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OXYGEN-ISOTOPE GEOCHEMISTRY OF THE GRANITOID ROCKS IN THE WINNIPEG RIVER PEGMATITE DISTRICT, SOUTHEASTERN MANITOBA

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Abstract

We have measured the oxygen-isotope ratio $^{18}O/^{16}O$ in 131 samples (rocks or coexisting minerals) taken from Archean granite stocks and their host rocks in southeastern Manitoba. Four chemically similar small bodies (< 6 km^2) of late- to post-tectonic peraluminous pegmatitic granites that intrude the Bird River greenstone belt vary considerably in their $\delta^{18}O$ values: Greer Lake 8.1–8.7‰; Eaglenest Lake, 8.6–9.3‰; Tin Lake, 10.3–10.9‰; Osis Lake, 11.1–12.4‰. Coexisting minerals from all samples of granitoid rocks analyzed are near oxygen-isotope equilibrium, thus confirming that low-temperature alteration has not affected the bulk $\delta^{18}O$ values of these granites. Likewise, fractional crystallization is not responsible for the variations in $^{18}O/^{16}O$ among the pegmatitic granites; in contrast to the expected effect, the most fractionated granites are the least enriched in ^{18}O. The high $\delta^{18}O$ values (> 10‰) suggest involvement of ^{18}O-rich metasediments in the genesis of the Osis Lake and Tin Lake granites, either by anatexis or by isotopic exchange with, and(or) assimilation of, host rocks. The lower $\delta^{18}O$ values of the more fractionated but otherwise similar Greer Lake and Eaglenest Lake granites are most simply explained by isotopic exchange with metabasalt host rocks. Neither igneous nor sedimentary parentage for the granitic magmas can be ruled out by the oxygen-isotope data. Granitoid rocks from the Lac-du-Bonnet batholithic complex and the Winnipeg River batholithic belt have still lower whole-rock $\delta^{18}O$ values (7.0–8.6‰); their origin is probably due to partial melting of mafic rocks from the upper mantle or lower crust.

Keywords: peraluminous Archean granites, oxygen isotopes, isotope exchange, anatexis of metasedimentary rocks, greenstone belt, Bird River, Manitoba.

Sommaire

On a mesuré le rapport isotopique $^{18}O/^{16}O$ dans 131 échantillons (roches ou minéraux coexistants) de massifs granitiques archéens et des roches encaissantes du sud-est du Manitoba. Quatre petits massifs (< 6 km^2) de granite pegmatitique hyperalumineux, tous chimiquement semblables, tardi- ou post-tectoniques et intrusifs dans la ceinture de roches vertes Bird River, diffèrent beaucoup l'un de l'autre par la valeur du paramètre $\delta^{18}O$ (en ‰): Greer Lake 8.1–8.7, Eaglenest Lake 8.6–9.3, Tin Lake 10.3–10.9, Osis Lake 11.1–12.4. Les minéraux en présence sont presque en équilibre isotopique, preuve que l'altération à basse température n'a pas perturbé les valeurs globales de $\delta^{18}O$ dans ces granites. De même, la cristallisation fractionnée ne peut expliquer la variabilité du rapport isotopique dans les pegmatites, vu que l'effet prévu ne se produit pas: les granites les plus évolués sont en fait les plus pauvres en ^{18}O. Pour les granites à valeur élevée ($\delta^{18}O$ > 10‰), on pourrait invoquer, pour expliquer la genèse des granites Osis Lake et Tin Lake, l'intervention, par anatexie, échange isotopique ou assimilation, de roches métasédimentaires. Dans les granites plus évolués (Greer Lake, Eaglenest Lake), les valeurs plus basses de $\delta^{18}O$ refléteraient tout simplement un échange isotopique avec roches encaissantes métabasaltiques. Pour les magmas granitiques, les données isotopiques ne permettent d'exclure ni une filiation ignée, ni une filiation sédimentaire. Les roches granitoïdes du complexe batholitique Lac-du-Bonnet et de la ceinture batholitique Winnipeg River possèdent des valeurs $\delta^{18}O$ encore plus basses (7.0–8.6‰); leur origine remonterait à une fusion partielle de roches mafiques de la partie supérieure du manteau ou de la base de la croûte.

(Traduit par la Rédaction)

Mots-clés: granite archéen hyperalumineux, isotopes d'oxygène, échange isotopique, anatexie de roches sédimentaires, ceinture de roches vertes, Bird River, Manitoba.

INTRODUCTION

This paper contains the results of an oxygen-isotope study of Archean muscovite- and garnet-bearing Al-rich pegmatitic granites, associated batholithic granites and their host rocks from the Superior province of the Canadian Shield in southeastern Manitoba. These peraluminous granitoid rocks are highly fractionated and parental to one of the most diversified rare-element pegmatite districts in Canada.

The study of Silverman (1951) showed that granites ($\delta^{18}O \geqslant 7‰$) are more ^{18}O-rich than mantle rocks (5–6‰). The problem has been to determine whether this ^{18}O enrichment results from crystal fractionation or from interaction with pre-existing ^{18}O-rich supracrustal rocks. Differentiation from a basaltic parent can, at most, produce a granite with $\delta^{18}O$ of about 8‰, because the oxygen-isotope fractionation factors between silicate melts and coexisting minerals are very small at magmatic temperatures (Garlick 1966). Most Phanerozoic granites have higher $\delta^{18}O$ values (8–10‰), and muscovite-bearing, Al-rich granites have $\delta^{18}O$ values greater than 10‰ (*e.g.*, Matsuhisa *et al.* 1972, Sheppard 1977, O'Neil & Chappell 1977, O'Neil *et al.* 1977, Longstaffe *et al.* 1980b). The enrichment in ^{18}O can come from supracrustal rocks (*e.g.*, clastic sedimentary and weathered volcanic rocks); these have high $\delta^{18}O$ values, mostly because of their content of clay minerals formed at low temperatures (Savin & Epstein 1970). Most studies of ^{18}O-rich granites indicate that partial melting or assimilation of ^{18}O-rich supracrustal rocks or isotopic exchange with such rocks has occurred. For example, Turi & Taylor (1971) showed that interaction between granites and hydrothermal fluids that have equilibrated with ^{18}O-rich host rocks can cause ^{18}O enrichment of granitoid plutons, either during the magmatic stage or immediately following crystallization. The effects of such exchange are most easily recognized if the contrast between the initial $\delta^{18}O$ values of the granite and its host rock is large, and where the volume ratio of country rock to granite is high. High temperatures and hydrous mesozonal environments also favor exchange (Turi & Taylor 1971). Not surprisingly, small, fluid-rich pegmatites invariably completely exchange oxygen isotopes with their hosts (Turi & Taylor 1971, Taylor & Friedrichsen 1978, Taylor *et al.* 1979). On a larger scale, Shieh *et al.* (1976) showed that the high ^{18}O content of the Loon Lake pluton ($\simeq$50 km^2) results from isotopic exchange with ^{18}O-rich country rocks or direct mixing of granitoid magma and mobilizate derived from the country rocks. On a still larger scale, Matsuhisa *et al.* (1972) proposed that the high $\delta^{18}O$ (10–13‰) values of some Japanese batholithic granites reflect isotopic exchange between the granitic magma and ^{18}O-rich host rocks during regional metamorphism. In contrast, O'Neil & Chappell (1977) demonstrated that the high $\delta^{18}O$ values (> 10‰) of some peraluminous Australian granites require an origin by anatexis of ^{18}O-rich supracrustal parent rocks.

Archean granitoid batholiths, plutons and gneisses from the Superior province have lower average $\delta^{18}O$ values (7.5–8.0‰) than Phanerozoic granitic rocks (Longstaffe 1979, Longstaffe & Birk 1981). A logical interpretation is that the role of ^{18}O-rich supracrustal rocks in granitoid rock formation was much less important during the Archean than in post-Archean times. Granitic rocks from the Superior province are also generally very low in normative corundum and only moderately fractionated. Thus the peraluminous and highly fractionated nature of the Archean pegmatitic granites described here is rather unusual. Consequently, a study of their $\delta^{18}O$ values, and of the possible involvement of metasedimentary rocks in the formation of these granites, is of special interest.

GEOLOGICAL SETTING

The Archean Winnipeg River pegmatite district is located in southeastern Manitoba about 200 km east of Winnipeg. It occupies about 1500 km^2 within the Superior province of the Canadian Shield. Radiometric dating places the area in the Kenoran orogeny (2.7–2.5 *Ga*: Farquharson & Clark 1971, Penner & Clark 1971, Farquharson 1975). The study area can be divided broadly into four units, the Bird River greenstone belt, the Winnipeg River batholithic belt, the Lac-du-Bonnet batholithic complex and, intrusive into the greenstone belt, the pegmatitic granites (Fig. 1). Details of the geology are summarized from McRitchie (1971), Beakhouse (1977), Ermanovics *et al.* (1979), Trueman (1980) and Černý *et al.* (1980).

The metavolcanic and metasedimentary units of the Bird River greenstone belt form a broad synclinorium and generally have a subvertical attitude within the study area. Metamorphic

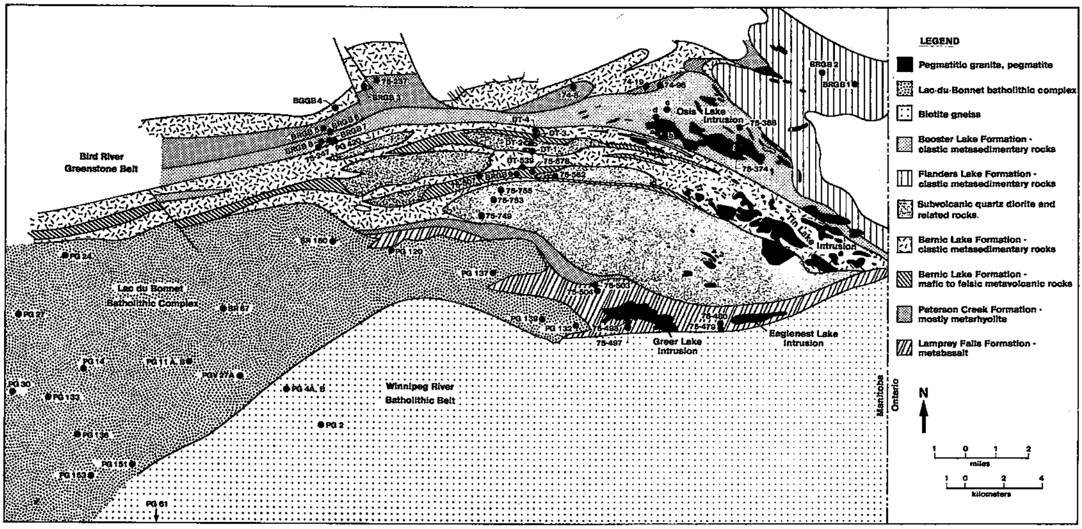

FIG. 1. Simplified geological map of the study area, after Černý *et al.* (1980). Sample locations are shown (see Tables 1, 2 and 3) except for specimens from the pegmatitic granites. In the latter case, representative samples were taken from the main body of each intrusion.

grade ranges from greenschist to amphibolite facies. The oldest unit, the Eaglenest Lake Formation (too small to show on Fig. 1), is composed of clastic metasedimentary rocks and crops out intermittently along the contact with the Winnipeg River batholithic belt. Figure 1 illustrates the lithology and stratigraphy of the remaining units.

The Lac-du-Bonnet batholithic complex is composed of muscovite–biotite leucogranite, biotite granite and localized patches of biotite–hornblende quartz diorite. Leucogranite constitutes the eastern extremity of the complex and is intruded by the biotite granite. Černý *et al.* (1980) concluded, from a study of rare-earth elements, that the quartz diorite and the biotite granite formed by partial melting of mafic rocks from the lower crust or upper mantle. The leucogranite is much more highly fractionated than the younger biotite granite; the two units cannot be related by a simple process of fractional crystallization. Černý *et al.* (1980) suggested that the leucogranite was differentiated from an acid precursor magma which crystallized at a shallow level; a similar parent material is proposed for the chemically similar metarhyolite of the Peterson Creek Formation.

Four major bodies of pegmatitic granite (Greer Lake, Eaglenest Lake, Tin Lake and Osis Lake) have been intruded along, or flank, early subvertical faults and shear zones within the Bird River greenstone belt. These plutons represent the last intrusive event in the area. Both the Greer Lake ($\simeq$2 km^2) and the Eaglenest Lake ($\simeq$1.5 km^2) plutons are hosted by Lamprey Falls metabasalt, with which they display sharp contacts. Inclusions of host material and wall-rock reaction are virtually nonexistent. The Tin Lake intrusion ($\simeq$5 km^2) is a roughly circular body divided into two lobes; the southern lobe is intruded into quartz diorite and the northern one, into clastic metasedimentary rocks of the Bernic Lake Formation. Host-rock inclusions are ubiquitous in the southern lobe. Smaller bodies of pegmatitic granite extend northwest and southeast of the main intrusion. The Osis Lake intrusion ($\simeq$6 km^2) interfingers with, includes and has partially digested its metaturbidite host (Booster Lake Formation). Wall-rock alteration is extensive; tourmalinization and the growth of albite and muscovite are ubiquitous in the host rock at the contact with granite. At depth, the Osis Lake granite may also be in contact with metarhyolite of the Peterson Creek Formation.

The intrusions are composed of pegmatitic leucogranite, fine-grained leucogranite transitional into sodic aplite, and potassic pegmatite, commonly interlayered in alternating banded sequences. The primary cause of internal diversity is probably separation of supercritical fluids from a volatile-oversaturated melt (Goad & Černý 1981, Černý *et al.* 1980). Medium grained biotite granite also is present at Osis Lake. Muscovite is present and can be abundant in these granites. Garnet is a ubiquitous accessory, especially in the sodic aplite. Tourmaline is present but only in the Osis Lake and Tin Lake bodies.

The pegmatitic granites are peraluminous (CIPW-normative corundum from 0.1 to 5.0%) and highly fractionated ($SiO_2 \simeq$ 74–78%, CaO < 0.4%, MgO < 0.2%). Of the four bodies, the Greer Lake and the Eaglenest Lake intrusions are the most fractionated. For example, they are enriched in Rb (560 ± 310 ppm, 540 ± 260 ppm, respectively) and depleted in Sr (18 ± 6 ppm, 10 ± 5 ppm) relative to the Tin Lake (Rb 245 ± 75 ppm, Sr 35 ± 14 ppm) and Osis Lake bodies (Rb 140 ± 50 ppm, Sr 26 ± 11 ppm).

Analytical Methods

The silicate samples were analyzed by the BrF_5 method of Clayton & Mayeda (1963). The data are reported in the usual δ notation with respect to standard mean ocean-water (SMOW: Craig 1961) using a CO_2–H_2O fractionation factor of 1.0407. The standard deviation calculated from the pooled residual variance of replicate analyses performed during the course of this study is ±0.09‰. The partitioning of ^{18}O between two minerals A and B is given by $1000 \ln\alpha \simeq \Delta_{AB} = \delta_A - \delta_B$, where α is the oxygen-isotope fractionation factor between A and B.

TABLE 1. $\delta^{18}O$ RESULTS, GRANITOID ROCKS FROM SOUTHEASTERN MANITOBA

Sample No.	Rock type	$\delta^{18}O$ ‰ (SMOW)
	Lac-du-Bonnet batholithic complex	
SR 67	quartz diorite (porphyroblastic)	7.8
SR 150	quartz diorite (porphyroblastic)	7.4
PG 151	quartz diorite (spotty)	8.1
PG 120	gneissic leucogranite	8.3
PG 137	gneissic leucogranite	7.8
PG 139	gneissic leucogranite	7.4
PG V27A	gneissic leucogranite	8.2
PG 132	gneissic leucogranite	7.8
PG 4A	biotite granite	7.6
PG 11A	biotite granite	8.4
PG 136	biotite granite	7.5
PG 133	biotite granite	8.3
PG 11B	biotite granite	9.6
PG 14	biotite granite	8.2
PG 24	biotite granite	8.1
PG 30	biotite granite	8.6
PG 27	biotite granite	8.2
PG 153	biotite granite	7.9
	Greer Lake intrusion	
GLW 38A	fine-grained leucogranite	8.1
GL 1002	pegmatitic leucogranite	8.2
GL 1003	pegmatitic leucogranite	8.4
GLW 6a-P	potassic pegmatite	8.7
GLW 6d-P	potassic pegmatite	8.5
GLW 6a-A	sodic aplite	8.3
GLW 6d-A	sodic aplite	8.6
	Eaglenest Lake intrusion	
GLE 6b	fine-grained leucogranite	9.3
GLE 29c	fine-grained leucogranite	9.0
ENL 1001	pegmatitic leucogranite	8.6
GLE 25c-P	potassic pegmatite	9.1
GLE 25c-A	sodic aplite	9.0
GLE 10-1	sodic aplite	8.8
GLE 20c	sodic aplite	8.8
	Tin Lake intrusion	
TNL 31	fine-grained leucogranite	10.9
TNL 27	fine-grained leucogranite	10.9
TNL 34	fine-grained leucogranite	10.9
TNL 34-HG	fine-grained leucogranite	10.7
TL 1006	pegmatitic leucogranite	10.3
	Osis Lake intrusion	
FD 8	fine-grained leucogranite	11.9
FD 10	fine-grained leucogranite	11.8
FD 23	medium-grained biotite granite	11.1
FD 23a	medium-grained biotite granite	11.1
FD 23b	medium-grained biotite granite	11.2
FD 23c	medium-grained biotite granite	11.1
OL 1004	pegmatitic leucogranite	11.7
OL 1005	pegmatitic leucogranite	11.7
OL 1005-1	pegmatitic leucogranite	11.5
FD 11	sodic aplite	11.8
FD 12	sodic aplite	12.4

Results

The whole-rock $\delta^{18}O$ values for samples of the pegmatitic granites and the Lac-du-Bonnet batholithic complex are listed in Table 1, whole-rock $\delta^{18}O$ analyses of samples from the Bird River greenstone belt and the Winnipeg River batholithic belt in Table 2 and $\delta^{18}O$ analyses of separated minerals from the granitic rocks in Table 3. Except for one anomalous value (9.6‰), the whole-rock $\delta^{18}O$ values of the Lac-du-Bonnet batholithic complex show only limited variation (7.4–8.6‰). Whole-rock $\delta^{18}O$ values for the biotite gneiss of the Winnipeg River batholithic belt are slightly lower (7.0–7.5‰; Fig. 2). The whole-rock $\delta^{18}O$ values for the pegmatitic granites range from 8.1 to 12.4‰; however, a limited range of values characterizes each intrusion regardless of rock type: Greer Lake 8.1–8.7‰, Eaglenest Lake 8.6–9.3‰, Tin Lake 10.3–10.9‰, Osis Lake 11.1–12.4‰ (Fig. 2).

Coexisting minerals from the granitic rocks are enriched in ^{18}O in the usual order: quartz > alkali feldspar > muscovite > biotite (Taylor & Epstein 1962, Taylor 1968; Fig. 3). In fact, despite the >4‰ variation in whole-rock $\delta^{18}O$, the oxygen-isotope fractionations between coexisting mineral phases are remarkably uniform from sample to sample (Δ quartz–alkali feldspar = 2.3 ± 0.4‰, 13 pairs; Δ quartz–muscovite = 3.8 ± 0.3‰, 7 pairs; Δ quartz–biotite = 6.9 ± 0.4‰, 9 pairs).

The whole-rock $\delta^{18}O$ values for samples from the Bird River greenstone belt are, for the most part, typical of their lithologies (Fig. 4). The greenschist-facies metabasalt from both the Lamprey Falls and Bernic Lake Formations has $\delta^{18}O$ values (6.7–9.2‰) higher than fresh basalt (5.5–6.0‰). Such enrichment in ^{18}O is typical of basalts modified by submarine weathering (Muehlenbachs & Clayton 1972a, b). Similar data have been reported for other Archean greenschist-facies mafic metavolcanic rocks from the Superior province (Longstaffe *et al.* 1977,

TABLE 2. $\delta^{18}O$ ROCK RESULTS, BIRD RIVER GREENSTONE BELT AND WINNIPEG RIVER BATHOLITHIC BELT

Sample No.	Rock Type	$\delta^{18}O$ ‰ (SMOW)
	Winnipeg River batholithic belt	
PG 61	biotite gneiss, 5 km southeast of Lac-du-Bonnet plutonic complex	7.5
PG 2	biotite gneiss, 2 km southeast of Lac-du-Bonnet plutonic complex	7.5
PG 4b	biotite gneiss, <0.5 km southeast of Lac-du-Bonnet plutonic complex (transitional zone)	7.0
	Bird River greenstone belt	
75-498	metabasalt, Lamprey Falls Formation, <0.5 km south of Greer Lake intrusion	7.5
75-497	metabasalt, Lamprey Falls Formation, as above	6.7
75-504	metabasalt, Lamprey Falls Formation, >1 km northwest of Greer Lake intrusion	8.9
75-503	metabasalt, Lamprey Falls Formation, as above	8.6
75-480	metabasalt, Lamprey Falls Formation, <0.5 km west of Eaglenest Lake intrusion	7.5
75-479	metabasalt, Lamprey Falls Formation, as above	8.2
BRGB 9	metabasalt, Bernic Lake Formation, >10 km northwest of Greer Lake and Eaglenest Lake intrusions	9.2
75-507	metabasalt, Bernic Lake Formation, as above	9.0
75-578	metabasalt, Bernic Lake Formation, as above	8.7
DT 1	metabasalt, Bernic Lake Formation, as above	8.0
BRGB 5	metarhyolite flow, Peterson Creek Formation, ≈15 km west of Osis Lake intrusion	15.8
BRGB 6	metarhyolite tuff, Peterson Creek Formation, as above	14.8
75-237	metarhyolite tuff, Peterson Creek Formation, as above	9.7
74-8	metarhyolite flow, Peterson Creek Formation, ≈5 km northwest of Osis Lake intrusion	11.6
74-19	metarhyolite flow, Peterson Creek Formation, ≈2 km northwest of Osis Lake intrusion	14.7
74-96	metarhyolite tuff, Peterson Creek Formation, as above	15.1
75-749	subvolcanic quartz diorite intrusion, ≈12 km west of Tin Lake intrusion	7.2
75-755	subvolcanic quartz diorite intrusion, as above	7.5
75-753	subvolcanic quartz diorite intrusion, as above	7.0
BRGB 3	metaconglomerate, Bernic Lake Formation, ≈15 km west of Osis Lake intrusion	9.3
BRGB 4	metaconglomerate, Bernic Lake Formation, as above	10.8
DT 539	volcanoclastic metasandstone, Bernic Lake Formation, ≈10 km west of Tin Lake intrusion	10.6
75-562	volcanoclastic metasandstone, Bernic Lake Formation, as above	11.1
DT 2	volcanoclastic metasandstone, Bernic Lake Formation, as above	10.9
DT 3	volcanoclastic metasandstone, Bernic Lake Formation, as above	10.1
DT 4	volcanoclastic metasandstone, Bernic Lake Formation, as above	12.0
BRGB 1	metasandstone (lithic arenite), Flanders Lake Formation, ≈4 km northeast of Osis Lake intrusion	7.7
BRGB 2	metasandstone (lithic arenite), Flanders Lake Formation, as above	9.9
BRGB 7	metasandstone (greywacke), Booster Lake Formation, ≈15 km west of Osis Lake intrusion	10.1
BRGB 8	metasandstone (greywacke), Booster Lake Formation, as above	10.7
c	metasandstone (greywacke), Booster Lake Formation, ≈1 km from Osis Lake intrusion	10.9
b	metasandstone (greywacke), Booster Lake Formation, <0.5 km from Osis Lake intrusion	13.0
a	metasandstone (greywacke), Booster Lake Formation, as above	11.4
d	metasandstone (greywacke), Booster Lake Formation, as above	11.1
75-247	metapelite, Booster Lake Formation, ≈15 km from Osis Lake intrusion	13.1
PG 420	metapelite, Booster Lake Formation, as above	9.8
75-368	metapelite, Booster Lake Formation, <0.5 km from Osis Lake intrusion	11.2
75-374	metapelite, Booster Lake Formation, <0.2 km from Osis Lake intrusion	12.1

TABLE 3. $\delta^{18}O$ MINERAL VALUES FOR GRANITOID ROCKS FROM SOUTHEASTERN MANITOBA

	$\delta^{18}O$ ‰ (SMOW)			
Sample No. and rock type	Quartz	Alkali feldspar	Muscovite	Biotite
	Lac-du-Bonnet batholithic complex			
Sr 67, quartz diorite (porphyroblastic)	10.2	8.0		3.9
PG 151, quartz diorite (spotty)	9.3	7.0		3.0
PG 120, gneissic leucogranite	9.2	7.5		
PG 27, biotite granite	9.6	7.5		2.8
	Greer Lake intrusion			
GLW 38A, fine-grained leucogranite	10.2	7.5	6.4	3.4
GL 1002, pegmatitic leucogranite	9.8	7.4	6.4	
	Eaglenest Lake intrusion			
GLE 6b, fine-grained leucogranite	10.6	7.9	6.3	
ENL 1001, pegmatitic leucogranite	10.4	7.7	6.5	3.3
	Tin Lake intrusion			
TNL 34, fine-grained leucogranite	12.1	10.1	8.4	5.3
TL 1006, pegmatitic leucogranite	12.0	9.7		4.9
	Osis Lake intrusion			
FD 10, fine-grained leucogranite	13.4	11.0	9.6	5.8
OL 1004, pegmatitic leucogranite	13.0	10.7		
FD 23b, medium-grained biotite granite	13.0	11.4	9.5	5.9

1980a, Beaty & Taylor 1979). The whole-rock $\delta^{18}O$ of the subvolcanic quartz diorite (7.0–7.5‰) is similarly not unusual (Taylor 1968, Longstaffe 1979). Except for one low value (7.7‰), the whole-rock $\delta^{18}O$ values of the clastic metasedimentary rock (9.3–13.1‰) are typical of greywacke-type sediments that have experienced ^{18}O enrichment during sedimentary processes (Magaritz & Taylor 1976, Longstaffe & Schwarcz 1977, Longstaffe *et al.* 1980b). In contrast, the high whole-rock $\delta^{18}O$ values of the metarhyolite (9.7–15.8‰) are quite unusual; normal values for such rocks range from 6 to 10‰ (Taylor 1968).

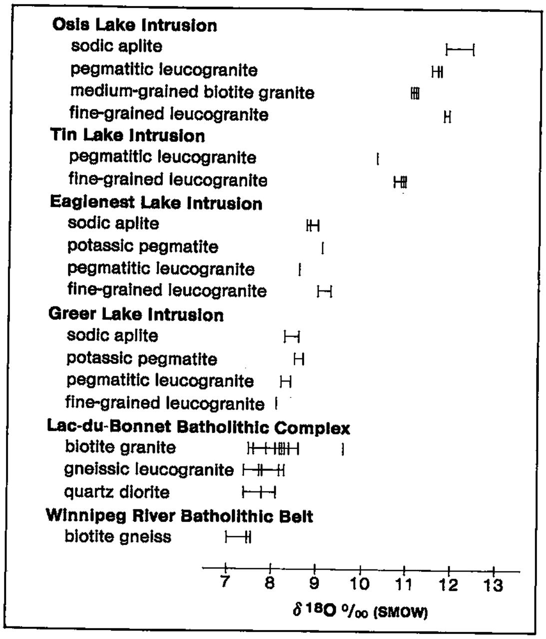

FIG. 2. Whole-rock $\delta^{18}O$ values for the granitoid rocks from southeastern Manitoba.

DISCUSSION

The most important observations of this study are the differences in whole-rock $\delta^{18}O$ among the granitic rocks and, in particular, the ^{18}O-rich character of the Tin Lake and Osis Lake intrusions. The 7-to-9‰ $\delta^{18}O$ values of the Winnipeg River biotite gneiss, the Lac-du-Bonnet batholithic complex and the Greer Lake and Eaglenest Lake pegmatitic granites lie within the range hitherto reported for granitoid rocks from the Superior province (Longstaffe & Schwarcz 1977, Shieh & Schwarcz 1977, Longstaffe 1979). In contrast, the Tin Lake and Osis Lake intrusions ($\delta^{18}O$ = 10.3–12.4‰) are the most ^{18}O-rich granitoid rocks yet reported from the Superior province, except for one 11.2‰ granite near Red Lake, Ontario (Taylor 1968).

It is not difficult to show that the $\delta^{18}O$ values of the granitic rocks reflect ^{18}O contents attained during magmatic processes, rather than during secondary, low-temperature alteration. Secondary alteration can drastically raise or lower the ^{18}O contents of igneous rocks. Such alteration is easily recognized, since it causes reversals in the order that minerals fractionate ^{18}O during igneous processes. For example, depending upon the temperature of meteoric water–rock interaction, feldspars can attain $\delta^{18}O$ values lower than coexisting mafic minerals (Taylor & Forester 1971) or higher than coexisting quartz (Wenner & Taylor 1976). Isotopic reversals were *not* found in the analyzed Manitoban granitic rocks (Fig. 3). Instead, the fractionations of ^{18}O between coexisting minerals are normal and remarkably uniform from specimen to specimen despite variations in whole-rock $\delta^{18}O$ values. These fractionations correspond to average temperatures of 380 ± 45°C (Δ quartz–feldspar), 435 ± 30°C (Δ quartz–muscovite) and 430 ± 20°C (Δ quartz–biotite), based on the equations of Bottinga & Javoy (1975). Such concordant temperature estimates would be extremely unlikely for oxygen-isotope systems disturbed by low-temperature alteration. That the temperatures are subsolidus is not troublesome; retrograde re-equilibration of oxygen isotopes characteristically occurs during the cooling of hydrous plutonic rocks (Javoy 1977, Deines 1977).

The generally low average $\delta^{18}O$ values of the Winnipeg River biotite gneiss (7.3‰) and the Lac-du-Bonnet batholithic complex (8.0‰) preclude extensive interaction with ^{18}O-rich supracrustal rocks. Instead, these rocks are probably juvenile additions to the continental crust; the very low initial Sr-isotope ratio of the Lac-du-Bonnet biotite granite (0.7001 ± 0.0014, Farquharson 1975) supports this model. This interpretation is also favored for the leucogranite of the Lac-du-Bonnet complex. However, the possibility of isotopic exchange with the large volume of biotite granite that intrudes it cannot be completely eliminated.

The variation in whole-rock $\delta^{18}O$ values among the pegmatitic granites can be explained by two models: (A) differentiation from a juvenile (igneous) source followed by isotopic exchange with (or assimilation of) greenstone-belt host rocks, or (B) anatexis of metasedimentary rocks of the greenstone belt followed by isotopic exchange with (or assimilation of) host rocks during subsequent differentiation. A third possibility, isotopic variation due to fractional crystallization, can be eliminated, as the relatively ^{18}O-poor granites (<9.5‰) are more fractionated than the ^{18}O-rich granites. This is opposite to the variation in $^{18}O/^{16}O$ ratios expected for progressive fractionation of residual magmas.

Model A

Conditions are favorable for ^{18}O enrichment of the Tin Lake – Osis Lake intrusions by large-scale isotopic exchange with their ^{18}O-rich clastic

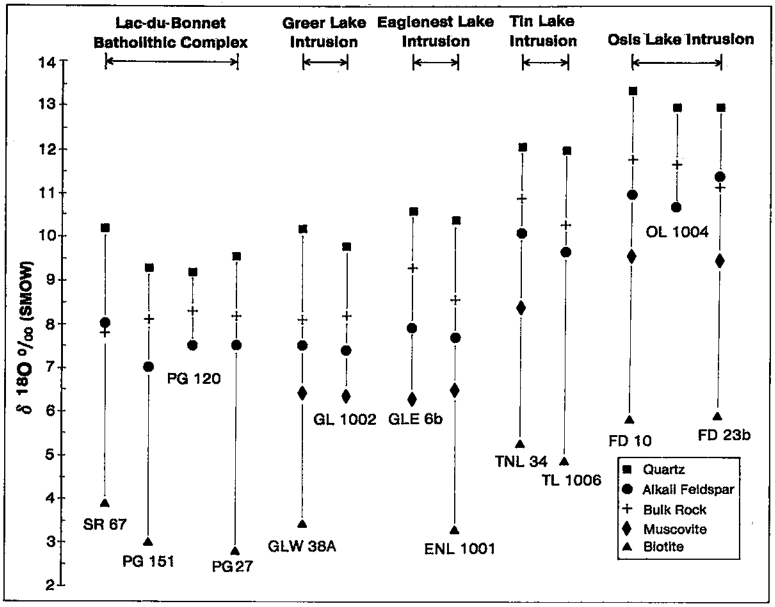

FIG. 3. $\delta^{18}O$ mineral results for the granitoid rocks from southeastern Manitoba.

metasedimentary host rocks: (1) The $^{18}O/^{16}O$ ratios of the putative granitic magmas and the country rocks (9.3–13.1‰) differed by up to 5‰ if the magmas initially had $\delta^{18}O$ values typical of "normal" granites derived from igneous sources (~7–8‰). (2) The volumes of the intrusions are very small compared with those of the host rocks (Fig. 1). (3) The late- to post-tectonic environment was probably still quite hot. (4) Abundant volatiles were associated with the intrusions.

In addition, bulk assimilation of clastic metasedimentary rocks would contribute to the ^{18}O-rich character of the Tin Lake and Osis Lake plutons, and would also partly account for their peraluminous chemistry. Both bodies are tourmaline-bearing, and the latter contains numerous partly digested metasedimentary inclusions. Thus, the Osis Lake pluton is probably most affected by sediment assimilation; notably, it is about 1‰ richer in ^{18}O than the Tin Lake body (Table 1). The Osis Lake body may also have assimilated and exchanged with chemically similar but much more ^{18}O-rich Peterson Creek metarhyolite at depth (Figs. 1, 4).

The applicability of model A to the pegmatitic granites at Greer Lake ($\delta^{18}O$ = 8.1–8.7‰) and Eaglenest Lake ($\delta^{18}O$ = 8.6–9.3‰) is much more equivocal. The $\delta^{18}O$ values of these rocks could reflect a dominantly igneous parentage. However, both intrusions are more ^{18}O-rich than other juvenile granitoid rocks from the Superior province (average $\delta^{18}O$ = 7.5–8.0‰: Longstaffe 1979). This enrichment could reflect derivation from sialic crust rather than from mafic rocks of the lower crust or upper mantle. Enrichment in ^{18}O due to exchange with the metabasaltic host rocks ($\delta^{18}O$ = 6.7–8.9‰) is also possible. Physical interaction between the country rocks and the granites, however, appears to have been limited, as contacts are sharp and host-rock inclusions virtually absent.

The peraluminous chemistry of the plutons lower in ^{18}O is difficult to explain solely as a function of igneous processes. However, at least the Eaglenest pluton may have assimilated clastic sedimentary material at depth; the Eaglenest Lake Formation crops out less than 1 km southeast of the intrusion, and may be in contact with it at depth. Such assimilation (or exchange) might explain why the Eaglenest Lake granite is about 0.5‰ richer in ^{18}O than its counterpart at Greer Lake (Fig. 4, Table 1).

The mechanism of the postulated isotopic exchange between the granites and their host rocks is difficult to define uniquely. Advecting currents within the small magma-chambers could assist isotopic equilibration of the granitic liquids with the enclosing country rocks. Isotopic exchange would be facilitated by reaction with country-rock inclusions such as are ubiquitous in the Osis Lake body. Exchange with upward-

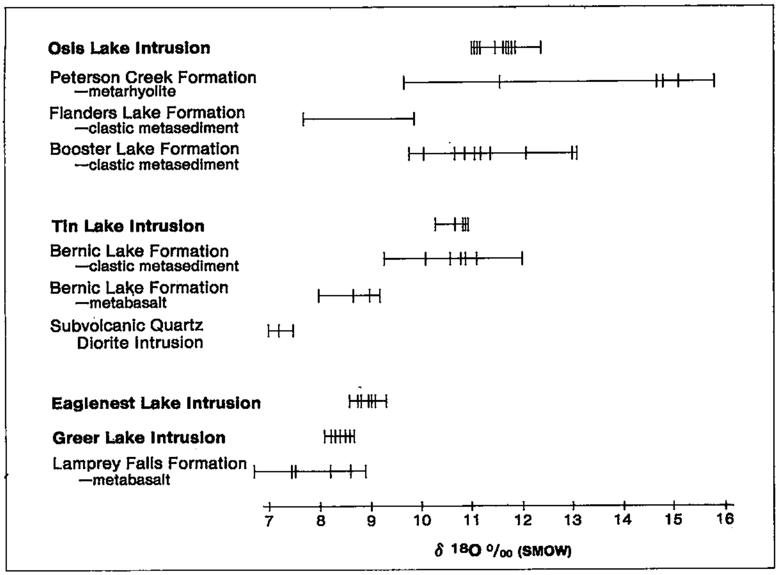

FIG. 4. Whole-rock $\delta^{18}O$ values for the formations of the Bird River greenstone belt. Whole-rock $\delta^{18}O$ values for the pegmatitic granites are also shown together with the formation(s) that they intrude.

circulating ^{18}O-rich metamorphic fluids immediately following crystallization is also possible (Turi & Taylor 1971, Taylor 1978), especially given the subsolidus temperatures calculated for these granites from mineral fractionations. The major problem with subsolidus exchange, and perhaps with the exchange model in general, is that the ^{18}O contents of the country rocks close to the intrusions appear to be unchanged. For example, the whole-rock $\delta^{18}O$ values of samples from the Booster Lake Formation collected near to and far from the Osis Lake pegmatitic granite (Fig. 1, Table 2) do not differ greatly.

Model B

An origin for the pegmatitic granites by anatexis of greenstone-belt clastic metasedimentary material is immediately suggested by the peraluminous, muscovite- and garnet-rich nature of all four plutons. O'Neil & Chappell (1977) showed that such rocks have unusually high $\delta^{18}O$ values ($>10‰$). Thus, this model is particularly appealing for the Tin Lake and Osis Lake plutons ($\delta^{18}O$ = 10.3–12.4‰). Their clastic metasedimentary host rocks have $\delta^{18}O$ values of 9.3–13.1% (Table 2); granitic magmas produced at depth from the subvertical extension of these host rocks would have the requisite peraluminous chemistry and high whole-rock $\delta^{18}O$ values. Subsequent assimilation of, and exchange with, ^{18}O-rich host rocks at higher levels could still occur, as described in model A, but in this case with an already ^{18}O-rich magma. In this scheme, the absence of ^{18}O depletion in the country rocks about the plutons presents no problem.

The lower $\delta^{18}O$ values of the Greer Lake and Eaglenest Lake plutons can be interpreted according to model B in one of two ways: (1) The granitic magmas were generated at depth from ^{18}O-rich clastic metasediments, but subsequently exchanged ^{18}O with their metabasaltic host-rocks at a higher tectonic level. (2) The clastic metasedimentary parent rocks became depleted in ^{18}O prior to, or during, anatexis. Fourcade & Javoy (1973), Shieh & Schwarcz (1974) and Longstaffe (1979) have shown that some mesozonal to catazonal Precambrian clastic metasedimentary rocks are depleted in ^{18}O, probably by exchange with a mafic reservoir, during high-grade metamorphism.

CONCLUSIONS

The ^{18}O-rich character of the small, late- to post-tectonic peraluminous Tin Lake and Osis Lake plutons ($\delta^{18}O$ = 10.3–12.4‰) was acquired either by anatexis of ^{18}O-rich greenstone-belt metasedimentary rocks or by isotopic exchange with, and assimilation of, host-rock metasediments. In either case, the involvement of metasedimentary material in the genesis of these pegmatitic granites has been established. Whether the parental granitic melts were of juvenile or supracrustal origin remains uncertain.

The $\delta^{18}O$ values (8.1–9.3‰) of the more fractionated but otherwise similar Greer Lake and Eaglenest Lake pegmatitic granites are higher than those typical for juvenile granitoid rocks of the Superior province but lower than expected for granites derived from ^{18}O-rich metasedimentary rocks. This ambiguity in the data cannot be uniquely resolved. The simplest explanation is isotopic exchange between these small plutons and their metabasalt host-rocks. Thus, neither igneous nor sedimentary origins for the parental granitic melts can be eliminated.

The still lower average $\delta^{18}O$ values of the nearby but older Lac-du-Bonnet batholithic complex ($\simeq$8.0‰) and the biotite gneiss of the Winnipeg River batholithic belt ($\simeq$7.3‰) suggest that interaction with sialic crust has been minimal and that, for the most part, these rocks constitute new additions to the continental crust.

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