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AN INTEGRATED FLUID–MINERAL STABLE-ISOTOPE STUDY OF THE GRANITE-HOSTED MINERAL DEPOSITS OF THE NEW ROSS AREA, SOUTH MOUNTAIN BATHOLITH, NOVA SCOTIA, CANADA: EVIDENCE FOR MULTIPLE RESERVOIRS

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Abstract

Granitic rocks in the New Ross area, Nova Scotia, host polymetallic (Cu, Mo, Mn, Sn, U, W), aplite–pegmatite, greisen, and vein-type mineral deposits. The host granitic rocks belong to the Late Devonian (380 Ma) peraluminous South Mountain Batholith (SMB), which consists of several coalesced plutons. Among these, the New Ross Pluton, composed of monzogranite, leucomonzogranite, and leucogranite, is one of the youngest. White mica separates from aplite–pegmatite, greisen, and veins have $\delta^{18}O$ values between +4.0 and +10.0‰, and δD values between –42 and –108‰. Fluid-inclusion extracts from quartz in samples from the area of the same deposits have δD values between –42 and –97‰. The isotopic composition of fluids in equilibrium with the white mica and isotopic data for the fluid-inclusion extracts record the transition from an early, orthomagmatic stage, in which magmatic fluid dominated (*i.e.*, aplites, pegmatites), to a subsequent hydrothermal stage (*i.e.*, greisens, veins) where an increasing amount of another fluid, inferred to be meteoric water, infiltrated the systems and mixed with the magmatic fluid. In addition, the low δD values for fluid-inclusion extracts compared with the δD_{fluid} values calculated for coexisting white mica samples are considered to record infiltration of a second meteoric water much later in the evolution of the SMB. Integration of the new stable isotope data with our previous results on fluid inclusions suggests the following history of fluid evolution within the New Ross area: (1) exsolution of a magmatic fluid at ~600°C, (2) incursion of a low-$\delta^{18}O$, high-δD meteoric fluid at *ca.* 380 Ma through faults, and subsequent mixing with magmatic and metamorphic fluids through circulation by convection before cooling to ~400–500°C, and (3) incursion at a much later time of a low-δD meteoric fluid, mostly affecting the composition of fluid-inclusion extracts.

Keywords: stable isotopes, granite, polymetallic mineral deposits, meteoric water, New Ross pluton, South Mountain Batholith, Nova Scotia.

Sommaire

Les granites de la région de New Ross en Nouvelle-Écosse comprennent plusieurs zones minéralisées à caractère polymétallique (Cu, Mo, Mn, Sn, U, W) sous forme d'aplite et de pegmatite, de greisen et de veines. Les roches hôtes font partie du batholite de South Mountain, d'âge dévonien supérieur (~380 Ma), fait d'un ensemble de plutons distincts, tous hyperalumineux. Parmi ceux-ci, le pluton de New Ross appartient aux derniers stades d'intrusion du granite et se compose de monzogranite, leucomonzogranite et leucogranite. De nouvelles données sur les isotopes stables (δD, $\delta^{18}O$), obtenues à partir de micas blancs séparés des roches, donnent des valeurs entre +4 et +10.0‰ ($\delta^{18}O$), et entre –42 et –108‰ (δD), alors que les extraits d'inclusions fluides (piégées dans le quartz) ont des valeurs de δD comprises entre –42 et –97‰. La composition

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isotopique du fluide en équilibre avec les micas blancs, ainsi que celle des extraits d'inclusions fluides, témoignent de la transition d'un régime magmatique à un régime hydrothermal, avec l'infiltration progressive d'une phase fluide d'origine météorique durant l'emplacement et la cristallisation du batholite de South Mountain et durant la formation des veines et des greisens. L'infiltration de fluides météoriques après la cristallisation de la batholite explique l'abaissement des valeurs de δD_{eau} dans les inclusions fluides en comparaison des valeurs de δD_{eau} calculées à partir du mica blanc coexistant dans le même échantillon. En intégrant les données isotopiques et les résultats d'une étude antérieure des inclusions fluides, l'évolution des fluides d'un régime magmatique à un régime hydrothermal dans la région de New Ross se serait déroulée de la façon suivante: (1) un fluide magmatique se sépare du magma à une température d'environ 600°C. (2) Un fluide d'origine météorique avec une valeur de $\delta^{18}O$ plus faible et une valeur de δD plus élevée que celles du fluide magmatique s'infiltre le long de larges failles rejoignant la surface et se mélange avec le fluide magmatique, qui se mélange à son tour avec un fluide métamorphique issu des sédiments métamorphisés du supergroupe de Meguma, grâce à sa circulation par l'intermédiaire de cellules de convection, avant de se refroidir jusqu'à 400°–500°C. (3) Une incursion tardive d'un fluide d'origine météorique ayant une composition de δD plus faible que précédemment affecte principalement la composition des extraits d'inclusions fluides.

Mots-clés: isotopes stables, granite, minéralisation polymétallique, fluide météorique, batholite de South Mountain, pluton de New Ross, Nouvelle-Écosse.

INTRODUCTION

In an effort to investigate the nature and origin of the fluids involved in the granite-hosted mineralization of the New Ross area (~500 km^2) of the South Mountain Batholith (SMB), Nova Scotia (Fig. 1a), we have coupled stable isotopic analyses (δD, $\delta^{18}O$) of white mica and fluid-inclusion extracts with results of a previous fluid-inclusion study (Carruzzo *et al*. 2000). The granite-hosted polymetallic (Cu, Mo, Mn, Sn, U, W) mineralization of the study area shows a large variety of silicate, phosphate, oxide, and sulfide minerals, with each deposit displaying unique assemblages (O'Reilly *et al.* 1982). In an earlier fluid-inclusion study, we demonstrated that at least three different fluid reservoirs (magmatic, metamorphic, meteoric) were involved in the mineralized environments (Carruzzo *et al.* 2000). In addition, mineralization and crystallization of the SMB have been shown to be coincident based on $^{40}Ar/^{39}Ar$ laser-probe dating of white mica and Re/Os dating of molybdenite (Carruzzo *et al.* 2004). In this paper, we present stable-isotope data pertaining to the white mica (δD, $\delta^{18}O$) and quartz-hosted fluid-inclusion extracts (δD) from eight mineral occurrences in the New Ross area of the SMB.

The objective of this study is to characterize the isotopic composition of fluids involved in mineralization of a variety of environments in the New Ross area of the SMB (Fig. 1). Integration of the fluid-inclusion and isotopic data indicate that variable proportions of both magmatic and non-magmatic fluids (*i.e.,* metamorphic, meteoric) were involved in the mineralizing systems, but the non-magmatic component is partly masked by the intrinsically low fluid:rock ratios that characterize the systems.

REGIONAL GEOLOGICAL SETTING

The Meguma Lithotectonic Terrane (MLT), the easternmost terrane of the northern Appalachians, is separated from the Avalon Terrane by the Cobequid–Chedabucto Fault (Fig. 1a; Williams & Hatcher 1983). Two distinct lithologies dominate the bedrock geology of the MLT: (1) Cambrian–Ordovician metawackes and metapelites of the Goldenville and Halifax groups, respectively, of the Meguma Supergroup (Schenk 1995), which underwent regional deformation and greenschist- to amphibolite-grade metamorphism during the ~390 Ma Acadian Orogeny (Hicks *et al.* 1999), and (2) approximately twenty-five Late Devonian peraluminous granitic plutons, including the SMB batholith, which intruded the Meguma Supergroup (Clarke *et al*. 1997). MacDonald *et al.* (1992) subdivided the SMB into early Stage-1 plutons (granodiorite, biotite monzogranite), and more evolved Stage-2 plutons (leucomonzogranite, leucogranite) (Fig. 1a).

There are several mineral occurrences in the SMB, and those central to this study include six locations in a Stage-2 pluton (New Ross pluton, NRP) and two locations in a Stage-1 pluton (Salmontail Lake pluton, SLP) (Fig. 1b). The NRP contains aplite–pegmatite-type deposits (Reeves, Keddy, Morley's, Walker, Long Lake) and greisen-type deposits (Turner, Walker, Long Lake), whereas the SLP hosts vein-type deposits (Millet Brook, Mn Mines) (Fig. 1b). MacDonald (2001) gave a detailed history of exploration for base metals and lithophile elements in the SMB, including the New Ross area. Producing deposits in the past include East Kemptville (Sn–Zn–Cu–Ag), located in the southwestern part of the SMB (Kontak 1994), and the Mn Mines in the New Ross area (O'Reilly 1992), whereas uranium occurs in the undeveloped Millet Brook deposit (Chatterjee & Strong 1984). Table 1 contains information about the host granitic rock, mineralization, alteration, and samples from each locality used in this study, whereas Carruzzo *et al.* (2000) described the geology and mineralogy of each mineral occurrence.

Radiometric age determinations (U–Pb, Re–Os, $^{40}Ar/^{39}Ar$) indicate that the SMB intruded and crystallized at *ca.* 380 Ma, with a protracted cooling history to 370 Ma (Reynolds *et al.* 1981, 1987, Harper 1988, Clarke *et al.* 1993, Kontak *et al.* 2003, Carruzzo *et al.*

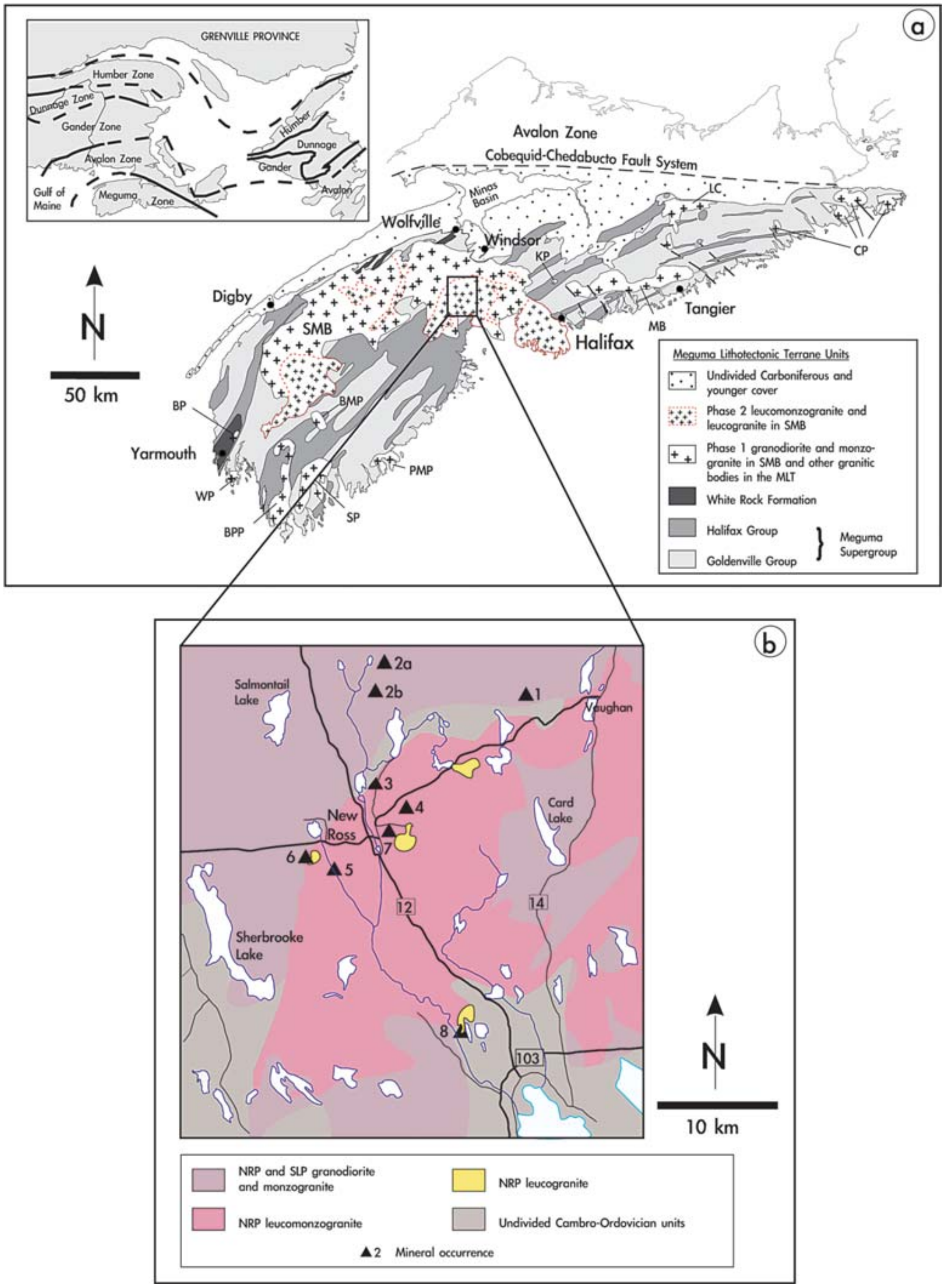

FIG. 1. (a) Simplified geological map of the Meguma Lithotectonic Terrane (MLT) of the Canadian Appalachians, with inset map of the terranes of the Canadian Appalachians. Abbreviations: SMB: South Mountain Batholith, BMP: Bald Mountain pluton, WP: Wedgeport pluton, BPP: Barrington Passage pluton, PMP: Port Mouton pluton, BP: Brenton pluton, KP: Kinsac pluton, MB: Musquodoboit Batholith, CP: Canso pluton, LC: Liscomb Complex. (b) Detailed geological map of the New Ross area with the location of the mineral occurrences: (1) Millet Brook, (2a) Dean and Chapter mine (part of the Mn mines), (2b) Cain and Riddle mine (part of the Mn mines), (3) Turner, (4) Walker, (5) Keddy, (6) Reeves, (7) Morley's, and (8) Long Lake.

2004). Younger ages reported for the granites are the consequence of post-crystallization tectonothermal activity (Reynolds *et al.* 1981, 1987, Kontak & Cormier 1991).

The Lower Devonian (Lochkovian to Lower Emsian) Torbrook Formation is the youngest formation intruded by the SMB. Clastic sedimentary rocks and marine carbonates of the Horton and Windsor groups, respectively, unconformably overlie the SMB. Miospores of Late Devonian age (Late Famennian) in the basal part of the Horton Group provide a minimum age of *ca.* 360 Ma [time scale of Tucker *et al.* (1998)] for exhumation of the northeastern part of the SMB (Martel *et al.* 1993). Thus emplacement, crystallization, uplift, and partial erosion of the SMB occurred after the Lower Emsian and before the Late Famennian.

On the basis of a study of the contact-metamorphic aureole of the SMB, Raeside & Mahoney (1996) obtained pressures of 3.2 to 3.8 kbars for batholith emplacement, corresponding to a depth of emplacement of 10–12 km. Such a pressure is in agreement with those inferred from mineral equilibria (Clarke *et al.* 1976, Ham & Kontak 1988, Halter & Williams-Jones 1999).

TABLE 1. DESCRIPTION OF HOST GRANITIC ROCK, MINERALIZATION*, ALTERATION, AND SAMPLES CHOSEN FOR ISOTOPIC ANALYSES FROM EIGHT MINERAL DEPOSITS IN THE NEW ROSS AREA, NOVA SCOTIA

Deposit (type)	Host granitic rock	Metal association, mineralization	Type of alteration	"Isotope samples"	Ref.
1. Reeves (pegmatite)	Keddy-Reeves leucogranite (intrudes the Salmontail Lake monzogranite to the north)	**Sn**: cassiterite; F: fluorite; W: ferberite, scheelite, tungstite; Li: lepidolite, Nb, Ta: columbite		SC-6-R: host rock U2: quartz vein SC-10-R: pegmatite	1, 2, 3
2. Morley's (pegmatite)	New Ross leuco-monzogranite	**W**: scheelite, ferberite; Sn: cassiterite; F: fluorite; Na, Ta: columbite	hematization brecciation	SC-1-M: host rock SC-7-M: pegmatite	1, 2, 3
3. Keddy (pegmatite, greisen)	New Ross leuco-monzogranite	**Mo**: molybdenite; W: ferberite, scheelite; F: fluorite; Nb, Ta: columbite	greisenization (qtz, ms, toz)	SC-2-K: host rock SC-3-K: pegmatite SC-5-K: greisen	1, 2, 3
4. Long Lake (pegmatite, aplite, greisen)	Long Lake leucogranite	**Mo**: molybdenite; Cu: chalcopyrite; W: ferberite, scheelite; F: fluorite	greisenization (qtz, ms, chl)	SC-1-L: host rock SC-12-L: pegmatite SC-13-L: banded aplite SC-5-L: greisen	2, 3
5. Walker (pegmatite, aplite, greisen)	Panuke Lake leuco-monzogranite	**Mo**: molybdenite; Cu: chalcopyrite, bornite; F: fluorite; Zn: gahnite; W: ferberite; Sn: cassiterite	greisenization (qtz, ms, bt, chl, tur)	SC-13-W: host rock SC-9-W: pegmatite SC-2-W: aplite SC-7/14-W: greisen	1, 2, 3
6. Turner (greisen)	Panuke Lake leuco-monzogranite	**Sn**: cassiterite; Cu: chalcopyrite, chalcocite, covellite; As: arsenopyrite; Zn: sphalerite; U: autunite	greisenization (qtz, ms, chl) chloritization hematization	SC-9-T: host rock SC-2-T: greisen SC-14-T: elvan dyke	1, 2, 3, 4
7. Millet Brook (vein)	Salmontail Lake monzogranite and biotite granodiorite	**U**: uraninite, autunite, torbernite; Cu: chalcopyrite, bornite, covellite, chalcocite; Pb: galena; Zn: sphalerite W: ferberite	hematization carbonitization albitization biotitization	SC-5/6-MB: host rock, SC-3-MB: U-irradiated samples (smoky quartz)	3, 5, 6
8. Mn Mines (vein)	Salmontail Lake monzogranite	**Mn**: pyrolusite, manganite, massive Mn oxides, rhodochrosite; Fe: hematite, goethite, hydrous Fe oxides	episyenitization hematization	SC-3-D: host rock SC-1-D: episyenite DC 86-3 847: hematized host-rock	2, 3, 7

* The main metal is shown in bold. References: 1: Charest (1976), 2: O'Reilly *et al.* (1982), 3: Logothetis (1985), 4: Farley (1978), 5: Chatterjee *et al.* (1982), 6: Chatterjee & Strong (1984), 7: O'Reilly (1992). The use of ferberite is not intended to exclude the presence of hübnerite.

METHODOLOGY

Sampling, preparation of material, and analytical techniques

Samples were collected from eight mineral deposits in the New Ross area, whereas samples for the Millet Brook and Mn Mines deposits (except SC–1/3–D) come from drill core stored by the Nova Scotia Department of Natural Resources. In all cases, we selected samples of mineralized zones and their host granitic rock.

Quartz and white mica grains (1–3 mm diameter) were hand-picked from crushed and sieved material under a binocular microscope. For finer-grained samples (*e.g.,* aplite), conventional magnetic and heavy-liquid techniques were used to prepare separates.

White mica grains from host granitic rocks and pegmatite–aplite (with the exception of a pegmatite at the Reeves deposit) reveal chemical and textural evidence of crystallization at the magmatic stage of the SMB evolution, whereas grains of white mica from greisen represent either remnant grains of the magmatic stage, or newly crystallized grains formed during the subsequent hydrothermal stage (Carruzzo 2003). Grains of white mica from a pegmatite at the Reeves deposit are texturally stellate aggregates, and have a chemical enrichment in Fe + Mg + Mn similar to white mica grains from greisens. The origin of white mica grains from vein deposits is ambiguous, because of their presence in altered samples, but yet they display a chemical composition similar to magmatic white mica (*i.e.,* limited tschermakitic substitution and solid-solution toward biotite).

Twenty-six samples of white mica were analyzed for their $\delta^{18}O$ and δD values at the Isotope Laboratory of the Department of Geological Sciences and Geological Engineering at Queen's University (Kingston, Ontario). Twenty-two samples of fluid-inclusion extracts from quartz grains were also analyzed for their δD values at the same laboratory. These extracts represent sampling of the fluids during thermal decrepitation *in vacuo* and obviously may represent a mixture of fluid-inclusion types. In Table 2, we summarize the δD and $\delta^{18}O$ data for white mica and δD data for the fluid-inclusion extracts. The data include analytical results for all three types of deposits (aplite–pegmatite, greisen, and vein) from the New Ross area and their various host-rocks (*i.e.,* granodiorite, monzogranite, leucomonzogranite, leucogranite). The analytical protocol follows the methods and procedures given in Kyser *et al.* (1998).

We use the standard $\delta^{18}O$ and δD notation in per mil (‰) relative to the reference V–SMOW. Reproducibility of the data is ±0.2‰ for $\delta^{18}O$ on silicates and ±5‰ for δD values for fluid-inclusion extracts.

ANALYTICAL DATA

The δD data determined on fluid-inclusion extracts

The δD values for fluid-inclusion extracts from quartz grains in the deposits range from –42 to –97‰ (mean: –62‰), with 19 of the 22 values between –40 and –75‰ (Fig. 2a). There is no consistent pattern as a function of the style of mineralization; the data for all types overlap. The data for fluid-inclusion extracts from the granitic host-rocks range from –55 to –97‰, and clearly do not include in the weighting samples with the heavier $\delta^{18}O$ signature, as in the mineralized samples (*cf.* Figs. 2b, c). These results compare well with previous findings in the MLT (Kontak *et al.* 1999, 2001, Kontak & Kyser 2002).

The $\delta^{18}O$ data determined on white mica separates

The values determined on separates of white mica range from +4 to +10‰ (mean: +9‰), with all but one (a sample of greisen from the Turner deposit) within a range of +7.2 to +10‰ (Fig. 3a). No trend is apparent in the data arranged by deposit, deposit types, or host rock (Figs. 3b, c). These values compare well with other values of $\delta^{18}O$ for white mica from granitic rocks of the MLT (Longstaffe *et al.* 1980, Kontak *et al.* 1988, 1991,

TABLE 2. HYDROGEN AND OXYGEN ISOTOPE DATA (‰) FOR MINERALIZED ROCKS AND HOST ROCKS IN THE NEW ROSS AREA, NOVA SCOTIA, AS WELL AS CALCULATED COMPOSITIONS OF FLUID*

Sample no.	Location (deposit)	Rock type	δD fluid inclusions	$\delta^{18}O_{mica}$	$\delta^{18}O_{fluid}$	δD_{mica}	δD_{fluid}
SC-2-W	Walker	aplite	-58	9.9	10.7	-52	-42
SC-9-W		pegmatite	-48	7.8	8.6	n.d.	n.d.
SC-7-W		greisen	n.d.	9.5	8.1*	-47	-17*
SC-13-W		monzogranite (host)	-76	9.4	10.2	-56	-46
SC-14-W		greisen	-55	9.2	7.8*	-52	-22*
SC-3-MB	Millet Brook	U-irradiated sample	-71	9	7.6*	-69	-39*
SC-5-MB		granite (host)	-68	9.4	10.2	-57	-47
SC-6-MB		granite (host)	-72	8.9	9.7	-68	-58
SC-2a-T	Turner	greisen	-46	4	2.6*	-45	-15*
SC-9-T		monzogranite (host)	-55	7.9	8.7	-50	-40
SC-14-T		elvan	-83	7.2	8.0	-56	-46
SC-13-L	Long Lake	aplite	n.d.	9.4	10.2	-60	-50
SC-12-L		pegmatite	-42	10	10.8	-57	-47
SC-5-L		greisen	-49	9.2	7.8*	-42	-12*
SC-1-L		granite (host)	-97	9.8	10.6	-75	-65
DC 86-3 847	Mn Mines	hematized monzogranite	-53	9	7.6*	-54	-24*
SC-1-D		episyenite	n.d.	9.3	7.9	-53	-23
SC-3-D		monzogranite (host)	-61	9.8	10.6	-59	-49
SC-7-M	Morley's	pegmatite	-48	8.5	9.3	-57	-47
SC-1-M		granite (host)	-57	9.3	10.1	-61	-51
SC-3-K	Keddy	pegmatite	-46	9.5	10.3	-82	-72
SC-5-K		greisen	-91	9.5	8.1*	-63	-33*
SC-2-K		monzogranite (host)	-68	9.3	10.1	-85	-75
SC-10-R	Reeves	pegmatite	-60	9.5	10.3	-108	-98
U2		pegmatite	-40	n.d.	n.d.	n.d.	n.d.
SC-6-R		leucogranite (host)	-72	9.7	10.5	-48	-38

* on the basis of muscovite–H_2O oxygen and hydrogen isotopic equilibria at temperatures of 600°C for host granitic rocks, aplite, and pegmatite samples, and 400°C (*) for vein and greisen samples. n.d.: not determined.

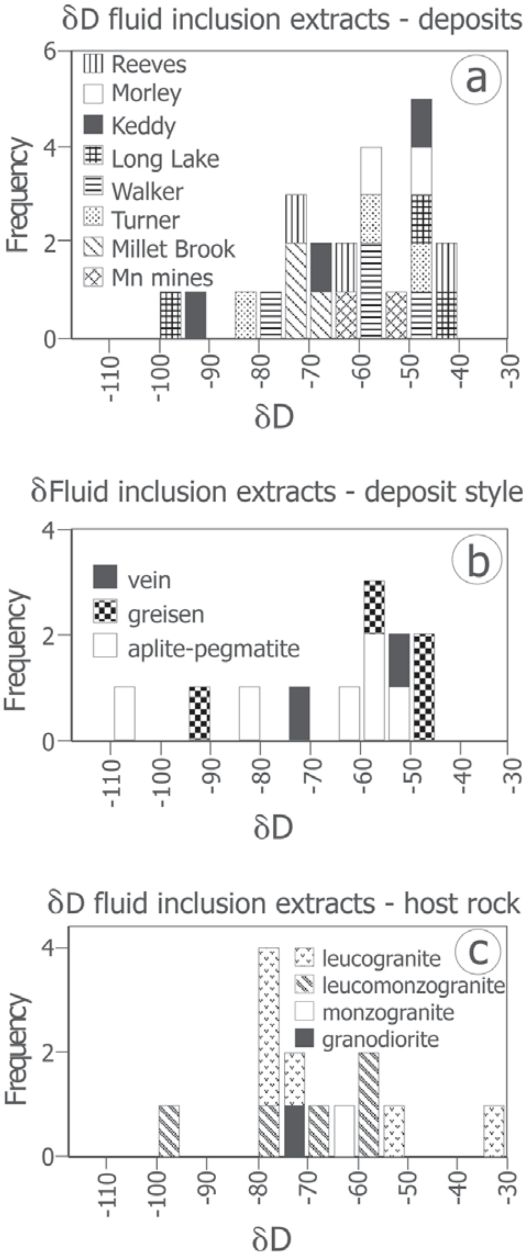

FIG. 2. Values of δD for fluid-inclusion extracts from quartz grains organized by: (a) deposit, (b) deposit style, and (c) host-rock type.

1999, 2001, 2002, Poulson *et al.* 1991, Dinnett 1995). The $\delta^{18}O$ values of whole-rock samples from the SMB range from +7.7 to +15‰ (Longstaffe *et al.* 1980, Chatterjee *et al.* 1985, Kontak 1988, Clarke *et al.* 1993, Kontak *et al.* 1988, 1991, 2002), and are typical of peraluminous, crust-derived granites worldwide (Sheppard 1986).

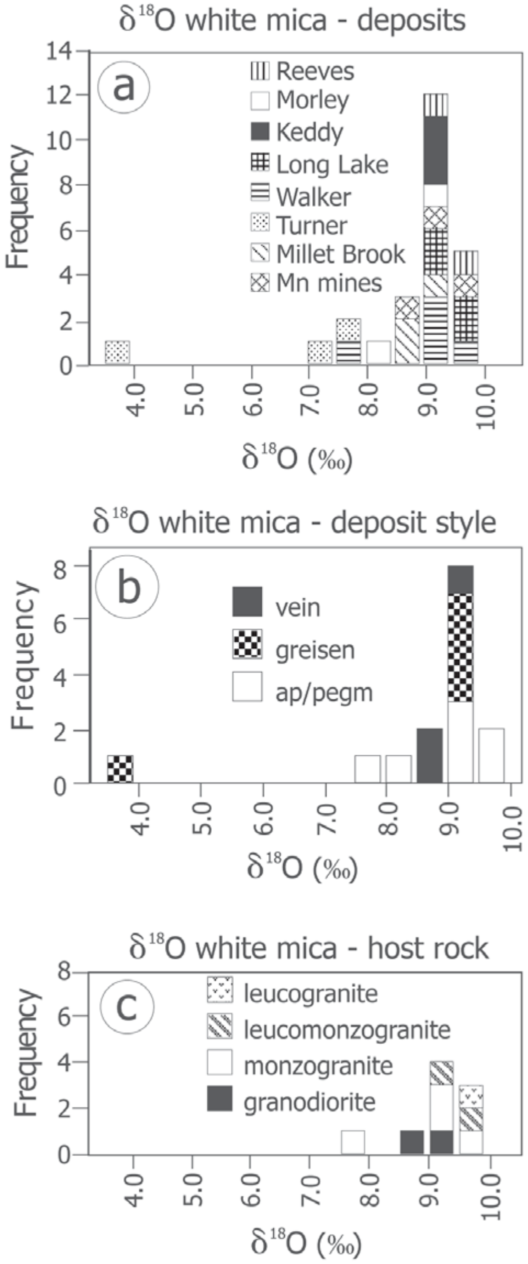

FIG. 3. Values of $\delta^{18}O$ for white mica separates organized by: (a) deposit, (b) deposit style, and (c) host-rock type.

We calculated the $\delta^{18}O$ composition of H_2O in equilibrium with the white mica at temperatures of 600° and 400°C [Table 2; see Carruzzo *et al.* (2000) for discussion of these temperatures] using the equations of O'Neil & Taylor (1969), with the former representing a magmatic temperature close to the granite solidus (host rock, pegmatite–aplite), and the latter, a minimum temperature for hydrothermal alteration (greisen, vein). Calculated $\delta^{18}O$ values for the H_2O in equilibrium with the white mica from the host granitic rocks and pegmatites–aplites range between +8.0 and +10.8‰, whereas those for the H_2O in equilibrium with the white mica from greisen and vein deposits range between +7.6 and +8.1‰, except for a single sample of greisen from the Turner deposit at +2.6‰. Interestingly, Kontak *et al.* (1988) also noted a low $\delta^{18}O$ value for an elvan dyke-rock from this same locality.

The δD data determined on white mica separates

The δD values determined on white mica separates range from –42 to –108‰ (mean: –61‰), with 20 of 24 values between –42 and –69‰ (Fig. 4a). No systematic pattern emerges from arrangement of the data by deposit or host-rock type (Figs. 4b, c). Few δD data exist for white mica in the SMB (–30 to –59‰; Dinnett 1995, Kontak *et al.* 2001, 2002), but they are close to the main range of values obtained in the present study.

We calculated the δD composition of H_2O in equilibrium with the white mica using the equations of Suzuoki & Epstein (1976) at temperatures of 600° and 400°C (Table 2). The calculated δD_{H2O} values range between –38 and –58‰ for white mica from the host granite and pegmatite–aplite samples, whereas they range between –12 and –39‰ for white mica from greisen and vein samples. The four samples of white mica with anomalously low values of δD pertain to host rock and pegmatite–aplite samples, and these have calculated δD_{H2O} values of –65 to –98‰ at 600°C.

DISCUSSION

Integration with previous data

Results of stable isotope investigations of the mineral deposits of the New Ross area were earlier provided by Kontak *et al.* (1988, 1991), and isotopic values for minerals other than white mica are available, thus permitting tests of isotopic equilibrium among coexisting minerals. The three main criteria used are: (1) absence of isotopic reversals, (2) absence of unusually large values of fractionation between coexisting phases, and (3) concordance of temperature derived using different pairs of minerals (O'Neil 1986). The order of $\delta^{18}O$ values for minerals in a granite, assuming isotopic equilibrium retained from crystallization, should be quartz > feldspar > white mica (O'Neil 1986). In Figure 5, we show all available isotopic data from the mineral deposits of the New Ross area. Although δ^{18}O values are not all in the expected order for quartz and feldspar, most samples of white mica have lower δ^{18}O values than coexisting quartz and alkali feldspar.

Kontak *et al.* (1991) calculated temperatures of 600° ± 65°C for the SMB using the data on quartz – white mica from Longstaffe *et al.* (1980), whereas $\delta^{18}O_{qtz-ms}$ data from pegmatites in the New Ross area provide a range of apparent temperatures (1200° to 230°C), indicating post-crystallization isotopic exchange. Using the white mica values obtained in this study and integrating them with quartz values for samples from the same locations, the Δ_{qtz-ms} (*i.e.,* $\delta^{18}O_{quartz} - \delta^{18}O_{muscovite}$) ranges from 0.4 to 8.1. The expected Δ_{qtz-ms} value for magmatic conditions in a granite is 3.8 ± 0.3‰ (Longstaffe 1982). With the exception of two samples, the calculated Δ_{qtz-ms} for samples from the New Ross area are below this range and provide evidence of isotopic disequilibrium (Fig. 6). The $\Delta_{qtz-kfs}$ values range from 0 to 7, whereas Longstaffe (1982) provided a $\Delta_{qtz-kfs}$ range of 1.6 to 2.4‰ for granitic rocks equilibrated under magmatic conditions. Pegmatites from Reeves and Morley's show appropriate $\Delta_{qtz-kfs}$ values, and temperatures calculated for these samples are between 480 and 520°C. Thus, a temperature range in close agreement with pressure-corrected temperature of entrapment derived from fluid-inclusion thermometry of pegmatites is determined. Disequilibrium values are not unexpected within the SMB, as slowly cooling igneous rocks (plutonic environment) are susceptible to isotopic exchange at both at high and low temperatures (Giletti 1986).

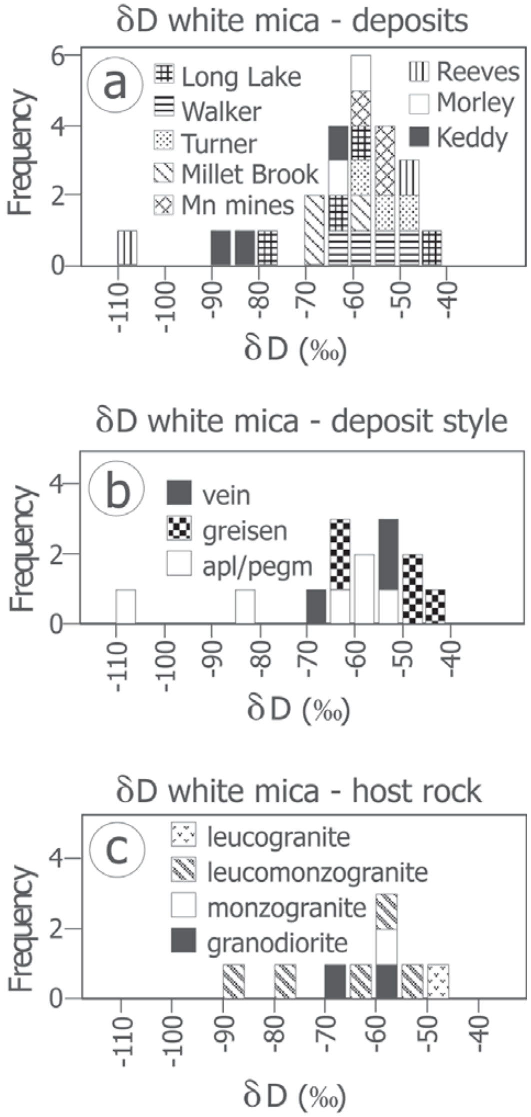

FIG. 4. Values of δD for white mica separates organized by: (a) deposit, (b) deposit style, and (c) host-rock type.

Isotopic signature of fluid reservoirs

Carruzzo *et al.* (2000) used salinity measurements and decrepitate analysis of fluid inclusions in quartz from each of the eight mineral deposits of the New Ross area to identify three different reservoirs (magmatic, metamorphic, meteoric). Such fluids may be isotopically distinct (Sheppard 1986); if they are, their signature within the SMB should be discernable. Other sources of fluids at the time of fluid circulation (*i.e.,* 380–370 Ma), such as evaporated brines or seawater, are not considered because of the absence of evaporites from the stratigraphic record at the time of SMB emplacement. The isotopic signature of magmatic and metamorphic fluids depends mainly on the isotopic composition of the source rock, whereas the isotopic signature of the meteoric fluid depends on terrane-specific parameters such as latitude and altitude, and also local variables such as temperature. The following summaries present the isotopic signature of the three fluids previously recognized in the New Ross area.

Magmatic fluid: Sheppard *et al.* (1969) defined the isotopic composition of magmatic fluid as having δD values from –40 to –80‰ and δ^{18}O values from +5.5 to +9.5‰. However, a magmatic fluid in equilibrium with high-δ^{18}O peraluminous granites might plot outside this range [*i.e.,* shows enrichment of a few ‰ in δ^{18}O, Sheppard (1986)], as in the case of peraluminous granites of the Cornubian Batholith of southwestern England (Sheppard 1977). Using δ^{18}O data ($n = 24$) from unaltered local granitic rocks (Longstaffe *et al.* 1980, Dinnett 1995), we define $\delta^{18}O_{magmatic\ fluid}$ for the SMB to be +10 ± 2‰ at 600°C. Similarly, using $\delta D_{mineral}$ data ($n = 2$) from Dinnett (1995), we estimate the δD range for the

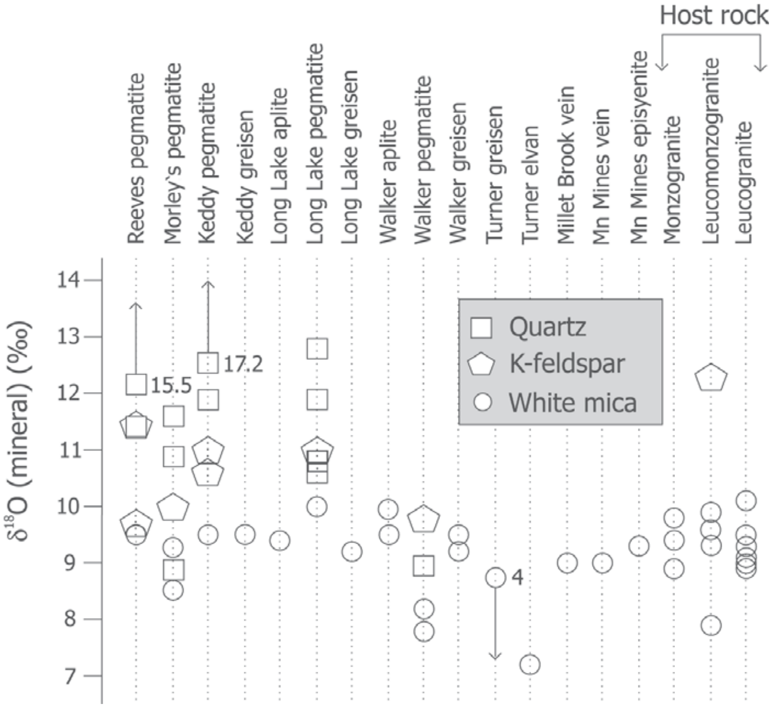

FIG. 5. Integrated isotopic data for mineral separates from the mineral deposits of the New Ross area using data from Kontak *et al.* (1988, 1991) and this study.

SMB to be –40 ± 10‰, also at 600°C. These data define a magmatic field for the SMB that compares well for $\delta^{18}O$, but is slightly lighter for δD compared to the field defined by Kontak *et al.* (2002) for magmatic fluids in the Peggys Cove area of the SMB (Fig. 7). Thus, the δD values used here are within the range proposed by Taylor (1987) for H_2O exsolved from felsic melts.

Metamorphic fluid: Sheppard (1986) gave values of δD from 0 to –70‰ and $\delta^{18}O$ from +3 to +20‰ for a metamorphic fluid in equilibrium with minerals of its host rock. To narrow this large range reflecting the diversity of possible protoliths, we calculated a composition of hypothetical metamorphic fluid specific to the Meguma Supergroup [psammite and pelite, average compositions in Kontak & Dostal (1992)] using the only whole-rock isotopic data existing for the Meguma Supergroup (Longstaffe *et al.* 1980). Assuming a certain $\delta^{18}O_{\text{metamorphic fluid}}$ at 600° and 400°C, and knowing the fractionation factor between each mineral of the sample (quartz, chlorite, albite, white mica, and biotite) and H_2O, we obtained the $\delta^{18}O$ for each mineral [using the equations of Bottinga & Javoy (1973) for quartz – albite – muscovite – biotite – H_2O fractionation, and of Savin & Lee (1988) for chlorite–water fractionation]. We then assigned a weighted $\delta^{18}O$ value for each mineral according to the amount of oxygen it contains; the sum of weighted $\delta^{18}O$ value for each mineral represents $\delta^{18}O_{\text{whole rock}}$. By comparing the $\delta^{18}O_{\text{whole rock}}$ calculated to the values measured by Longstaffe *et al.* (1980), we varied the $\delta^{18}O_{\text{metamorphic fluid}}$ to obtain matching values (Table 3). Between 600° and 400°C, $\delta^{18}O_{\text{metamorphic fluid}}$ ranges from +7.5 to +10.8‰; this range overlaps the $\delta^{18}O_{\text{magmatic fluid}}$, suggesting that the Meguma metasedimentary units are probably the result of denudation of a crust similar compositionally to that of the SMB.

Meteoric fluid: The meteoric water line (MWL) describes the systematic isotopic attributes of present-day meteoric water; ancient meteoric water may present a similar relationship (Sheppard 1986). Paleoreconstructions of the position of Laurentia (Kent & Van der Voo 1990, Scotese & McKerrow 1990) reveal that the Meguma Terrane was located at ~20° South latitude during the Late Devonian (Famennian) and show the closure of the Theic Ocean between Laurentia and Gondwana by Late Carboniferous – Early Permian (Scotese & McKerrow 1990), leaving the MLT close to open seawater until ~300 Ma. Using isostatic calculations (Airy model), assuming that the actual thickness of the crust is similar to that in the Late Devonian minus erosional effects (estimated from geobarometry to be ~12 km, the depth of emplacement of the SMB,

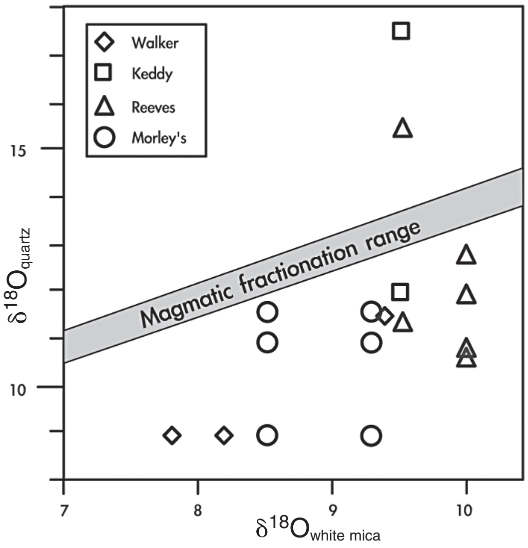

FIG. 6. Values of $\delta^{18}O_{white\ mica}$ *versus* $\delta^{18}O_{quartz}$ for mineral pairs from mineral deposits of the New Ross area. The magmatic fractionation field (Longstaffe 1982) represents the area where data plot if equilibrium between both minerals exists. Values of $\delta^{18}O_{quartz}$ from Kontak *et al.* (1988, 1991), and values of $\delta^{18}O_{white\ mica}$ from this study.

Raeside & Mahoney 1996) and a compensation depth of ~100 km, we obtained a paleo-altitude of ~1700 m at 380 Ma.

A possible model for the paleo-isotopic composition of the MLT, considering both latitude (20°S) and altitude (~1700 m) restrictions, lies in the present-day Central Andes, where a $\delta^{18}O_{H2O}$ value of –6 ± 2‰ exists (Taylor 1997). We obtained a δD_{H2O} value of –38 ± 16‰ for that range of $\delta^{18}O_{H2O}$ by applying the equation for the meteoric water line. Recent measured isotopic values from precipitations in South America (Los Molinos station in Argentina, latitude of ~24°S and altitude of 1300 m) are $\delta^{18}O$ = –8 to 0‰ and δD = –40 to –5‰ [www.isohis.iaea.org].

Nature and origin of fluids based on white mica data

The mineral deposits of the New Ross area represent a variety of mineralization styles and, as shown by Carruzzo *et al.* (2000), their formation involved three roughly coeval reservoirs of fluid. In the conventional $\delta^{18}O$ – δD diagram (Fig. 7), the standard fields for different types of water (Sheppard 1986, and references therein, Ohmoto 1986) are outlined and compared to (i) fields for meteoric and magmatic fluids within the Meguma Supergroup, (ii) the calculated $\delta^{18}O$ range for a metamorphic fluid in equilibrium with the Meguma metasedimentary rocks; and (iii) the data, calculated at 600° and 400°C, for mineral deposit environments of the New Ross area. From this diagram, we note the following points: (1) most mineral deposit data overlap the field of metamorphic fluid in terms of δD, and are in the $\delta^{18}O$ range defined for H_2O equilibrated with metasedimentary rocks of the Meguma Supergroup; (2) most data for granite host-rock and pegmatite–aplite fall in the field for magmatic H_2O equilibrated with peraluminous granitic rocks at 600°C (except for a single point with an anomalously low-δD value), and at the low δD end of the SMB magmatic H_2O field (except for five points with lower δD values), and (3) the greisen and vein data fall at the low end of the $\delta^{18}O$ field for magmatic H_2O and have generally higher δD values compared with the peraluminous magmatic range for δD. Thus, there appears to be a shift toward lower $\delta^{18}O$ and higher δD values of the fluids from the magmatic through to the hydrothermal regime.

An erroneous estimation of temperature alone cannot account for the range in isotopic composition of the fluids shown in Figure 7. Therefore, we explore other reasons for such variations below. A progressive lowering of $\delta^{18}O$ values from about +10.0 to about +8.0‰ can result from interaction with a low-$\delta^{18}O$ fluid. We suggest that the low-$\delta^{18}O$ fluid is of meteoric origin, and the proportion of this fluid increased with time, concomitant with a change from the magmatic regime to that favoring greisen and vein formation. This interpretation is supported by the following data. First, low-salinity fluids are the dominant type of fluid inclusion in quartz from both greisen and vein deposits, whereas

TABLE 3. CALCULATIONS OF $\delta^{18}O_{metamorphic\ water}$ IN EQUILIBRIUM WITH A SANDSTONE AND SILTSTONE OF THE MEGUMA SUPERGROUP

	Mineral proportion (%)	Proportion of O in each mineral	Proportion of O in each sample	1000lnα at 600°C	1000lnα at 400°C	$\delta^{18}O_{mineral}$ at 600°C	$\delta^{18}O_{mineral}$ at 400°C	Weighted $\delta^{18}O_{mineral}$ at 600°C	Weighted $\delta^{18}O_{mineral}$ at 400°C
Sandstone									
quartz	30	0.53	15.9	1.59	3.42	12.39	12.27	4.29	4.25
feldspar	20	0.49	9.8	0.99	2.41	11.79	11.26	2.52	2.40
chlorite	30	0.39	11.7	-5.32	-2.35	5.48	6.5	1.40	1.66
biotite	10	0.39	3.9	-2.54	-2.08	8.26	6.77	0.70	0.58
muscovite	10	0.46	4.6	-0.77	1.36	10.03	10.21	1.01	1.02
Total oxygen in sample:			45.9			$\delta^{18}O_{whole\ rock}$*		9.91	9.91
Siltstone									
quartz	48	0.53	25.44	1.59	3.42	10.89	10.92	5.68	5.69
feldspar	28	0.49	13.72	0.99	2.41	10.29	9.91	2.89	2.79
chlorite	15	0.39	5.85	-5.32	-2.35	3.98	5.15	0.48	0.62
biotite	5	0.39	1.95	-2.54	-2.08	6.76	5.42	0.27	0.22
muscovite	4	0.46	1.84	-0.77	1.36	8.53	8.86	0.32	0.33
Total oxygen in sample:			48.8			$\delta^{18}O_{whole\ rock}$**		9.64	9.65

The calculated values are close to the mean $\delta^{18}O$ of 9.9‰ in the sandstone samples (*) and of 9.6‰ in the siltstone samples (**) of the Meguma Supergroup (Longstaffe *et al.* 1980). For the sandstone, the $\delta^{18}O_{mineral}$ is calculated at 600°C with $\delta^{18}O_{fluid}$ equal to 10.8‰, and at 400°C, with a $\delta^{18}O_{fluid}$ equal to 8.8‰. For the siltstone, the $\delta^{18}O_{mineral}$ is calculated at 600°C with $\delta^{18}O_{fluid}$ equal to 9.3‰, and at 400°C, with a $\delta^{18}O_{fluid}$ equal to 7.5‰.

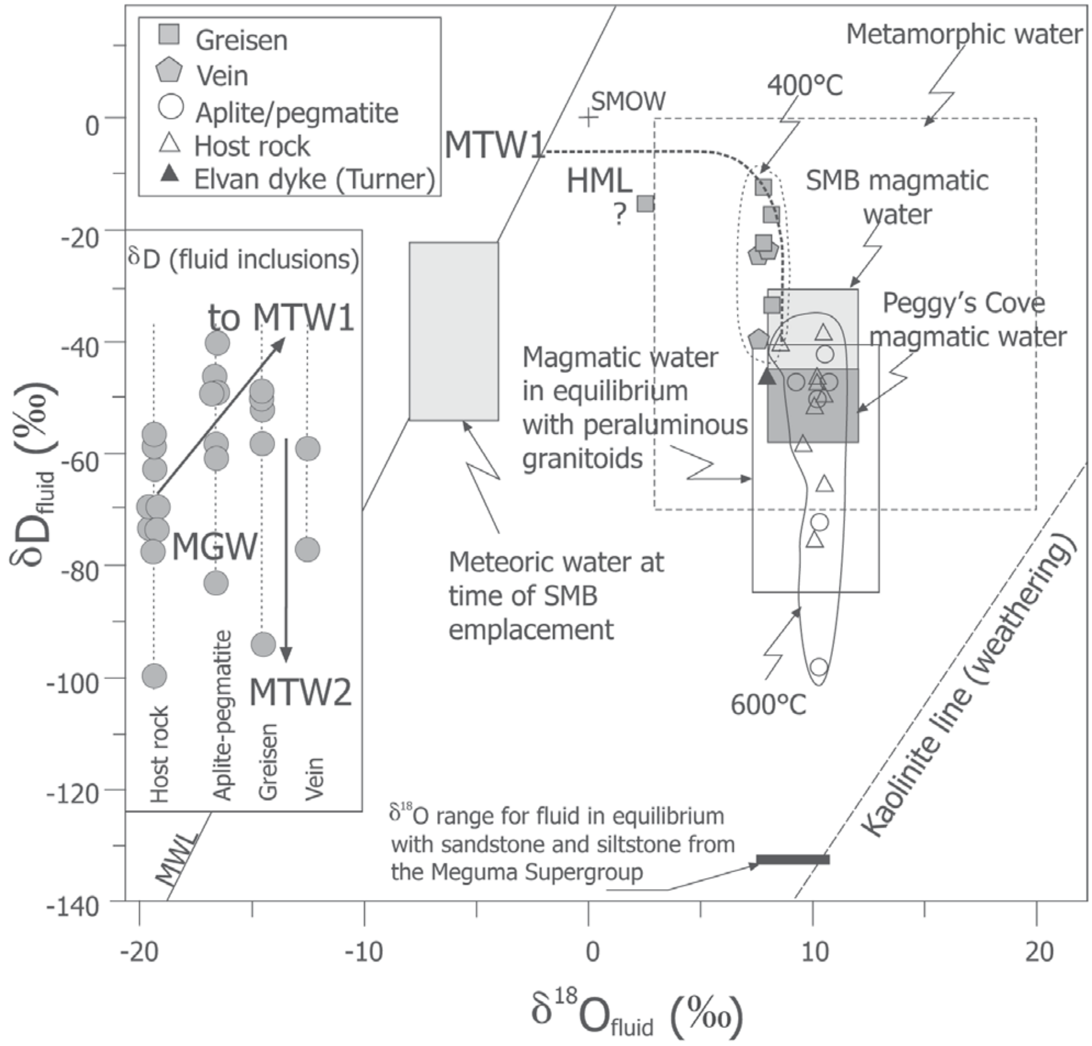

FIG. 7. Plot of δD *versus* $\delta^{18}O$ showing fluid reservoirs (Sheppard 1986, and references therein); the magmatic field for water in equilibrium with "S-type" granitic rocks is taken from from Ohmoto (1986). The meteoric water line is that of Craig (1961). The meteoric water field at the time of SMB emplacement, SMB and Peggy's Cove magmatic water, and range of $\delta^{18}O$ values for H_2O in equilibrium with Meguma metasedimentary rocks are defined in the text. Inset diagram shows δD values for fluid-inclusion extracts from quartz. The unfilled symbols represent δD values for a fluid in equilibrium with white mica at 600°C, whereas the filled symbols represent δD values for a fluid in equilibrium with white mica at 400°C. The terms MGW (magmatic fluid or water of host rock), MTW1 and MTW2 (meteoric water 1 and 2, respectively), and HML (hypothetical mixing line) are discussed in the text.

high-salinity fluids dominate the fluid inclusions from host granitic rocks and pegmatite–aplite (Carruzzo *et al.* 2000). These data are consistent with the lowering of $\delta^{18}O_{H2O}$, as inferred from the $\delta^{18}O_{H2O}$ for fluid in equilibrium with the white micas (Fig. 8), and suggest incursion of a meteoric fluid. Secondly, with decreasing temperatures, the isotopic composition of white mica in equilibrium with the hydrothermal fluid should evolve toward lower values of δD (see isopleths in Fig. 9). However, instead of finding this expected trend if the system were closed (*i.e.,* dominantly magmatic), we observe higher δD values for the greisen and vein samples (Fig. 7). Therefore, mixing with a fluid of higher δD composition appears as the most likely explanation. This shift to higher δD values may also reflect the presence of meteoric water in the system, if the appropriate paleolatitude and altitude existed.

Therefore, on the basis of isotopic data for the white micas, we suggest that an evolution of increasing fluid:rock ratio through time occurred, as the system changed from a magmatic to a hydrothermal regime. This transition may well account for the distribution of the data in Figures 7 and 8. Simultaneously with this evolution, the nature of the fluid changed from early magmatic to later meteoric. A hypothetical mixing line (HML) between the presumed end-member fluids illustrates this evolution of the isotopic signature in $\delta^{18}O$ – δD space (Fig. 7). Note that the composition of the end-member meteoric fluid (MTW 1) has a significantly higher δD value (~–6‰) than the range calculated for the MLT at *ca.* 380 Ma. Although the incursion of a metamorphic fluid may also explain the distribution of the data, we again note that our fluid-inclusion thermometric study indicates the dominance of a meteoric fluid rather than a metamorphic fluid in the system. Kontak & Kyser (2002) also reported the presence of low-$\delta^{18}O$ fluids in a vein-hosted silica–clay mineralized zone in the southwestern part of the SMB (*i.e.,* northeast of Yarmouth, Fig. 1a) and, in addition, the fluid inclusions in the quartz record the dominance of a low-salinity (<0.5 wt.% equiv. NaCl) fluid, similar to that in the greisen and vein deposits of the New Ross area.

Nature and origin of fluids based on fluid-inclusion extracts

The δD_{fluid} values for fluid-inclusion extracts arranged according to host rock are shown inset in Figure 7. We note that there is a progressive trend toward higher values of δD_{fluid} from host granitic rocks to greisen, with a few anomalously lower values. We interpret the data to reflect two stages of incursion of an exotic water, for which we infer a meteoric origin. The first stage (MTW1) of incursion occurred late in the orthomagmatic–hydrothermal stage of the SMB, when meteoric water mixed with primary magmatic fluid (MGW) of the host rock. This mixing resulted in a trend toward progressively higher values of δD_{fluid}. The second stage of incursion of meteoric water (MTW2) occurred later, after crystallization and mineralization within the SMB. In addition, this second stage of infiltration must have occurred when the signature of meteoric water was considerably lower, thus shifting the fluid-inclusion extract data to much lower values of δD. This interpretation of the fluid-inclusion extract data on δD is based in part on comparison with the δD_{fluid} values calculated for white mica in the same samples.

Figure 9 consists of four diagrams of $\delta D_{white\ mica}$ *versus* temperature (°C), with isopleths for δD_{fluid} calculated using the muscovite–H_2O fractionation equation of Suzuoki & Epstein (1976). These diagrams also show the δD data from fluid-inclusion extracts (closed circles) and the calculated δD data for fluids in equilibrium with the white mica (open circles). The solid lines in these diagrams link the δD datum from white mica to the δD datum from fluid-inclusion extracts for the same sample. As discussed above, a temperature of 600°C is used for isotopic data of the white mica in host granitic rocks and aplite–pegmatite, *versus* 400°C for white mica in veins and greisens. Pressure-corrected temperatures of entrapment of 500°C are used for fluid inclusions in samples of host granitic rock and aplite–pegmatite, *versus* 400°C for vein and greisen samples [see Carruzzo *et al.* (2000) for discussion]. With decreasing temperature, fluids in equilibrium with the white mica evolve toward higher values of δD for a white mica with the same δD value. If fluid inclusions in quartz contained the same H_2O as that structurally bound in the white mica, then both δD_{fluid} compositions should lie on the

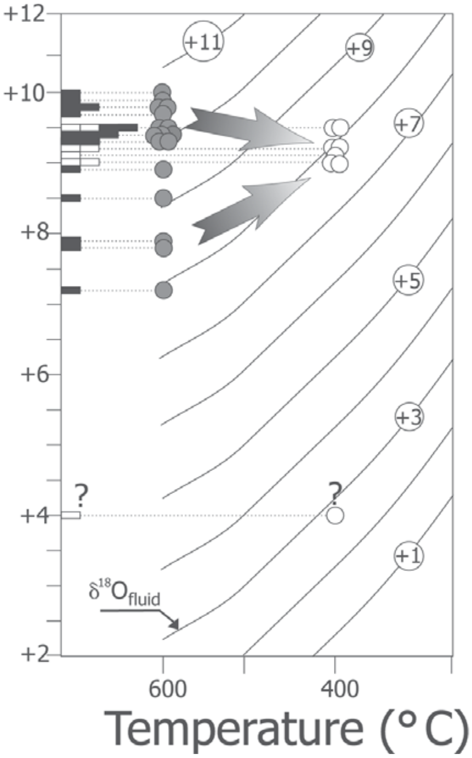

FIG. 8. Plot of $\delta^{18}O_{white\ mica}$ *versus* temperature (°C), with isopleths for $\delta^{18}O_{water}$ calculated using the muscovite–H_2O fractionation equation of O'Neil & Taylor (1969). Rectangles represent $\delta^{18}O_{white\ mica}$, with closed symbols for granite aplite and open symbols for greisen–vein setting. The circles equate to the $\delta^{18}O$ values of a fluid inferred to be in equilibrium with the micas at 600° and 400°C, respectively, for open and closed symbols. Arrows indicate the inferred direction of fluid evolution, *i.e.*, toward lighter $\delta^{18}O_{fluid}$ values.

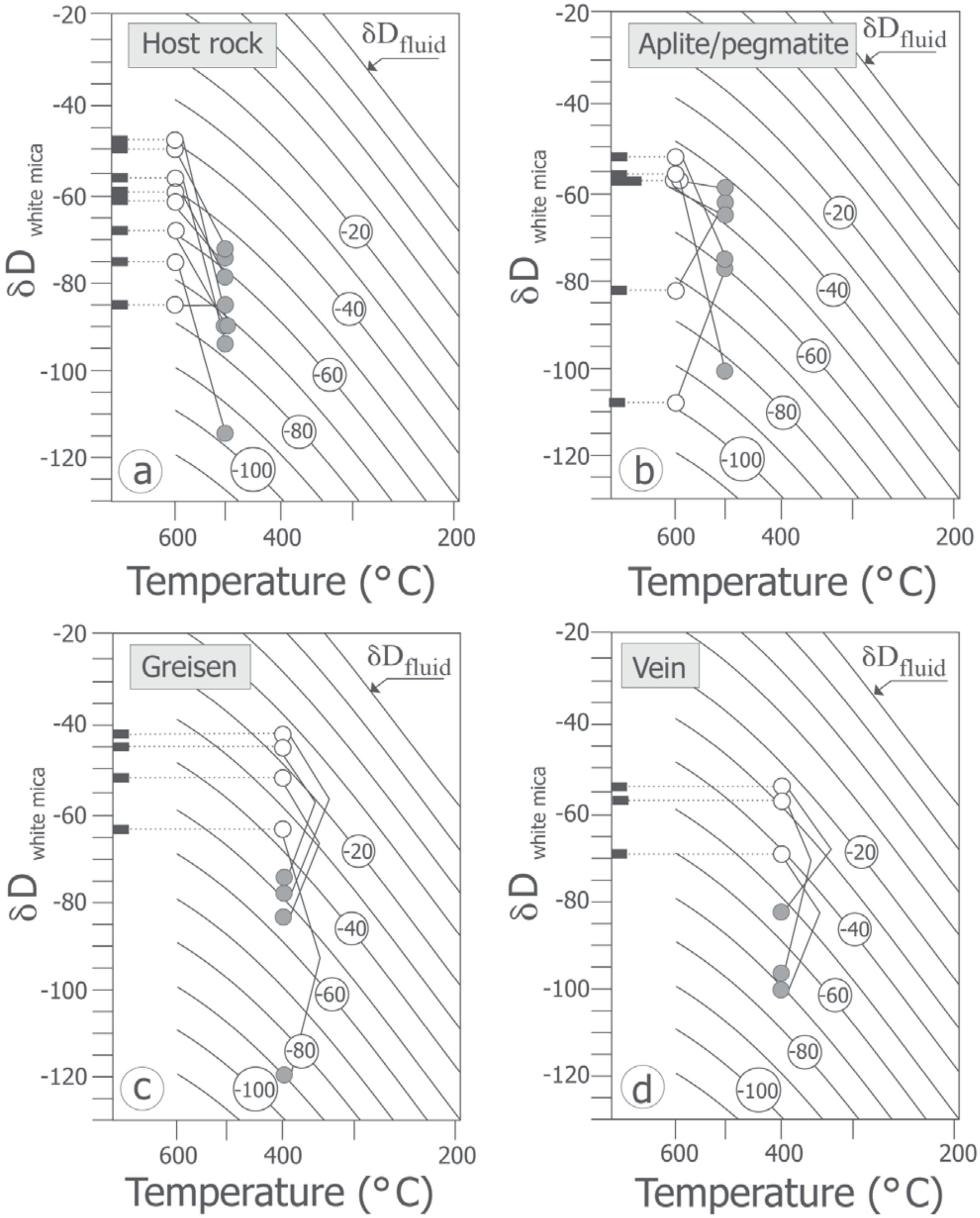

FIG. 9. Diagrams of $\delta D_{white\ mica}$ *versus* temperature (°C) with isopleths for δD_{fluid} calculated using the muscovite–H_2O fractionation equation of Suzuoki & Epstein (1976). Rectangles represent $\delta D_{white\ mica}$, open circles represent fluid in equilibrium with the white mica at the inferred temperatures (600°C for host rock and aplite–pegmatite, 400°C for greisen–vein), and closed circles represent fluid-inclusion extracts plotted at the inferred temperatures of entrapment (500°C for host rock and aplite–pegmatite, 400°C for greisen–vein).

same isopleth. Instead, the fluid-inclusion extracts generally have lower δD values than the calculated δD_{H2O} of white mica in the same hand sample (*i.e.,* do not lie on same isopleth). We also note that the discrepancy of the two datasets increases from host granitic rocks, through aplite–pegmatite, to the greisen and vein samples, such that the difference is greatest for the greisen and vein samples. Thus, there has clearly been incursion of an exotic fluid at some point after crystallization of the white mica in the samples, as recorded by the fluid-inclusion extract data. It would appear, therefore, that the fluid inclusions have the ability to record this change in the isotopic composition of the fluid by preserving the fluid as secondary inclusions, whereas the white mica retains the δD value of the fluid at the time of crystallization.

Particularly intriguing in the dataset are the anomalously low δD values measured in several fluid-inclusion extracts, for which we ascribe an origin due to later meteoric water. Such values are not unique to this study (*e.g.,* Linnen 1995, Kontak *et al*. 1999). Given their unusual occurrence, we discuss several possible explanations below.

1. There may be an analytical problem: This possibility relates to release of molecular water or H-balanced defects in quartz during *in vacuo* thermal decrepitation of the sample (Gleeson *et al*. 2002). In their study, Gleeson *et al.* (2002) found that samples with anomalously low values of δD are characterized by a low abundance of fluid inclusions. We presently discount this possibility as a viable explanation for two reasons: (1) given that the samples used in the present study are "inundated" with fluid inclusions, the aforementioned fraction of molecular water would be relatively small; (2) the above mechanism requires further documentation before it can be considered applicable in the present study.

2. Incursion of a low-δD fluid after formation of white mica and before entrapment of the fluid, still during the hydrothermal stage: this process requires involvement of a low-δD fluid during hydrothermal activity. Given that a much higher value of δD characterizes the signature of meteoric water at this paleo-latitude and altitude during hydrothermal activity at *ca.* 380 Ma (*i.e.,* see discussion above), we exclude such a fluid reservoir. An alternative fluid is that of organic origin, as originally suggested by Sheppard & Charef (1986) and used by Linnen (1995) to explain low-δD fluids in Sn-mineralized pegmatites of Thailand. This explanation is difficult to assess given the lack of documentation of such fluids in organic-matter-rich sedimentary basins; therefore, we also discount this possibility.

3. Incursion of a low-δD fluid during a fluid-circulation event subsequent to SMB emplacement: in this case, we appeal to trapping of a fluid of probably meteoric origin much later in time than emplacement of the SMB (MTW2). Relevant to this case are the following: (i) the presence of low-temperature (*i.e.,* 100–150°C), low-salinity fluid inclusions and the presence of monophase liquid inclusions in most samples (Carruzzo *et al.* 2000), and (ii) the documentation of a low-δD_{fluid} phase by other investigators and attributed to trapping of later, younger meteoric water [*e.g.,* Sun & Eadington (1987) in tin-mineralized granite from New South Wales in Australia; Criss & Taylor (1983) in the Idaho Batholith].

The δD values for a fluid in equilibrium with white mica from two samples hosted by the Keddy–Reeves leucogranite (pegmatites from Keddy and Reeves deposits) are considerably lower than the corresponding δD for fluid-inclusion extracts from these samples, but we note that the $\delta^{18}O$ values for these micas are similar to the others (Table 2). We have at present no satisfactory explanation for this disparity between the $\delta^{18}O$ and δD data, except to note that the micas have clearly equilibrated with a fluid with a low-δD signature.

Isotopic evolution of fluids upon cooling

In an integrated model for fluid circulation within the New Ross area (Fig. 10), we suggest the following succession of fluid involvement. A magmatic fluid, characterized by salinities of 19–25 wt.% equiv. NaCl, and Na–K chlorides as major solutes (Carruzzo *et al.* 2000), a $\delta^{18}O$ isotopic composition of +8 to +12‰, and a δD of –40 to –80‰, exsolves from the magma at a temperature of ~600–650°C (Stage 1). A low-salinity fluid with a significant component of meteoric water and inferred end-member composition of $\delta^{18}O \equiv -2$‰ and $\delta D \equiv -6$‰ enters the system (MTW1 in Fig. 7; Stage 2). Fluid inclusions from meteoric-water-dominated mineral deposits (mainly vein deposits) record a temperature of ~400°C. Meteoric fluid can reach such an elevated temperature by travelling to a depth of ~10–12 km (emplacement depth for the SMB) through faults created either by thermal contraction of the cooling batholith, by overpressure release by potential explosive unroofing of the SMB (Clarke & Bogutyn 2003), or by active faulting related to regional tectonics at the time of SMB emplacement (Horne *et al.* 1992). Also, convection cells driven by heat in the batholith allow for some mixing of magmatic fluid with a Ca-rich, high-salinity (29–43 wt.% equiv. NaCl) metamorphic fluid originating within the Meguma metasedimentary sequence (Stage 3). Such mixing is recorded by the cation content of fluid-inclusion decrepitates (Carruzzo *et al.* 2000).

After crystallization and cooling of the SMB, a second generation of meteoric fluid (MTW2) with a lower δD signature than MTW1 infiltrated the system and was trapped as secondary fluid inclusions along healed fractures in quartz. This second generation of meteoric water thus accounts for the difference between the calculated δD_{fluid} values of white mica and the δD values of fluid-inclusion extracts in coexisting quartz. The greater disparity of these data for the vein and greisen

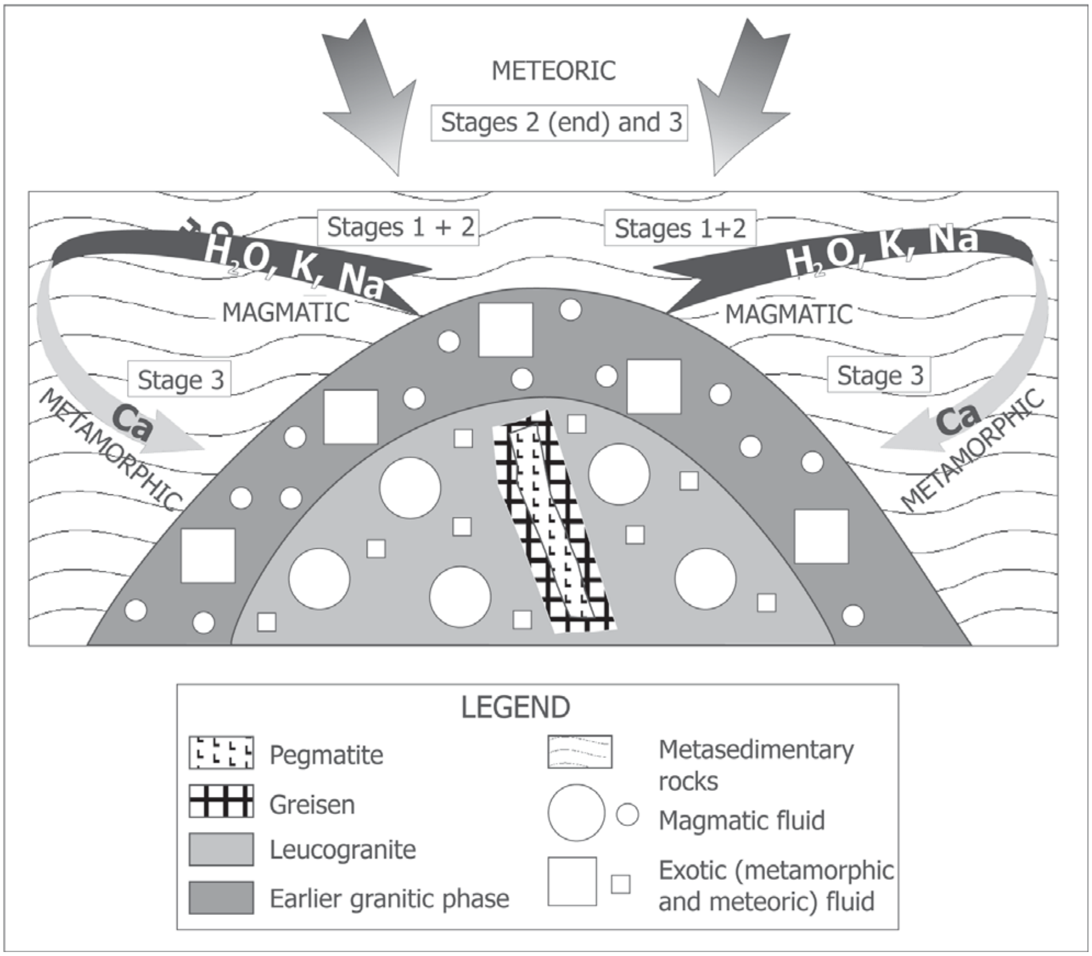

FIG. 10. Integrated model for fluid circulation within the New Ross area involving magmatic fluid from the crystallization of the host intrusion and exotic fluid of both meteoric and metamorphic origin. The large and small squares and circles schematically represent the amount of fluid interacting with the samples of granite.

samples *versus* the granitic host and aplite–pegmatite samples suggests that more of this fluid is present in the vein and greisen samples.

CONCLUSIONS

Isotopic data for white mica and fluid-inclusion extracts from eight mineral deposits of the New Ross area of the South Mountain Batholith, Nova Scotia, permit the following conclusions:

1. As noted in previous studies (Kontak *et al.* 1988, 1991), coexisting silicate minerals have not preserved isotopic equilibrium, thus impeding further determination of formation temperature using isotopic fractionation. Variable degrees of mineral–fluid re-equilibration at the subsolidus stage are probably the reason for the isotopic disequilibrium and relate to the protracted thermal history of the SMB.

2. The isotopic compositions ($\delta^{18}O$, δD) of white mica from host granitic rocks, pegmatite–aplite, greisen, and vein samples, along with δD values for fluid-inclusion extracts from quartz in the same samples, record a transition from an initial fluid-dominated orthomagmatic stage to a later hydrothermal-fluid-dominated system into which an increasing amount of meteoric water infiltrated during greisen and vein formation. We propose a hypothetical mixing line between magmatic and meteoric waters for the SMB during its emplacement and subsequent crystallization at 380 Ma.

3. Fluid-inclusion extracts from quartz record lower values of δD_{fluid} than calculated for coexisting samples of white mica using appropriate fractionation factors and temperatures of crystallization. We attribute these differences to post-crystallization entrapment of later meteoric water and emphasize the potential significance of integrating δD values obtained from hydrous silicates with those from fluid-inclusion extracts in coexisting minerals. Anomalous depleted values of δD for white mica samples from two pegmatite suites also reflect fluid-mediated recrystallization of these phases at the same time as the infiltration of a low-δD water.

4. Integration of our earlier fluid-inclusion results for the same samples with the new isotopic data provides corroborating support for incursion of meteoric water during crystallization and subsequent hydrothermal alteration of the SMB. A set of low-temperature, low-salinity fluid inclusions plausibly records the incursion of a second meteoric fluid, as suggested from the isotope data.

5. Although we documented the participation of a metamorphic fluid reservoir in the hydrothermal fluid history of the New Ross area through the fluid-inclusion study, our isotopic data do not allow the discrimination of this fluid, given the lack of δD data for such rocks in the study area.

In summary, the evolution of fluids from a magmatic to hydrothermal regime within the New Ross area is as follows: (i) exsolution of a magmatic fluid at ~600–650°C, (ii) subsequent cooling to ~400–500°C with incursion of a meteoric fluid *via* faults and mixing with magmatic and metamorphic fluids through fluid circulation by convection, and (iii) incursion at a much later time of a low-δD meteoric fluid involved in cooling in the area.

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