & Andersen thermometer gives a better approximation of the opx–cpx equilibrium temperature.

B. Kinzigitic gneisses

Two mineral associations were used to estimate the PT conditions: garnet–cordierite and garnet–plagioclase, both in the presence of sillimanite and quartz. The compositions of grains in contact have been determined. Repeated measurements do not reveal either zoning or significant compositional variations from grain to grain. Representative compositions of mineral pairs are given in Table 3.

We have determined the equilibrium temperature between garnet and cordierite using the Thompson (1976) thermometer, which, according to Martignole & Sisi (1981), gives a better temperature estimate than that of Hensen & Green (1973). Gt–cord pairs yield a temperature of about 600°C. This temperature approximates those

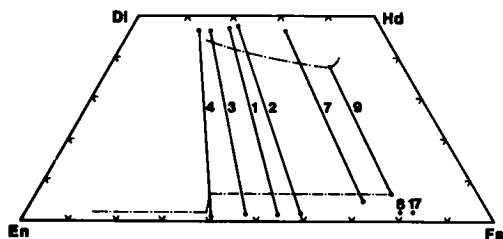


Fig. 3. Composition of representative opx and opx–cpx pairs plotted in the pyroxene quadrilateral: nos. 1 to 4 are from metabasites; no. 7 is from the Breimyrknutan charnockite; no. 8 is from a Hidra charnockitic dyke; no. 9 is from a BSKS mangerite; no. 17 is the most Fe-rich opx in BSKS quartz mangerite. The BSKS trend is also represented.

Table 3. Representative garnet-cordierite and garnet-plagioclase pairs in kinzigitic gneisses.

| Specimen no. | 5(*) | | 6 | | 5 | | 6 | |
|--------------------------------|--------|-------|--------|-------|--------|--------|--------|--------|
| | Gt | Cd | Gt | Cd | Gt | Pl | Gt | Pl |
| SiO ₂ | 38.57 | 48.82 | 38.23 | 48.49 | 38.56 | 57.82 | 39.05 | 59.43 |
| TiO ₂ | 0.06 | 0.03 | 0.09 | | 0.06 | 0.04 | 0.04 | 0.03 |
| Al ₂ O ₃ | 22.06 | 34.69 | 22.68 | 34.66 | 22.07 | 26.90 | 23.59 | 26.01 |
| FeOtot | 33.34 | 5.86 | 33.74 | 6.47 | 31.78 | 0.17 | 33.03 | 0.46 |
| MnO | 1.13 | 0.12 | 0.70 | 0.05 | 1.34 | 0.08 | 0.63 | |
| MgO | 6.09 | 9.98 | 4.76 | 9.35 | 6.20 | | 6.58 | |
| CaO | 0.78 | 0.02 | 0.55 | 0.03 | 0.90 | 7.68 | 1.01 | 6.90 |
| Na ₂ O | | 0.11 | | | | 7.05 | 0.08 | 7.43 |
| K ₂ O | 0.01 | 0.02 | 0.04 | 0.05 | | 0.13 | | 0.31 |
| Total | 102.04 | 99.65 | 100.79 | 99.10 | 100.91 | 99.87 | 104.01 | 100.57 |
| Si | 2.987 | 4.911 | 3.015 | 4.911 | 3.009 | 10.352 | 2.951 | 10.564 |
| Al ^{IV} | 0.013 | 1.089 | | 1.089 | | 5.685 | 0.049 | 5.445 |
| Al ^{VI} | 1.998 | 3.018 | 2.121 | 3.049 | 2.029 | | 2.050 | |
| Ti | 0.003 | | 0.005 | | 0.004 | 0.011 | 0.002 | 0.004 |
| Fe ³⁺ | | | | | | | | |
| Fe ²⁺ | 2.152 | 0.495 | 2.218 | 0.548 | 2.073 | 0.022 | 2.080 | 0.008 |
| Mn | 0.074 | 0.012 | 0.047 | 0.006 | 0.089 | 0.011 | 0.040 | |
| Mg | 0.708 | 1.498 | 0.563 | 1.412 | 0.721 | | 0.746 | |
| Ca | 0.065 | | 0.046 | 0.006 | 0.075 | 1.475 | 0.082 | 1.313 |
| Na | | 0.024 | | | | 2.455 | | 2.560 |
| K | | | | 0.006 | | 0.022 | | |
| Pyr | 24.2 | | 19.9 | | 25.1 | | 25.7 | |
| Alm | 73.6 | | 78.4 | | 72.3 | | 71.5 | |
| Gro | 2.2 | | 1.6 | | 2.6 | | 2.8 | |
| An | | | | | | 37.3 | | 33.3 |
| Ab | | | | | | 62.1 | | 64.9 |
| Or | | | | | | 0.6 | | 1.8 |

* Al-spinel in this rock has also been analysed. The structural formula is: (Fe_{5.7}Mg_{2.3}) (Al_{15.9}Si_{0.1})O₃₂ (hercynite).

obtained with the Lindsley & Andersen opx-cpx thermometer in neighbouring metabasites. Though the blocking temperatures cannot be expected to be similar in the various retrograde equilibria, the difference is small, which increases the consistency of the model. A 3.5 kb pressure corresponding to this 600°C temperature is inferred from the P-T-X_{Fe} grid for garnet and cordierite presented by Martignole & Sisi (1981), assuming that cordierite is anhydrous, as indicated by a sum of oxides near 100% and the occurrence of optically positive crystals (Martignole & Nantel 1982; Armbruster & Bloss 1982). Aranovitch & Podlesskii (1983) geothermobarometer also gives 600 ± 10°C and a pressure between 2.7 ± 0.5 kb (assuming the fluid phase is pure CO₂) and 4.4 ± 0.4 kb (assuming the fluid is pure H₂O). The lower pressure estimate is preferred because of the anhydrous character of cordierite.

Different geobarometers based on the garnet-

plagioclase equilibrium in the presence of sillimanite and quartz have also been applied. Since the petrographic observations do not allow us to reject the hypothesis that the gt-cord and gt-pl equilibria were arrested at different conditions, we have calculated the pressure at a 600°C temperature, estimated by the gt-cord equilibrium. The pressures calculated according to the models of Wood & Fraser (1976) and Newton & Haselton (1981) are 3.5 kb and 2.7 kb respectively. Ganguly & Saxena (1984) have suggested adding +0.6 kb to the Newton & Haselton value, which then yields a value of 3.3 kb. In conclusion it is suggested that the kinzigitic gneisses reflect equilibrium pressures of 3 ± 0.5 kb at about 600°C.

Armoured relics of spinel + quartz in garnet however indicate that the temperature was reached as a retrograde effect. Assuming an anhydrous cordierite, the position of the gt-cord-sill-q-sp univariant reaction curve, suggested by Hensen & Green (1973), can be constrained as

follows: (1) the curve passes through the invariant point (in the system FeO–Al₂O₃–SiO₂) where the association gt–cord–sill–sp–q is stable (Fig. 4) (Holdaway & Lee 1977; Bohlen et al. 1986); (2) a chemographic approach (Vielzeuf 1983) shows that the univariant reaction curve must have a steeper slope than the $gt + sill \rightleftharpoons sp + q$ reaction curve determined by Bohlen et al. (1986). These considerations point to peak temperatures in the envelope, in excess of 800°C to 900°C.

C. Charnockitic rocks

Charnockitic rocks from the upper part of the Bjerkreim-Sokndal lopolith (BKSK) permit an accurate estimation of temperatures and pressures prevailing during crystallization. The graphical thermometer of Lindsley & Andersen (1983) applied to the Ca-poor and Ca-rich clinopyroxenes analysed by Duchesne (1972b) leads to values between 800°C and 830°C (specimen nos. 9 and 16, Table 4), close to the interval of 750–800°C estimated by Duchesne (1972a) by means of the Fe–Ti oxide minerals. It must be

emphasized that the pyroxene compositions were determined either via chemical analysis of a separated mineral fraction (specimen no. 16) or via microprobe analysis on grains with fine-scale exsolutions, using a defocused beam (specimen no. 9). The compositions thus considered are, or closely approach, solidus compositions. The use of the same thermometer on opx from specimen nos. 10 and 12 yields lower temperatures (600°C). This can be explained by exsolution processes, difficult to evaluate under the microscope, and also to a lesser precision of the thermometer, due to high sensitivity to small variations in composition.

Microprobe analyses from Rietmeijer (1979) using the Lindsley & Andersen (1983) thermometer give a wide temperature range from ca. 600°C to ca. 880°C (Table 4). However, the high temperature range is obtained on recalculated pigeonites with high wollastonite content, which are less likely to have been re-equilibrated due to granule exsolution processes. On the other hand, as clearly shown by Rietmeijer (1984), the low temperature range is found in subsolvus re-

Table 4. Application of Lindsley & Andersen geothermometer to rocks of the upper part of the Bjerkreim-Sokndal lopolith.

| Rock type | Specimen no. (this study) | Thermometer used | T(°C) | Note |
|------------------|---------------------------|------------------|-------|---|
| Mangerite | 9 | Ca-poor cpx | 830 | rock no. TII in Duchesne (1972b) |
| | | Ca-rich cpx | 800 | Table 2 (anal.8) idem (anal.8a) |
| Monzonite | 10 | Ca-poor cpx | 880 | rock no. R40 in Rietmeijer (1979) Tables 4.1 and 4.5 (anal. 40.1,2) |
| | | Ca-rich cpx | 680 | the highest Wo cont. of all recalculated pigeonites. idem (anal.40.13): possibly external granule exsolution |
| | 12 | primary opx | 600 | idem (anal. 40.3) |
| | | opx | 600 | rock no. R217: idem (anal. 217.1) |
| Quartz mangerite | 13 | Ca-rich cpx | 700 | idem: recal. pigeon. (anal. 217.2,4) |
| | | Ca-rich cpx | 700 | rock no. R98: possibly external granule exsolution (anal. 98.10) |
| | 15 | id | 700 | rock no. R392: idem (anal. 392.3) |
| | 16 | id | 830 | rock no. 66.261 (anal. 9a) in Duchesne (1972b) |

Table 5. Application of Bohlen & Boettcher geobarometer to quartz mangerites of the Bjerkreim-Sokndal and Hidra massifs.

| Mineral association | Specimen no. (this study) | Mineral composition | Note |
|----------------------------------|---------------------------|---|---|
| Fayalitic olivine (+ quartz) | 13 | Fa _{96.4} Fe _{0.6} | rock no. R98 in Rietmeijer (1979) Table 4.7 |
| | 14 | Fa _{95.6} Fe _{0.4} | rock no. R95 idem |
| | 15 | Fa _{95.8} Fe _{0.2} | rock no. R392 idem |
| | 16 | Fa ₉₄ Fe _{0.6} | rock no. 66261 in Duchesne (1972b) Table 2 |
| Primary orthopyroxene (+ quartz) | 8 | Fs ₈₀ En ₁₈ Wo ₂ | rock no. 283 2/2 in Demaiffe (1977) p. 75 |
| | 11 | Fs ₈₀ En ₁₇ Wo ₃ | rock no. R247 in Rietmeijer (1979) Tables 4.1 and 4.7 |
| | 17 | Fs ₈₀ En ₁₈ Wo ₂ | rock no. 314 in Wiebe (1984) Table 2 |
| | 12 | Fs ₈₃ En ₁₅ Wo ₂ | rock no. R217 in Rietmeijer (1979) |

equilibrated pyroxenes. Therefore, *solidus* temperatures in the upper part of BKSK can be estimated in the range 800–850°C.

Bohlen & Boettcher (1981) have experimentally determined the PT conditions of the reaction $\text{opx} \rightleftharpoons \text{olivine} + \text{quartz}$. The equilibrium is strongly influenced by pressure, the value of which increases with the Fe content of the opx. Given the range of crystallization temperatures in the upper part of BKSK, application of this barometer permits the calculation of upper and lower limits for the pressure of crystallization (Table 5). Fa₉₄Fe_{0.6} + quartz (in specimen no. 16) would react to opx + olivine + quartz at a pressure of 7 kb at 800°C, and 7.7 kb at 850°C. The maximum Fe enrichment found in primary opx (Fs₈₃En₁₅Wo₂) (specimen no. 12) implies a minimum pressure of 6 kb at 800°C and 6.5 kb at 850°C. In the Breimyrknutan body, 17 pairs of opx–cpx have been measured between cpx exsolution lamellae and host opx (see Table 1, specimen no. 7): averages of 800 ± 24°C (Wells) and 550°C (Lindsley & Andersen) give *solvus* temperatures. Overall similarity in whole-rock chemical composition of the Hidra charnockitic dykes with the late stage olivine quartz mangerite of BKSK points to the same range of crystallization temperatures, i.e. 800–850°C. The stability of the Fe-rich opx (Fe₈₀En₁₈Wo₂) (specimen no. 8) would thus indicate a minimum pressure range 1 kb lower than that of the BKSK crystallization.

Discussion

Two distinct stages of PT conditions have thus been determined here and are shown schematically in Fig. 4. Stage A corresponds to the crystallization of the upper part of BKSK and

probably also of Hidra (800–850°C, 6.0–7.5 kb), while stage B represents the re-equilibration of the metapelites and possibly of the metabasites (600°C, 2.5–3.5 kb), attained by retrogression of an assemblage stable at higher temperatures and pressures. It is therefore suggested that the two sets of PT conditions correspond to two successive stages on a PT path. The upper part of BKSK and the Hidra massif clearly show characteristics of post-tectonic intrusions. They obviously post-date the regional metamorphism M1 phase of Jansen et al. (1985) and the subsequent emplacement of the Åna-Sira massif. Accordingly, the emplacement of all anorthositic massifs as high-temperature crystal mushes (Duchesne et al. 1985), immediately followed by the intrusion of the BKSK parental magma and the other monzonitic intrusions (Apophysis, Hidra), provided extra heat and produced peak temperatures and a large 'contact aureole' in the metamorphic envelope. Cooling and differentiation of BKSK ended with products of acidic composition, the crystallization of which has been recorded in stage A conditions. Further cooling and retrogression of metapelites led to stage B. Therefore, the PT path linking stages A to B characterizes the *post-magmatic evolution*.

These views on the PT evolution are quite consistent with the independent approach used by Touret and coworkers with fluid inclusions (Figs. 2 and 4). Intrusion of igneous rocks and subsequent crystallization (our stage A) took place in PT conditions compatible with high density CO₂ fluids in the metamorphic envelope. The anorthosites and related magmas can conveniently be considered as the source of these fluids, as already suggested by Touret (1974). The progressive evolution towards stage B conditions of re-equilibration at lower PT values also appears

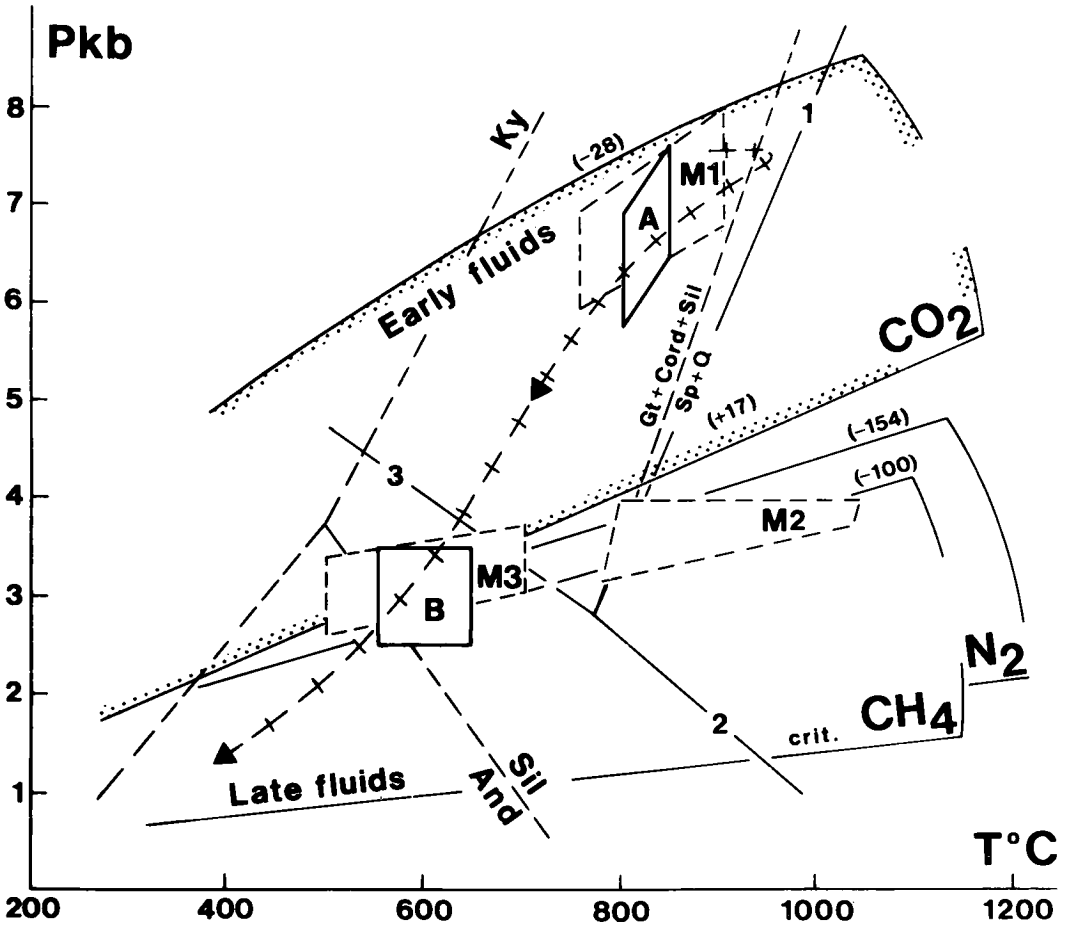


Fig. 4. PT diagram showing evolution stages A and B reconstructed in this paper and M1 to M3 metamorphisms of Jansen et al. (1985). Post-magmatic PT trajectory indicated by a broken ticked line. Range of isochores (labelled with homogenization temperature) for pure CO₂ (stippled), N₂ and CH₄ fluid inclusions (after Touret & Van den Kerkhof, pers. comm.). Al₂SiO₅ stability relations after Holdaway (1971). Reaction curves after Bohlen et al. (1986): (1) $gt + sil \rightleftharpoons sp + q$; (2) $Fe-cord \rightleftharpoons sp + q$; (3) $Fe-cord \rightleftharpoons gt + sil + q$.

consistent with the change in fluid compositions towards CH₄- and N₂-rich fluids.

Our stage B conditions are also consistent with the M3 retrograde metamorphism of Jansen et al. (1985); the relatively large PT interval characterizing retrogression can be ascribed partly to the fact that re-equilibration of the various mineral pairs took place at different PT conditions, depending on the minerals and the controlling processes. The lack of deformation after the emplacement of the igneous bodies did not assist the attainment of uniform PT conditions of re-equilibration on a regional scale. Our results however significantly differ from the Jansen et al. (1985) model in the pressure conditions attributed to the emplacement of BKSK and to the associ-

ated M2 metamorphism. Instead of the 4 kb pressure claimed by these authors, we support much higher values (6–7.5 kb) in the range of those of the M1 regional metamorphism.

The M2 pressure estimates are mainly based on the occurrence of osumilite in the contact aureole of the igneous masses, north of BKSK (Maijer et al. 1981). Grew (1982) predicted the stability of osumilite at temperatures above 750°C and total pressure less than 8–9 kb. He showed on a qualitative Schreinemaker's petrogenetic grid that the Rogaland osumilite is stable at somewhat higher pressures than the osumilite occurring in the contact aureole of the Nain anorthosite, emplaced at about 3.5 kb pressure (Bohlen et al. 1983). Sillimanite inclusions in osumilite, mentioned by

Tobi et al. (1985), tend to indicate minimum pressures of 5 kb at 900–1000°C. Our ignorance of the influence of volatiles (H₂O, F, Cl, etc.) on the stability of osumilite is a further handicap for an accurate determination of PT conditions. It must therefore be concluded that the pressure conditions at the peak temperature in the envelope, indicated by the present PT path, are not incompatible with the occurrence of osumilite in the contact aureole of BSKS.

Conclusions

Temperatures measured in metabasites near the igneous bodies vary considerably according to the geothermometer used. Calculated values of around 840°C (Wells), 800°C (Wood & Banno), and 650°C (Lindsley & Andersen) corroborate similar differences observed in other metamorphic terranes. The temperature obtained by the Lindsley & Andersen thermometer approaches those measured in neighbouring metapelites. In the latter rocks, various thermometers and barometers point to 600°C and 2.5–3.5 kb for equilibrium conditions, which were attained during decreasing temperature, as evidenced by armoured relics of spinel + quartz in garnet.

PT conditions for the crystallization of the upper part of the BSKS massif are 800–850°C and ca. 6–7.5 kb according to the Lindsley & Andersen thermometer and the Bohlen & Boettcher barometer. Similarly, the minimum pressure conditions for the end of the Hidra crystallization are about 5–5.5 kb. These igneous bodies are clearly atectonic and were emplaced later in the evolution of the province than the massif-type anorthosites.

The two stages of PT conditions thus reconstructed correspond to two successive stages along a PT path characterizing the post-magmatic evolution. The first stage is obtained by temperatures decreasing from a peak due to the intrusion of the igneous masses. The second stage is attained by retrograde metamorphism. The PT path is in good agreement with fluid inclusion data. It is concluded that the depth of emplacement of the Åna-Sira massif-type anorthosite is not less than 22–28 km (6–7.7 kb) along its contact with the neighbouring BSKS massif, and not less than 18–20 km (>5–5.5 kb) near the contact with the Hidra body. It follows that the Rogaland massif-

type anorthosites can still be considered deep-seated intrusions.

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Appendix: Analytical method and calculations

The analyses were performed on a 4 spectrometer Camebax microprobe. Matrix effects were corrected by means of Cameca's ZAF program according to the Philibert models and with the Heinrich coefficients. This method of correction gives good results for most of the minerals analysed, but not for garnet, the analysis of which generally sums to more than 100%. This problem, met with by various users of the Cameca probe, has not been entirely solved (Laduron & Wautier, pers. comm.). However, by modifying the value of some coefficients, it has been possible to reduce the difference to 100% for most of the garnet analyses. This factor is of minor importance in the present study, however, since the proportion of the elements in the calculations remains virtually unchanged by these modifications. In the 58 analysed pyroxenes, the summation is better (99.75 with $s = 0.75$). It has also been verified that analyses made with a defocused beam are not significantly different from those made in the usual way.

The structural formula of the pyroxenes is calculated on the basis of 4 cations and 12 charges and the trivalent iron is recalculated from the charge deficit. The structural formulae of garnet, cordierite and plagioclase have been calculated on the basis of 12, 18 and 32 oxygens, respectively.

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