Fluid Participation in Deep Fault Zones: Evidence From Geological, Geochemical, and $^{18}O/^{16}O$ Relations

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Fluid incursion into fault zones and their deeper level counterparts, brittle-ductile shear zones, is examined in a number of different crustal environments. At the Grenville front, translation was accommodated along two mylonite zones and an associated boundary fault. The high- (MZ II) and low-temperature (MZ I) mylonite zones formed at 580 to 640°C and 430 to 490°C, respectively, in the presence of fluids of metamorphic orgin indigenous to the immediate rocks. A population of posttectonic quartz veins occupying brittle fractures were precipitated from fluids with extremely negative $\delta^{18}O$ at 200 to 300°C. The water may have been derived from downwards penetration into fault zones of low-^{18}O precipitation on a mountain range induced by continental collision with, uplift accommodated at deep levels by the mylonite zones coupled with rebound on the boundary faults. At Lagoa Real, Brazil, Archaean gneisses overlie Proterozoic sediments along thrust surfaces and contain brittle-ductile shear zones locally occupied by uranium deposits. Following deformation at 500 to 540°C, in the presence of metamorphic fluids and under conditions of low water to rock ratio, shear zones underwent local intense oxidation and desilicification. All minerals undergo a shift of $-10‰$, indicating discharge of meteoric water recharged formation brines in the underlying Proterozoic sediments up through the Archaean gneisses during overthrusting: about 1000 km^3 of solutions passed through these structures. At Yellowknife, a series of large-scale shear zones developed by brittle-ductile mechanisms, involving volume dilation with migration of ~ 5 wt % volatiles into the shear zone from surrounding metabasalts. This early deformation involved no departures in redox state or whole rock $\delta^{18}O$ from background states of $Fe^{2+}/\Sigma Fe = 0.72$ and 7 to 7.5‰, respectively, attesting to conditions of low water/rock. Shear zones subsequently acted as high-permeability conduits for pulsed discharge of >9 km^3 of reduced metamorphic hydrothermal fluids at 360–450°C. The West Bay fault, a late major transcurrent structure, contains massive bull quartz that grew at 200–300°C from fluids of 2 to 6‰ (possibly formation brines). In general, flow regimes in these fault and shear zones follow a sequence from conditions of high temperature and pressure with locally derived fluids at low water to rock ratios (during initiation of the structures) to high fluxes of reduced formation or metamorphic fluids along conduits as the structures propagate and intersect hydrothermal reservoirs. Later in the tectonic evolution and at shallower crustal levels there was incursion of oxidizing fluids from near-surface reservoirs into the faults.

INTRODUCTION

Fluid involvement in fault zones and their deeper level counterparts, brittle-ductile shear zones, is an established phenomenon both from contemporary observations of seismic pumping, the transient surface effusion of hydrothermal fluids following moderate to large earthquakes in consolidated rocks [*Sibson*, 1982], and from the presence of mineralization in old fault zones exhumed by erosion.

Fault and shear zones have been extensively studied from a mechanical standpoint [*Sibson*, 1977, 1981], with reference to their crustal setting [*Nicholas et al.*, 1977; *Berthe et al.*, 1979; *Wintsch*, 1979], and in terms of detailed structural analysis [*Ramsay and Graham*, 1970; *Ramsay and Allison*, 1979]. The mechanical and chemical role of fluids in fault and shear zones has been examined by a number of authors for ancient structures [*Beach and Fyfe*, 1972; *Beach*, 1976; *Kerrich et al.*, 1977a, 1980; *Fyfe et al.*, 1978], as well as in their seismically active modern counterparts [*Irwin and Barnes*, 1975; *Sibson*, 1982]. However, there is little quantitative information on the

fluid origin, temperature, or quantities of fluids transported through fault systems or of the possible evolution of fluid sources through progressive tectonic development of the structures.

This paper sets out to examine fluid regimes in fault zones, to identify the source, temperature, and magnitude of reservoirs involved, and specifically to consider the time relations of fluid incursion to fault translation. These questions have been addressed with reference to structures in three regions: (1) the Grenville front fault zone, Ontario, of Proterozoic age (~ 1 Ga), (2) the Lagoa Real district, Minas Gerais, Brazil, deformed at 820 Ma, and (3) the Yellowknife granite-greenstone belt of Archaean age (~ 2.8 Ga). Selection of the regions was based on existing documentation of the faults as major structural features of tens to hundreds of kilometers in lateral extent, which encompass a variety of crustal settings, protoliths, and styles of deformation. Furthermore, the presence of different varieties of mineralization sporadically developed within the faults provides the opportunity to study fluid regimes associated with the structures.

The source, temperature, and magnitude of fluid reservoirs participating in fault zones has been examined by measuring oxygen isotope compositions, redox state, and fluid inclusion

filling temperatures. Oxygen isotope exchange between rocks and an aqueous reservoir is reflected in shifts in the $\delta^{18}O$ rock or mineral from the initial value by a magnitude that depends on the $\delta^{18}O$ fluid, the ambient temperature, and the fluid to rock ratio [see *Taylor*, 1974]. Five distinct terrestrial fluid reservoirs have been identified on the basis of their oxygen and hydrogen isotope compositional ranges. These are marine, meteoric, magmatic, and metamorphic fluids, plus formation brines which generally contain a large component of meteoric recharge water [see *Taylor*, 1974]. Oxygen isotope studies also permit estimates of ambient mineral and fluid temperatures at equilibrium: this is because the fractionation of oxygen isotopes between coexisting silicates or metal oxides is a function of temperature alone [*Javoy*, 1977; *Friedman and O'Neil*, 1977].

In general, ferrous-ferric partitioning undergoes little change during regional metamorphism or deformation where rock is greater than water. However, in environments where high water/rock ratios are involved, departures of Fe^{2+}/Fe^{3+} from the initial state may be significant and are a function of the total mass of oxidant or reductant introduced by fluids [see *Eugster and Skippen*, 1967; *Beach and Fyfe*, 1972; *Spooner et al.*, 1977; *Kerrich et al.*, 1977].

ANALYTICAL METHODS

Mineral separates for oxygen isotope analysis were prepared by standard magnetic and heavy liquid techniques. Oxygen was extracted from minerals and powders using BrF_5 followed by quantitative conversion to CO_2 using the methods described by *Clayton and Mayeda* [1963]. Isotopic data are reported as $\delta^{18}O$ values in per mil relative to SMOW (standard mean ocean water). The overall reproducibility of $\delta^{18}O$ analyses is about 0.18‰ (two standard deviations).

Fractionations or differences in $\delta^{18}O$ among minerals are quoted as Δ values, defined as

$$\Delta_{A-B} = 1000 \ln \alpha_{A-B} \approx \delta_A - \delta_B$$

where α_{A-B} is the fractionation factor for the coexisting minerals A and B.

Double polished plates $\sim 50 \ \mu m$ in thickness were prepared for microscopy on fluid inclusions. Filling temperature measurements were made using a Reichart Kofler microscope heating stage [cf. *Roedder*, 1967]. Temperatures were recorded on a mercury thermometer calibrated against the fusion point of high-purity substances located adjacent to the sample. A heating rate of 2°C/minute was used throughout the homogenization studies. Replicate determinations on individual inclusions were generally reproducible to ±2°C. The overall accuracy of filling temperature data is about ±5°C.

THERMAL AND FLUID REGIMES DURING DEFORMATION AT THE GRENVILLE FRONT

Geological Setting

The Grenville front near Coniston, Ontario, is defined by the Grenville front boundary fault and associated mylonite zones, which form the demarkation between metasediments of the Southern Province and Grenville gneisses [*La Tour*, 1981a]. Motion on the fault zone is thought to be of Grenville age (~ 1 Ga), although the magnitude and direction of displacement are not precisely known. Bordering these structures immediately to the north is the Wanapitae fault, the displacement on which is of unknown magnitude and age (Figure 1).

Two discrete and distinct mylonite zones have been dis-

tinguished at the Grenville front near Coniston, Ontario (Figure 1); these zones have been designated as MZ I and MZ II and have been differentiated on the basis of several criteria by *La Tour* [1981a]. Specifically, metamorphic reactions associated with mylonitic deformation in MZ II are prograde in nature, whereas those in MZ I are exclusively retrograde. For instance, garnet-biotite and garnet-amphibole geothermometry indicates that mylonitic deformation in MZ II occurred at approximately 610°C, whereas mylonitic deformation in MZ I occurred at temperatures below 540°C [*La Tour*, 1981b] Protoliths of mylonitic rocks in MZ II include quartzofeldspathic gneisses with interlayered amphibolite, metadiabase unit Md 4, and minor pelitic rocks. Mylonite zone MZ I traverses similar lithologies as well as metasediments and granite (Figure 1; see also *La Tour* [1981a, b]).

Oxygen Isotope Relations

Estimates of the ambient temperature of deformation in MZ II from quartz-muscovite fractionations (2.6 to 2.8‰) are 580 to 640°C (Table 1), in close accord with the temperatures deduced from garnet-biotite and garnet-amphibole geothermometry. The $\delta^{18}O$ of quartz in schists is relatively uniform at 11.0 to 11.8‰: syntectonic quartz veins in these schists have oxygen isotope compositions within the same range as quartz in the host rock, signifying growth of the vein in isotopic equilibrium with the schist and probably from the same fluid. Fluids present during deformation and veining have a calculated $\delta^{18}O$ of 9.4 to 9.8‰ (Table 1).

In deformed granites of MZ I, quartz-albite fractionation (1.8 to 2.0‰) correspond to isotopic temperatures of 430 to 490°C, broadly commensurate with estimates from mineral chemical data. The $\delta^{18}O$ of quartz in the syntectonic veins is within the same range of 10.1 to 11.0‰ as quartz in the host granites and, based on the ambient temperatures, signifies growth in the presence of fluids estimated at +6.4 to +7.6‰ (Table 1). Hence the fluids present during deformation along MZ II and MZ I were probably of metamorphic origin, indigenous to the immediate host rocks.

The difference in estimated fluid isotopic composition for MZ I and MZ II is in accord with the higher temperature, and therefore smaller rock-water fractionations, in the latter: a corollary to this is that the fluid $\delta^{18}O$ was probably rock buffered, with minimal isotopic communication between MZ I and MZ II. Low fluid/rock ratios are also inferred to pertain during mylonitic deformation, based on the observation that the minerals occupying syntectonic veins generally correspond to those in the immediate host rock.

The albitite body located in the region of MZ I has a restricted range of $\delta^{18}O$ albite at 6.7 to 7.1‰ (Figure 1, sample location 5). A single quartz separate yields a quartz-albite fractionation of 1.8‰, corresponding to an isotopic temperature of 500°C and a calculated fluid isotopic composition of +5.5‰ (Table 1). The origin and precise timing of albitite formation relative to MZ I is not known, but given its low overall state of deformation, emplacement probably occurred during the late stages of translation on MZ I. Precipitation of albite may have resulted from focused streaming of high-Na fluids through the Grenville front zone.

Rock units transected by the Wanapitae fault have undergone extensive brittle fracturing, accompanied by local quartz-albite-carbonate veining (Figure 1, samples 14 and 15). Metadiabase unit Md 1 shows no shift in $\delta^{18}O$ whole rock or actinolite between massive precursor and fractured counterparts in the fault zone. Isotopic compositions of quartz and
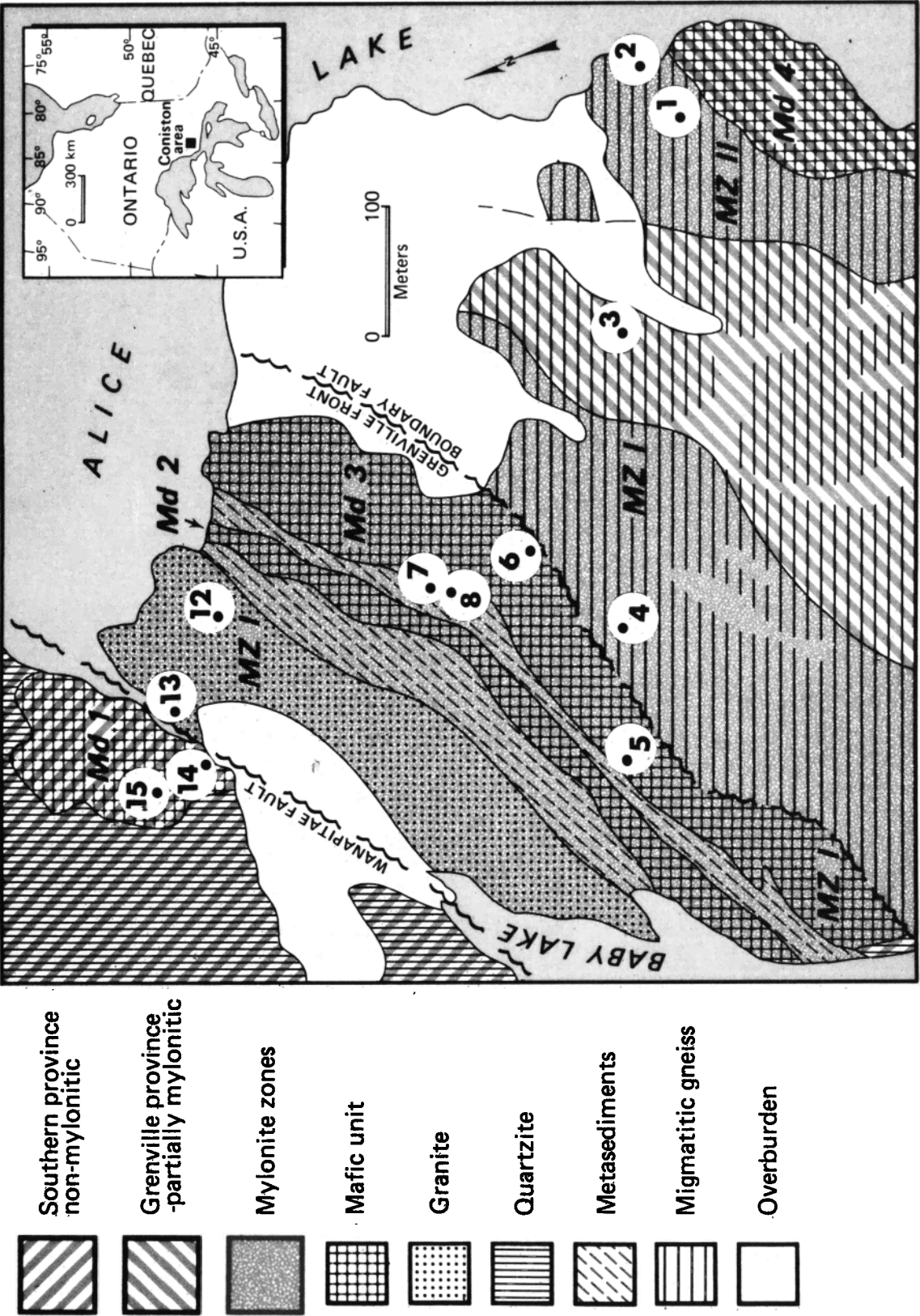
Fig. 1. Location map and simplified geology of the Grenville front, near Coniston, Ontario. Metadiabases prefixed "Md." Sampling sites for oxygen isotope analysis are indicated 1 through 15.

TABLE 1. Oxygen Isotope Compositions of Minerals and Whole Rocks From the Grenville Front Zone

Sample Description	Sample Number	$\delta^{18}O$ Quartz	$\delta^{18}O$ Albite	$\delta^{18}O$ Muscovite	$\delta^{18}O$ Actinolite	$\delta^{18}O$ Whole Rock	$\Delta^{18}O$ Mineral Pair	$T°C$	$\delta^{18}O$ Fluid
				MZ II					
Quartz-muscovite	GF 1	11.4		8.8			2.6 (q-m)	640	9.8
schists	GF 1a	11.7		9.0			2.7 (q-m)	610	9.8
	GF 1b	11.0		8.4			2.6 (q-m)	640	9.4
	GF 2	11.8		9.0			2.8 (q-m)	580	9.6
								[610]*	
Syntectonic	GF 1v	11.3							
quartz veins	GF 2v	11.1							
	GF 2av	11.8							
Posttectonic	GF 2cv	7.7							
quartz veins	GF dv	7.9							
	GF 2ev	7.6							
				MZ I					
Granite mylonite	GF 12a	10.9	8.9				2.0 (q-a)	430	6.4
	GF 12b	11.0	9.2				1.8 (q-a)	490	7.6
	GF 12c	10.8						[≤540]*	
Syntectonic	GF 12av	10.7							
quartz veins	GF 12bv	10.9							
	GF 12cv	10.1							
Albitite body	GF 5a	8.7	6.9				1.8 (q-a)	500	5.5
	GF 5b		7.1						
			6.7						
Post tectonic	GF 13av	−0.9						†200	−14
	GF 13bv	−1.0						to	to
	GF 13cv	−0.8						350	−8
	GF 13dv	−1.3							
				Wanapitae Fault					
Metadiabase	GF 14				5.3	5.9			
Fractured									
metadiabase	GF 15				5.5	6.2			
Quartz-albite-carbonate-									
epidote veins	GF 15v	12.1	9.3				2.8 (q-a)	280	3.2

*Temperature estimates from garnet-biotite and garnet-amphibole thermometry [La Tour, 1981b].

†Temperatures estimated from pressure corrected fluid inclusion homogenisation determinations. Isotopic temperatures calculated from the mineral-water fractionation equations reported by Clayton et al. [1972] (quartz-water), O'Neil and Taylor [1969] (muscovite-water), and O'Neil and Taylor [1967] (plagioclase-water). (q-m) = $\Delta^{18}O$ (quartz-muscovite), (q-a) = $\Delta^{18}O$ (quartz-albite).

albite from fracture fillings signifies temperatures of 280°C and a fluid $\delta^{18}O$ of +3.2‰. Hence fluid involvement in the Wanapitae fault was at relatively low temperatures, the fluid isotopic composition being consistent with a reservoir of formation brines [cf. Taylor, 1974].

Deformed granites of MZ I have a second population of quartz-hematite veins related to fractures postdating the mylonite fabric, and characterized by $\delta^{18}O$ quartz of −0.8 to −1.3‰. Primary fluid inclusions in quartz are two-phase aqueous inclusions which homogenize in the liquid phase, yielding filling temperatures of 200 to 250°C. The depth at which fracturing took place is not known; based on a range of 2 km to 10 km, assuming a hydrostatic pressure gradient and fluids with a salinity of 10 wt. percent NaCl equivalent, temperature corrections for pressure would amount to +20°C (18 MPa) to +60°C (110 MPa). Under a lithostatic pressure gradient the corrections would be +30°C to +180°C [cf. Lemmlein and Klevstov, 1961]. From the $\delta^{18}O$ quartz and temperature estimates, fluids implicated in these late veins would have an oxygen isotope composition of −8‰ (350°C) to −14‰ (200°C).

Continental meteoric water is the only terrestrial fluid reservoir with a negative $\delta^{18}O$. The fluid δ values calculated above are commensurate with those of contemporary precipitation on high mountain ranges at mid- to northerly latitudes, such as the Sierra Nevadas and Rocky Mountains [cf. Taylor,

1974]. The low-$\delta^{18}O$ quartz veins could represent fracture filling during circulation of meteoric water through the granite during its cooling and emplacement [cf. Taylor, 1974], but this is not considered likely in this instance, given that the $\delta^{18}O$ of quartz (10.8 to 11.0‰) and readily exchangeable feldspar (8.9 to 9.2‰) in the granite do not also reflect such an event (Table 1). This episode of fracturing accompanied by quartz precipitation is interpreted to have resulted from downward penetration of isotopically light meteoric water at relatively low temperatures and high redox into the Grenville front zone after ductile deformation on the higher temperature mylonite zones had terminated.

Given the extremely low fluid $\delta^{18}O$, the water may have been derived from precipitation on a mountain range induced by continental collision, with deformation accompanying crustal thickening accommodated at deep levels by translation on the high-temperature mylonite zone MZ II, followed by isostatic rebound on the low temperature zone MZ I and the associated Grenville front boundary fault [cf. La Tour, 1981a].

SHEAR ZONES IN A DEEP CRUSTAL ENVIRONMENT: LAGOA REAL, BRAZIL

Geological Setting

A series of large scale ductile shear zones transect Archaean basement gneiss at Lagoa Real, northeast of Caetité, State of

TABLE 2. The Oxygen Isotope Composition of Mineral Separates and Whole Rocks From Country Rock Gneiss and Uranium Mineralized Counterparts Within Shear Zones, Lagoa Real, Brazil

Sample Number	$\delta^{18}O$ Quartz	$\delta^{18}O$ Feldspar	$\delta^{18}O$ Pyroxene	$\delta^{18}O$ Magnetite	$\delta^{18}O$ Whole Rock	Δ (Quartz-Feldspar)	$\Delta^{18}O$ (Quartz-Pyroxene)	$\Delta^{18}O$ (Quartz-Magnetite)	Temperature, °C	$\delta^{18}O$ Fluid
					*Gneiss**					
0.1	12.15	9.85†		2.54	10.12	2.3		9.6		
0.2	10.87	8.50		1.03		2.4		9.8		
0.3	11.35	8.64		1.18	8.28	2.7		10.2		
0.4	8.76	6.99	6.04	0.40		1.8	2.7	8.4		+6
0.5	9.05	7.16			7.19	1.9				
0.6	9.86	7.67				2.2				
				Zone of U Mineralization‡						
M1	−1.20	−2.63§		−9.47		1.4		8.3	540	−4
M2	+0.53	−1.37		−8.44		1.9		9.0	500	−3
M3	−0.02	−1.75	−4.15	−8.70		1.7	4.1	8.7	520	−3

*Quartz, oligoclase, kspar, pyroxene, biotite ± magnetite.
†$\delta^{18}O$ of oligoclase in the gneiss.
‡Albite, pyroxene, quartz, magnetite.
§$\delta^{18}O$ of albite in uraniferous rocks.

Bahia, Brazil. The gneisses, dated at 2.6–3.0 Ga [*Inda and Barbosa*, 1978], are at amphibolite to granulite facies of metamorphism. The shear zones are 10 to 120 m in width, being exposed for distances in excess of 80 km. Deformed gneisses within the shear zones are characterized by an amphibolite facies mineral assemblage. Locally within the shear zones there are domains of hydrofracturing occupied by U mineralization, with attendant hydrothermal alteration [*Forman and Waring*, 1980; *Geisel et al.*, 1980].

Both to the east and west of Lagoa Real the Archean gneisses overlie Proterozoic metasediments and metavolcanics of the Espinhaco supergroup along a thrust surface. Other major thrusts are reported along the N-S trending Proterozoic belts [*Geisel et al.*, 1980; *Inda and Barbosa*, 1978]. Limited U-Pb dating of the uranium deposits indicates an age of mineralization at 820 Ma, placing the U-enrichment process in the Brazilian or Pan African period.

In proximity to domains of U-mineralization in shear zones, hydrothermal metasomatism has overprinted the regional amphibolite grade metamorphism of the gneisses on a massive scale and the tectonic fabric within the shear zones: this alteration occurs sporadically 40 km across and 100 km along strike. The principal mineral reactions accompanying alteration in shear zones are as follows: (1) potassium feldspar is hydrolyzed to albite, (2) hornblende is replaced by ferrohastingsite or removed entirely, (3) magnetite and aegirine-augite are formed, (4) quartz is dissolved, and (5) hematite, calcite, epidote-allanite, and garnet appear as late phases. Qualitatively, the alteration has involved enrichment of Na and Fe^{3+}, with a concomitant depletion of K and Si.

Removal of silica via solution of quartz requires aqueous fluids to be travelling up thermal gradient, and oxidation could then be in response to the mechanism discussed by *Beach and Fyfe* [1972]. Silica dissolution could also have occurred by CO_2 immiscibility from an aqueous hydrothermal fluid, since dissolved CO_2 acts to suppress the solubility of quartz. Alternatively, the quartz removal could have been induced by increase of fluid pH into the alkaline range. However, there is no mineralogical evidence that either of these two processes operated to an appreciable extent.

Constraints on pressure are sparse. Kyanite coexisting with muscovite in metarhyolites of the Espinhaco supergroup indi-

cate pressures of at least 3 kbar (0.3 GPa) at temperatures of <650°C [*Holdaway*, 1971; *Turner*, 1981]. Higher pressures are most unlikely in view of the low jadeite content of the pyroxenes, which are compositionally aegirine-acmite [cf. *Newton and Fyfe*, 1976].

Uranium ore bodies occupy hydrofractures transecting the ductile shear zone fabric, and are locally caught up in renewed movement along the structures. These relations are interpreted to signify that the shear zones acted as conduits of high permeability for periodic release, via hydrofracturing, of an overpressurized fluid reservoir under conditions where $P_{fluid} \geq \sigma_3 + T$ (where σ_3 is the minimum compressive stress and T is the tensile strength of gneiss) [cf. *Kerrich and Allison*, 1978].

Several elements are enriched along with U in the shear zones, including V (mean 590 ppm ± 360 1σ), Pb (460 ± 316 1σ), La (1530 ± 1140 1σ), and Ce (546 ± 230 aσ) [*Stein et al.*, 1980]. However, typical high-grade gneisses are specifically characterized by low abundances of these elements [*Heier*, 1973; *Atal et al.*, 1978], raising the question as to the cause of their anomalous enrichment in the Lagoa Real gneisses.

Oxygen Isotope Data

The oxygen isotope composition of country rock gneisses, their deformed equivalents, and U-mineralized counterparts within shear zones is reported in Table 2. In undeformed gneisses the $\delta^{18}O$ of quartz is relatively uniform at 11.3 to 12.1‰, as are the $\delta^{18}O$ of plagioclase (8.5 to 8.8‰) and magnetite (1.0 to 2.5‰). Quartz-oligoclase and quartz-magnetite fractionations are not concordant, indicating variable disturbance of the isotopic equilibrium attained at maximum metamorphic temperatures in the gneisses during subsequent cooling.

At the shear zone boundaries, coincident with the first development of a ductile tectonic fabric, the $\delta^{18}O$ of quartz diminishes to 8.8 to 9.9‰ (samples 04, 05, 06) representing a shift of about −2‰ relative to quartz in the gneissic precursors. In one sample (04) the quartz-oligoclase and quartz magnetite fractionations are triply concordant, signifying a deformation temperature of 500 to 540°C (Table 2). The calculated isotopic composition of water in equilibrium with quartz under the ambient conditions of ductile deformation is 5.8 to 6.4‰. These results are consistent with the range of $\delta^{18}O$
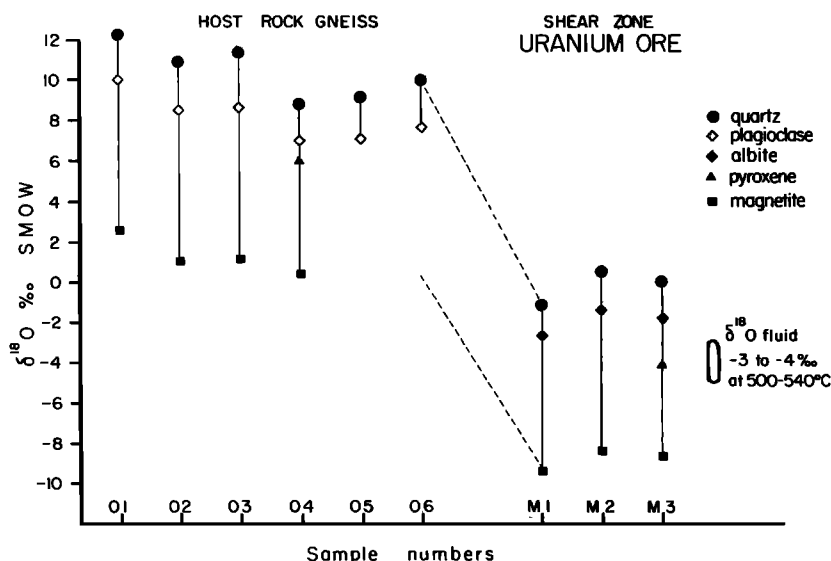
Fig. 2. Oxygen isotope composition of mineral separates from gneisses and their deformed and U-mineralized counterparts, Lagoa Real, Brazil.

values for metamorphic fluids reported by *Taylor* [1974], and an interpretation in terms of metamorphic fluids indigenous to the gneisses is consistent with the geology.

Within U-mineralized shear zones, quartz, feldspar, pyroxene, and magnetite all undergo a shift of about $-10\permil$ relative to their δ values in the parental gneiss (Figure 2). Quartz-magnetite fractionations are in a narrow interval of 8.3 to $9.0\permil$, corresponding to isotopic temperatures of $500-540°C$. In samples M1 and M2, coexisting quartz, albite, and magnetite are triply concordant, and in sample M3, quartz, albite, pyroxene, and magnetite are quadruply concordant, providing a rigorous criterion for retention of isotopic equilibrium in the mineralized rocks (Table 2).

Following *Bottinga and Javoy* [1973], we assign concordant isotopic temperatures to those coexisting minerals for which $T°C$ (quartz-albite), $T°C$ (quartz-pyroxene), and $T°C$ (quartz-magnetite), etc., agree to within an uncertainty in Δ(quartz-albite), Δ(quartz-pyroxene), etc., where the error in mineral pair fractionations $\varepsilon_A = \pm(\varepsilon_\delta{}^2 + \varepsilon_\delta{}^2)0.5 = \pm0.23\permil$, and the reproducibility of δ values, ε_δ, is $\pm0.16\permil$. Errors in temperatures calculated from quartz-magnetite fractionations amount to about $\pm10°C$. Isotopic temperatures have been calculated from mineral water fractionations given in the following sources [*O'Neil and Taylor*, 1967; *Bottinga and Javoy*, 1973; *Clayton et al.*, 1972; *Friedman and O'Neil*, 1977; *Javoy*, 1977]. Based on the temperature estimates and mineral δ values, fluids implicated in the uranium mineralization have a $\delta^{18}O$ of -3 to $-4\permil$, an isotopic signature consistent only with meteoric water.

Based on the isotopic evidence, fluids passing through the shear zone are interpreted to have been derived by dewatering of sedimentary formation brines that originated from earlier meteoric water recharge of the Proterozoic sedimentary basin aquifers. Modern formation brines such as in the Illinois and Alberta sedimentary basins typically possess a low $\delta^{18}O$ that essentially reflects the isotopic composition of ambient continental meteoric water recharge [*Taylor*, 1974].

Alternatively, the low ^{18}O fluids could have been derived from downwards circulation of surface meteoric water (hydrostatic conditions), but this is not consistent with the inferred

high fluid pressures accompanying mineralization in shear zones or the thrust regime. It is possible that a low ^{18}O fluid reservoir could have evolved from an isotopically heavier precursor either via exchange reactions with isotopically light rocks at high temperatures and low water/rock or by CO_2 immiscibility, which acts to drive the residual aqueous fluids to lower δ values [see *Higgins and Kerrich*, 1982]. The former is not considered likely in view of the presence of a sedimentary sequence and relatively high ^{18}O gneisses (Table 2). The latter possibility cannot be evaluated in the absence of fluid inclusion studies which could potentially reveal the presence or absence of CO_2 separation.

Discussion

Any viable hypothesis for the origin of the shear zones and uranium deposits in high-grade gneisses must account for the following geological, geochemical, and hydrological constraints: (1) fluids capable of carrying significant U, V, Pb, and REE (rare earth elements), (2) a fluid reservoir characterised by high Na/K and low $\delta^{18}O$, (3) discharge of the fluid reservoir under conditions of $P_f \geq \sigma_3 + T$, with fluids likely moving up thermal gradient to $\sim540°C$ and inducing oxidation, (4) mineralization and hydrofracturing proceeding at depths of ~15 km, which requires $P_f \geq \sigma_3 + T$, and (5) mineralization located within ductile shear zones, bounded by faults, and lying above rocks of lower metamorphic grade.

A possible model for the shear zones and uranium mineralization arises from consideration of the regional structure of the area. The Archaean basement is either faulted up or thrust over the Proterozoic sediments 30 km to the west, and this general structure appears to be common for something like 600 km along a N-S trend in Bahia and south into the state of Minas Gerais. These observations, along with the general problem of bringing basement to the surface, are all suggestive of a thrust mechanism of the type described by *Cook et al.* [1979, 1980]. Their thin-skinned thrust model, developed in the Southern Appalachians, involves the thrusting of sheets (10-20 km thick) of older crystalline basement over younger sedimentary formations for distances of over 260 km [see also *Meissner et al.*, 1981]. An alternative tectonic regime that
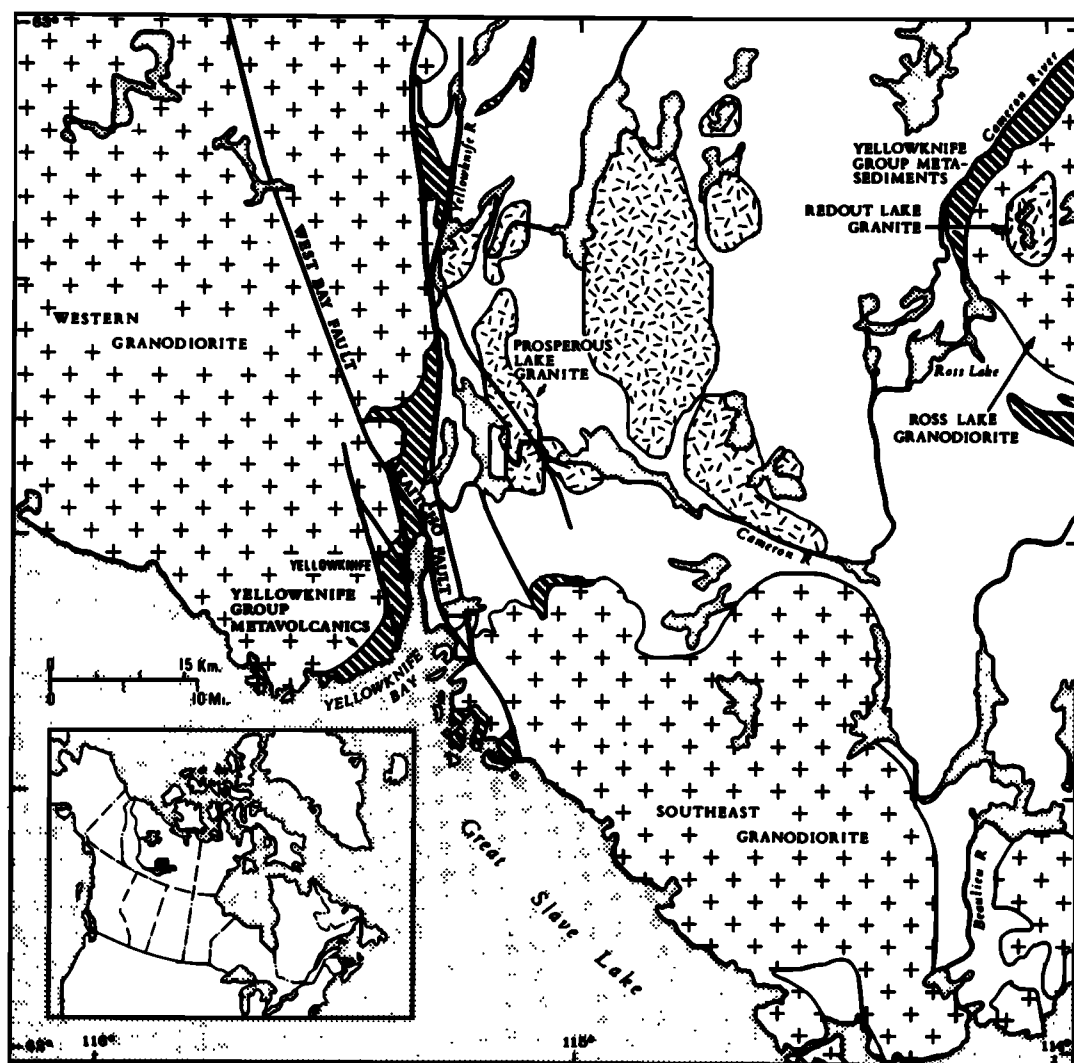
Fig. 3. Generalized geological map of the Yellowknife area.

would also be consistent with the constraints outlined above involves underthrusting of continental crust beneath continental basement, followed by isostatic rebound, as discussed for the Southern Alps of New Zealand [*Allis*, 1981].

Downward penetration of oxidized, saline, and U-bearing meteoric water to temperatures and depths of 500°C, 15 km could account for the observed silica removal coupled with oxidation. However, in extant geothermal fields and in continental ore deposits which involve thermally driven circulation of meteoric water, fluid temperatures and depths in excess of 400°C, 10 km have not been recorded [*Taylor*, 1974]. In a hydrological regime involving circulation of surface reservoirs the maximum fluid temperature is probably constrained by the ability of rocks to maintain fracture permeability at elevated temperatures and under hydrostatic fluid pressures. However, if fluid pressure is greater than or equal to lithostatic pressure, as discussed above for fluids generated at depth, there is no limit on crustal depth for fluid motion through hydrofractures.

Based on an upper limit of 100 ppb for U solute concentrations in deep groundwaters [*Langmuir*, 1978], in excess of 1000 km³ of hydrothermal solutions would be required to account for the 100 ktonne of U present in the shear zones. This estimated fluid volume is in the same range as the

volume of fluid generated by dewatering the 10-km-thick Proterozoic sequence over an area occupied by the thrust sheets, given a decrease in porosity of about 1%.

FLUID REGIMES IN FAULT AND SHEAR ZONES, YELLOWKNIFE

Geological Setting

The Yellowknife greenstone belt, juxtaposed against the Western granodiorite (an intrusion of batholithic proportions) is notable for a series of large-scale brittle-ductile shear zones locally containing gold lodes, displaced by a NNW-trending array of mineralized major faults (Figures 3 and 4). This section addresses questions of fluid flow and chemical transport within these large-scale structures, which traverse both the mafic volcanic sequence and margins of the batholith.

The largest shear zones, the Con, Campbell, and Giant, are exposed for distances up to 70 km, are 10-150 m wide, and extend to depths in excess of 1-2 km (Figures 3 and 4). These shear zones transect metabasic volcanic rocks and interflow metasediments at angles of 20°-70° in the vertical plane. The geometrical relations of minor structures indicate that the shear zones conform to the simple shear model of *Ramsay and*

LEGEND

ARCHEAN

DIVISION A

META-BASALT AND META-ANDESITE

DACITE, AGGLOMERATE AND TUFF

CHERTY TUFFS

DIVISION B

GREYWACKE, TUFF, AND CONGLOMERATE

PORPHYRITIC META-DACITE AND META-RHYOLITE

META-GABBRO SILLS

META-GABBRO AND META-DIORITE

CONGLOMERATE

GRANODIORITE

DIABASE, GABBRO

SHEAR ZONES

FAULTS

/80 BEDDING

▲ MINE

0 _____ 1500 METRES
0 _____ 5000 FEET

GIANT

GIANT SHEAR ZONE

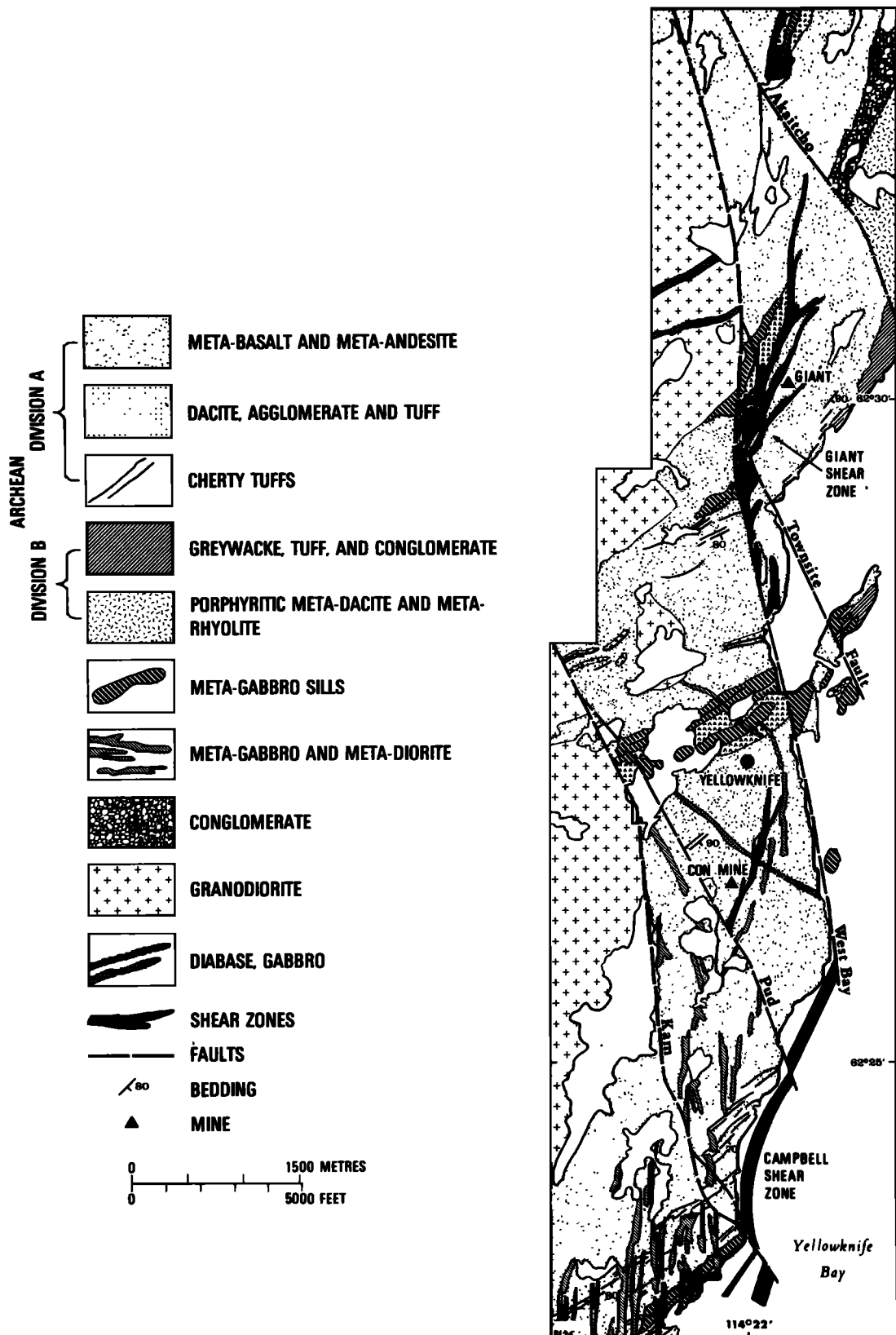YELLOWKNIFE

CON MINE

CAMPBELL SHEAR ZONE

Yellowknife Bay

Fig. 4. Simplified geological map of the Yellowknife greenstone belt, illustrating the disposition of major shear zones and faults (modified after *Henderson and Brown* [1966]).

TABLE 3. Oxygen Isotope Compositions of Metabasalts and Granodiorites Along With Their Deformed and Mineralized Counterparts, Yellowknife Greenstone Belt

Sample Description	Sample Number	$\delta^{18}O$ Quartz	$\delta^{18}O$ Chlorite	$\delta^{18}O$ Muscovite	$\delta^{18}O$ Whole Rock	$\Delta^{18}O$ Mineral Pair	$T\,°C$	$\delta^{18}O$ Fluid
Metabasalts: Unmineralized Shear Zone								
Epidote-amphibolite	J 6				7.2			
metabasalts	J 3				7.6			
Chlorite schists	J 11.5	13.9	6.7		7.1	7.2 (q-c)	340	7.2
	J 11				7.0			
	J 12.5	14.2	6.6		7.3	7.6 (q-c)	320	6.8
Stage 1 quartz-albite vein	J 12.5 v	14.0	6.7			7.3 (q-c)	340	7.2
Metabasalts: Mineralized Shear Zones								
Au-quartz veins in quartz-	229	11.9	6.1	8.6	8.9	5.8 (q-c)	450	7.8
carbonate-chlorite-muscovite	188	11.9		8.6		3.3 (q-m)	460	7.8
schist	177	13.0	6.9		9.1	6.1 (q-c)	420	8.3
	185	11.4		7.7		3.7 (q-m)	410	6.5
Campbell shear	234	12.7	7.0		9.5	5.7 (q-c)	460	8.8
	118	12.5		8.7		3.8 (q-m)	400	7.5
	4983	12.0	6.1	8.5	10.0	5.8 (q-c)	450	7.9
						3.5 (q-m)	430	7.5
	250	10.4						
	252	10.8						
	255	10.8						
	261 G	10.9						
	120	11.6	5.5			6.1 (q-c)	430	7.1
Con shear	143	13.2	6.4			6.8 (q-c)	370	7.3
Granodiorite								
Interior of batholith	C109	9.7			7.8			
traversing to	C110	10.9			8.4			
contact with greenstones	C113	11.4						
	C114	13.0			11.0			
	C115	13.7						
Quartz-orthoclase pegmatites	J16-CZ	10.7						
in granodiorite	WO	9.9						
Granodiorite: Unmineralized Shear Zone								
	C178 U	10.4						
	C178 D	10.1						
Granodiorite: Unmineralized Shear Zone								
Granodiorite, undeformed	7526 U	12.2	6.1			6.1 (q-c)	420	7.5
Granodiorite, deformed	7526 D	12.4						
Quartz vein in shear zone	7526 v	12.1						
West Bay Fault Zone								
	J16 C	12.2					250–350*	2–6
	J16 C1	11.4						
	J16 E1	11.6						

*Temperatures estimated from pressure-corrected fluid inclusion homogenization determinations. Isotopic temperatures calculated from the mineral-water fractionation equations reported by *Clayton et al.* [1972] (quartz-water), *O'Neil and Taylor* [1969] (muscovite-water), and *Wenner and Taylor* [1971] (chlorite-water). (q-m) = $\Delta^{18}O$ (quartz-muscovite); (q-c) = $\Delta^{18}O$ (quartz-chlorite).

Graham [1970]. Utilizing this model and assuming an average width for the Campbell zone of 80 m, the maximum computed displacement is about 300 m, west side up [cf. *Kerrich and Allison*, 1978].

After initial development of the shear zones, these structures acted as permeable conduits for episodic discharge of hydrothermal fluids which precipitated Au-bearing quartz-carbonate veins. The structures and ore bodies contained in them have been described by *Boyle* [1961], *Henderson and Brown* [1966], *Breakey* [1975], and *Kerrich and Allison* [1978]. Shear zones in the granodiorite have a geometry and attitude which are essentially congruent with that of shear zones in the metabasalts and for which a similar sense of displacement and timing is indicated.

Oxygen Isotope Data:
Unmineralized Structures

In traverses across unmineralized sections of the Campbell and Con shear zones, from isotropic epidote-amphibolite facies metabasalts through to deformed chlorite schists, the whole-rock $\delta^{18}O$ values remain approximately constant at +7 to +7.5‰, and $Fe^{2+}/\Sigma Fe$ is uniform at ~0.7 (Tables 3 and 4). These data are interpreted to indicate that deformation took place under conditions of low water/rock ratio [see also *Kerrich and Fyfe*, 1981]. The overall hydration accompanying initial development of shear zones is here interpreted to result from volume dilation during an early phase of brittle fracturing preceding development of a ductile fabric [cf. *Kerrich*

TABLE 4. Summary of $Fe^{2+}/\Sigma Fe$ Determinations Across Shear Zones in the Metabasalts and Granodiorite, Yellowknife Greenstone Belt

Group	Lithology	Number of Determinations	Arithmetical Mean	Standard Deviation
I	wall rock 1: epidote-amphibolite metabasalt	11	0.76	0.028
II	shear zone 1: chlorite-sericite schist	15	0.78	0.056
III	wall rock 2: epidote-amphibolite metabasalt	9	0.75	0.076
IV	shear zone 2: chlorite-sericite schist	10	0.79	0.058
V	mineralized shear zone: chlorite-sericite-pyrite (quartz-Au) schist	24	0.95	0.041
VI	wall rock adjacent to mineralization: epidote-amphibolite metabasalt	9	0.81	0.061
VII	granodiorite: plagioclase-quartz-microcline-biotite	12	0.72	0.060
VIII	deformed granodiorite in shear zones: albite-quartz-chlorite-epidote-muscovite	8	0.68	0.060
IX	mineralized shear zones in granodiorite: albite-quartz-chlorite-epidote-muscovite	8	0.87	0.072

and Allison, 1978], with migration of pore fluids indigenous to the surrounding rocks into the structures.

Taylor [1968] has shown that fresh, unaltered submarine basalts have whole-rock $\delta^{18}O$ values of $+5.0$ to $+6.0$‰. The slight enrichment in ^{18}O recorded for these volcanic rocks relative to primary submarine basalts is attributed to oxygen isotope exchange with seawater at low temperatures during thermally driven convective cooling of the ocean crust prior to development of the shear zones.

Mineralized Shear Zones

Along discrete conduits within the shear zones, chlorite schists have transformed to chlorite-carbonate-muscovite-quartz-pyrite schists in envelopes around massive gold-bearing quartz veins. In proximity to veins the whole-rock $\delta^{18}O$ and $Fe^{2+}/\Sigma Fe$ ratios exhibit a major perturbation from background values of $+7$‰ and 0.7, respectively, (Au < 4 ppb) to $+9$ to $+10$‰ and an average $Fe^{2+}/\Sigma Fe$ of 0.92 in carbonate-sericite schists enveloping Au-bearing quartz veins (Au 1–40 ppm). The $\delta^{18}O$ of vein quartz is equal to that of quartz extracted from the carbonated mafic schists to within the limits of precision (Table 3), implying oxygen isotope equilibrium between hydrothermal fluids and altered wall rocks under conditions of high water/rock.

In general, studies of metamorphic processes indicate that ferrous/ferric ratios undergo little change during regional metamorphism, but this is not the case in environments where high water/rock ratios were involved. Oxidation-reduction reactions under high-temperature metamorphic conditions occur dominantly in response to the dissociation of water, according to the reaction $2H_2O \rightarrow 2H_2 + O_2$. Note that for an aqueous fluid at high temperatures, the equilibrium $P_{H_2} \gg 2P_{O_2}$ on account of reactions of the type $(Fe^{2+}$-silicate$)$ $+ H_2O \rightarrow (Fe_3O_4$-silicate$) + H_2$. The predominance of Fe^{2+} in mineralized wall rocks is thus believed to result from the reduction of Fe^{3+} in silicates and metal oxides by hydrogen during interaction with a large flux of H_2-bearing aqueous solutions ascending along a temperature-pressure gradient through conduits within the shear zones. Hydrogen removal corresponds to the difference in equilibrium hydrogen concentration at two different temperature and pressure states (represented by the metamorphic source region and shear

zones, respectively [see Eugster and Skippen, 1967; Kerrich et al., 1977]). The arguments involved are discussed by Kerrich et al. [1977].

Minimum integrated water/rock ratios of greater than $3:1$ are indicated by the shift in redox state (Table 4), and values in excess of $30:1$ may be deduced from the abundance of quartz and quartz solubility relations. Considerations of gold concentration from background into the veins reveals that a minimum of 9 km^3 of hydrothermal solutions discharged through the structures.

Ambient thermal conditions during mineralization may be estimated from two independent lines of evidence. The coexistence of abundant chlorite with almost pure albite (Ab_{92}) in wall rocks adjacent to veins provides an upper limit of $\sim 470°C$ for the ambient temperature [Liou et al., 1974]. Temperatures of fluid transport through the shear zones have also been calculated from the oxygen isotope data. Quartz-muscovite fractionations in precious metal–bearing quartz veins exhibit a narrow range from $+3.8$ to $+3.4$‰ (Table 3) corresponding to calculated isotopic temperatures of 360–460°C. Quartz-chlorite fractionations are $+5.7$ to $+6.8$‰, representing isotopic temperatures of 370–450°C. Errors in the isotopic temperatures, arising from the analytical uncertainty of ± 0.18‰, amount to about $\pm 30°C$. These calculated temperatures must be treated with caution because concordant isotopic equilibrium among coexisting mineral triplets is a necessary condition for obtaining reliable temperature estimates. Quartz, muscovite, and chlorite separated from one sample (4983) are triply concordant. However, the isotopic results are consistent with independent estimates of temperature discussed above.

If the ambient mineralization temperatures of 400–450°C derived above are broadly correct, then the calculated $\delta^{18}O$ of fluids from which the veins were precipitated is $+7$ to $+9$‰. These values are consistent with the range in $\delta^{18}O$ of fluids implicated in metamorphism and overlap the primary magmatic range $(+6$ to $+8$‰) [Taylor, 1974]. A magmatic origin of the fluids is considered unlikely, inasmuch as quartz veins in the nearby Western Granodiorite batholith are characterized by $\delta^{18}O$ quartz of 9–11‰. Quartz δ values are in general ~ 1.5‰ lighter and temperatures $\sim 100°C$ hotter in the high water/rock mineralized domains compared to barren low wa-

ter/rock counterparts, commensurate with transport of high temperature fluids from depth up the structures.

Western Granodiorite

In the internal regions of the Western granodiorite batholith, rocks are generally characterized by a primary igneous mineralogy. The $\delta^{18}O$ whole rock values of 7.8 to 8.4‰ and $\delta^{18}O$ quartz 9.7 to 10.8‰ places these rocks (C109–C113, Table 3) within the class of isotopically "normal" granitic rocks of Taylor [1978], defined as 6‰–10‰. Pegmatites within the batholith have $\delta^{18}O$ quartz in compliance with that of quartz in the host rock, signifying common retention of a magmatic signature.

Traversing towards the batholith periphery, near the greenstone contact, the $\delta^{18}O$ whole rock and quartz progressively increase to 11‰ and 13.7‰ respectively (C113–C115, Table 3). The enrichment in ^{18}O correlates with appearance of a retrograde greenschist facies mineralogy together with higher precious metal contents (Au 8 to 120 ppb). Where shear zones formed in the interior regions of the batholith, no discernable shift in isotopic composition accompanying deformation was detected. However, toward the pluton boundary, undeformed granodiorite, its deformed product in shear zones, together with quartz veins emplaced in these structures all record $\delta^{18}O$ quartz of 12.1 to 13.7‰. These elevated values relative to the batholith interior, are close to that of second-generation vein quartz in shear zones transecting the neighboring metabasalts. The results are collectively interpreted in terms of a single-fluid reservoir with highly focused discharge through shear zone conduits within the metabasalts but lower fluxes and less focused streaming through the adjacent batholith margin.

West Bay Fault

The West Bay Fault is a major NNW trending structure of unknown magnitude or direction of displacement (for a discussion see Boyle [1961]). Locally the fault zone contains white "bull" quartz-hematite veins along with their brecciated products. Quartz δ values of the veins range from 11.4 to 12.6‰. It is difficult to establish the temperature of quartz precipitation in the fault. Preliminary evidence from primary fluid inclusions in quartz reveal two phase aqueous inclusions which homogenise in the liquid phase at temperatures of 220 to 260°C. The depth of faulting is not known, but based on a range of 2 to 10 km and assuming a hydrostatic pressure gradient, temperature corrections for pressure would amount to +20°C (18 MPa) to +60°C (110 MPa) [Lemmlein and Klevstov, 1961]. From the $\delta^{18}O$ quartz and temperature estimates, fluids streaming through the fault zone and precipitating quartz would have an oxygen isotope composition of +2 to +6‰. The isotopic results are compatible with flow of several possible reservoirs, including formation brines or mixtures of low temperature metamorphic fluids, formation brines, and surface waters. The presence of hematite signifies relatively high redox conditions, implying that a component of oxidizing surface waters was involved.

Conclusions

In summary, early transformation of epidote-amphibolite metabasalts to chlorite schists within shear zones is considered to have taken place at ~340°C under conditions of low water/rock, with the chemical system essentially closed except for minor addition of volatiles. Volatiles were probably indigenous to the immediately surrounding metabasalts, migrating into shear zones during volume dilation accompanying early

brittle fracturing. After hydration the shear zones changed to a ductile deformation mode dominated by pressure solution.

Subsequently, the shear zones acted as high-permeability conduits for discharge of greater than 9 km^3 of reducing fluids at 360 to 460°C and of probable metamorphic origin. Fluid discharge occurred in repeated pulses controlled either by successive build up and decay of fluid pressure in a reservoir at depth or, alternatively, by episodic relaxation of the horizontal stress. Structural, thermal, and fluid relations in the adjacent granodiorite were generally similar to those in the metabasalts, except that the fluid flux was lower and less focused, such that large volumes of granodiorite underwent isotopic and chemical exchange with fluids, in addition to the shear zone conduits.

The West Bay fault acted as a site for venting of substantial fluid volumes at temperatures of 200 to 300°C, the reservoir possibly being surface water derived formation brines.

DISCUSSION AND CONCLUSIONS

In general, fluid regimes in the fault and shear zones studied follow a sequence from conditions of high temperature and pressure with locally derived fluids at low water/rock ratios during initiation of the structures to high fluid fluxes along discrete conduits as the structures propagate, develop large-scale permeability, and penetrate hydrothermal reservoirs. At this second stage, discharging fluids were at high temperature and pressure and generally reduced due to their transport upwards from deeper crustal levels: such fluids were likely of metamorphic origin, their characteristic high $\delta^{18}O$ reflecting equilibrium of the reservoir with crustal rocks at elevated temperatures in the source region.

Later in the tectonic evolution, as deeper crustal levels were exposed by erosion such that conditions of lower ambient temperature and pressure pertained, incursion of near-surface waters into faults occurred: these fluids were either formation brines in aquifers intersected by the fault or from direct downwards penetration of surface fluids into the structures. Such fluids were oxidized and of lower $\delta^{18}O$, the former reflecting near-surface environments and the latter commensurate with a component of low-^{18}O meteoric continental surface water and/or shallower level metamorphic fluids where the $\delta^{18}O$ is buffered to smaller values by larger rock-water fractionations at the lower ambient temperatures. The high-temperature, high-^{18}O and low-temperature, low-^{18}O fluid regimes of faults may proceed coevally, the former at deep and the latter at shallow crustal levels, with juxtaposition of the two taking place by subsequent vertical displacements on the structure. Flow of both deep (metamorphic) and shallow level (meteoric) fluids has been implicated in contemporary seismic activity on the San Andreas fault by Irwin and Barnes [1975] and Sibson [1982], respectively.

At Lagoa Real, discharge of the low-^{18}O fluids of surface water origin probably took place at ~15 km depth in the crust and relatively high temperatures of 550°C. This situation appears to have taken place during overriding of a sedimentary sequence with meteoric water–recharged aquifers by high-grade gneisses during overthrusting.

The role of metamorphic fluids in shear zones at deep crustal levels has been emphasized by Beach and Fyfe [1972], Beach [1976], and Kerrich and Fyfe [1981]. Etheridge et al. [this issue] present evidence for high pore fluid pressures during prograde regional metamorphism, with resultant low effective confining stress leading to enhanced permeabilities and, in turn, to the local dominance of advective flow. All of

these studies, including the present one, emphasize that large-scale fluid flow may occur at deep crustal levels under appropriate conditions. On the other hand, some shear zones appear to have been essentially chemically and hydrologically closed systems. This is the case for the Mieville shear zone transecting Hercynian granodiorite of the Aiguilles Ranges, Switzerland, where no significant excursions of $\delta^{18}O$ rock, oxidation state, or rock composition occur from the protolith across the structure [Kerrich et al., 1980].

In addition to the mechanical effects of transient high fluid pressures in faults [Sibson, 1977, 1981; Kerrich and Allison, 1978], modification of ambient chemical conditions by fluids streaming through faults may play a role in silicate deformation. As discussed above, the oxidation state of iron in some structures indicates significant reduction by fluids, probably due to elevated partial pressures of hydrogen, and it is possible that such high P_{H_2} may influence the cracking behavior of minerals in the fault zone. For instance, hydrogen embrittlement is a well-known phenomenon in metals and ceramics and is thought to occur by chemisorption of hydrogen at fracture surfaces which promotes nucleation of dislocations at crack tips [Lynch, 1979; Michalske and Freiman, 1982]. Hobbs [this issue] notes that at an intracrystalline level, incorporation of a hydrogen defect that is capable of acting as an acceptor into silicates leads to a strong dependence of point defect chemistry on the fugacities of both water and oxygen.

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