THE TIMEDJELALEN ALKALINE RING-COMPLEX AND RELATED N–S DYKE SWARMS (ADRAR DES IFORAS, MALI) — A Pb–Sr–O ISOTOPIC STUDY

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Abstract

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The Iforas alkaline province is composed of spectacular N–S-trending acid dyke swarms, plateaux of ignimbritic and rhyolitic lavas, and silica-oversaturated ring-complexes. This province was emplaced just after the Pan-African collision which occurred between the Tuareg shield and the West-African craton 600 Ma ago. It is devoid of Sn mineralization in comparison with similar Paleozoic–Mesozoic complexes in Niger and Nigeria.

The Pb–Sr–O data on the Timedjelalen ring-complex and on contemporaneous N–S dykes demonstrate the cogenetic mantle source for their parental magma, with a contamination by the lower continental crust from 10% to 50%, on the basis of Pb isotopes. There is no evidence for incorporation of old upper continental crust. These alkaline rocks show little alteration except for hydrothermal activity at ring- or dyke-country-rock contacts, as shown by lower $\delta^{18}O$-values. The hydrothermal water involved appears isotopically heavy (in the range 0 to $-3‰$) and may have originated in a residual sea or lake. Lack of mineralization can be explained by the fact that the primary source is economically barren (mantle and lower crust) and that the interaction with country-rocks during the intrusion process was weak and mainly with barren mantle-derived Pan-African rocks.

1. Introduction

The Cambrian alkaline province of Adrar des Iforas includes (Figs. 1 and 2) early plutons, major dyke swarms, lava plateaux and ring-complexes, all following a subduction–collision regime (Black et al., 1979; Liégeois and Black, 1984; Ba et al., 1985). The present study is focused on the dyke swarms of the Tiralrar–Ichoualen area and on the Timedjelalen ring-complex, situated just southwest of the dykes. Both are contemporaneous (around 540 Ma) and have comparable geochemical characteristics (Liégeois and Black, 1984). Except for some turbid feldspars and chloritized bio-

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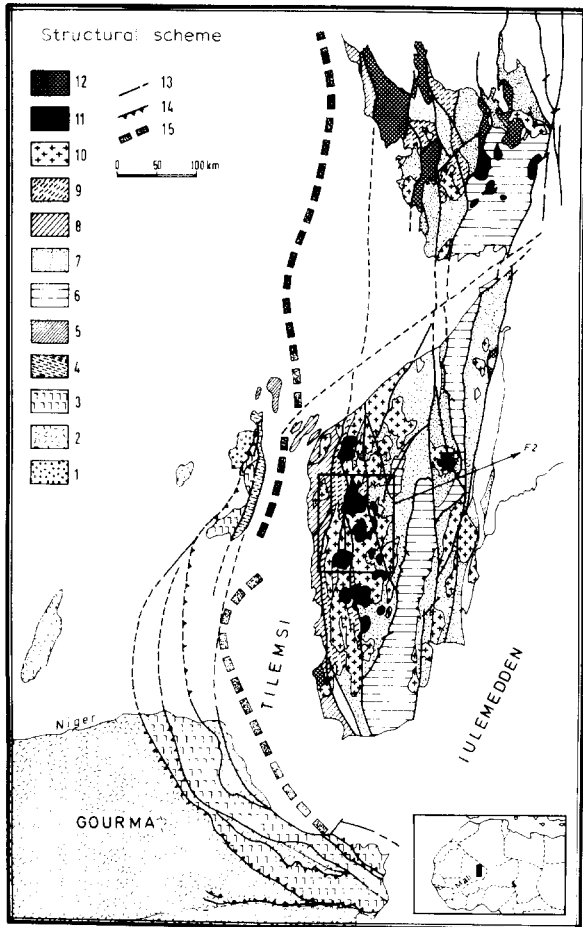
Fig. 1. Structural scheme modified after Fabre et al. (1982): *West African craton* (west of the suture): *1* = unreactivated Eburnean basement; *2* = Gourma aulacogen (autochthonous passive margin); *3* = Timetrine-Gourma nappes; *4* = Permian Tessofi graben; *5* = Permian undersaturated ring-complexes (Tadhak area) and *Iforas mobile belt* (east of the suture): *6* = slightly reactivated basement and shelf cover; *7* = reactivated basement; *8* = island-arc volcanosedimentary assemblage; *9* = cordillera volcanosedimentary assemblage; *10* = calc-alkaline batholith; *11* = alkaline oversaturated ring-complexes and lava plateaux; *12* = Cambrian molasse; *13* = shear zones and transcurrent faults; *14* = thrusts; *15* = gravimetric suture zone.

Fig. 2. Geological map of the Tiralrar-Timedjelalen area representing the alkaline rocks (*1* = Tahrmert early alkaline pluton; *2* = N-S dykes; *3* = rhyolitic and ignimbritic plateaux; T*1*-T*6* — different intrusive phases of the Timedjelalen ring-complex: *4* (T*1*) = syenitic and microgranitic ring-dykes; *5* (T*2*) = medium-grained perthite-aegirine-augite-Na-Ca-amphibole-biotite granite; *6* (T*3*) = coarse-grained perthite-aegirine-augite-Na- and Na-Ca-amphiboles; *7* (T*4*) = perthite-oligoclase-amphibole-biotite granite; *8* (T*5*) = fine-grained perthite-oligoclase-biotite granite; *9* (T*6*) = medium- to fine-grained aegirine-arfvedsonite granite.

tites, there is little alteration, especially in the ring-complex.

The alkaline Iforas province, with typical anorogenic features, represents the last post-tectonic event related to the late Precambrian intercon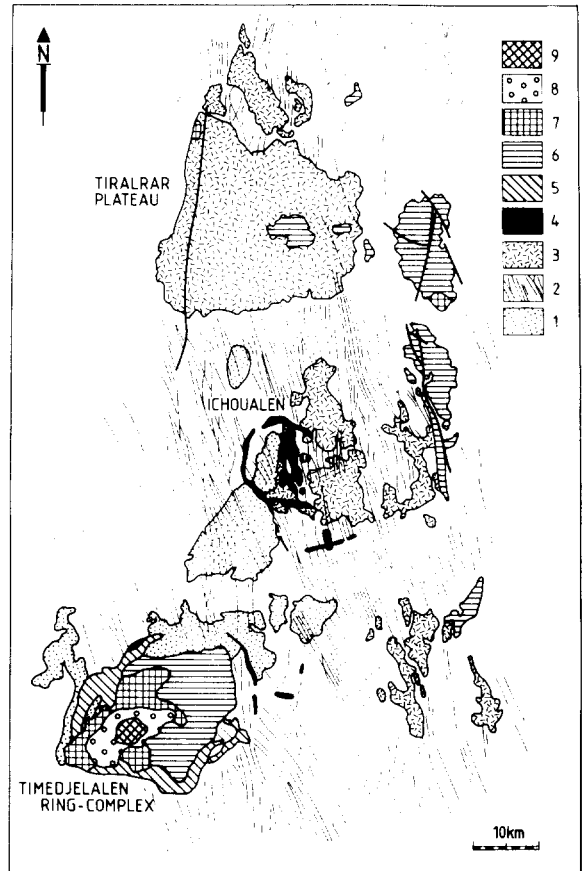tinental collision between the Tuareg shield and the West African craton (Black et al., 1979). We have undertaken a combined Pb-Sr-O isotopic study on these dykes and on the ring-complex in an attempt to evaluate the nature and intensity of the hydrothermal processes as well as to characterize the source of this alkaline magmatism. This study should put new constraints on the source of these alkaline

rocks linked to plate-tectonics and on subvolcanic intrusion processes. This should be especially interesting as alkaline acid magmatism with comparable characteristics in both oceanic and continental environments is inferred to have a mantle origin (Bonin and Lameyre, 1978; Bonin and Giret, 1985). These Malian alkaline rocks are devoid of mineralization, in contrast to the mineralized Niger–Nigeria alkaline province whose other characteristics are very similar.

2. Geological setting

The Adrar des Iforas, the Malian part of the Tuareg shield, has been integrated in a late Precambrian plate-tectonic model consisting of a complete "Wilson cycle" (Black et al., 1979; Fabre et al., 1982). An extensional period led, around 850 Ma ago, to the formation of an ocean between the West African craton and the Tuareg shield. In the Iforas, the Pan-African active margin consists of two well-developed environments elongated in a N–S direction along the trench: (1) to the west, an island arc root composed of deep volcanosedimentary sequences intruded by basic plutonic rocks and with no sign of old continental basement (Caby, 1981; Dupuy et al., in prep.). There is evidence to suggest that subduction started around 720 Ma ago (R. Caby and U. Andreopoulos-Renaud, unpublished data); and (2) to the east of this island arc, a cordillera with Archean and Proterozoic basement was intruded essentially by tonalites and diorites and covered by andesites (Chikhaoui, 1981). This subduction period, with easterly dipping Benioff zone, ended 620–590 Ma ago during collision between the West African craton and the Tuareg shield (Ducrot et al., 1979; Liégeois and Black, 1984). At present, the two paleo-continents are separated by a suture marked by a string of positive gravimetric anomalies (Bayer and Lesquer, 1978) and the two paleo-environments (island arc and cordillera) of the Iforas are separated by the N–S mega-shear zone (0°50′E) of Tessalit–Anefis (Fig. 1). Collision between the two continents has probably been oblique, corresponding to a tectonic phase with important N–S stretching (constriction) (Ball and Caby, 1984). The intensity was weak and has not induced a Himalayan-type confrontation. This oceanic closure has produced major amounts of calc-alkaline granitoids mainly of mantle origin (Bertrand and Davison, 1981; Liégeois and Black, 1984; Dupuy et al., in prep.). Pre- and syn-tectonic granitoids constitute a large composite batholith (size: 200×50 km; Liégeois et al., 1986, fig. 1). Rapid uplift occurred soon after collision contemporaneously with the intrusion of a volumetrically small post-tectonic calc-alkaline group. This suite is considered as the last manifestation of the previous syn-tectonic magmatic event (Liégeois and Black, 1984).

Around 560 Ma ago the alkaline group which is well represented in the Adrar des Iforas was emplaced. An early intrusion was the Tahrmert-type biotite hypersolvus granite, which occurs in high-level plutons. After this intrusion all the units (pre-, syn- and post-tectonic) were eroded, leading to unroofing of the batholith. Rhyolitic and ignimbritic lavas were extruded on this eroded surface, and their relics constitute extrusive plateaux (Nigritian; Karpoff, 1960). Because the late molasses of the belt are composed essentially of rhyolitic and ignimbritic material (Fabre, 1982), these acid lavas must have covered all the Iforas. They were fed by impressive N–S dyke swarms (commonly more than 100 km long), which are studied in this paper. The last magmatic event in the Iforas consisted of the intrusion of the alkaline ring-complexes between 560 and 540 Ma (Kidal and Timedjelalen ring-complexes, 25- and 21-WR Rb–Sr isochron, respectively; Liégeois and Black, 1984; Ba et al., 1985). They are aligned along the N–S axis of the calc-alkaline batholith in a direction parallel to the suture, and locally intrude the acid volcanic pile. The country-rocks of the ring-complexes are then essentially of Pan-African age.

This study is focused on the alkaline group

represented by the N–S dykes in the Tiralrar area and the space- and time-related Timedjelalen ring-complex. The lavas were not included in this study because they have been severely altered.

3. The Tiralrar N–S dykes and the Timedjelalen ring-complex

The studied N–S dyke swarms, with a length of more than 100 km, are situated NE of the Timedjelalen complex in the Tiralrar–Ichoualen area and include the feeder dykes of the ignimbritic and rhyolitic Tiralrar lava plateaux. The dykes are composed of brownish devitrified rhyolites, granophyric and porphyritic microgranites and quartz-microsyenites often containing basic inclusions. The phenocrysts are rhyolitic quartz, perthitic K-feldspar often deuterically altered and a few ferromagnesian aggregates (mostly biotite but also clinopyroxene and amphiboles). As in the case of the alkaline granites, both metaluminous and peralkaline varieties occur. The groundmass is composed of quartz and feldspars. The N–S dykes are SiO_2 rich, most of them having around 75% SiO_2.

Adrar Timedjelalen is one of the largest ring-complexes in the Iforas and is very typical with its six principal concentric phases (Fig. 2). The field survey has shown the following succession from the border to the centre (Ba et al., 1985):

T1: Porphyritic microgranite (quartz, malgachitic perthitic alkali feldspar, microcrystalline groundmass)

T2: Medium-grained granite (quartz, perthite, aegirine–augite, Na- and Ca-Na-amphiboles)

T3: Coarse-grained granite (quartz, perthite, aegirine–augite, Na- and Ca-Na-amphiboles)

T4: Heterogeneous granite (quartz, perthite, oligoclase, amphibole, biotite)

T5: Fine-grained granite (quartz, perthite, oligoclase, biotite)

T6: Fine- to medium-grained granite (quartz, perthite, aegirine, arfvedsonite, astrophyllite)

Compositions of the perthites vary as a function of the hypersolvus or subsolvus character of the granites and there are no relations

between the purity of phases and the different rings (Ba et al., 1985): indeed, the orthoclase proportions in the perthites (N_{or}) are: hypersolvus-group perthite (T1, T2, T3, T6): K-feldspar: $N_{or} = 95.2$–98.2, albite: $N_{or} = 0.2$–1.4; and subsolvus-group perthite (T4, T5): K-feldspar: $N_{or} = 89$ (T4), 73 (T5), albite: $N_{or} = 2.25$ (T4), 13.6 (T5). Volcanic structures are absent. The aluminous series (T4 and T5) occurred in the middle of the magmatic cycle: it is flanked by the peralkaline series (preceded by T1, T2 and T3 and followed by T6). As Ba et al. (1985) have pointed out, the Iforas ring-complexes are mainly characterized by the following features: (1) all the rocks are SiO_2 rich (more than 65% SiO_2); (2) the hydrothermal alteration is weak and has not modified the chemistry; this is a distinctive feature of the Iforas rocks when compared to the Younger Granites of Nigeria (Bowden, 1979). The origin of the two magmatic trends (aluminous and peralkaline) in ring-complexes is problematical: either derivation of the two series by fractionation from a common mantle source (Jacobson et al., 1958; Black, 1965; Upton, 1974; Bonin, 1980, 1982; Gadalia and Varet, 1983) or mantle origin for the peralkaline series by differentiation from a basic magma which in turn melts the continental crust producing the aluminous series (Bailey and Schairer, 1966; Bowden, 1970; Wones, 1974; Barker et al., 1975).

4. Analytical techniques

4.1. Lead

About 100 mg of sample powder (crushed to < 200 mesh in an agate mortar) were dissolved in a Teflon® bomb with a 7 : 1 : 1 HF–HNO_3–$HClO_4$ mixture. Pb was separated on anion-exchange columns in HBr–HCl conditions following a method derived from Nunès et al. (1973) and Manhès et al. (1978). Pb and U concentrations were measured on the same sample solution (split before loading on columns and spiked with a mixed spike ^{235}U–^{206}Pb)

and U was separated in HNO_3 medium. All the operations were carried out in an overpressurized (> 5 mm Hg) ultraclean laboratory and in laminar-flow cabinets. Total Pb blank values were lower than 2 ng for the whole chemical procedure. Pb isotope compositions were measured on single-zone-refined Re filament with a Finnigan® Mat 260 mass spectrometer at the Belgian Centre for Geochronology (B.C.G., M.R.A.C.–U.L.B.–V.U.B.) by the silica gel–phosphoric acid method (Cameron et al., 1969). All results were corrected for mass fractionation ($1 \pm 0.34\%_0$ per a.m.u.) on the basis of 34 analyses of the NBS 981 Pb standard (Catanzaro et al., 1968) in a temperature range of 1100–1200°C. Within-run precision for Pb isotopic analyses was better than $\pm 00.5\%_0$ ($2\sigma_m$) and between-run precision better than $\pm 0.1\%$ for the $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios, and better than $\pm 0.15\%$ for the $^{208}Pb/^{204}Pb$ ratios. Pb and U concentrations were measured by the isotope dilution technique on a Varian® Mat TH5 mass spectrometer (B.C.G.) and are given with a precision better than $\pm 2\%$.

4.2. Oxygen

Just before loading in Ni tubes, the samples were crushed in an agate mortar to obtain the best homogeneity. The samples were reacted overnight at 550–600°C with bromine pentafluoride (Clayton and Mayeda, 1963). Oxygen was quantitatively converted to CO_2 on a carbon filament. Isotopic compositions were measured on a VG Micromass® 602C double collector mass spectrometer (Laboratoire de Géochimie des Isotopes Stables, Paris VII), with a CO_2 standard calibrated against silicate and carbonate oxygen international reference materials. Analyses of the silicate SF3 working standard gave an $\delta^{18}O$ average value of $+11.62 \pm 0.06\%_0$ ($2\sigma_m$). NBS 28 was measured at $+9\%_0$ during this period. Each sample was analysed at least twice, especially the alkaline granites with turbid K-feldspars to avoid reproducibility problems due to inhomogeneity of post-magmatic alteration. The mineral separations were done with a Franz® magnetic separator and heavy liquids (M.R.A.C.). Finally, quartz was purified by selective feldspar dissolution [HF treatment (3–5 min.) and HCl washing; Fourcade and Javoy, 1984]. The K-feldspar $\delta^{18}O$-value was obtained from mass-balance calculations. For each sample the mineral purity was checked with a binocular microscope. $^{18}O/^{16}O$ ratios are expressed conventionally in $\delta^{18}O$-values relative to SMOW.

4.3. Strontium

The Sr data have been previously published in a general paper on the Iforas magmatism (Liégeois and Black, 1984). The reader is referred to this paper for analytical procedure and sample petrography.

All age values given in this paper have been calculated with the decay constants recommended by the I.G.C.P.–I.U.G.S. Subcommission of Geochronology (Steiger and Jäger, 1977).

5. Results

5.1. Oxygen

Nine whole-rock oxygen analyses were made in each case while four quartz–K-feldspar pairs were analysed for the Timedjelalen ring-complex and three for the N–S dykes. The results are reported in Table I and plotted vs. SiO_2 in Fig. 3.

For the *Timedjelalen ring-complex*, most whole-rock $\delta^{18}O$-values are between $+5.7$ and $+7.1\%_0$ [Table I, (b)] and only two samples show distinctly lower $\delta^{18}O$-values. Quartz gives homogeneous $\delta^{18}O$-values with an average at $+6.8 \pm 0.4\%_0$ while the K-feldspars show more variable $\delta^{18}O$-values. In fact, the low whole-rock $\delta^{18}O$-values correspond to low K-feldspar $\delta^{18}O$-

TABLE I

δ^{18}O-values of the N–S dykes and of the Timedjelalen ring-complex (the ring number is in brackets), for whole rock (WR), quartz (Qz) and K-feldspars (Kf); the fourth column gives the difference between quartz and K-feldspars δ^{18}O-values (Δ_{Qz-Kf})

Samples	δ^{18}O (‰)			Δ_{Qz-Kf} (‰)
	WR	Qz	Kf	
(a) N–S dykes:				
JPL357a	+6.3±0.2			
JPL358	+3.3±0.1			
JPL366	+3.1±0.3	+6.5	+0.6	+5.9
JPL375	+4.5±0.3			
JPL376	+5.0±0.5			
JPL387	+2.8±0.1	+6.3	+1.6	+4.7
JPL399	+2.6±0.2	+6.4	+1.3	+5.1
JPL403	+4.1±0.1			
JPL419b	+2.0±0.3			
(b) Timedjelalen ring-complex:				
JPL306 (T2)	+6.2±0.1	+7.0	+5.7	+1.3
RB360 (T2)	+5.7±0.1			
JPL334 (T3)	+5.7±0.3			
JPL325 (T4)	+7.1±0.2			
JPL341 (T4)	+3.1±0.3	+6.3	+1.3	+4.7
JPL327 (T5)	+6.5±0.3			
JPL340 (T5)	+3.8±0.3	+7.1	+1.3	+5.8
JPL337 (T6)	+6.1±0.3	+6.7	+5.7	+1.0
JPL339 (T6)	+6.6±0.2			

Fig. 3. δ^{18}O vs. %SiO$_2$ for the N–S dykes and for the Timedjelalen ring-complex. Timedjelalen whole-rock (♦ = T1; ● = T2; ■ = T3; ▼ = T4; ● = T5; ▲ = T6; corresponding *horizontally hatched symbols* for quartz and *vertically hatched symbols* for K-feldspar. N–S dykes: * = whole-rock; + = quartz; × = K-feldspar.

values ($< +2$‰) with large Δ_{Qz-Kf}-values ($> +4.5$‰).

For the *N–S dykes*, whole-rock δ^{18}O-values are more scattered, between +6.3 and +2.0‰. The δ^{18}O in quartz are again very homogeneous with an average at +6.4±0.1‰ while all the analysed feldspars have lower δ^{18}O-values (~ +1.1‰) with corresponding large Δ_{Qz-Kf}-values. All N–S dyke samples plot far away from the magmatic equilibrium line in the δ^{18}O$_{Qz}$ vs. δ^{18}O$_{Kf}$ diagram (Fig. 4).

5.2. Strontium

A Rb–Sr isochron based on 14 whole-rock samples from different dykes (in an area of 200

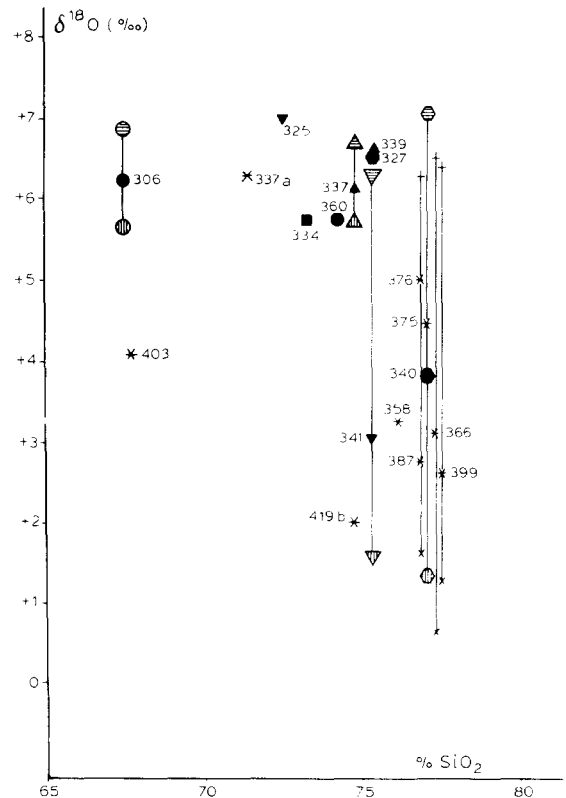km^2) yields an age of 543±9 Ma (initial ^{87}Sr/^{86}Sr (IR) = 0.7050 ± 0.0003, MSWD (mean square of weighted deviates) = 2.0, Fig. 5, calculated by the Williamson (1968) method). Twenty-one whole-rock samples from the Timedjelalen ring-complex yield a concordant Rb–Sr age of 549 ± 6 Ma (21 WR, IR = 0.7052 ± 0.0005, MSWD = 3.0, Fig. 5). The earliest alkaline member, the Tahrmert pluton also gives a comparable age and initial ratio: 541 ± 7 Ma (IR = 0.7061±0.0004) (Liégeois and Black, 1984).

5.3. Lead

Eight samples from the Timedjelalen ring-complex have been analysed for whole-rock Pb

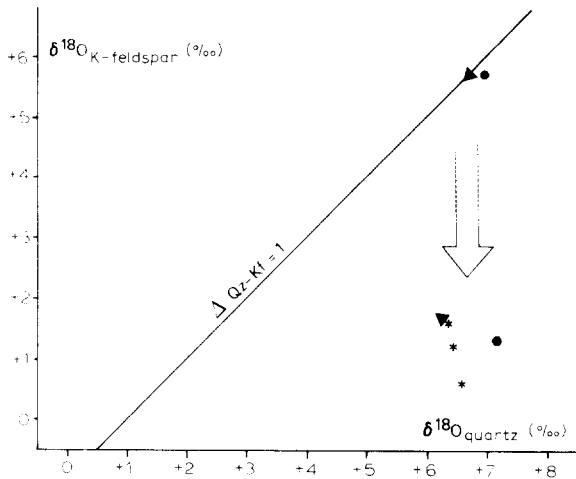Fig. 4. $\delta^{18}O_{K\text{-feldspar}}$ vs. $\delta^{18}O_{quartz}$ diagram. Timedjelalen ring-complex (same *symbols* as in Fig. 3; * = N-S dykes).

isotopic composition and Pb and U concentrations (Table II). In the Pb–Pb diagram (Fig. 6), these samples define a linear array whose slope, if considered as a secondary isochron, corresponds to an age of 698 ± 145 Ma (MSWD $= 0.5$). This value is consistent with the Rb–Sr age within analytical errors (the errors of the Pb–Pb method are much larger for relatively young ages because of the relative small variations of the $^{207}Pb/^{204}Pb$ ratios in recent times, due to the small fraction of ^{235}U left compared with ^{238}U). The $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ diagram shows a bigger scatter. The U–Pb systems are, as usual, much more disturbed, but give fairly concordant results (Fig. 7). The $^{206}Pb-^{238}U$ system gives an isochron only if the four WR samples *299, 334, 360* and *429* are considered (the other samples are also the ones which are scattered in the $^{208}Pb-^{206}Pb$ diagram). In this case it yields an age of 573 ± 26 Ma, $(^{206}Pb/^{204}Pb)_0 = 16.73 \pm 0.04$ (4 WR, MSWD $= 3.9$). This value is very close to that obtained by the Rb–Sr method (549 ± 6 Ma) and implies that these four samples do not show any significant U loss. The $^{207}Pb-^{235}U$ chronometer leads to the same conclusions: the same four WR samples, *299, 334, 360* and *429* give: 599 ± 164 Ma, $(^{207}Pb/^{204}Pb)_0 = 15.36 \pm$
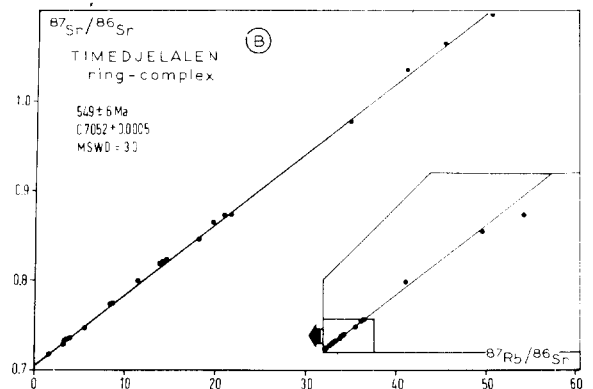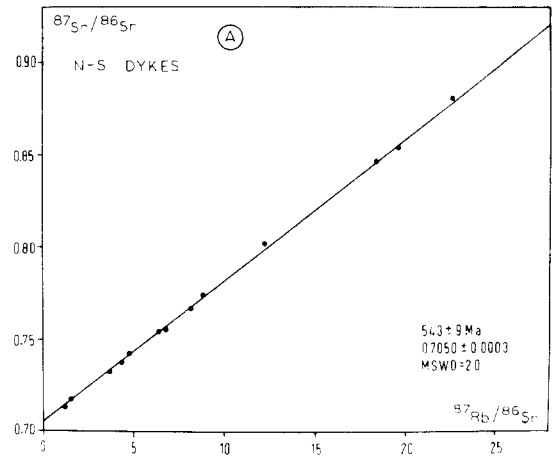

Fig. 5. Rb–Sr isochrons of the N-S dykes (A) and of the Timedjelalen ring-complex (B) after Liégeois and Black (1984).

0.03 (MSWD $= 0.0$). It is noteworthy, that for each sample, the relative deviation from each U–Pb isochron is similar (Fig. 7). The $^{207}Pb/^{204}Pb$ ratios are relatively low and in general, all the Pb ratios are not very radiogenic for granitic rocks of that age. The Pb initial ratios can be deduced in two different ways: either with the Pb–Pb system or with the U–Pb isochron diagrams. Considering that the Timedjelalen complex crystallized 550 Ma ago (Rb–Sr age), the Pb–Pb system provides model initial values of 16.77 and 15.37 for the $^{206}Pb/^{204}Pb$ and $^{207}Pb/^{204}Pb$ ratios, respectively (Fig. 6). These values are not significantly different from those given by the U–Pb systems (see above). Consequently, the initial Pb ratios for the Timed-

TABLE II

Whole-rock Pb isotopic compositions and U and Pb concentrations for the Timedjelalen ring-complex (the ring number is in brackets)

Sample	$^{208}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{206}Pb/^{204}Pb$	Pb(*) (ppm)	U(*) (ppm)	$^{238}U/^{204}Pb$
JPL 299 (T2)	40.182	15.503	19.054	10.4	3.8	24.25
RB 360 (T2)	37.988	15.415	17.547	16.9	2.5	9.31
JPL 334 (T3)	37.280	15.381	17.052	31.7	1.7	3.22
RB 429 (T4)	38.029	15.420	17.682	17.2	2.9	10.45
RB 446 (T4)	38.020	15.452	18.051	23.4	4.4	11.95
JPL 340 (T5)	38.761	15.486	18.610	18.0	3.6	12.68
JPL 337 (T6)	38.959	15.498	18.869	48.1	13.8	18.49
RB 393 (T6)	37.790	15.454	18.043	72.3	14.0	12.08

*Isotope dilution method.
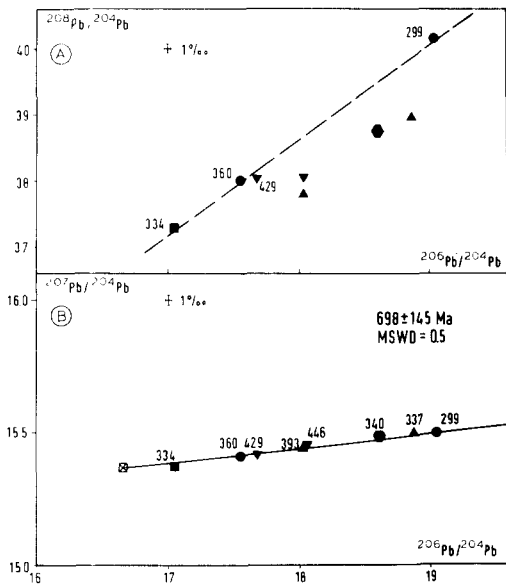
Fig. 6. $^{208}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ (A) and $^{207}Pb/^{204}Pb$ vs. $^{206}Pb/^{204}Pb$ (B) diagrams for the Timedjelalen ring-complex. The *cross in a square* represents the calculated initial ratios for the $^{206}Pb/^{204}Pb-^{207}Pb/^{204}Pb$ pair. *Symbols* as in Fig. 3. See text for explanation.

Fig. 7. $^{207}Pb-^{235}U$ (A) and $^{206}Pb-^{238}U$ (B) isochron diagrams for the Timedjelalen ring-complex. *Symbols* as in Fig. 3.

6. Discussion

6.1. Interpretation of the low $\delta^{18}O$-values

jelalen ring-complex are taken to be those of the U–Pb isochrons:

$$(^{206}Pb/^{204}Pb)_0 = 16.73 \pm 0.04$$
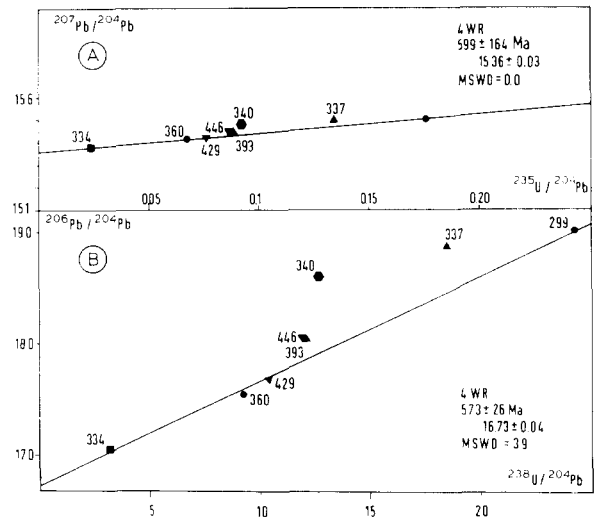
and

$$(^{207}Pb/^{204}Pb)_0 = 15.36 \pm 0.03$$

All but two of the Timedjelalen complex samples, as well as one N–S dyke, have O isotopic compositions at the lower end of the normal-^{18}O group ($\delta^{18}O$ = +6 to +8‰) of granites as defined by H.P. Taylor (1968). On the other hand, two samples from the ring-com

plex and most of the N–S dyke whole-rock samples show lighter O isotopic compositions ($\delta^{18}O$ = +2.0 to +5.0‰). This could reflect either interaction with a low-$\delta^{18}O$ fluid (meteoric water at high temperature or previously hydrothermally altered country-rocks) or an origin by direct partial melting of pre-existing, low-$\delta^{18}O$ rocks.

Both the low-$\delta^{18}O$ Timedjelalen and the N–S dyke samples show typical geological and petrological features of igneous complexes characterized by ^{18}O depletion as described by H.P. Taylor (1977). The intrusions are subvolcanic complexes, the feldspars are turbid and more depleted in ^{18}O than the other minerals, whilst the occurrence of granophyric intergrowhts between K-feldspar and quartz is common. All these features, together with the very homogeneous $\delta^{18}O_{qz}$-values at around +6.5‰, as well as the majority of whole-rock $\delta^{18}O$-values in the Timedjelalen ring-complex above +5.7‰, indicate that the low $\delta^{18}O$-values reflect a secondary process, i.e. hydrothermal alteration. This alteration must have occurred at relatively low temperatures (below the stability fields of the OH-bearing minerals) and does not result from direct crystallization from a low-$\delta^{18}O$ magma. This is confirmed by the position of the low-$\delta^{18}O$ samples in the $\delta^{18}O_{Qz}$ vs. $\delta^{18}O_{Kf}$ diagram (Fig. 4) which fall on the trend of rocks whose isotopic disequilibrium has been attributed to hydrothermal alteration (Criss and Taylor, 1983; Fourcade and Javoy, 1984). As quartz is a very resistant mineral, whilst K-feldspar is very sensitive to isotope exchange during hydrothermal alteration, this diagram is very useful for testing the isotopic equilibrium within magmatic rocks. Indeed, the Timedjelalen samples with normal $\delta^{18}O$-values fall close to the line defined by a fractionation factor, Δ_{Qz-Kf}, of +1‰ corresponding to magmatic equilibrium values (Garlick, 1966; H.P. Taylor, 1968, 1977; Anderson et al., 1971).

All the altered K-feldspars have low $\delta^{18}O$-values, giving an average at +1.3 ± 0.4‰ (+0.6 to +1.6‰, Fig. 3). For usual water/rock ratios and for a mean temperature of alteration of ~400°C, this homogeneity of the $\delta^{18}O_{Kf}$-values implies that the water/rock ratios have not varied by more than a factor of 2. From the balance and fractionation values in the equilibria equations (Bottinga and Javoy, 1975), we derive $\delta^{18}O_{Kf}$-values for the meteoric water between 0 and −3‰, with water/rock ratios between 0.5 and 2, respectively. These values are relatively heavy by comparison with the typical negative values of meteoric water (Craig, 1961; H.P. Taylor, 1977) in the range −15 and −5‰. An explanation could be found in the existence of residual basins subjected to variable evaporation rates. Preliminary studies of contemporaneous molasses (Fabre et al., 1982) seem to confirm the existence of such kind of basins in the Iforas at that time. However, more geological and isotopic data, particularly hydrogen systematics, are necessary to confirm this suggestion.

Another question which arises from the oxygen results concerns the "engine" of the hydrothermal circulation: is it driven by the intrusions themselves or by another heat source? The Timedjelalen ring-complex could easily provide the heat necessary to generate a convective hydrothermal system, but this is doubtful in the case of the dykes. Nevertheless, another possible heat source could be the thick rhyolitic and ignimbritic pile which is thought to have covered the entire region at the closure of the Pan-African period (Liégeois and Black, 1984). Such hydrothermal systems are known under flood basalt sequences (Walker, 1960) and are likely to have produced similar effects in the In Zize complex (Hoggar; Fourcade and Javoy, 1984). Moreover, the fact that only two closely-spaced samples from the Timedjelalen ring-complex have been affected (within 500 m, but from two different intrusions) implies that the hydrothermal alteration, which was nevertheless intense, was only a local phenomenon along contacts between successive ring–dykes just as in the case of the dyke–country-rock contacts for the N–S dykes. In this way, the

respective size and shape of the intrusions easily explain why the dykes are more affected than the ring-complex.

6.2. The source of the parental magma

The initial $^{87}Sr/^{86}Sr$ ratios of 0.7050 ± 0.0003 and 0.7052 ± 0.0005, identical within analytical error for the N–S dykes and the ring-complex are slightly higher than the Pan-African mantle values in subduction environment under the Iforas (0.7025–0.7030; Dupuy et al., in prep.). They are nevertheless low enough to preclude an origin by anatexis of old upper-crustal rocks. The role of the lower crust remains uncertain. Although its participation in the genesis of the alkaline Iforas Pan-African rocks could be invoked to explain their high silica contents, it is difficult to derive the volume of rocks observed in the Adrar des Iforas from an unradiogenic and residual lower crust only. Moreover, the Archean granulitic unit of the Iforas, which may represent the lower crust in this area, was shown to have a mean of $^{87}Sr/^{86}Sr$ around 0.709 during the Pan-African epoch (Ferrara and Gravelle, 1966; Liégeois et al., 1986).

O-isotopic compositions with whole-rock $\delta^{18}O$-values between +5.7 and +7.0‰ agree with the Sr isotope model. These values are in the low-^{18}O end of the normal-^{18}O group of granites defined by H.P. Taylor (1968), i.e. within the range of values which could have been formed by direct differentiation of mantle-derived magmas. Unaltered mid-ocean ridge and ocean island basalts have values between +5.4 and +7.3‰ with an average at +5.8 ± 0.3‰ (Garlick, 1966; H.P. Taylor, 1968; Javoy, 1970; Muehlenbachs and Clayton, 1972; Pineau et al., 1976). Granitic rocks usually have values between +7 and +16‰. Many of the high values have been ascribed to interaction with water at low temperature, although values as high as +12‰ have also been considered as magmatic (H.P. Taylor et al., 1984). On the other hand, $\delta^{18}O_{SR}$-values of ~ +6‰ have seldom been reported for granitic rocks. Only rhyolites from Easter Island (H.P. Taylor, 1968), acidic rocks from Ethiopia (Javoy, 1970) and alkaline granite inclusions in Ascension lavas (Weis et al., 1987) show values below +7‰. They confirm that fractional crystallization alone cannot produce values higher than +7‰. This is also reflected by the values (from +5.9 to +6.3‰) of the only "acidic" sample found on the Moon (12013; H.P. Taylor and Epstein, 1973).

The most likely ultimate origin for low-$\delta^{18}O$ granitoids is the mantle with or without contribution from the lower continental crust. Sr and O isotopes cannot differentiate these two reservoirs because many of the deep-basement rocks have $^{87}Sr/^{86}Sr$ and $\delta^{18}O$ within the mantle range (Fourcade and Javoy, 1973; Shieh and Schwarcz, 1974). Nevertheless, both Sr and O isotopic compositions indicate a common parental magma, with similar geochemical characteristics for the ring-complex and the dykes. This is confirmed by the remarkable homogeneity of $\delta^{18}O$ of quartz in all alkaline granite anorogenic complexes (Javoy and Weis, submitted).

The Pb isotopic system provides new insight into the problem of the origin of the parental magma. The Pb initial isotopic ratios of the Timedjelalen complex fall between the evolution curve of the mantle and the lower crust as defined by the plumbotectonics model (Zartman and Doe, 1981). Additional and more precise constraints can be added to infer the origin of the alkaline rocks of the Iforas as it has been shown that the parental magma of the Tadhak complex originated in an ocean island basalt (OIB) type mantle, under northern Mali (Weis et al., submitted). The Tadhak alkaline pluton, of Permian age (Liégeois et al., 1983) is situated just west of the suture, 200 km NW of the Timedjelalen ring-complex (Fig. 1). It is the main plutonic complex of an undersaturated province intruding the Eburnean (2000-Ma-old African orogeny) West African craton, composed of old gneisses and granites. The geo-

chemical and isotopic characteristics of the Tadhak pluton compared with those of the country-rocks are very different. The parental magma of this complex derived from the subcontinental mantle whose geochemical and isotopic characteristics are comparable to those of the OIB source mantle, with no evidence of continental crust influence (Weis et al., submitted). Its initial Pb and Sr isotopic ratios can then be used as a reference for characterizing the OIB mantle under that region. For the purpose of this study, the values were corrected back from 270 Ma (age of intrusion) to 540 Ma. They are: $^{87}Sr/^{86}Sr = 0.70426-0.70445$ ($^{87}Rb/^{86}Sr$ of 0.08 and 0.03, respectively), $^{206}Pb/^{204}Pb = 18.11$ and $^{207}Pb/^{204}Pb = 15.56$.

Like the Sr and O isotopic results, the initial Pb isotopic ratios ($^{206}Pb/^{204}Pb = 16.73$ and $^{207}Pb/^{204}Pb = 15.36$) in the ring-complex rule out a participation of old upper crust. Because of the similar geochemical and isotopic characteristics of the N–S dykes and the Timedjelalen ring-complex, we will assume that it is the same for the Pb isotopes. The latter indicate that the parental magma of the Timedjelalen pluton can derive from the mantle. Nevertheless, the comparison of the initial Pb ratios with those of the mantle at 540 Ma (Tadhak-type mantle, $^{206}Pb/^{204}Pb = 18.1-18.7$ and $^{207}Pb/^{204}Pb = 15.56-15.59$, or plumbotectonics model mantle, 17.24 and 15.42, respectively) suggests that another component was also involved in the genesis of these rocks. Indeed, these initial ratios are significantly lower than the mantle values (either OIB or MORB values). The lower crust is the only reservoir which has Pb ratios less radiogenic than the mantle and it appears to be the best candidate to fulfil all the constraints imposed by Pb, Sr and O isotopic systems. An approximate estimate of the amount of Pb of lower crust incorporated in the alkaline magmatism in the Iforas can be obtained by modelling the Pb isotope systematics in the Timedjelalen complex. For that purpose, we assume that the mantle values correspond to the Tadhak isotopic ratios recalculated at 540 Ma. The Pb content of the mantle component is estimated at ~ 0.15-0.2 ppm ("primary mantle"; O'Nions et al., 1979; Zartman and Doe, 1981; Chen and Pallister, 1981) while that of the lower crust is estimated at 7.5 ppm (S.R. Taylor and McLennan, 1981). The Pb isotopic compositions of the latter can be estimated from the plumbotectonics model values (Zartman and Doe, 1981; lower crust 540 Ma ago: $^{206}Pb/^{204}Pb = 16.71$, $^{207}Pb/^{204}Pb = 15.29$). The calculation gives a range of values from 10% to 50% for the lower-crust participation in the genesis of the Timedjelalen complex on the basis of the Pb ratios, a more realistic value being $\sim 30\%$. A change in some of the assumed parameters, i.e. Pb isotopic compositions of the two extreme components, has only a very small influence on the calculated amount of lower-crust contamination because of the drastically lower Pb contents in the mantle compared to the lower continental crust. This indicates that the primary magma source was the mantle, even in the case where the alkaline rocks have been formed just after an important tectonic event (intercontinental collision). On the other hand, a change in the assumed concentration values, for both the mantle and the lower-crust components, has a much larger effect on the estimated proportions of contamination: for instance, assuming 15 ppm Pb in the lower crust (which is not especially unrealistic as the Pb contents depend strongly on the chemical composition of the crust, i.e. felsic or mafic) reduces the contamination values by a factor of 2. This indicates also that these values have to be taken as a general indication only. The mean $^{87}Sr/^{86}Sr$ ratio 600 Ma ago of the Iforas Archean granulites, susceptible to represent the Iforas lower crust, has been measured at 0.709 from a representative set of samples (Liégeois et al., 1986). With the same mantle reference (Tadhak, $^{87}Sr/^{86}Sr = 0.7043$, 600 Ma ago), the Sr isotopes indicate a much weaker lower-crust contamination. The Pb isotopic system appears to have a different behaviour than Sr during magmatic processes and is

much more sensitive to crustal contamination. In fact, modelling of the isotope systematics provides only selective contamination values. Bulk amounts of lower crust incorporated are probably much lower.

The results of this study coupled with those obtained on the basement (Liégeois and Black, 1984) may help to explain some economic geological features. Indeed, a striking characteristic of the Iforas alkaline province is the lack of mineralization (except in Th), as also observed in the oceanic ring-complexes of the Kerguelen Islands (Lameyre et al., 1976) whose petrographical and structural characteristics are very similar to the well-known mineralized Niger–Nigeria province (Van Breemen et al., 1975). However, in the Adrar des Iforas, three features could account for this lack of mineralization:

(1) the main primary source of the alkaline Iforas magmatism is the mantle, which has very low concentrations of elements of economic interest;

(2) the contaminant is the lower crust, which is also relatively poor from an economic point of view;

(3) the hydrothermal alteration, which was nevertheless intense, has a very limited extension and corresponds to low temperatures. Moreover, the country-rocks are essentially barren because they are constituted by the calc-alkaline Pan-African composite batholith of the Iforas (Liégeois et al., 1986) which is rooted in the mantle and shows only a weak lower-crust contamination (Liégeois and Black, 1984, 1986). Black (1984) suggests that the chances of finding deposits of economic interest (particularly Sn) in alkaline anorogenic complexes are considerably enhanced when, like in Nigeria, they cut Pan-African crustally derived granitoids with higher $^{87}Sr/^{86}Sr$ initial ratios which have already redistributed and concentrated the metal.

7. Conclusions

(1) In the Iforas, the parental magma of the oversaturated alkaline rocks has a mantle origin with lower-crust contamination (Sr, O and Pb isotopic evidence). The amounts of contamination range, on the basis of Pb isotopes, from 10% to 50% (a more likely value being around 30%) and correspond probably to selective values for Pb. They depend on Pb concentrations and, to a smaller extent, on Pb compositions assumed for the two end-members. Sr isotopes indicate a much weaker lower-crust incorporation.

(2) The participation of the upper crust can be entirely excluded in the alkaline Iforas province on the basis of the Sr, O and Pb isotope systematics in the Timedjelalen ring-complex and the N–S dyke swarms. This is interesting because the upper crust has often been considered as a possible source or as a possible contaminant in the genesis of alkaline complexes in general, and in some cases it has been involved in the genesis of these rocks (Nigerian Younger Granites, Van Breemen et al., 1975; Seychelles granites, Weis and Deutsch, 1984).

(3) $\delta^{18}O_{Qz}$-values are very homogeneous and in the magmatic range ($+6.6 \pm 0.3‰$) both in the dykes and in the ring-complexes. K-feldspars from the dykes and two samples from the ring-complex show lower values indicative of either late or even post-magmatic hydrothermal alteration which must have been intense but is spatially limited at dyke and ring contacts.

(4) The lack of typical mineralization appears characteristic of the Iforas alkaline province and could be explained by an origin within economically-barren reservoirs (mantle and lower crust) of both the alkaline rocks and their calc-alkaline country-rocks, more especially as the hydrothermal alteration was of very limited extension.

is now "chercheur qualifié" of the same institution.

References

Anderson, A.T., Clayton, R.N. and Mayeda, T.K., 1971. Oxygen isotope thermometry of mafic igneous rocks. J. Geol., 79: 715–729.

Ba, H., Black, R., Benziane, B., Diombana, D., Hascoet-Fender, J., Bonin, B., Fabre, J. and Liégeois, J.P., 1985. La province des complexes alcalins sursaturés de l'Adrar des Iforas, Mali. J. Afr. Earth Sci., 3: 123–142.

Bailey, D.K. and Schairer, J.F., 1966. The system $Na_2O-Al_2O_3-Fe_2O_3-SiO_2$ at 1 atmosphere and the petrogenesis of alkaline rocks. J. Petrol., 7: 114–170.

Ball, E. and Caby, R., 1984. Open folding and constriction synchronous with nappe tectonics along a mega-shear zone of Pan-African age. In: J. Klerkx and J. Michot (Editors), African Geology, Mus. R. Afr. Cent., Tervuren, p. 75–90.

Barker, F., Wones, D.R., Sharp, W.N. and Desborough, G.A., 1975. The Pikes Peak batholith, Colorado Front Range, and a model for the origin of the gabbro-anorthosite-syenite-potassic granite suite. Precambrian Res., 2: 97–160.

Bayer, R. and Lesquer, A., 1978. Les anomalies gravimétriques de la bordure orientale du craton ouest-africain: géométrie d'une suture pan-africaine. Bull. Soc. Géol. Fr., 20: 863–876.

Bertrand, J.M.L. and Davison, I., 1981. Pan-African granitoids emplacement in the Adrar des Iforas mobile belt (Mali) — A Rb/Sr isotope study. Precambrian Res., 14: 333–362.

Black, R., 1965. Sur la signification pétrogénétique de la découverte d'anorthosites associées aux complexes annulaires sub-volcaniques du Niger. C.R. Acad. Sci., Paris, 260: 5829–5832.

Black, R., 1984. The Pan-African event in the geological framework of Africa. Pangea, 2: 6–16.

Black, R., Caby, R., Moussine-Pouchkine, A., Bayer, R., Bertrand, J.M.L., Boullier, A.M., Fabre, J. and Lesquer, A., 1979. Evidence for Precambrian plate tectonics in West Africa. Nature (London), 278: 223–227.

Bonin, B., 1980. Les complexes acides alcalins continentaux: l'exemple de la Corse. Thesis, Pierre and Marie Curie University, Paris, 756 pp.

Bonin, B., 1982. Les granites des complexes annulaires. Bur. Rech. Géol. Min., Man. Méth. No. 4, 250 pp.

Bonin, B. and Giret, A., 1985. Contrasting roles of rock-forming minerals in alkaline ring-complexes. J. Afr. Earth Sci., 3: 41–50.

Bonin, B. and Lameyre, J., 1978. Réflexions sur la position et l'origine des complexes magmatiques anorogéniques. Bull. Soc. Géol. Fr., 20: 45–59.

Bottinga, Y. and Javoy, M., 1975. Oxygen isotope partitioning among the minerals in igneous and metamorphic rocks. Rev. Geophys. Space Phys., 13: 401–418.

Bowden, P., 1970. Origin of the Younger Granites of northern Nigeria. Contrib. Mineral. Petrol., 25: 153–162.

Bowden, P., 1979. Geochemical aspects of the evolution and mineralization of the Nigerian Mesozoic anorogenic granites. Stud. Geol., 14: 183–188.

Caby, R., 1981. Associations volcaniques et plutoniques prétectoniques de la bordure de la chaîne pan-africaine en Adrar des Iforas (Mali): un site de type arc-cordillère au Protérozoïque supérieur. Abstr. 11th Colloq. on African Geology, Milton Keynes, p. 30.

Cameron, A.E., Smith, D.H. and Walker, R.L., 1969. Mass spectrometry of nanogram-size samples of lead. Anal. Chem., 41: 525–526.

Catanzaro, E.J., Murphy, T.J., Shields, W.R. and Garner, E.L., 1968. Absolute isotopic abundance ratios of common, equal atom, and radiogenic lead isotope standards. J. Res. Natl. Bur. Stand., 72A: 261–267.

Chen, J.H. and Pallister, J.S., 1981. Lead isotopic studies of the Samail ophiolite, Oman. J. Geophys. Res., 86: 2699–2708.

Chikhaoui, M., 1981. Les roches volcaniques du Proterozoïque supérieur de la chaîne pan-africaine du NW de l'Afrique (Hoggar, Anti-Atlas, Adrar des Iforas) — Caractérisation géochimique et minéralogique implications géodynamiques. Thesis, University of Montpellier II, Montpellier, 195 pp.

Clayton, R.N. and Mayeda, T.K., 1963. The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis. Geochim. Cosmochim. Acta, 27: 43–52.

Craig, H., 1961. Standard for reporting concentrations of deuterium and oxygen-18 in natural waters. Science, 133: 1833–1834.

Criss, R.E. and Taylor, Jr., H.P., 1983. An $^{18}O/^{16}O$ and D/H study of Tertiary hydrothermal systems in the southern half of the Idaho batholith. Geol. Soc. Am. Bull., 94: 640–663.

Ducrot, J., de La Boisse, H., Renaud, U. and Lancelot, J.R., 1979. Synthèse géochronologique sur la succession des évènements magmatiques pan-africains au Maroc, dans l'Adrar des Iforas dans l'Est-Hoggar. 10e Colloq. on African Geology, Montpellier, p. 40.

Dupuy, C., Liégeois, J.P. and Caby, R., in prep. The Tilemsi magmatic arc and the Pan-African suture zone in Mali — A geochemical and isotopic study.

Fabre, J., 1982. Pan-African volcano-sedimentary formations in the Adrar des Iforas. Precambrian Res., 19: 201–214.

Fabre, J., Ba, H., Black, R., Caby, R., Leblanc, M. and Lesquer, A., 1982. Notice explicative de la carte géologique de l'Adrar des Iforas. Dir. Natl. Géol. Mines, Bamako, 85 pp.

Ferrara, G. and Gravelle, M., 1966. Radiometric ages from the western Ahaggar (Sahara) suggesting an eastern limit for the West African craton. Earth. Planet. Sci. Lett., 1: 319-324.

Fontes, J.C. and Gonfiantini, R., 1967. Fractionnement isotopique de l'hydrogène dans l'eau de cristallisation du gypse. C.R. Acad. Sci. Paris, 265: 4-6.

Fourcade, S. and Javoy, M., 1973. Rapport $^{18}O/^{16}O$ dans les roches du vieux socle catazonal d'In Ouzzal (Sahara Algérien). Contrib. Mineral. Petrol., 42: 235-244.

Fourcade, S. and Javoy, M., 1984. Preliminary investigations of $^{18}O/^{16}O$ and D/H compositions in rhyo-ignimbrites in the In Hihaou (In Zize) magmatic center, central Ahaggar, Algeria. Contrib. Mineral. Petrol., 89: 285-295.

Gadalia, A. and Varet, J., 1983. Les rhyolites miocènes de l'Est de l'Afar. Bull. Soc. Géol. France, 25: 139-153.

Garlick, G.D., 1966. Oxygen isotope fractionation in igneous rocks. Earth Planet. Sci. Lett., 1: 361-368.

Jacobson, R.R.E., MacLeod, W.N. and Black, R., 1958. Ring-complexes in the Younger Granite province of northern Nigeria. Mem. Geol. Soc. London, No. 1, 72 pp.

Javoy, M., 1970. Utilisation des isotopes de l'oxygène en magmatologie. Thesis, University of Paris VI, Paris, 196 pp.

Javoy, M. and Weis, D., submitted. Oxygen isotopic composition of alkaline anorogenic granites as a clue to their origin. — The problem of crustal oxygen.

Karpoff, R., 1960. La géologie de l'Adrar des Iforas. Publ. Bur. Rech. Géol. Min., Dakar, No. 30, 256 pp. (Thesis, University of Paris, 1958).

Lameyre, J., Marot, A., Zimine, S., Cantagrel, J.M., Dosso, L. and Vidal, P., 1976. Chronological evolution of the Kerguelen Islands syenite-granite ring-complex. Nature (London), 263: 306-307.

Liégeois, J.P. and Black, R., 1983. Preliminary results on the geology and geochemistry of the late Pan-African composite batholith of western Iforas (Mali). 12th Colloq. on African Geology, Mus. R. Afr. Cent., Tervuren, p. 62.

Liégeois, J.P. and Black, R., 1984. Pétrographie et géochronologie Rb-Sr de la transition calco-alcaline-alcaline fini-panafricaine dans l'Adrar des Iforas (Mali): accrétion crustale au Précambrien supérieur. In: J. Klerkx and J. Michot (Editors), African Geology. Mus. R. Afr. Cent., Tervuren, pp. 115-145.

Liégeois, J.P. and Black, R., 1986. Alkaline magmatism subsequent to collision in the Pan-African belt of the Adrar des Iforas (Mali). In: G.J. Fitton and G. Upton (Editors), The Alkaline Igneous Rocks. Geol. Soc. London, Spec. Publ., (in press).

Liégeois, J.P., Bertrand, H., Black, R., Caby, R. and Fabre, J., 1983. Permian alkaline undersaturated and carbonatite province, and rifting along the West African craton. Nature (London), 305: 42-43.

Liégeois, J.P., Bertrand, J.M. and Black, R., 1986. The subduction- and collision-related Pan-African composite batholith of the Adrar des Iforas. — A review. In: J. Kinnaird and P. Bowden (Editors), African Geology Review. Wiley, London (in press).

Manhès, G., Minster, J.F. and Allègre, C.J., 1978. Comparative uranium-thorium-lead and rubidium-strontium study of the Saint-Séverin amphoterite: consequences for early solar system chronology. Earth Planet. Sci. Lett., 39: 14-24.

Muehlenbachs, K. and Clayton, R.N., 1972. Oxygen isotope studies of fresh and weathered submarine basalts. Can. J. Earth Sci., 9: 172-184.

Nunès, P.D., Tatsumoto, M., Knight, R.M., Unruh, D.M. and Doe, B.R., 1973. U-Th-Pb systematics of some Apollo 16 lunar samples. Proc. 4th Lunar Sci. Conf., Geochim. Cosmochim. Acta, Suppl. No. 4, 2: 1797-1822.

O'Nions, R.K., Evensen, N.M. and Hamilton, P.J., 1979. Geochemical modelling of mantle differentiation and crustal growth. J. Geophys. Res., 84: 6091-6101.

Pineau, F., Javoy, M., Hawkins, J.W. and Craig, H., 1976. Oxygen isotope variations in marginal basin and ocean-ridge basalts. Earth Planet. Sci. Lett., 28: 299-307.

Shieh, Y.N. and Schwarcz, H.P., 1974. Oxygen isotope studies of granite and migmatite, Grenville Province of Ontario, Canada. Geochim. Cosmochim. Acta, 38: 21-45.

Steiger, R.H. and Jäger, E., 1977. Subcommission on Geochronology: Convention on the use of decay constants in geo- and cosmochronology. Earth Planet. Sci. Lett., 36: 359-362.

Taylor, Jr. H.P., 1968. The oxygen isotope geochemistry of igneous rocks. Contrib. Mineral. Petrol., 19: 1-71.

Taylor, Jr., H.P., 1977. Water/rock interactions and the origin of H_2O in granitic batholiths. J. Geol. Soc. London, 133: 509-558.

Taylor, Jr., H.P. and Epstein, S., 1973. Oxygen and silicon isotope ratios of the Luna 20 soil. Geochim. Cosmochim. Acta, 37: 1107-1109.

Taylor, Jr., H.P., Turi, B. and Cundari, A., 1984. $^{18}O/^{16}O$ and chemical relationships in K-rich volcanic rocks from Australia, East Africa, Antartica, and San Venanzo-Cupaello, Italy. Earth Planet. Sci. Lett., 69: 263-276.

Taylor, S.R. and McLennan, S.M., 1981. The composition and evolution of the continental crust: rare earth element evidence from sedimentary rocks. Philos. Trans. R. Soc. London, Ser. A 301: 381-399.

Upton, B.G.J., 1974. The alkaline province of South-West Greenland. In: H. Sørensen (Editor), The Alkaline Rocks. Wiley, London, pp. 221-237.

Van Breemen, O., Hutchinson, J. and Bowden, P., 1975. Age and origin of the Nigerian Mesozoic granites: a Rb-Sr isotopic study. Contrib. Mineral. Petrol., 50: 157–172.

Walker, G.P.L., 1960. Zeolite zones and dykes distribution in relation to the structure of the basalts of Eastern Iceland. J. Geol., 68: 515–528.

Weis, D. and Deutsch, S., 1984. Nd and Pb isotope evidence from the Seychelles granites and their xenoliths: mantle origin with slight upper crust interaction for alkaline anorogenic complexes. Isot. Geosci., 2: 13–36.

Weis, D., Demaiffe, D., Cauët, S. and Javoy, M., 1987. Sr, Nd, O and H isotopic ratios in Ascension Island lavas and plutonic inclusions: cogenetic origin. Earth Planet. Sci. Lett. (in press).

Weis, D., Liégeois, J.P. and Black, R., submitted. Whole rock U–Pb isochrons and deep mantle origin for a continental alkaline ring-complex (Tadhak, Mali).

Williamson, J.H., 1968. Least-square fitting of a straight line. J. Can. Phys., 46: 1845–1847.

Wones, D.R., 1974. Igneous petrology of some plutons in the northern part of the Penobscot Bay area. N.E.I.G.C. (Northeastern Inst. Geol. Chronol.), Orono, Maine, A7: 99–114.

Zartman, R.E. and Doe, B.R., 1981. Plumbotectonics — the model. Tectonophysics, 75: 135–162.