

$^{18}\text{O}/^{16}\text{O}$ Ratios of Anorthosites and Related Rocks From the Rogaland Complex (SW Norway)

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Abstract. The South Rogaland Complex (South Western Norway) consists of several anorthositic intrusions emplaced in granulite facies metamorphic rocks. The anorthosites and related norites and jotunitites have $\delta^{18}\text{O}$ values of 5.2 to 7‰ suggesting a mantle origin for these rocks, in agreement with the strontium isotopic evidence. The acidic rocks, mostly charnockitic, associated with the anorthosites have similar $\delta^{18}\text{O}$ values and thus a comagmatic relation between these two rock types is inferred. Small departures from mantle values are explained in terms of crustal contamination by surrounding gneisses that have $\delta^{18}\text{O}$ values between 4.3 and 10‰. Locally, this corresponds to important anatexis as has been suggested for the Farsund charnockite on the basis of strontium isotope and REE geochemistry. The isotopic temperatures calculated from the isotopic fractionations are in the range 500°–700° C, lower than the orthomagmatic temperatures and probably due to subsolidus isotopic exchange during the slow cooling of these plutonic rocks, either during a late magmatic deuteric stage or during a slow, postorogenic ascent under wet conditions.

I. Introduction

The genesis of anorthosites has interested many petrologists, i.e. Bowen, Barth, Buddington, ... but important questions still remain unanswered: (e.g., the nature and the source region of the parental magmas) and the relation between these rocks and the spatially related acidic rocks, mostly of charnockitic type¹. Geochemical and isotopic data on anorthosites are unfortunately rather scarce with the exception of the

¹ Streckeisen's (1974) nomenclature of the charnockitic rocks has been adopted in this paper

isotopic work of Taylor (1969). However, since the discovery of anorthosites on the Moon (i.e., Wood et al. 1971; Taylor 1975), interest in terrestrial anorthosites has been renewed. This paper is focused on the oxygen isotope geochemistry of the South Rogaland complex of Norway (Michot and Michot 1969; De Waard et al. 1974), a typical example of massif-type anorthosite (Anderson and Morin 1969). This paper complements the previous ones dealing with the trace element geochemistry, especially the REE, and the Sr isotopic composition of this complex (Duchesne et al. 1974; Demaiffe 1977; Duchesne and Demaiffe 1978; Demaiffe et al. 1979).

II. Geological Setting

The Precambrian of the Southern tip of Norway (Fig. 1) is commonly subdivided into two major lithologic units (Michot 1960):

1. A supracrustal gneissic series of granulite facies metamorphism. *P–T* conditions of these rocks have been calculated from the Fe–Mg distribution between coexisting garnet and cordierite at equilibrium or between ortho- and clinopyroxene. 20 determinations lead to a mean of 750° C, 6.3 kbar total pressure (Henry 1974; Demaiffe 1977; Jacques de Dixmude 1978).

To the North and the East, a transition towards amphibolite facies is observed.

2. The South Rogaland anorthositic complex (Michot and Michot 1969; de Waard et al. 1974) which consists of a sequence of massif-type andesine anorthosite plutons. The following intrusive sequence is deduced mainly from field relationships:

- the large Egersund-Ogna anorthositic body;
- the thin Lakssevelefjellet noritic sheet;
- the layered anorthosito-noritic lopolith of Bjerkrem-Sogndal giving rise to charnockitic differentiates (the so-called mangerites and quartz mangerites) at the end of the differentiation process (Michot 1965; Duchesne 1972, 1979).

An anatexis process (Michot 1961) giving rise to the Haaland-Helleren and Aana-Sira massifs takes place before the final stage of the Bjerkrem-Sogndal differentiation.

To the East, the main complex is bordered by Outliers emplaced within the granulite facies metamorphic rocks:

- the Hidra and Garsaknatt leuconoritic bodies with their jotunitic border facies and associated charnockitic dykes (Demaiffe 1977);

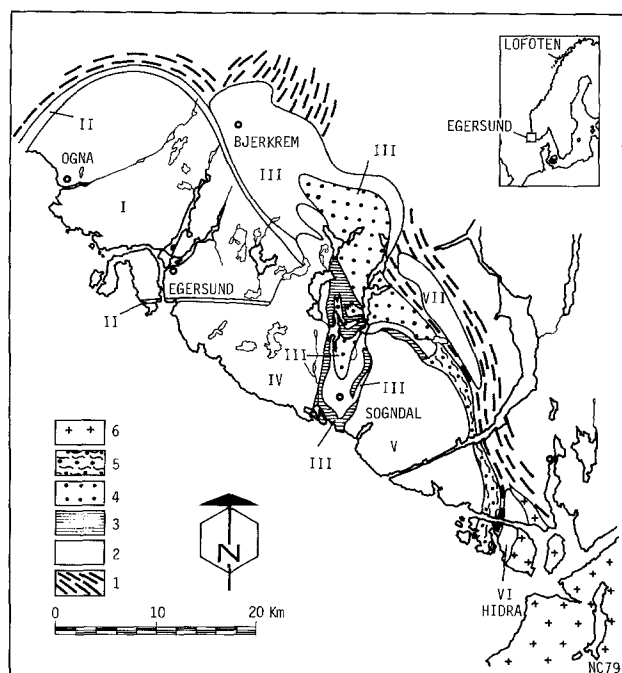


Fig. 1. Schematic geological map of the Rogaland igneous complex (after Michot and Michot, 1969; slightly modified). 1: gneisses of the envelope; 2: anorthosite and leuconorite; 3: jotunite; 4: mangerite and quartz mangerite; 5: noritomangeritic complex; 6: Farsund charnockite. I: Egersund-Ogna massif; II: Laksvelefjellet sheet; III: Bjerkrem-Sogndal lopolith; IV: Haaland-Helleren massif; V: Aana-Sira massif; VI: Hydra massif; VII: Garsaknatt massif

– the Farsund charnockite and the Lyngdal hornblende granodiorite (Falkum et al. 1972).

Pasteels et al. (1979) have shown, by U-Pb age determinations on uraniferous accessory minerals (zircon, sphene, monazite, uraninite) of the acidic rocks, that the magmatic activity of the South-Eastern part of the Rogaland Complex took place in a short time span, between 955 m.y. and 910 m.y. Although the intrusion age of the anorthosite bodies cannot be determined because of their inadequate chemical and mineralogical composition (absence of zircons, very low and constant Rb/Sr ratios), field evidence suggests that the whole Rogaland Complex was emplaced during a single orogenic cycle, the Sveconorwegian cycle, so that the first intrusion is probably not older than 1,200 m.y.

Results and Discussion

(a) Whole Rock Values and Source Materials

The $\delta^{18}\text{O}$ values of the anorthosites and related norites and jotunites (Table 1A) fall in the range 5.2‰ to 6.6‰ (except for the quartz jotunite 66125 with a $\delta^{18}\text{O}=7.1$ ‰). This is very close to the 'magmatic anorthosite' range defined by Taylor (1969) for unmetamorphosed anorthosites. In particular the Bjerkrem-Sogndal lopolith rocks are very similar to the Haaland-Helleren and Aana-Sira rocks analyzed by Taylor while the Hydra body has somewhat lower

values on the average. These values cannot be directly compared to those reported for mantle materials such as M.O.R. basalts or peridotites because the anorthosites are cumulate plagioclases of intermediate (An 45–50) composition. In order to make a valid comparison with fresh M.O.R.B. tholeiites, which correspond to the largest number of analyses of magmatic rocks of mantle origin (Muehlenbachs and Clayton 1972; Pineau et al. 1976; Taylor 1967) we have calculated the $\delta^{18}\text{O}$ of a model tholeiite (60% plagioclase An 85+30% Cpx+10% Ol) in equilibrium with the Rogaland rocks at 1,000°C using the equations of Bottinga and Javoy (1975). The results of these calculations are given in the second column of Table 1A and in Fig. 2. The mean fractionation between the anorthosites and the model tholeiites is +0.5‰ but some particular modal compositions or results on separated minerals finally give about the same total range as the original results (5.2‰ to 6.6‰). About 75% of the results range between 5.2‰ and 6.0‰, which compare very well with the results for fresh M.O.R.B. tholeiites or high temperature peridotites (5.5‰ to 6.0‰; Muehlenbachs and Clayton 1972; Pineau et al. 1976; Javoy 1979). Most of the results below 5.5‰ belong to the Hydra body, which hence appears a little depleted on the average relative to mantle values. Thus the oxygen isotope data clearly suggest a mantle origin for Rogaland anorthosites, norites and jotunites with minor modifications for a few of them. The strontium isotope results are in agreement with this statement. The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio calculated from a mean age of 1 b.y. varies in the range 0.7035–0.7055 (Demaiffe 1977; Duchesne and Demaiffe 1978) which is close to but larger and higher than the range of values of 0.7022–0.7035 found for fresh M.O.R.B. tholeiites (Hofmann and Hart 1978). This might suggest a contamination by a more radiogenic source, although mantle-derived basalts from oceanic islands have higher ratios, in the range 0.7030–0.7065 (Hofmann and Hart 1978). This possible contamination could have been selective but not necessarily very small since the Sr concentrations of the parental magmas of the anorthosites are rather large (380–450 ppm). We shall return to this subject in the discussion of the acidic rocks. Finally, the REE geochemistry by the same authors shows that the jotunites (charnockitic diorites) are the best representatives of the parental magma for the anorthosite bodies (absence of europium anomaly). The $\delta^{18}\text{O}$ values of these rocks are +6.4‰ for the Bjerkrem Sogndal lopolith and 5.2‰ for the Hydra body. Given the chemical composition if these rocks the $\delta^{18}\text{O}$ value is 'normal' for Bjerkrem Sogndal ($\delta^{18}\text{O}$ normalized to model tholeiite ≈ 6.2 ‰) and low for Hydra ($\delta^{18}\text{O}$ normalized to model tholeiite ≈ 5.0 ‰).

Table 1. Oxygen and strontium isotope analyses of whole rocks and minerals from the Rogaland Complex. Taylor's values (1969) are indicated with an asterisk (*)

	$\delta^{18}\text{O}$ (‰)	$\delta^{18}\text{O}$ normalized to mean tholeiitic composition (1)	$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ (2)		$\delta^{18}\text{O}$ (‰)	$\delta^{18}\text{O}$ normalized to mean tholeiitic composition (1)	$\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ (2)
<i>A. Anorthosites and Related Rocks</i>				<i>B. Acidic rocks associated with anorthosites</i>			
Egersund-Ogna body				Hidra charnockitic dykes			
64157 Plagioclase	6.0	5.5	0.7036	443-4/1	6.8 ± 0.3		0.7086
6646 Olivine leuconorite	6.4			283-2/2	7.4		
Plagioclase*	6.9	6.4		Bjerkrem-Sogndal			
66-119/2 Orthopyroxene				Mangerites and quartz			
Phenocryst	6.3	6.0	0.7044	Mangerites			
Bjerkrem-Sogndal lopolith				66261 Mangerite			
66-190 Anorthosite	6.6	6.2		Whole rock	4.0	3.86	
64-190 Plagioclase	6.1	5.6		Mesoperthite	3.9		
46-74 Olivine leuconorite	6.0	6.0	0.7057	Clinopyroxene	2.0		
66/0 Jotunite	6.4	5.7		Magnetite	-3.3		
66125 Jotunite	7.1			66209 Mangerite			
Haaland-Helleren body				Whole rock	6.3		
Plagioclase*	6.2	5.7		Orthopyroxene	4.6	6.13	
Plagioclase megacryst*	6.3	5.8		Magnetite	0.3		
Matrix plagioclase*	6.7	6.2	0.7059	6687 Quartz mangerite	6.1		0.7105
Plagioclase*	7.1	6.6		73776 Quartz mangerite	6.5		
Aana-Sira body				Pa66P Quartz mangerite	6.4		
Plagioclase	5.8	5.3		TII mesoperthite	6.2		
from anorthosite*				mesoperthite*	6.0		
Plagioclase	6.5	6.0	0.7056	Lyngdal granodiorite			
from noritic pegmatite*				137-1/1	4.4		0.7054
Orthopyroxene	5.7	6.0		134-1/1	7.0		
from noritic pegmatite*				Farsund charnockite			
Hidra body				Pa 70A	6.4		
259-1/1 Jotunite				0055-1/4	5.4		0.7128
Whole rock	5.2			<i>C. Country rock gneisses</i>			
Plagioclase	5.7			295-2/1 Noritic gneiss	4.3		
Ilménite	1.0 ± 0.2			SO-3/70 Noritic gneiss	5.9		
Orthopyroxene	4.3 ± 0.2	4.6		389-1/1 Noritic gneiss	5.0		
249-1/1 Leuconorite-norite				75-62 Mangeritic gneiss (3)	5.5		
Whole rock	5.2	5.2		73-14-11 Granitic gneiss (3)	5.8		
Plagioclase	5.7	5.2	0.7055	Pa 66L Charnockitic gneiss	8.2		
Ilménite	1.8 ± 0.2			Pa 70b Charnockitic gneiss	7.1		
Orthopyroxene	5.0 ± 0.1	5.2		0068-11 Granitic gneiss	9.8		
0058-1.1 Anorthosite							
Whole rock	5.7	5.1					
Plagioclase	5.9	5.4					
K-feldspar	6.4	5.6					
Magnetite	1.0						
0306-1.1 Leuconorite	6.4	6.4					
9200/73 Olivine leuconorite	6.0	6.0					

(1) See text for explanation

(2) from Duchesne and Demaiffe (1978) and Demaiffe et al. (1979)

(3) Xenoliths in the Bjerkrem-Sogndal mangerite

This again suggests some contamination by a low ^{18}O material, which appears also in some of the acidic differentiates.

These acidic, mainly charnockitic rocks, spatially associated with the anorthosites have $\delta^{18}\text{O}$ values close to those of anorthosites but the range is somewhat larger (4.0‰ to 7.5‰, Table 1 B) with 75% between 6.1‰ and 7.5‰ and three low values (4.0, 4.4 and 5.4‰). It is rather difficult to relate these

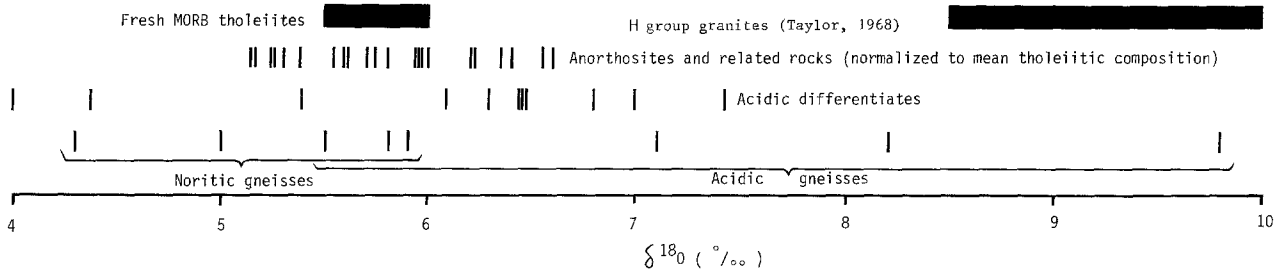


Fig. 2. $\delta^{18}\text{O}$ values (in ‰) of the Rogaland rocks: anorthosites and related rocks, acidic differentiates and gneisses. The ^{18}O compositions of the anorthosites have been recalculated for an average tholeiite in equilibrium with these rocks (see text) for a more accurate comparison with MORB (the correction amounts to 0.2–0.6‰)

values to a model mantle material since the temperatures of formation are lower than 1,000°C and probably rather variable. However, the average difference between them and anorthosites and leuconorites, 0.9‰ can probably largely be attributed to differences in chemical composition. But the low $\delta^{18}\text{O}$ values definitely call for another explanation. We do not have enough $\delta^{18}\text{O}$ values to compare accurately the two groups defined by Demaiffe et al. (1979) on the basis of REE contents and Sr isotopic composition. However, some features appear similar: group I, defined by a high REE content ($\Sigma\text{REE} \approx 400$ ppm) and a well defined negative europium anomaly, includes the Lyngdal granodiorite, the Hidra charnockitic dykes and the Bjerkrem Sogndal quartz mangerites. It is supposed to correspond to the final differentiation products of the anorthosite bodies, whereas group II, with a flat REE pattern, represented by the Farsund charnockite is considered as an anatectic crustal melt. $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios are high in group II (0.7128, Pedersen and Falkum 1975) lower and variable (0.7054 to 0.7108) for group I. $\delta^{18}\text{O}$ values are also very variable for group I corresponding to a total range of 4.0‰ to 7.5‰. For the Farsund charnockite, we have only two values, rather comparable and close to mantle values.

Hence we find in the magmatic rocks of Rogaland an overall feature of ^{18}O homogeneity around mantle values with small departures towards high (6 to 6.6‰) and low (4 to 5.5‰) $\delta^{18}\text{O}$ (recalculated for a tholeiitic composition). The positive departures are so small that they may be explained almost entirely in terms of magmatic phenomena: the fractionations between high silica minerals and magma may be a little higher than postulated, the temperatures a little lower. Finally, about half of the values corresponding to $\delta^{18}\text{O} > 6$ come from Taylor's work of 1969 and the Caltech $\delta^{18}\text{O}$ scale for this period is considered to be around 0.2‰ higher than the one we use in this paper ($\delta^{18}\text{O}$ of NBS 28 quartz = 9.6‰).

On the other hand, the negative departures probably require another explanation: the fractionation be-

tween plagioclase An 45 and a jotunitic magma should be close to +0.3‰ at 1,000°C, a depletion from 5.5‰ to 4.0‰ would require essentially total crystallization as plagioclase, which is impossible.

Hence the cause of the ^{18}O depletion has to be looked for in the environment of the massif. We have analyzed light catazonal gneisses whose $\delta^{18}\text{O}$ are reported in Table 1 C. The sampling is not quantitatively representative as noritic gneisses are only subordinate. The five acidic rocks analyzed have $\delta^{18}\text{O}$ values ranging from 5.5‰ to 9.8‰, whereas the three basic rocks (noritic) range from 4.3‰ to 5.9‰. These basic gneisses are considered to result from the metamorphism of basic intrusions. In this respect, the values of 5.0 and especially 4.3‰ are definitely too low and these rocks could be candidates for the explanation of low ^{18}O magmatic rocks. However, they are not very low, in fact not lower than the low $\delta^{18}\text{O}$ magmatic rocks themselves. Hence to produce the observed $\delta^{18}\text{O}$ they would have to melt essentially completely. We surely need a more extensive sampling of the catazonal rocks to support this explanation but it appears that part of the magmas could be the result of remelting of basic to 'semi-basic' parts of the lower crust.

The acidic rocks display a large variation in ^{18}O content with the highest $\delta^{18}\text{O}$ values comparable to normal granites and also to $\delta^{18}\text{O}$ of granulitic rocks of the Bamble area (Pineau and Javoy, to be published) whereas the lowest value of 5.8‰, which also corresponds to a granitic composition is in the mantle range. Small amounts of contamination by these rocks could explain the slight increase in $\delta^{18}\text{O}$ for certain parts of the anorthositic bodies.

Hence we don't have in this region a large scale homogeneity of the crust such as that found in the 3.5 b.y. catazone of the In Ouzal (Fourcade and Javoy 1973), in Australia (Wilson et al. 1970), or in the Grenville province of Ontario (Shieh and Schwarcz 1974). The lowest value for granitic gneiss could be explained by some amount of exchange with the mantle-derived material, but since some of the

magmatic rocks seem modified in ^{18}O by the country rocks, this means that there is no real 'reservoir effect' from either side. The most reasonable description of the Rogaland genesis is that essentially non contaminated mantle material of jotunitic composition was emplaced in a lower crust of rather variable ^{18}O composition. Some anatectic events lead to the production of minor amounts of magmas variable in both chemistry and isotopic composition. The explanation of the low ^{18}O rocks is probably to be found in high temperature water-rock interaction at low depth during intrusion of mafic material in the source material of the granitic gneiss series. Some of this ^{18}O -depleted material was eventually incorporated as part of the roof material during the emplacement of the jotunitic magma. It should be interesting to measure Rb and Sr concentrations and the isotopic composition of strontium in the surrounding catazonal rocks to delineate the possible relationships. Contamination is frequently discarded on the basis that the average sialic crust is very much enriched in trace elements such as U or Th and it is shown not to be necessarily the case for deep crustal environment. If the ^{18}O depletions are primary (i.e., originated in the mantle or during the magmatic event without any contamination) this calls for isotopic effects which have already been imagined but for very different occurrences (Garlick et al. 1971).

(b) Isotopic Temperatures

Mineral fractionations look rather consistent and can be used for the calculation of isotopic temperatures according to the thermometric equations developed by Bottinga and Javoy (1975) and Javoy (1977). The best average temperature is obtained by the 'isotherm' method described in Javoy (1977) and the quality of the concordancy is measured by the linear correlation coefficient, ρ , in a $\Delta_{A-X} - a_{A-X}$, b_{A-X} diagram where A is a reference mineral, $\Delta_{A-X} = \delta_A - \delta_X$ and the thermometric equation is $\Delta_{A-X} = a_{A-X} + b_{A-X} (10^6/T^2)$. The results are given in Table 2. We can see that the isotopic temperatures are low, from 620° to 710° C for the anorthosites and leuconorites, and 520° to 560° C for the mangerites. This is much less than the temperatures estimated by the Kudo-Weill plagioclase thermometer for the porphyritic jotunites (1,050–1,100° C; Demaiffe 1977) and by the magnetite-ilmenite thermometer for the quartz mangerites (750°–800° C; Duchesne 1972). It probably indicates subsolidus isotope exchange during slow cooling. Deuteric readjustment through a fluid phase is apparent at the contact of ilmenite and magnetite grains (Duchesne 1972) and this fluid could be responsible

Table 2. Oxygen isotopic fractionations between coexisting minerals and the corresponding isotopic temperatures calculated using the equations of Bottinga and Javoy (1973 and 1975)

	Minerals	$\Delta^0/1000^a$ now calculated	T (°C)	Isotherm temperatures	
				T (°C)	ρ (2)
259-1/1	Plag-Ilm	4.78	637	620 ± 70	0.998
	Plag-Opx	1.44	725		
	Opx-Ilm	3.35	537		
249-1/1	Plag-Ilm	3.87	720	710 ± 200	0.967
	Plag-Opx	0.66	1,120		
	Opx-Ilm	3.21	616		
0058-1/1	Plag-KFeld	0.54	688	650 ± 20	0.999
	Plag-Magn	4.84	647		
	KFeldsp-Magn	5.38	651		
66209 (1)	Mesop-Magn	6.50	560	560 ± 100	0.975
	Mesop-Opx	1.59	754		
	Opx-Magn	4.91	485		
66261	Mesop-Magn	7.24	515	520 ± 80	0.980
	Mesop-Cpx	1.88	670		
	Cpx-Magn	5.36	439		

(1) The mesoperthite of this sample has not been analyzed; we have taken the mesoperthite TII for the calculations

(2) Linear correlation coefficient of the isotherm

$$^a \Delta_{A-B} = \delta_A - \delta_B$$

for the isotopic reequilibration. (It is to be noted that Duchesne calculated his temperatures by using isolated grains considered as relicts of the orthomagmatic stage). It should be possible to relate the isotopic temperature to the grain size of the minerals and the rate of cooling which, after cooling of the magma to catazonal conditions, is related to the speed of ascent (Bottinga and Javoy 1975; Javoy 1977). For a mean grain size of 1 mm and an initial temperature of 800° C, the speed of ascent would vary between 1 (520° C) and 5 (720° C) mm/year if the rate limiting step is diffusion in magnetite under hydrous conditions. Of course, the model is complicated as the various minerals have different grain sizes and diffusion coefficients. It is apparent that some pairs give a temperature very different from the average 'isotherm' temperature and this could be just the result of these different characteristics.

Conclusions

1. Both the strontium and oxygen isotopic compositions are consistent with a mantle origin for the anorthosites. This is also true for most of the charnockites which appear co-magmatic with the anorthosites,

in agreement with other geochemical and isotope data.

2. However, small departures from the 'mantle range' ($5.5 < \delta^{18}\text{O} < 6.0\text{‰}$), especially the lower values call for some kind of crustal contamination. The oxygen isotope survey of surrounding rocks reveals a large heterogeneity with both high and low values in the range $4.3\text{--}10\text{‰}$. As the low values do not differ much from the mantle range, the magmatic rocks with low $\delta^{18}\text{O}$ values could correspond to a globally minor but locally important contamination phenomenon, that is the melting of ^{18}O depleted roof rocks. This must be tested by a more detailed examination of the trace element geochemistry of the catazonal gneisses.

3. The slow cooling of all the plutonic rocks investigated in this study has allowed some subsolidus isotopic exchange to occur; the oxygen isotope system becomes closed at a temperature around $500^{\circ}\text{--}700^{\circ}\text{C}$. This corresponds either to important deuteric activity during the late magmatic stage or to a slow post-orogenic ascent under wet conditions.

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