New constraints on the age and origin of the Dunbrack Pb-Cu-Zn-Ag deposit, Musquodoboit Batholith, southern Nova Scotia

Daniel J. Kontak¹, Kevin Ansdell², and Douglas Archibald³

¹Nova Scotia Department of Natural Resources, P.O. Box 698, Halifax, Nova Scotia B3J 2T9, Canada
²Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan S5W 0W0, Canada
³Department of Geological Sciences, Queen's University, Kingston, Ontario K7L 3N6, Canada

Date Received: October 8, 1998
Date Accepted: January 10, 1999

Mineralization at the Dunbrack Pb-Zn-Cu-Ag deposit occurs within a ca. 1 m wide quartz vein characterized by comb, cockade, crustiform, and breccia textures. The mineralized vein, NW-trending and steeply dipping, has a medium- to coarse-grained, biotite-muscovite-cordierite monzogranite phase of the 370 Ma Musquodoboit batholith as its hanging wall and a fine-grained felsic dyke rock as its footwall. Silicate mineralogy in the vein is dominated by quartz with trace amounts of K-feldspar (Or56-100), muscovite (±2 wt. % FeO, ±0.8 wt. % F), chlorite, and kaolinite, whereas sulphides include Fe-poor sphalerite (±4 wt. % Fe), galena, chalcopyrite, and a variety of secondary Cu sulphides. Dating (⁴⁰Ar/³⁹Ar) of vein muscovite and the footwall dyke rock indicate similar ages of 370 Ma; therefore, vein formation and dyke injection are interpreted to be coincident with emplacement of the 370 Ma Musquodoboit Batholith.

Fluid inclusion studies of aqueous, L-V inclusions indicate homogenization temperatures of ca. 140 ± 5°C and salinities of 20 ± 2 wt. % eq. NaCl, but minor amounts of inclusions with higher (24 to 27 wt. %) and lower (8-16 wt. %) salinities occur. First melting temperatures and analysis of decrepitate mounds indicate two fluid types, a more abundant NaCl-H₂O fluid and a less abundant NaCl-CaCl₂-KCl-H₂O fluid.

Sulphur isotopes for hypogene galena, chalcopyrite, and sphalerite are uniform and equate to δ³⁴S of +4.2 to +6.6‰ (at 300°C), which is similar to data for other granite-related sulphide mineralization in the Meguma Zone. Vein quartz has an δ¹⁸O value of +15.3 ± 1.2‰ (n=10), which equates to δ¹⁸Owater of +9 to +11.5‰ at 300°C, whereas fluid inclusion extracts indicate 8D values of -90 to -113‰ (n=6).

Collectively, the timing of mineralization (i.e., 370 Ma) and the dominantly magmatic isotopic and geochemical signature for the vein-forming fluid suggest a genetic association with the Musquodoboit Batholith. However, the Ca-rich nature of some fluid inclusions and low 8D values of fluid extracts suggest involvement of another reservoir, possibly reflecting a fluid that equilibrated with the country rock of the Meguma Group.

La minéralisation au dépôt de Pb-Zn-Cu-Ag Dunbrack se produit dans une veine de quartz d’une largeur c. 1 m caractérisée par des textures de peignes, cockade, croûteuses et de brèches. La veine minéralisée en direction nord-ouest a un pendage a pente forte, et une phase a grain grossier, de biotite-muscovite-cordierite monzogranite provenant du batholite Musquodoboit de 370 Ma en tant que lèvre supérieure et à un dyke a grain fin de roche felsique en tant que sa lèvre inférieure. La minéralogie de silicate dans la veine est dominée par le quartz avec des traces de K-feldspar (Or56-100), muscovite (±2 poids en pour-cent de FeO, ±0.8 poids en pour-cent de F), chlorite, et kaolinite, tandis que les sulfures incluent la sphalérite pauvre en Fe (±4 poids en pour-cent de Fe), la galène, la chalcopyrite, et une variété de sulfures secondaires de Cu. La datation (⁴⁰Ar/³⁹Ar) de la veine de muscovite et du dyke de roche a lèvre inférieure indiquent des âges semblables de 370 Ma ; donc, la formation de veine et l’injection de dyke sont interprétées pour être coincidentes avec la mise en place Batholite Musquodoboit de 370 Ma.

Les études d’inclusions fluides « L-V » et aquauses, indiquent les températures d’homogénéisation de c.140 ± 5°C et salinités de 20 % équivalent poids en pour-cent ± 2 NaCl, mais des quantités mineures d’inclusions avec plus haut (24 à 27 poids %) et abaissement (8-16 poids %) des salinités produites. Les premières températures et analyse de fonte des monticules décrits indiquent deux types de liquides, un fluide plus abondant de NaCl-H₂O et un fluide moins abondant de NaCl-CaCl₂-KCl-H₂O.

Les isotopes de soufre pour la galène hypogée, la chalcopyrite, et la sphalérite sont uniformes et égalisent à δ³⁴S de +4.2 à +6.6 (à 300°C), semblable a des données pour d’autre minéralisation de sulfure jumeler au granite dans la zone de Meguma. Le quartz de veine a une valeur de δ¹⁸O de +15.3 ± 1.21 (n=10), qui égale à δ¹⁸Owater de +9 à +11.5 à 300°C, tandis que les extraits liquides d’inclusion indiquent des valeurs de 8D de -90 à -113 (n=6).

Collectivement, la synchronisation de la minéralisation (c.-à-d., 370 Ma) et la signature isotopique et géochimique principalement magmatique pour le fluide provenant de veine suggèrent une association génétique avec le Batholite de Musquodoboit. Cependant, la nature riche en calcium de quelques inclusions liquides et valeurs basses 8D d’extraits liquides suggèrent la participation d’un autre réservoir, reflétant probablement un fluide qui a équilibré avec la roche de la région du Groupe de Meguma.

[Traduit par la rédaction]
INTRODUCTION

The Dunbrack Pb-Cu-Zn-Ag deposit, located 4.8 km northwest of the village of Musquodoboit Harbour east of Halifax, represents vein-type mineralization within the 370 Ma Musquodoboit Batholith (Fig. 1). The deposit was discovered in 1888 and worked intermittently over the next 40-50 years via two shafts sunk in order to evaluate the extent of the mineralization (i.e., several 100 t of a few % Pb and minor Cu and Ag). The deposit is unusual because of (1) the association of tridymite - pyromorphite - meneghinite (Friedlaender 1968, 1970; Chatterjee 1977) and other minerals (see below), which are unique to this part of the Musquodoboit Batholith and mineralized granites within the Meguma Zone, (2) vein textures (e.g., comb, cockade) reminiscent of epithermal veins, and (3) a $^{40}$Ar/$^{39}$Ar date on hydrothermal sericite(?) indicating an apparent age of ca. 304 Ma for mineralization and associated alteration (MacMichael 1975), which contrasts with the 370 Ma age for the host granite. Previous work on the deposit focused on vein mineralogy, textures, and fluid inclusions (MacMichael 1975; Dickie 1978). The present study was initiated because of the apparent similarity in time of formation between Dunbrack and the nearby, carbonate-hosted Gays River Zn-Pb deposit (Fig. 1) and the presence of similar mineralization elsewhere in the Meguma Zone, such as Tobetic Lake (Corey 1995; Corey and Graves 1996) and Tangier Grand Lake (Corey 1993) (Fig. 1). In this paper we summarize the paragenesis and mineralogy of the vein, mineral and whole-rock geochemistry, fluid inclusion studies, stable isotope systematics, and $^{40}$Ar/$^{39}$Ar geochronology.

GEOLOGICAL SETTING

The Dunbrack deposit is located in the central part of the 370 Ma Musquodoboit Batholith, the second largest intrusion in the Meguma Zone (Fig 1). The batholith intruded metasedimentary rocks of the Lower Paleozoic Meguma Group. A narrow contact metamorphic aureole containing cordierite porphyroblasts is developed within a few hundred metres of the contact (Horne et al. 1998). The central part of the batholith is characterized by the presence of coarse cordierite grains in a two-mica leucomonzogranite host (MacDonald and Clarke 1985; Ham 1993), whereas the flanking parts of the batholith contain lesser amounts of cordierite. In the vicinity of the Dunbrack deposit (i.e., within 1 km) there occurs a fine-grained and texturally variable leucogranite, one of several that occur throughout the batholith. Fine-grained felsic dykes and pegmatites, with variable trends including NW (Ham 1998) are also present, one of which occurs adjacent to the mineralized vein (see below) (Fig. 2). Airborne radiometric surveys indicate that local areas of the batholith are underlain by fine-grained and/or texturally variable phases with high eU/eTh ratios (Ford 1993; Ham 1993; Fig. 1).

Numerous NW-trending faults cut the batholith (Jones and MacMichael 1976; MacDonald and Clarke 1985; Ham 1993, 1994). However, faults strongly influence the shape of the Musquodoboit Batholith and the adjacent South Mountain Batholith (Horne et al. 1992), and thus were probably active at the time of batholith emplacement. The veins at Dunbrack may be related to left-lateral offset along a fault transecting Tangier Grand Lake (Jones and MacMichael 1976).

GEOLOGY OF THE DUNBRACK DEPOSIT

The following description of the Dunbrack deposit was extracted from summaries in assessment reports in Nova Scotia Department of Natural Resources files, theses of MacMichael (1975) and Dickie (1978), and observations by the senior author based on examination of material in dump piles. The mineralization occurs within a ±1 m wide quartz-rich vein which pinches out vertically (inset Fig. 2), has a coarse leucomonzogranite hanging wall (HW) and fine grained, 1.2 m-wide dyke as a foot wall (FW) (Fig. 2). The vein, oriented 110° and dipping 62°N, was traced on surface for a distance of several 100 m and the ore grade decreases across the vein towards the HW. The dyke at the FW contact with the vein is very fine grained, highly altered, and reddish coloured as a result of hematite stain. The contact of the vein with the HW granite is a zone of silicification 1 to 3 cm wide. Based on the presence of chloritic and sericitic alteration of biotite and feldspar, respectively, in the granite, MacMichael (1975) contoured areas of increasing alteration, which showed a strong correlation with proximity to the mineralized vein (Fig. 2). This alteration suggests that the vein-forming fluids infiltrated outwards from the vein. The dyke rock is intensely altered to chlorite, hematite and sericite adjacent to the vein.

The character of the vein can be ascertained from the debris in the dump pile. Numerous boulders of brecciated material (Fig. 3) contain (1) angular fragments of altered dyke rock (red, hematitic, fine-grained, felsic rock), either in isolation or forming the nucleus of comb quartz overgrowths (i.e., cockade texture); (2) multiple layers of variably textured quartz, and (3) coarse galena occurring as massive clumps, as coarse to fine aggregates within a quartz matrix, or within altered fragments with other sulphides (e.g., chalcopyrite, sphalerite). Several generations of vein formation occurred with repeated brecciation of earlier material. Well-developed comb quartz, cockade and crustiform layers are all indicative of open-space infilling and epizonal conditions (e.g., Stanton 1972; Vearcombe 1993; Dong et al. 1995; Dong and Zhou 1996), but Dowling and Morrison (1988) noted that such textures may also occur in plutonic environments beneath epithermal settings.

MINERALOGY AND TEXTURES OF VEIN SAMPLES

The vein at Dunbrack consists of the following minerals: quartz, including tridymite(?), calcite, fluorite, galena, chalcopyrite, chalcocite, sphalerite, bornite, malachite, azurite, pyromorphite [Pb(PO$_4$)$_2$Cl], meneghinite (Pb$_{13}$Cu$_7$Sb$_7$S$_{26}$), pyrrhotite, cerussite, limonite, djurleite (Cu$_{19}$S), and digenite (Cu$_8$S). Assays of the vein indicate the presence of anomalous Au (to 30 ppm) and Ag (10's of ppm), but their mineralogic expression has not yet been determined. Details of the mineralogy are described below.
PALEOZOIC - MESOZOIC
sedimentary and volcanic rocks

DEVONIAN - CARBONIFEROUS
granitoid rocks

ORDOVICIAN - SILURIAN
sedimentary and volcanic rocks

CAMBRIAN - ORDOVICIAN
MEGUMA GROUP
sedimentary rocks

Fault . . . . . . . ~~~~

Fig. 1. Geological map of southern Nova Scotia showing the Musquodoboit Batholith (MB) and the location of the Dunbrack showing and other places mentioned in the text. Bottom diagram shows the general geology of the Musquodoboit Batholith (after Ham 1993, 1994, pers. commun., 1996).
Fig. 2. Location of the Dunbrack deposit with contours for percentage of alteration of biotite to chlorite and feldspar to sericite in the Musquodoboit Batholith (after MacMichael 1975). The inset diagram of the deposit area shows a north-south cross-section (modified from Dickie 1978).
Vein Silicates

Silicate mineralogy, in descending order of abundance, includes quartz, K-feldspar (Kf), muscovite and chlorite (Fig. 4). Quartz occurs in several different habits, the most common being (1) flamboyant, (2) radially arranged acicular prisms in comb texture, with growth banding defined by numerous milky white zones containing abundant fluid and silicate inclusions, and (3) equant crystals in places occurring with Kf (Fig. 5). Trace amounts of muscovite occur in areas of quartz inundated with inclusions, whereas hematite is sporadically distributed throughout. Banding is defined by equant quartz succeeded by acicular flamboyant comb quartz which widens outwards from the base. Multiple layers or generations of quartz may occur within a single sample. In areas of brecciation, angular pieces of quartz suggest fragmentation or comminution of massive quartz vein material. K-feldspar occurs as equant to anhedral grains intergrown with quartz and may account for 30 to 40% of some areas of the vein (Fig. 5); it is commonly altered to hematite or kaolinite. In some glomeroclasts (see below) granophyric textures occur. Muscovite occurs as coarse, equant to tabular grains that may be bent or kinked (Fig. 5), within the quartz vein or as part of altered fragments coexisting with quartz and Kf. Muscovite may be partially replaced by quartz and also occurs as small grains disseminated in cores of quartz grains. Chlorite is rare and probably represents altered biotite. Fragments of altered granite (?) also occur in the vein (Fig. 5).

Also present in the vein are oval, equant, and irregularly-shaped areas of coarser-grained (i.e., relative to the matrix material of the vein), equigranular, glomeroclastic aggregates of quartz, Kf, muscovite, chlorite/biotite, in which hematite and kaolinite, after Kf, are common (Fig. 5). The quartz in this occurrence may contain small euhedral zircon.

The matrix to the coarse vein material consists of very fine-grained (i.e., grain size of a few \( \mu \text{m} \)) material with a banded texture that is perpendicular to the comb quartz fabric (Fig. 4b, c, d, e). This material is reddish in hand specimen as a result of hematitic alteration, perhaps after feldspar. Fragments of this material are common in the vein.

Ore Mineralogy of the Vein

The ore mineralogy is dominated by hypogene galena, chalcopyrite and sphalerite with lesser amounts of supergene Cu sulphides (Fig. 6). The primary sulphides occur infilling original porosity where quartz terminates as euhedral crystals in open spaces, or as disseminations in the matrix of fine-
grained vein material. The supergene minerals rim chalcopyrite or form along fractures with the alteration widest at grain margins and tapering inwards. The assemblage of secondary minerals is variable and several minerals may be intergrown with complex textural relationships.

**Analytical Techniques**

Major and trace element compositions were obtained on a sample of dyke rock using wet chemical techniques at DalTech, formerly TUNS, in Halifax, and also at Memorial University of Newfoundland using solution ICP-MS (Jenner et al. 1990). Mineral chemistry was done at Dalhousie University using a JEOL Superprobe (details in Kontak and Smith 1993). Details of stable isotope data (at the University of Saskatchewan) and fluid inclusion studies (at the Nova Scotia Department of Natural Resources) are fully described in Kontak et al. (1996) and Kontak (1998), respectively. The procedures for \(^{40}\)Ar/\(^{39}\)Ar dating (at Queen's University) are given in Erdmer et al. (1998).

**\(^{40}\)Ar/\(^{39}\)Ar Dating**

A sample of altered, fine-grained, reddish felsic dyke rock occurring as fragments in vein breccia and a hand-picked separate of coarse-grained muscovite from a silicate-sulphide vein sample were dated using the \(^{40}\)Ar/\(^{39}\)Ar method. The age spectrum for the whole-rock sample (Fig. 7a; Table 1) is characterized by ages of ca. 360 to 370 Ma for the first 35% of the gas, but with large errors related to the low potassium content and "dirty" nature of the sample. The central part of the age spectrum has ages of 330 to 355 Ma and the high temperature steps have ages of ca. 280 Ma. The integrated age for this sample is 335 ± 17 Ma and the isotope correlation age is 354 ± 56 Ma.

The age spectrum for the muscovite separate (Fig. 7b; Table 1) is characterized by a hump for the low temperature steps which give ages of ca. 385 Ma, but the majority (94%) of the release spectrum consists of steps that fall in the narrow interval of 369 to 376 Ma. The highest temperature analyses (54% of total gas) define a plateau age of 368.5 ± 3.0 Ma. The integrated age for the sample is 370.5 ± 2.5 Ma.

Seventeen spot dates on muscovite (Table 1) from the mineral separate indicate apparent ages of 366 to 373 Ma, except for one anomalous age of 404 Ma, and integrated and correlation ages identical to the age spectrum (Fig. 7b).

Fig. 4. Photomicrographs (all cross nicols) of vein samples from the Dunbrack deposit. (a) Vein sample illustrating series of events which start with crystallization of fine-grained matrix material (stage 1), growth of comb quartz layers after vein opening to generate open space (stage 2), and finally a cross vein (stage 3) which cross cuts all the above. Width of photo is 1.2 cm. (b) Brecciated quartz fragments and fine-grained matrix material infilling space lined with coarse quartz euhedra. Note the bimodal grain size for vein fill, lack of undulose extinction in the quartz, and absence of any comb or flamboyant textures in the quartz fragments. Width of photo is 2.0 cm. (c) Vein sample with textures indicating multiple stages of vein growth both in the left and right of photo. Note that the quartz in the centre layer is truncated by an unconformity surface. Width of photo is 2.0 cm. (d) Detail of part of the sample shown in Fig. 4c to illustrate the banding in the matrix material (flowage feature?) and the comb quartz layers, the textures of which indicate infilling of open space as in a vein with the matrix material at top and bottom forming the substrate. Width of photo is 1 cm. (e) Comb quartz in detail showing the presence of primary growth faces as outlined by areas inundated by fluid inclusions. The dark areas in the growth bands are void spaces. Width of is photo 1.6 mm.

Fig. 5. Back-scattered electron image of vein material illustrating a hypidiomorphic texture for material consisting of quartz (Qtz), K-feldspar (Kf), and muscovite (Ms). Other abbreviations: fragment (Frg).
Fig. 6. Reflected light (a, b) and transmitted light (c) photomicrographs of ore minerals at Dunbrack. (a, b) Chalcopyrite (cpy) filling open space between euhedral quartz grains. The chalcopyrite is cut and rimmed by secondary Cu-Fe sulphides (cc), including covellite, bornite, djurleite and chalcocite, and surrounded by galena (gal). Widths of photos in a and b are 1.2 mm and 0.3 mm, respectively. (c) Sphalerite (centre of photo) between comb quartz (top) and fine-grained matrix material containing disseminated galena (black). Width of photo is 1.2 mm.
Table 1. Summary of data for $^{36}$Ar/$^{40}$Ar step-heating and laser dating of Dunbrack samples.

<table>
<thead>
<tr>
<th>Step</th>
<th>$^{36}$Ar/$^{40}$Ar (X 0.00001)</th>
<th>$^{39}$Ar/$^{40}$Ar (X 0.001)</th>
<th>Ca/K %40Ar (atm)</th>
<th>%39 Ar 40Ar * / 39Ar</th>
<th>Age (±) Ma</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>110.66(85.5)</td>
<td>44.64(2.30)</td>
<td>0.00</td>
<td>32.63</td>
<td>0.38</td>
</tr>
<tr>
<td>2</td>
<td>64.02(26.2)</td>
<td>29.55(0.70)</td>
<td>0.03</td>
<td>18.90</td>
<td>1.04</td>
</tr>
<tr>
<td>3</td>
<td>18.88(16.7)</td>
<td>29.88(0.50)</td>
<td>0.40</td>
<td>5.57</td>
<td>1.55</td>
</tr>
<tr>
<td>4</td>
<td>4.34(10.1)</td>
<td>30.99(0.42)</td>
<td>1.39</td>
<td>1.10</td>
<td>5.74</td>
</tr>
<tr>
<td>5</td>
<td>3.73(8.00)</td>
<td>32.04(0.35)</td>
<td>1.69</td>
<td>1.16</td>
<td>8.23</td>
</tr>
<tr>
<td>6</td>
<td>3.92(8.20)</td>
<td>32.18(0.34)</td>
<td>1.83</td>
<td>1.30</td>
<td>9.27</td>
</tr>
<tr>
<td>7</td>
<td>4.4(9.60)</td>
<td>32.42(0.24)</td>
<td>2.92</td>
<td>1.83</td>
<td>8.74</td>
</tr>
<tr>
<td>8</td>
<td>6.19(10.0)</td>
<td>32.24(0.36)</td>
<td>3.48</td>
<td>2.44</td>
<td>7.33</td>
</tr>
<tr>
<td>9</td>
<td>7.65(10.6)</td>
<td>32.23(0.40)</td>
<td>3.15</td>
<td>26.00</td>
<td>14.00</td>
</tr>
<tr>
<td>10</td>
<td>8.26(10.2)</td>
<td>32.43(0.40)</td>
<td>3.27</td>
<td>2.60</td>
<td>30.76(0.04)</td>
</tr>
<tr>
<td>11</td>
<td>9.55(11.2)</td>
<td>32.44(0.43)</td>
<td>3.04</td>
<td>3.75</td>
<td>30.95(0.33)</td>
</tr>
<tr>
<td>12</td>
<td>10.03(11.1)</td>
<td>32.14(0.41)</td>
<td>3.36</td>
<td>2.96</td>
<td>27.51</td>
</tr>
</tbody>
</table>

DB-96-3 Muscovite; Laser Spot Dates; J Value = 0.007412

<table>
<thead>
<tr>
<th>Spot</th>
<th>0.004(154.6)</th>
<th>29.48(2.75)</th>
<th>0.01</th>
<th>13.59</th>
<th>2.71</th>
<th>30.69(0.79)</th>
<th>369.83(8.61)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>46.03(5.75)</td>
<td>28.14(0.46)</td>
<td>0.01</td>
<td>3.71</td>
<td>1.74</td>
<td>30.31(0.85)</td>
<td>365.69(9.33)</td>
</tr>
<tr>
<td>3</td>
<td>12.56(8.02)</td>
<td>32.12(0.57)</td>
<td>0.01</td>
<td>2.73</td>
<td>1.74</td>
<td>30.83(0.49)</td>
<td>371.33(5.39)</td>
</tr>
<tr>
<td>4</td>
<td>9.23(8.15)</td>
<td>32.08(0.60)</td>
<td>0.01</td>
<td>0.86</td>
<td>5.89</td>
<td>30.83(0.49)</td>
<td>369.36(6.12)</td>
</tr>
<tr>
<td>5</td>
<td>2.90(2.76)</td>
<td>32.15(0.46)</td>
<td>0.01</td>
<td>1.39</td>
<td>1.63</td>
<td>30.31(0.85)</td>
<td>366.26(5.39)</td>
</tr>
<tr>
<td>6</td>
<td>4.69(3.85)</td>
<td>32.17(0.45)</td>
<td>0.01</td>
<td>1.71</td>
<td>1.78</td>
<td>30.31(0.85)</td>
<td>369.36(6.12)</td>
</tr>
<tr>
<td>7</td>
<td>5.79(6.43)</td>
<td>32.11(0.63)</td>
<td>0.01</td>
<td>4.58</td>
<td>3.12</td>
<td>29.92(0.62)</td>
<td>361.36(6.84)</td>
</tr>
<tr>
<td>8</td>
<td>15.52(5.06)</td>
<td>31.88(0.52)</td>
<td>0.01</td>
<td>1.71</td>
<td>13.14</td>
<td>30.87(0.30)</td>
<td>371.77(3.31)</td>
</tr>
</tbody>
</table>
Fig. 7. $^{40}$Ar/$^{39}$Ar age spectrum plots for samples of fine-grained felsic dyke rock forming footwall to vein (DB-96-5 WR) and vein muscovite (DB-96-3 Ms) from Dunbrack.

**WHOLE-ROCK AND MINERAL CHEMISTRY**

**Dyke Rock**

A single sample (DB-96-5) of the fine-grained, reddish material used for dating purposes, possibly representing altered dyke rock that forms the hanging wall to the mineralized vein, was analysed. The analysis indicates 97.8 wt. % SiO$_2$, 2.5 wt. % Al$_2$O$_3$ and minor to trace Na$_2$O (0.14), CaO (0.35), K$_2$O (0.2), MgO (0.87), P$_2$O$_5$ (0.08), MnO (0.013) and TiO$_2$ (0.013). Thus, the data indicate that the dyke rock has been intensely silicified, at least locally. Rastering analysis of similar material using the electron microprobe (at Dalhousie University) also indicates that the material is highly altered, with a composition dominated by silica and alumina, but with up to 3-4 wt. % FeO and 1-2 wt. % K$_2$O locally.

Trace element analyses indicate the following (in ppm): Rb=42, Li=18, Sr=42, Mo<1, Cs=3, Ba=1, Pb=300, U=3.7, Th=1.2, Zr=12. Of note with respect to the trace element chemistry is that the U/Th ratio of 2.5 is similar to that of felsic rocks in general, and that the Zr content is high for a hydrothermal sample, possibly suggesting the presence of relict zircon. A chondrite-normalized plot for rare earth element (REE) data (Fig. 8) indicates a surprising enrichment in REE, in particular the LREE, given the high silica content of the sample. The REE profile is LREE enriched, strongly fractionated (La$_n$/Lu$_n$ = 250), and has a negative Eu anomaly (Eu$_n$/Eu$_{n^*}$ = 0.23). In contrast, dyke rocks and porphyries from the Musquodoboit Batholith (after MacDonald and Clarke 1985) are relatively depleted in REE and less fractionated (Fig. 8).

**Vein Minerals**

Alkali feldspar is homogeneous with no exsolution lamellae of albite observed, has compositions ranging between Or$_{86}$Ab$_{14}$ and Or$_{100}$ (Fig. 9a; Table 2), and may contain up to 0.4 wt. % P$_2$O$_5$. Muscovite is phengitic with up to 2 wt. % FeO and minor MgO (Table 2), and contains a maximum of 0.8 wt. % F with an inverse correlation between F and FeO (i.e., the Fe-F avoidance principle; Fig. 9b). Data for muscovite from the Musquodoboit Batholith indicate only trace amounts of F (L. Ham, personal communication, 1998). Chlorine was not detected in any of the muscovite grains analysed. Chlorite, after biotite, is epidotilistic in composition with Fe/(Fe+Mg) values of 0.6 to 0.7 (Table 2). Kaolinite was also analysed and stoichiometric chemistry obtained (Table 2).

Analyses of sulphides (Table 3) indicate that galena is generally pure with only trace amounts of Ag detected in a few analyses. Sphalerite is Fe-poor with a maximum of 4 wt. % Fe, but most analyses show Fe to be below the detection limit (i.e., <0.2 wt. % Fe). Grains may be optically zoned and the highest Fe is found in reddish as compared to yellow-red areas. Chalcopyrite and bornite are stoichiometric with no impurities. Secondary sulphides analysed include a variety of Cu sulphides (i.e., covellite (CuS), Cu$_{11}$S$_{9}$, Cu$_{11}$S$_{9}$, Cu$_{3}$FeS$_{4}$, Cu$_{5}$FeS$_{4}$) and variable mixtures of Cu-Fe-Pb-S, Cu-Pb-S, Zn-Cu-Fe-S. In the case of the latter phases (i.e., mixtures), some contamination from underlying phases is a possibility and further work will be required to clarify this mineralogy. The highest concentration of Ag (to 2.5 wt. %) is found in an unidentified Cu-Pb-S mineral (Table 3).

**FLUID INCLUSION STUDIES**

**Petrography**

Representative samples of vein quartz were cut to 150 μm and doubly polished for fluid inclusion petrography. The samples are similar, with three inclusion types dominant and a rare fourth type (Fig. 10a). The most abundant inclusions are irregular-shaped to equant aqueous inclusions of ≤5 to 30 μm with highly variable liquid:vapor (L:V) ratios; in fact, the dominance of monophase L- and V-rich types suggest post-entrapment modification (i.e., necking). The irregular-shaped inclusions are far more abundant (i.e., >90%) than the equant inclusions. This population of inclusions (Fig. 11a, b, c) occurs along growth zones in quartz (i.e., primary origin) and along healed fracture planes (i.e., secondary or pseudosecondary origin). The second type of inclusions are equant in quartz and elongate, two-phase L-V inclusions mostly ≥20–30 μm, but up to 150 μm, that occur along healed fracture planes in clear quartz or rarely cross-cutting milky white quartz (Figs. 10b, 11d, e, f, g). These inclusions are conservatively
classified as secondary, although some may in fact be pseudosecondary. The third type of inclusions includes equant to irregular-shaped monophase, L-type inclusions, ca. 10-20 μm in length, occurring along healed fracture planes (Fig. 11h). The fourth inclusion type, confined to a single locality in one sample (Fig. 10c), consists of liquid petroleum and its composition was confirmed by its fluorescence under UV light.

**Thermometric measurements**

Homogenization measurements (T_h) were determined for four quartz samples containing type 2 inclusions, some of which occur with type 4 inclusions, and very uniform results were obtained, as summarized in Figure 12. The range of T_h measurements for all inclusions was 124° to 151°C, but 95% of the data fall in the range of 134° to 142°C and in any group of inclusions the range is < 1°C (e.g., area in Figure 11e). These results contrast to the data of Dickie (1978; Fig. 12a) in which a much larger range of T_h was obtained. The reason for the discrepancy between the two studies is unclear, although one possibility is that necked inclusions may have been included in the population studied by Dickie (1978), as a similarly large range in T_h was observed for groups of inclusions with variable L:V ratios (i.e., necked inclusions), although such data are not presented in Figure 12a.

Freezing of inclusions indicated that first melting occurred at either -50° to -40°C or -28° to -21°C, which indicates that fluids contained additional cationic species in solution other than NaCl in the former group, probably Ca and Mg (Roedder 1984; Crawford 1981). Melting of hydrohalite occurred at three different intervals: -35° to -30°C, -23° to -21°C and rarely (n=4) -15° to -5°C; thus, variable NaCl/(NaCl+CaCl2) ratios are indicated. However, ca. 90% of inclusions have hydrohalite melting near -21°C, thus the fluids are best approximated by the NaCl-H2O system with minor amounts of a NaCl-CaCl2-H2O fluid. Last melting of ice and hydrohalite indicates that inclusion salinity (Fig. 12b) is 20 ± 2 wt. % eq. NaCl, with a few inclusions exhibiting both higher (i.e., 24 to 27 wt. %) and lower (i.e., 8 and 16 wt. %) salinities. Thus, at least two distinct fluid compositions are indicated by the freezing experiments, a dominant NaCl-H2O fluid of ca. 20 wt. % eq. NaCl, and a second, much less abundant (i.e., <5% of inclusions), of NaCl-CaCl2-H2O composition but with similar salinity.

It is also noted that type 2 inclusions occurring with type 1 and 3 inclusions have salinities of ca. 20 wt. % NaCl, which suggests that similar fluids are represented by these inclusion populations.
Table 2. Representative analyses of vein silicates, Dunbrack deposit, Nova Scotia.

<table>
<thead>
<tr>
<th>Sample Point</th>
<th>K-feldspar</th>
<th>Ceolite</th>
<th>Kaolinite</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>DB-96-8</td>
<td>DB-96-8</td>
<td>DB-96-8</td>
</tr>
<tr>
<td>Point</td>
<td>6</td>
<td>7</td>
<td>8</td>
</tr>
<tr>
<td>SiO₂</td>
<td>63.61</td>
<td>63.30</td>
<td>63.85</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>18.57</td>
<td>18.71</td>
<td>18.65</td>
</tr>
<tr>
<td>FeO</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>MnO</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>MgO</td>
<td>*</td>
<td>*</td>
<td>*</td>
</tr>
<tr>
<td>CaO</td>
<td>nd</td>
<td>nd</td>
<td>nd</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.06</td>
<td>0.90</td>
<td>0.95</td>
</tr>
<tr>
<td>K₂O</td>
<td>14.92</td>
<td>15.62</td>
<td>15.19</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>nd</td>
<td>nd</td>
<td>0.35</td>
</tr>
<tr>
<td>Total</td>
<td>98.16</td>
<td>98.53</td>
<td>98.99</td>
</tr>
</tbody>
</table>

Note: *, element not analyzed; nd, below detection limits (0.2%)

Fig. 9. Mineral chemistry for Dunbrack samples. (a) Histogram plot of mole % Or for K-feldspar grains. (b) Plot of wt. % F versus wt. % ΣFeO for muscovite.
### Table 3. Representative analysis of sulphide phases from Dunbrack deposit, Nova Scotia.

<table>
<thead>
<tr>
<th>Sample Point</th>
<th>DB-96-3</th>
<th>DB-96-3</th>
<th>Hypogene phases</th>
<th>DB-96-3</th>
<th>DB-96-1</th>
<th>DB-96-1</th>
<th>DB-96-1</th>
</tr>
</thead>
<tbody>
<tr>
<td>Phase</td>
<td>Cpy</td>
<td>Cpy</td>
<td>Gal</td>
<td>Gal</td>
<td>Sph</td>
<td>Sph</td>
<td>Sph</td>
</tr>
<tr>
<td>Sample</td>
<td>1</td>
<td>3</td>
<td>1</td>
<td>3</td>
<td>1</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>Weight % Cu</td>
<td>34.45</td>
<td>34.93</td>
<td>87.44</td>
<td>87.65</td>
<td>66.31</td>
<td>66.45</td>
<td>62.95</td>
</tr>
<tr>
<td>Pb</td>
<td>30.09</td>
<td>30.66</td>
<td>12.87</td>
<td>12.57</td>
<td>32.63</td>
<td>32.49</td>
<td>32.81</td>
</tr>
<tr>
<td>Zn</td>
<td>34.65</td>
<td>34.77</td>
<td>100.31</td>
<td>100.23</td>
<td>99.99</td>
<td>99.73</td>
<td>100.56</td>
</tr>
<tr>
<td>Fe</td>
<td>99.19</td>
<td>100.36</td>
<td>99.99</td>
<td>99.73</td>
<td>100.56</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Ag</td>
<td>25.08</td>
<td>25.18</td>
<td>51.25</td>
<td>51.89</td>
<td>49.45</td>
<td>49.72</td>
<td>46.41</td>
</tr>
<tr>
<td>S</td>
<td>49.99</td>
<td>49.67</td>
<td>48.74</td>
<td>48.10</td>
<td>49.62</td>
<td>49.58</td>
<td>49.33</td>
</tr>
<tr>
<td>Total</td>
<td>99.35</td>
<td>98.21</td>
<td>92.18</td>
<td>99.61</td>
<td>104.13</td>
<td>100.36</td>
<td>93.32</td>
</tr>
<tr>
<td>Atomic % Cu</td>
<td>42.06</td>
<td>41.12</td>
<td>49.38</td>
<td>48.26</td>
<td>54.31</td>
<td>19.90</td>
<td>18.40</td>
</tr>
<tr>
<td>Pb</td>
<td>11.38</td>
<td>11.62</td>
<td>2.94</td>
<td>9.00</td>
<td>15.51</td>
<td>12.13</td>
<td>1.08</td>
</tr>
<tr>
<td>Zn</td>
<td>46.55</td>
<td>47.01</td>
<td>49.10</td>
<td>48.26</td>
<td>48.24</td>
<td>48.29</td>
<td>47.82</td>
</tr>
<tr>
<td>Fe</td>
<td>30.88</td>
<td>31.03</td>
<td>31.02</td>
<td>32.25</td>
<td>29.06</td>
<td>31.12</td>
<td>31.97</td>
</tr>
<tr>
<td>Ag</td>
<td>55.32</td>
<td>53.81</td>
<td>61.16</td>
<td>63.93</td>
<td>33.44</td>
<td>34.78</td>
<td></td>
</tr>
<tr>
<td>S</td>
<td>62.95</td>
<td>66.45</td>
<td>87.65</td>
<td>66.31</td>
<td>52.39</td>
<td>34.32</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>99.19</td>
<td>100.36</td>
<td>99.73</td>
<td>100.56</td>
<td>98.04</td>
<td>96.49</td>
<td></td>
</tr>
</tbody>
</table>

**Abbreviations:** Cpy, chalcopyrite; Gal, galena; Sph, sphalerite; Born, bornite; Cov, covellite.

### Decrepitate analysis

Analysis of decrepitate mounds in areas of clear quartz were performed in order to determine the solute chemistry of the fluid inclusions (e.g., Haynes et al. 1988). The procedure involves decrepitating of fluid inclusions in the heating stage which results in formation of chemical precipitates (e.g., NaCl, CaCl₂, KCl) or mounds on the host surface, as illustrated with back-scattered electron images in Figure 13. Analysis of the mounds using a rastering technique with a defocused (10 μm) electron beam indicated that the majority of the mounds formed are equant (Fig. 13a) and consist of NaCl (Fig. 13d). Abundant voids within the quartz (Fig. 13b) may reflect evacuation of low-salinity inclusions. Skeletal mounds (Fig. 13c) are invariably K- and Ca-rich and form a discrete chemical population in Figure 13d. This last group may correspond to those inclusions with low eutectic temperature (i.e., -50°C) in which hydrohalite melted at -35°C to -30°C (i.e., elevated CaCl₂ chemistry). Thus, the decrepitate data suggest that two, and possibly three, compositionally distinct fluids may be present.

### Stable isotopes (S, O, D)

**Sulphur isotopes**

Sulphur isotopic data have been obtained for chalcopyrite (n=1), sphalerite (n=1), and galena (n=3) and δ⁴⁴S values range from +6.9‰ (sph) to -3.1‰ (Table 4). The results overlap most of the δ⁴⁴S data for sulphide mineralization in the South Mountain Batholith (Fig. 14). The sphalerite and chalcopyrite represent a sulphide pair from the same sample (DB-96-3), but their δ⁴⁴S value does not yield a reasonable temperature of vein formation (i.e., 250°C to 300°C, see below) and suggests disequilibrium. The presence of secondary
Quartz euhedra decorated with primary and pseudosecondary inclusions

Primary vapor- and liquid-rich aqueous inclusions; irregular L:V ratios

Pseudosecondary inclusions - \( L_{H_2O} - V \); uniform L:V ratios

Monophase aqueous

Liquid petroleum
Fig. 10. (a) Schematic illustration of the four types of fluid inclusions hosted by quartz at Dunbrack. (b) Plane of pseudosecondary, negative-shaped, two-phase (L-V) aqueous inclusions, in quartz at Dunbrack. These inclusions are isolated and occur in the same area as inclusions in Figure 10b. (c) Two-phase L-V petroleum inclusions in quartz from Dunbrack. These inclusions are isolated and occur in the same area as inclusions in Figure 10b.

Fig. 11. Photomicrograph of fluid inclusions hosted by vein quartz at Dunbrack. Photos taken in transmitted light. (a, b) Areas inundated with primary fluid inclusions along growth zones in comb quartz. (c) Close-up of area inundated with primary inclusions of irregular shape; in this case the inclusions are dominantly V-rich types and probably reflect post entrapment necking. (d) Two parallel fracture planes decorated by secondary or pseudosecondary, equant- to negative shaped L-V fluid inclusions (enlargement to the right) that homogenize at 140°C. (e) Abundant, elongate to negative-shaped, pseudosecondary, fluid inclusions decorating a healed fracture in comb quartz. These inclusions homogenized at 137 ± 0.5°C (n=20). (f) Isolated L-V fluid inclusions of somewhat irregular shape that occur with equant- and negative-shaped fluid inclusions with similar L-V ratios. (g) Area of irregular-shaped fluid inclusions that show evidence of necking, but note that L-V ratios are consistent which indicates necking occurred prior to separation of the V-phase. (h) Plane of equant-shaped monophase (L) inclusions.
Fig. 12. Histogram plots summarizing homogenization temperatures (a) and salinity measurements (b) on two-phase (L-V) aqueous inclusions in quartz from Dunbrack. Note the difference in homogenization temperatures between this study and those of Dickie (1978), with the latter having a much larger range of values compared to the results of the present study.
Fig. 13. Back-scattered electron images (a, b, c) and compositional data (d) for decrepitate mounds hosted by quartz from Dunbrack deposit. Note that the dendritic to skeletal shapes in Figure 13c are typical of K-rich salts and correspond to one of the populations in Figure 13d. In contrast, the equant mounds in Figure 13a are Na-rich and plot near the apex of the K-Na-Ca ternary in Figure 13d.

copper sulphides along the margins and internal fractures of chalcopyrite suggest, therefore, that the chalcopyrite $\delta^{34}$S data may not represent a primary signature.

In order to interpret the $\delta^{34}$S data in the context of source reservoirs, fractionation of $^{34}$S between mineral and fluid (i.e., $\Delta^{34}$S$_{\text{S-S}}$-sulfide) must be considered. Using the fractionation factors compiled by Ohmoto and Rye (1979), the $\delta^{34}$S$_{\text{H$_2$S}}$ values of the fluid compute to $+4.2$ to $+6.6\%$ for ca. 300°C, except for the anomalous galena sample (DB-96-12) which seems to have been deposited from a fluid with a lighter $\delta^{34}$S$_{\text{H$_2$S}}$ signature ($-1.2\%$), possibly reflecting involvement of another reservoir.

**Oxygen isotopes**

Oxygen isotopes have been determined in nine quartz separates, which represent the variety of textures and colours of quartz present, and in a single sample of kaolinite. The $\delta^{18}$O data for quartz range from $+13.0$ to $+17.6\%$ and average $+15.3 \pm 1.2\%$; there is no distinction of the data that would
Table 4. Summary of isotopic data, Dunbrack deposit.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Mineral</th>
<th>$\delta^{18}O$ ($%$)</th>
<th>$\deltaD$ ($%$)</th>
<th>$\delta^{34}S$ ($%$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>DB-96-1.1</td>
<td>Qtz</td>
<td>14.3</td>
<td>-90</td>
<td></td>
</tr>
<tr>
<td>DB-96-1.2</td>
<td>Qtz</td>
<td>15.7</td>
<td>-90</td>
<td></td>
</tr>
<tr>
<td>DB-96-3</td>
<td>Qtz</td>
<td>15.5</td>
<td>-91</td>
<td></td>
</tr>
<tr>
<td>DB-96-3</td>
<td>Kaol</td>
<td>17.1</td>
<td>-103</td>
<td></td>
</tr>
<tr>
<td>DB-96-3</td>
<td>Sph</td>
<td>17.1</td>
<td>-103</td>
<td></td>
</tr>
<tr>
<td>DB-96-3</td>
<td>Cpy</td>
<td>13</td>
<td>-94</td>
<td></td>
</tr>
<tr>
<td>DB-96-5.1</td>
<td>Qtz</td>
<td>15.2</td>
<td>-100</td>
<td></td>
</tr>
<tr>
<td>DB-96-6.2</td>
<td>Qtz</td>
<td>15.2</td>
<td>-100</td>
<td></td>
</tr>
<tr>
<td>DB-96-9</td>
<td>Qtz</td>
<td>14.8</td>
<td>-108</td>
<td></td>
</tr>
<tr>
<td>DB-96-9</td>
<td>Gal</td>
<td>17.1</td>
<td>-103</td>
<td></td>
</tr>
<tr>
<td>DB-96-10.1</td>
<td>Qtz</td>
<td>15.9</td>
<td>-100</td>
<td></td>
</tr>
<tr>
<td>DB-96-10.2</td>
<td>Qtz</td>
<td>17.6</td>
<td>-131</td>
<td></td>
</tr>
<tr>
<td>DB-96-10.3</td>
<td>Qtz</td>
<td>15.7</td>
<td>-113</td>
<td></td>
</tr>
<tr>
<td>DB-96-11</td>
<td>Gal</td>
<td>15.9</td>
<td>-100</td>
<td></td>
</tr>
<tr>
<td>DB-96-12</td>
<td>Gal</td>
<td>15.9</td>
<td>-100</td>
<td></td>
</tr>
</tbody>
</table>

Abbreviations: Qtz, quartz; Kaol, kaolinite; Sph, sphalerite; Cpy, chalcopyrite; Gal, galena.

correspond to a particular variety or colour of quartz. However, a variation of up to 2% is recorded within one sample from which 3 different generations of quartz were sampled (DB-96-10.1, 10.2, 10.3; Table 4).

Using the quartz-water fractionation equation, the $\delta^{18}O_{\text{water}}$ for a given temperature can be determined from the measured $\delta^{18}O_{\text{quartz}}$ value (Fig. 15). Assuming a similar temperature of ca. 300°C (see below for reasoning) for precipitation of all quartz samples, the equivalent $\delta^{18}O_{\text{water}}$ is +9 to +11.5%, but a $\delta^{18}O_{\text{water}}$ value of +14% would characterize the most enriched sample (DB-96-10.2).

The single kaolinite sample analysed for $\delta^{18}O$ has a value of +17.1%. This value compares, for example, to values of +13.8% for late-stage kaolinite at East Kemptville (Kontak 1994), and +17.2 and +18.7% for kaolinite from Cornwall, England (Jackson et al. 1982).

**Hydrogen isotopes**

Fluid inclusions from seven samples of quartz yielded $\deltaD$ values between -90 to -131% (avg. -100 ± 14%) and one kaolinite yielded a value of -103%. The fluid inclusion $\deltaD$ data have a narrow spread even though they represent sampling of mixed fluid inclusion types. The kaolinite sample has a $\deltaD$ value of -103%, which overlaps the composition of fluids extracted from quartz.

**DISCUSSION**

The nature of the vein mineralogy and textures at Dunbrack indicate that hydrothermal fluids exploited an open fracture system, previously used during magma injection to form the dyke. Examination of the vein textures, silicate and sulphide mineralogy, and fluid inclusions permit conclusions to be made regarding the origin of the deposit.

**Timing of vein formation**

The vein formed after the monzogranite and the footwall dyke, because both have been altered during vein formation. The age of the monzogranite is constrained by the age of the Musquodoboit Batholith, which is ca. 370 Ma as determined by $^{40}\text{Ar}^{39}\text{Ar}$ dating of mica (Reynolds et al. 1981; Keppie and Dallmeyer 1987). The $^{40}\text{Ar}^{39}\text{Ar}$ age spectrum on altered felsic dyke rock, despite being of low quality, does provide a minimum age of 366 ± 19 Ma for the hydrothermal activity, given that the sample constitutes part of the vein. Thus, the plateau age is interpreted as representing the timing of hydrothermal activity. The excellent plateau and spot ages of 370 Ma for vein-hosted muscovite are difficult to interpret unequivocally because the muscovite may represent grains...
Temperature (°C)

Fig. 15. Plot of $\delta^{18}O$ versus temperature (°C) with isopleths for $\delta^{18}O_{\text{water}}$, calculated using quartz-water fractionation equation of Clayton et al. (1972); the quartz data shown on left side of diagram. Note that the limiting values of +9 to +11.5% for $\delta^{18}O_{\text{water}}$, as discussed in the text, are for ca. 300°C.

Estimates of the temperature of vein formation from fluid inclusion homogenization temperatures ($T_h$) only provide a minimum, because a pressure correction may have to be applied if a homogeneous fluid was trapped. Based on the equant- to negative-crystal shape of the quartz-hosted fluid inclusions, and comparison of the inclusion morphology of Dunbrack samples to those from epithermal environments which formed at 200° to 240°C (e.g., Bodnar et al. 1985), higher temperatures must have prevailed and a minimum temperature of ca. 250°C is inferred (T.J. Reynolds, pers. communication, 1996). Based on the aforementioned discussion of whole-rock and mineral chemical data, minimum temperatures of vein formation are estimated at ca. 300° to 350°C. Using this temperature estimate, isochors for fluid salinities of 5 to 25 wt. % eq. NaCl, the dominant fluid composition (see above), indicate a confining pressure of ca. 3 kbars (Fig. 16), which is consistent with pressures inferred during emplacement of the 370 Ma granites of the Meguma Zone (Clarke et al. 1997). Because the vein exhibits open-space filling, it is probable that the pressure is some combination of lithostatic and hydrostatic, but the limited vertical extent of the system suggests that a hydrostatic component was minimal.
Fluid inclusion data indicate at least two moderately saline (ca. 20 wt. % eq. NaCl) fluids are present, a NaCl-H₂O fluid and a NaCl-CaCl₂-KCl-H₂O fluid, but the former is clearly the dominant fluid type based on the abundance of inclusions. These compositions are consistent with a magmatic reservoir (Roedder 1984), although the Ca-rich nature of the second fluid is also consistent with interaction of a fluid with or infiltration of a fluid from metamorphic rocks, as represented by the Meguma Group (e.g., Wilkinson 1990; Evans 1995; Pasteris et al. 1995). In addition, there is no indication from the inclusions that a meteoric fluid was involved in vein formation.

The sulphur isotope data are uniform in terms of the calculated $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ at +4.2 to +6.6‰, except for the single anomalous galena sample which gave a value of -1.2‰. The former values correspond to magmatic sulphur (Ohmoto and Rye 1979) and indicate, therefore, that sulphur was derived either directly from a magma or via dissolution of magmatic sulphides. These values are similar to the $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ for sulphides at the East Kemptville tin-base metal deposit (Kontak 1990, 1993) and intrabatholith sulphide mineralization in the South Mountain Batholith, as compiled in Figure 14. The anomalous $\delta^{34}\text{S}_{\text{H}_2\text{S}}$ value for one galena sample, despite being similar to the other two galena samples, might relate to weathering, disequilibrium processes, or indicate another reservoir for sulphur. The latter scenario is supported by the range of sulphur isotope data for the Meguma Group (Fig. 14) and, as discussed below, additional data that suggest possible incursion of an external fluid which equilibrated with the Meguma Group.

The nature of the fluid(s) responsible for vein formation is evaluated in terms of $\delta^{18}\text{O}$ signature in Figure 17. Whereas the $\delta^{18}\text{O}$ values overlap the field for magmatic and/or metamorphic fluids, the $\delta D$ values are clearly displaced from these fields, suggesting that another process or processes was/were involved. Firstly, the trend of the data in Figure 17 does not define a mixing line, suggesting that fluid compositions do not represent simple mixing of two fluids, but
instead water-rock interactions (Sheppard 1986). Although the low $\delta$D values suggest a possible model of interaction of meteoric waters at varying water:rock ratios (e.g., Sheldon et al. 1986) and elevated temperatures (W/R in Fig. 17), neither the fluid inclusion data (see above) nor the required paleolatitude of the Meguma Zone at ca. 370 Ma (Keppie 1977) support involvement of such a low-D (i.e., -130‰) meteoric water. Magmatic degassing can also generate low-D waters, but this model is not considered realistic given that it is usually documented in large, shallow-level magmatic-hydrothermal systems (e.g., Taylor 1988). Instead, the low $\delta$D values for the Dunbrack fluids are interpreted to reflect interaction of a fluid with organic (graphitic) material to produce organic water, the interaction of which is known to reduce the $\delta$D value of a fluid (see discussion of Sheppard 1986). Based on this interpretation, the $\delta$D data indicate interaction of a primary magmatic fluid with the country rock of the Meguma Group or alternatively, incursion of a metamorphically generated fluid, perhaps formed during contact metamorphism accompanying intrusion of the Musquodoboit Batholith. Regardless of the interpretation, more than one reservoir is required to account for the $\delta$D-$^{18}$O
data. A similar interpretation of low δD fluids has been made for granite-hosted Sn-W mineralization in Thailand (Linna and Williams-Jones 1994; Linnen 1998).

The kaolinite data shown in Figure 17 plot away from the quartz data, but are proximal to the magmatic field. However, most importantly the 18O-D data for kaolinite fall well away from the line defining kaolinite formed from weathering. Thus, the Dunbrack kaolinite is inferred to be of hypogene origin and probably formed during the waning stages of the hydrothermal system.

**SUMMARY**

Mineralogical, geochemical, stable isotopic, fluid inclusion, and geochronological data for a mineralized vein at the Dunbrack Pb-Zn-Cu-Ag deposit indicate that the mineralization occurred at ca. 370 Ma from dominantly magmatic fluids. However, both the compositions of fluid inclusions, as determined from thermometry and analysis of decrepitate mounds, and isotopic data indicate that whereas the magmatic fluid was responsible for mineralization, minor involvement of an externally-derived fluid probably occurred. The proximity of a fine-grained leucosome-granite body at the west end of the vein (i.e., near Pace Lake in Figure 2) and the presence of a fine-grained dyke rock adjacent to the mineralized vein are both reasons to infer a local source with generation of a late-stage fluid from an evolved, volatile-rich, felsic magma. That the vein is strongly controlled by structure indicates that the fluid exploited a pre-existing weakness. The exceptional quartz textures (e.g., cockade, comb; Figs. 3, 4) at Dunbrack are commonly associated with epithermal-type settings (e.g., Shimizu et al. 1998). However, Dowling and Morrison (1988) noted that such textures also occur in the plutonic environment at levels beneath the epithermal setting. In the case of Dunbrack, the nature of the mineralized quartz vein is suggestive of a fault-jog style vein (Hodgson 1989) which would generate a cavity for open space textures to develop, as originally proposed by MacMichael (1975) and reiterated here.

**ACKNOWLEDGEMENTS**

The project was funded by the Nova Scotia Department of Natural Resources. The analytical facilities at the University of Saskatchewan and Queen’s University are funded by NSERC Major Facilities Access and Research Grants. Discussions with A.K. Chatterjee and T.J. Reynolds contributed to our better understanding of the formation of the Dunbrack mineralization and constructive reviews of the paper by D.B. Clarke, A.J. Anderson and C. White are appreciated. This paper is published with permission of the Director, Nova Scotia Department of Natural Resources.


**Atlantic Geology**


1993 Sulphur enrichment in late stage veins, East Kemptville tin and base metal deposit, Nova Scotia, Canada: evidence for late incursion of metasedimentary-processed sulfur in a magmatic system. Economic Geology, 88, pp. 201-205.


Editorial responsibility: S.M. Barr