Fluid dynamics and fluid-structural relationships in the Red Lake mine trend, Red Lake greenstone belt, Ontario, Canada

Y. LIU¹, G. CHI¹, K. M. BETHUNE¹ AND B. DUBÉ²

¹Department of Geology, University of Regina, Regina, SK, Canada; ²Geological Survey of Canada, Quebec, QC, Canada

ABSTRACT

The Red Lake mine trend, a deformation zone in the Archean Red Lake greenstone belt that hosts the world-class Campbell-Red Lake gold deposit, is characterized by abundant foliation-parallel iron-carbonate ± quartz veins with banded colloform-crustiform structures and cockade breccias overprinted by silicification and gold mineralization. There is an apparent incompatibility between the cavity-fill structures of the veins and breccias (typically developed at shallow crustal depths) and the upper greenschist to lower amphibole facies metamorphic conditions recorded in the host rocks (indicating relatively deep environments). This, together with the development of veins along the foliation plane, represents an enigmatic problem that may be related to the interplay between fluid dynamics and stress field. We approach this problem through systematic study of fluid inclusion planes (FIPs) in the vein minerals, including the orientations of the FIPs and the pressure–temperature conditions inferred from fluid inclusion microthermometry. We find that fluid inclusions in the main stage vein minerals (pregold mineralization ankerite and quartz and syn-ore quartz) are predominantly carbonic without a visible aqueous phase, whereas many inclusions in the postore stage contain an aqueous phase. Most FIPs are subvertical, and many are subparallel to the foliation. High fluid pressure coupled with the high wetting angles of the water-poor, carbonic fluids may have been responsible for the abundance of brittle deformation features. The development of subvertical FIPs is interpreted to indicate episodic switching of the maximum principal compressive stress (σ1) from subhorizontal (perpendicular to the foliation) to subvertical (parallel to the foliation) orientation. The subvertical σ1 is favorable for the formation of foliation-parallel veins, as fractures are preferentially opened along the foliation in such a stress regime, the origin of which may be linked to the fluid source.

Key words: carbonate-quartz veins, CO₂ fluid, fluid inclusion planes, fluid inclusions, foliation-parallel, gold mineralization, Red Lake

INTRODUCTION

The Archean Red Lake greenstone belt in northwestern Ontario, Canada hosts a number of gold deposits mainly along deformation zones (Andrews et al. 1986; Dubé et al. 2004; Sanborn-Barrie et al. 2004). One of the deformation zones is the SE-trending Red Lake mine trend in the eastern part of the greenstone belt, which hosts the world-class Campbell-Red Lake gold deposit (total production plus remaining reserve = 840 t of gold) and the smaller Cochenour-Willans deposit (total production = 39 t of gold), as well as a number of gold occurrences (Dubé et al. 2004; Sanborn-Barrie et al. 2004; Chi et al. 2006, 2009; Lichtblau & Storey 2007). The gold deposits in the Red Lake mine trend consist of numerous barren to low-grade banded colloform-crustiform iron-carbonate ± quartz veins and cockade breccias overprinted by silicification and sulfidation accompanying gold mineralization (MacGeehan & Hodgson 1982; Mathieson & Hodgson 1984; Penczak & Mason 1997, 1999; Tarnocai 2000; Dubé et al. 2001, 2002, 2003, 2004). The cavity-filling structures in the veins and breccias are in
striking contrast with the development of penetrative foliation and the upper greenschist-lower amphibole facies metamorphic conditions recorded in the host rocks (Thompson 2003; Sanborn-Barrie et al. 2004). Furthermore, most of the carbonate ± quartz veins are parallel to the foliation (Zhang et al. 1997; Dubé et al. 2001, 2002), which is at odds with open-space filling style textures in the veins, as first noticed by MacGeenan & Hodgson (1982). This apparent contradiction between the vein-filling styles and structural environments has led to different genetic models for gold mineralization. Penczak & Mason (1997, 1999) and Damer (1997) proposed that the Campbell-Red Lake deposit is an epithermal deposit that was later deformed and reworked during peak regional metamorphism and deformation, whereas other studies suggest that gold mineralization is broadly contemporaneous with peak regional metamorphism and protracted deformation (Mathieson & Hodgson 1984; Andrews et al. 1986; Zhang et al. 1997; Parker 2000; Tarnocai 2000; Dubé et al. 2001, 2002, 2003, 2004; Chi et al. 2003, 2006, 2009). The latter scenario is comparable to that invoked for orogenic type gold deposits (Groves et al. 1998), although the unusual vein characteristics and abundance of carbonate veining make the Red Lake examples atypical (Dubé & Gosselin 2007).

Two lines of evidence support the second model described previously. While most carbonate ± quartz veins with colloform-crustiform structures and cockade breccias appear to predate gold mineralization, some carbonate breccia veins contain gold-mineralized rock fragments (Dubé et al. 2002), suggesting that the formation of at least some of these veins overlaps in time with the development of ductile structures (foliation) during the main phase of deformation, rather than during an early deformation event as would be expected for the epithermal model (Dubé et al. 2002). Fluid inclusion studies (Chi et al. 2002, 2003, 2006, 2009) yielded similar isochores for preore ankerite and quartz and syn-ore quartz, which indicate maximum fluid pressures in the range of 2.7–3.6 kbar for a temperature range of 350–550°C, further supporting that both carbonate ± quartz veins and overprinting silicification and gold mineralization took place in fairly deep environments, likely during the main phase of deformation.

It remains poorly understood why colloform-crustiform structures and cockade breccias, typically found in shallow crustal levels, are so abundant in the Red Lake mine trend if the veining events took place at great depths, and why most of the veins are parallel to the foliation. The purpose of this study is to tackle these problems through a systematic study of microfractures containing fluid inclusions, known as fluid inclusion planes (FIPs), in vein minerals, which can provide information on both the stress field (through measurement of orientation of the FIPs) and fluid pressure (through study of the fluid inclusions in the FIPs) (Lespinasse & Pecher 1986; Boullier & Robert 1992; Lespinasse 1999). The study also aims to provide new insights into the fluid dynamics and fluid-structural relationships in shear zone-controlled hydrothermal systems in general.

**GEOLOGICAL SETTING, SAMPLE LOCATIONS, AND METHODS OF STUDY**

**The Red Lake greenstone belt**

The Red Lake greenstone belt is located in the Uchi Subprovince of the western Superior Province (Fig. 1). The greenstone belt is made up of mafic-ultramafic to felsic volcanic rocks with subordinate sedimentary rocks surrounded by granitoid batholiths. The supracrustal rocks have been divided into two packages: the 2.99–2.89 Ga Mesoarchean and the 2.75–2.73 Ga Neoarchean packages, which are separated by a regional unconformity (Fig. 1; Sanborn-Barrie et al. 2001, 2004).

Two main episodes of penetrative deformation have been established in the Red Lake greenstone belt (Sanborn-Barrie et al. 2000, 2001, 2004). D1 deformation, which took place between 2742 and 2733 Ma, is interpreted to have occurred in response to W-E shortening, as reflected by a pervasive N–S, NNW–S, and NE-trending foliation (Sanborn-Barrie et al. 2001, 2004). D2 deformation, bracketed between 2723 and 2712 Ma, is more intensely developed and preserved, and is represented by a pervasive E–W, NE–W, and SE-trending foliation and related F2 folds resulting from an overall N-S shortening (Sanborn-Barrie et al. 2001, 2004; Dubé et al. 2004). According to Percival et al. (2006), D2 deformation is an expression of the Uchian orogeny, which involved collision between the North Caribou terrane to the north of the Uchi Subprovince and the Winnipeg River Subprovince to the south. Two post-tectonic deformation events (D3 and D4) are locally recorded.

The supracrustal rocks, which record greenschist- to amphibolite-facies metamorphism, were intruded by syn-to postvolcanic granitoid stocks and batholiths, as well as a series of mafic, lamprophyric, and feldspar porphyry dikes. Major granitoid intrusions were emplaced in three stages: pre-tectonic (2.74–2.73 Ga), syn-tectonic (2.72–2.71 Ga) and late- to post-tectonic (2.71–2.70 Ga) (Parker 2000; Sanborn-Barrie et al. 2004). Generally, the metamorphic grade increases from greenschist facies in the center of the greenstone belt to amphibolite facies toward the margins, suggesting that metamorphism was driven by heat supplied by surrounding batholiths (Andrews et al. 1986; Damer 1997; Menard & Pattison 1998). However, the timing of metamorphism is still debatable. While some suggest that peak regional metamorphism coincided with D2 deformation and was overprinted by amphibolite-facies contact
metamorphism (Menard et al. 1999), others postulate that the amphibole facies metamorphism represents part of the regional metamorphism (Thompson 2003).

The rocks of the Red Lake greenstone belt were variably affected by different types of hydrothermal alteration, including carbonate, potassic, and aluminous alteration as well as silicification (Parker 2000 and reference therein). Among these, carbonatization is the most extensive and consists of two types: calcite alteration and ferroan-dolomite alteration, the latter being generally proximal to gold mineralization (Parker 2000).

The Red Lake mine trend

The Red Lake mine trend is a SE-trending protracted deformation-alteration-veining zone in the eastern part of the Red Lake greenstone belt (Fig. 1). The main structures were formed in D2 and consist of SE-trending high-strain zones, antiforms, synforms, and penetrative foliation (Fig. 1), plus some associated E-W- and NNW-trending deformation zones (Dubé et al. 2001, 2002 and references therein). The shortening fabric (S2 foliation) is heterogeneously developed, strikes SE (110–130°), and dips steeply SW (60–80°). The attitude of the foliation is fairly consistent across the deformation zones, regardless of the intensity of deformation (Zhang et al. 1997).

While the E-W- and NNW-trending subvertical deformation zones have been interpreted as oblique- to strike-slip shear zones developed in response to NE-SW subhorizontal shortening (Zhang et al. 1997; Dubé et al. 2001, 2002), the mechanical nature of the SE-trending, generally foliation-parallel high-strain zones is debatable, including simple shear (normal, reverse-sinistral, and reverse-dextral) and pure shear. Based on the interpretation that some foliation-oblique veins are Riedel shear fractures (R and R’), and the observation that folded and boudinaged veins are asymmetrical relative to the foliation, Andrews et al. (1986) interpreted the SE-trending deformation zones as shear zones with normal displacement (SW side down), which is supported by the sense of rotation of synkinematic, metamorphic minerals documented in oriented thin sections from the Red Lake mine by Hugon & Schwerdtner (1984). However, based on mapping and apparent displacements of rock units, as well as observations of local C-S-type fabrics (C parallel to the shear zone and S oblique to the shear zone), it has been shown that the SE-trending deformation zones in the Campbell-Red Lake deposit accommodated either reverse-sinistral movement (the Campbell fault and the Red Lake or New Mine fault) or reverse-dextral movement (the Dickenson fault) (MacGeehan & Hodgson 1982; Mathieson & Hodgson 1984; Rogers 1992; Penczak & Mason 1997; Dubé et al. 2001, 2002).
Fluid dynamics and fluid-structural relationships in Red Lake 263

In contrast, Zhang et al. (1997), in studying the deformation zones in the Campbell mine, noted that the fabrics within the SE-trending high-strain zones are generally symmetrical and lack shear-sense indicators and suggested that they resulted from pure shear deformation related to NE-SW subhorizontal shortening.

Regardless of the different kinematic interpretations of the deformation zones, the SE-trending subvertical foliation indicates a dominant NE-SW subhorizontal maximum principal stress ($\sigma_1$) during D2. However, the intermediate ($\sigma_2$) and minimum ($\sigma_3$) principal stresses may have switched their orientations one or more times. Many of the foliation-parallel carbonate ± quartz veins have been boudinaged, and there are two different orientations of the boudin axis: subhorizontal and subvertical. According to Zhang et al. (1997) and Dubé et al. (2002), the subhorizontal boudinage indicates that $\sigma_2$ was subhorizontal and $\sigma_3$ subvertical, and the subvertical boudinage indicates that $\sigma_2$ was subvertical and $\sigma_3$ subhorizontal. The E-W- and NNW-trending shear zones, which form a conjugate shear system approximately bisected by the SE-trending deformation zones, also indicate that $\sigma_2$ was subvertical (Zhang et al. 1997; Tarnocai 2000). Zhang et al. (1997) documented that subhorizontal boudins are overprinted by subvertical ones and suggested that the minimum principal stress ($\sigma_3$) switched from subvertical to subhorizontal during D2.

Numerous centimeter- to meter-wide, banded colloform-crustiform carbonate (mainly ankerite) ± quartz veins (Fig. 2) and cockade breccias, which are barren to low grade, occur in the Red Lake mine trend and elsewhere in the district. These veins occur in varied orientations with respect to the S2 foliation, but are mainly parallel to S2 (Fig. 2A, B; Zhang et al. 1997; Dubé et al. 2001, 2002). Carbonate ± quartz veins of different orientations show complex crosscutting relationships and are variably deformed. The foliation-parallel veins were boudinaged, whereas the foliation-perpendicular veins were symmetrically folded and the foliation-oblique veins were asymmetrically folded or boudinaged (Fig. 2B, C). Gold mineralization is related to silicification and sulfidation of the earlier barren to low-grade carbonate±quartz veins and breccias and replacement of adjacent wall rocks (Dubé et al. 2001, 2002 and references therein). It also occurs in the form of veinlets within the carbonate ± quartz veins (Fig. 2D). Dubé et al. (2001, 2002) presented evidence that the auriferous silicic replacement and associated arsenopyrite were, at least in part, syn- to postdevelopment of asymmetrical boudinage of carbonate ± quartz veins.

The fact that most of the pre-ore carbonate ± quartz veins were controlled by D2 structures suggests that the emplacement of the veins likely took place during D2. On the other hand, the observation that most of these veins were deformed by D2 strain suggests that carbonate ± quartz veining mainly occurred before or in the early stages of D2, whereas silicification and associated gold mineralization mainly took place during the main phase of D2 (Dubé et al. 2001, 2002). Importantly, it has been documented that some barren carbonate ± quartz veins postdate gold mineralization (Dubé et al. 2002), suggesting that barren carbonate ± quartz veins and gold mineralization took place in similar structural conditions during protracted hydrothermal activity, rather than as two separate events in distinct environments.

The main phase of gold mineralization was followed by emplacement of 2714–2712 Ma feldspar porphyry dikes (Dubé et al. 2004). These dikes, crosscutting the S2 foliation, trending SE and E-W and containing weak SE-trending foliation, are considered to have intruded into the Red Lake mine trend in the late stages of D2 or early in D3 (Dubé et al. 2004). Numerous SE-trending lamprophyre dikes, dated as 2702–2699 Ma, are postore and clearly cut D2 structures (Fig. 2E). The weakly developed SE-trending foliation in these lamprophyre dikes is thought to have been formed in D3 (Dubé et al. 2004), when the stress field is interpreted to have been similar to that in D2.

The latest stage deformation (D4) was characterized by formation of the discrete brittle ‘black-line faults’. These faults, cutting and displacing auriferous veins and lamprophyre dikes, are barren, discrete, 2- to 5-mm slip planes composed of fine quartz, tourmaline, and dark chlorite (Penczak & Mason 1997; Tarnocai 2000; Dubé et al. 2003). Locally, conjugate NNW- and WNW-trending black-line faults show sinistral and dextral displacement, respectively (Fig. 2F), suggesting a NW-SE-trending maximum principal stress.

Samples and methods of study

A total of 32 oriented samples and 13 nonoriented samples of vein material were collected from the Campbell mine (underground), Red Lake mine (underground), Cochenour mine (outcrop), Redcon prospect (outcrop), and Sandy Bay–Woodland outcrops along Highway 125 (Fig. 1). Of these samples, 14 were analyzed for microthermometry and 22 were measured for FIP orientations. The sample locations, lithologies, and relationship to structures are listed in Appendix S1. The first four localities lie within the ‘proximal ferroan-dolomite alteration zone’ of Parker (2000), where the main carbonate mineral is ankerite, whereas the last locality is in the ‘distal calcite alteration zone’ (Parker 2000). Most samples are from within the Red Lake mine trend, except for those from Redcon (Fig. 1).

The samples from the Campbell, Red Lake, and Cochenour mines are ankerite ± quartz veins with or without overprinting gold silicification. The samples from the Sandy Bay – Woodland outcrops are centimeter-scale calcite–quartz veins without gold mineralization. The Redcon
prospect is situated about 4 km northeast of the Red Lake mine trend, in an area of relatively weak deformation, where a 1–2-m wide ankerite vein cuts weak foliation and is in turn cut by quartz-actinolite stringers associated with Au. The Redcon samples were studied to provide information about the fluid pressure and stress field outside the main deformation zones, for comparison with those inside the deformation zones.

Vein minerals are divided into three stages: pre-ore, syn-ore, and postore, based on their relationships with gold mineralization. The predominant mineral in the pre-ore barren or low-grade carbonate ± quartz veins is ankerite, whereas the main vein mineral during the syn-ore stage is quartz. In the Sandy Bay–Woodland outcrop, the main vein mineral is calcite, which is considered to be equivalent to pre-ore ankerite in other localities, and quartz overprinting calcite is considered to be equivalent to syn-ore quartz, although there is no gold mineralization at this locality. Postore quartz and calcite occur in the form of veins that crosscut the postore feldspar porphyry dikes and lamprophyre dikes, or fill fractures and residual pores in earlier carbonate veins.

Doubly polished thin sections were made from horizontally cut slabs of the oriented samples, and were used for...
measurement of the orientation of FIPs, and for microthermometric studies. The strike of an FIP is readily determined by rotating the manual stage to find the angle between the strike of the FIP and the marked azimuth direction on the section. The dip angle of the FIP is estimated by changing the focus on different depths of the FIP as follows. First, the stage is rotated so that the FIP is parallel to the crosshair and focus is made on the upper part of the FIP. Then readings of the FIP on the horizontal crosshair (H1) and the focusing screw (V1) are recorded. The focus is then changed to a deeper part of the FIP, and new readings of these parameters (H2, V2) are taken. The horizontal displacement of the focus of the FIP (ΔH) is equal to (H2−H1) multiplied by the length per unit for the objective used (e.g. for the ×50 objective the length per unit is 2 microns). The vertical displacement of the focus of the FIP (ΔV) is equal to (V2−V1) multiplied by the depth per unit reading, which can be obtained by using a slide of known thickness. The dip angle (θ) can then be calculated using the formula θ = atan (ΔV/ΔH).

The dip angle measurement depends on the vertical extent of the FIP and the accuracy of reading of focused objects, which is better than 1 micron. For an FIP with a dip angle of 45° running through the whole thickness of the doubly polished section (100 microns), an error of 1 micron reading would introduce an error of dip angle estimation of 0.3°, whereas if the vertical extent of the FIP is 10 microns, an error of 1 micron reading corresponds to an error of 3° in dip angle. The error for gently dipping fluid inclusion assemblages (FIAs) is generally larger than for steeper ones. For example, for an FIP with a dip angle of 10°, the error of dip angle estimation introduced by an error of 1 micron reading is up to 6°. For most of the FIPs studied, the error of dip angle is estimated to be <5°.

Microthermometry was carried out with a Linkam THMSG 600 heating/freezing stage. Fluid inclusions were studied from seven oriented samples as well as seven non-oriented samples. The stage was calibrated using synthetic fluid inclusions of H2O (ice-melting temperature = 0°C; critical temperature = 374.1°C) and H2O-CO2 (CO2-melting temperature = −56.6°C). The accuracy of CO2-melting temperature (TmCO2), CO2-homogenization temperature (ThCO2), clathrate-melting temperature (Tmclath), and ice-melting temperature (TmH2O) is about ±0.2°C, whereas the total homogenization temperature is around ±0.5°C. The fluid inclusion assemblage (FIA) analysis method of Goldstein & Reynolds (1994) was employed to constrain the validity of the microthermometric data.

Unlike most fluid inclusion studies, which focus on primary or pseudosecondary inclusions, this study is mainly concerned with secondary or pseudosecondary fluid inclusions along FIPs because they provide information on both fluid P-T-X conditions and stress regimes. A general problem with secondary fluid inclusions is that they may have been entrapped any time after the formation of the host minerals. However, it is possible that secondary inclusions entrapped in one crystal represent the same fluids that were entrapped as primary inclusions in another crystal within the same vein, because most veins experienced multiple deformation and incremental events. Two methods have been used in this study to evaluate whether or not the secondary inclusions in a given mineral phase represent the fluids present during precipitation of that mineral. One is to compare the microthermometric data between the FIP and non-FIP inclusions. The FIP inclusions are most likely secondary (some may be pseudosecondary), whereas the non-FIP ones (including isolated, random, and clustered inclusions and those along growth zones) are possibly primary or pseudosecondary. A similarity in microthermometric attributes between the FIP and non-FIP inclusions is taken to indicate that the secondary inclusions have trapped the same fluids as primary and pseudosecondary inclusions. The second method is to compare the fluid inclusion types between minerals of different stages; if a combination of fluid inclusion types is present in a relatively late mineral phase but absent in a relatively early mineral, it is inferred that the late fluids did not infiltrate the early mineral and were not entrapped as secondary inclusions in that mineral.

**FLUID INCLUSION TYPES, OCCURRENCES, AND MICROTHERMOMETRY**

**Types of fluid inclusions**

Four types of fluid inclusions were recognized in the samples studied (Table 1): (i) CO2 ± CH4, (ii) H2O-NaCl, (iii) CO2-H2O-NaCl, and (iv) CH4-CO2. Type 1 inclusions are CO2-dominated and are generally monophase at room temperature with no visible aqueous phase (Fig. 3A, E). Microthermometry and laser Raman spectroscopy indicate the presence of minor to trace amounts of CH4 and trace N2 in some cases (Chi et al. 2006). The homogenization temperatures within individual FIAs are fairly consistent (Liu 2010), indicating that type 1 inclusions did not result from H2O leakage from initially aqueous-carbonic inclusions (Chi et al. 2006, 2009; Liu 2010).

Type 2 inclusions are aqueous fluid inclusions with variable amounts of salt (Fig. 3B). Trace amounts of gas are present in some type 2 inclusions as indicated by clathrate formation in freezing runs (Fig. 3B). These inclusions consist of two phases at room temperature with vapor/total ratios generally <20%.

Type 3 inclusions are aqueous-carbonic inclusions, mostly consisting of two phases at room temperature: a liquid aqueous phase and a liquid CO2 phase (Fig. 3C),
with CO₂ phase/total ratios generally larger than 90%. The water phase is invisible in some but is indicated by the formation of clathrate during freezing.

Type 4 inclusions are monophase at room temperature and appear similar to type 1, but they have significantly lower melting temperatures and homogenization temperatures than type 1 (Fig. 3D). Raman spectroscopic analysis also indicates the presence of significant amounts of CH₄.

The types of fluid inclusions present in individual samples are listed in Appendix S1, and their abundance in minerals of different stages at different localities are summarized in Table 1. Type 1 inclusions are predominant in pre-ore and syn-ore minerals in all the localities studied, whereas the other three types mainly occur in the postore minerals.

### Occurrences of fluid inclusions

The four types of fluid inclusions occur both in FIPs and in various non-FIP modes. The non-FIP occurrences include clusters (Fig. 3B, C, D), isolated inclusions (Fig. 3E), randomly distributed inclusions (Fig. 3A), and growth zones (Fig. 3F). FIPs appear as intracrystal or intercrystal arrays consisting of numerous tiny (generally

---

**Table 1** Distribution of different types of fluid inclusions in different host minerals and study areas.

<table>
<thead>
<tr>
<th>Fluid type</th>
<th>Major components</th>
<th>Campbell-Red Lake</th>
<th>Cochenour</th>
<th>Redcon</th>
<th>Sandy Bay-Woodland</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type 1</td>
<td>CO₂ ± CH₄</td>
<td>+++</td>
<td>+++</td>
<td>+++</td>
<td>+++</td>
</tr>
<tr>
<td>Type 2</td>
<td>H₂O-NaCl</td>
<td>––</td>
<td>+</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Type 3</td>
<td>H₂O-CO₂-NaCl₂</td>
<td>––</td>
<td>–</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>Type 4</td>
<td>CH₄-CO₂</td>
<td>––</td>
<td>+</td>
<td>++</td>
<td>+</td>
</tr>
</tbody>
</table>

+++ predominant; ++, abundant; +, common; –, rare; blank, not observed.

---

**Fig. 3.** Photomicrographs of four different types of fluid inclusions. (A) Type 1, CO₂-dominated inclusions, monophase at room temperature (syn-ore quartz, Redcon prospect, sample no. GC–132); (B) type 2, H₂O-NaCl fluid inclusions, photograph taken at –1°C showing some gas hydrate crystals (postore calcite, Red Lake mine, sample no. GC–116); (C) type 3, CO₂-H₂O-NaCl inclusions (postore quartz, Cochenour outcrop, sample no. 23–8); (D) type 4, CH₄-CO₂ inclusions, photograph taken at –95°C, showing three coexisting phases (postore quartz, Cochenour outcrop, sample no. 23–8); (E) an isolated type 1 inclusion (syn-ore quartz, Red Lake mine, sample no. 21–6); (F) growth zones lined by type 1 fluid inclusions (pre-ore ankerite, Red Lake mine, sample no. KG 2000-50-5, from Chi et al. 2002).
Fluid inclusion microthermometry

Microthermometric measurements were made for fluid inclusions in pre-ore ankerite, syn-ore quartz, and postore quartz, both in FIPs and in non-FIPs. The results are summarized in Table 2 and Figs 5 and 6, which include data from Chi et al. (2002, 2003) in addition to those obtained in this study. The samples in which microthermometric measurements were taken in this study are listed in Appendix S1.

Pre-ore ankerite and quartz are dominated by type 1 fluid inclusions, with few type 2 and 3 inclusions. Microthermometry was only performed on relatively large non-FIP inclusions, and no data were obtained from fluid inclusions in FIPs. The melting temperature of the carbonic phase (\(T_{mCO_2}\)) of type 1 inclusions ranges from \(-61.2\) to \(-56.6^\circ C\) (Table 2), consistent with CO₂ being the dominant composition with minor other gases such as CH₄ and N₂. The temperatures of homogenization of the carbonic phase (\(T_{hCO_2}\)), mostly to the liquid phase, cover a wide range from \(-4.1\) to \(30.6^\circ C\) (Table 2 and Fig. 5A). The \(T_{mCO_2}\) values from Cochenour are generally lower than those from other localities (Table 2 and Fig. 5A). The few type 2 inclusions studied by Chi et al. (2002, 2003) yielded homogenization temperatures from 153 to 344°C (Table 2 and Fig. 5B) and ice-melting temperatures (\(T_{mH_2O}\)) from \(-22.7\) to \(-21.3^\circ C\) (Table 2).

Syn-ore quartz is also dominated by type 1 inclusions, with few type 2 and 3 inclusions. The type 2 and type 3 inclusions do not occur in the same FIP or cluster as type
Table 2. Microthermometric data for different types of fluid inclusions in different stages and study areas.

<table>
<thead>
<tr>
<th>FI type</th>
<th>Stage</th>
<th>Data</th>
<th>Campbell</th>
<th>Red Lake</th>
<th>Cochenour</th>
<th>Redcon</th>
<th>Sandy Bay-Woodland</th>
</tr>
</thead>
<tbody>
<tr>
<td>Type 1</td>
<td>Pre-ore</td>
<td>TmCO2 -57.5 to -56.6</td>
<td>-57.0 to -56.6</td>
<td>-58.5 to -57.1</td>
<td>-61.2 to -59.5</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 -2.4 to 28.1</td>
<td>1.9 to 30.6</td>
<td>-4.1 to 13.8(L)</td>
<td>2.0 to 26.5(L)</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Syn-ore</td>
<td>TmCO2 -57.8 to -56.9</td>
<td>-59.1 to -56.9 FIP</td>
<td>-61.7 to -56.8 FIP</td>
<td>-58.3 to -56.7 FIP</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 -7.6 to 12.7(L)</td>
<td>10.1 to 26.3(L)</td>
<td>-0.2 to 20.5</td>
<td>-26.0 to 31.0(L)</td>
<td>-16.5 to -14.0(L)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Postore</td>
<td>TmCO2 -57.9 to -57.7</td>
<td>-57.0 to -56.9</td>
<td>-60.9 to -56.9</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 0.9 to 28.8(L) #</td>
<td>22.0 to 23.4(L)</td>
<td>-15.3 to 22.7(L)</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Type 2</td>
<td>Pre-ore</td>
<td>TmCO2 171(L)</td>
<td>192 to 344(L)</td>
<td>153 to 173(L)</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 -252 to 380(L)</td>
<td>-</td>
<td>162 to &gt;250(L)</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Postore</td>
<td>TmCO2 72 to 326(L)</td>
<td>75 to &gt;200(L)</td>
<td>106 to 284(L)</td>
<td>-</td>
<td>168 to 298(L)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 58.3 to 104 to 388(L) # FIP</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Type 3</td>
<td>Syn-ore</td>
<td>TmCO2 -8.2(L)</td>
<td>-</td>
<td>-</td>
<td>-57.0 to -56.6</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tmclath 8.2(L)</td>
<td>-</td>
<td>7.8 to 8.9(L)</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 -253 to 380(L)</td>
<td>-</td>
<td>-8.4 to 28.2</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Postore</td>
<td>TmCO2 &gt;200</td>
<td>-</td>
<td>235 to &gt;255(V)</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tmclath -</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>THCO2 -2.0 to 8.9</td>
<td>-</td>
<td>-7.5 to 7.9 FIP</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>Type 4</td>
<td>Postore</td>
<td>TmCO2 -</td>
<td>-</td>
<td>-7.18 to -67.2</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Tmclath -</td>
<td>-</td>
<td>-69.8 to -48.2</td>
<td>-</td>
<td>-</td>
<td></td>
</tr>
</tbody>
</table>

*Fluid inclusion planes (FIP) data are shaded and non-FIP data are not shaded. All FIP data and most non-FIP data are from this study; some of the non-FIP data (underlined) are from Chi et al. (2002, 2003); TmCO2, CO2-melting temperature; THCO2, CO2 homogenization temperature; TmH2O, ice-melting temperature; ThH2O, aqueous inclusion homogenization temperature; Tmclath, clathrate-melting temperature; ThT total, total homogenization temperature of carbonic-aqueous inclusions; L, homogenized to liquid; V, homogenized to vapor; #, some homogenized to vapor.

1 inclusions, and their relative timing cannot be determined. TmCO2 of type 1 inclusions ranges from -61.7 to -56.6°C (Table 2), and THCO2 (to liquid) ranges from -26.0 to 31.0°C (Table 2 and Fig. 5A). Most of the low TmCO2 values (<0°C) are from Redcon and Sandy Bay-Woodland (Fig. 5A). In samples where both FIP and non-FIP inclusions were measured, their THCO2 values largely overlap, although those of non-FIP inclusions tend to extend to lower values (Fig. 6A). The few type 2 and 3 inclusions studied show homogenization temperatures from 162 to 380°C and 235 to >255°C, respectively (Table 2 and Fig. 5B, C). TmH2O of type 2 inclusions ranges from -6.2 to -5.6°C, and clathrate-melting temperatures (Tmclath) of type 3 inclusions range from 7.8 to 8.9°C (Table 2).

Postore quartz contains all four types of inclusions. Different types of fluid inclusions occur in separate FIPs or clusters. TmCO2 of type 1 and type 3 inclusions ranges from -60.9 to -56.9°C and from -61.2 to -56.9°C, respectively, and THCO2 from -15.3 to 28.8°C and from -5.8 to 22.5°C, respectively (Table 2). In samples where both FIP and non-FIP inclusions were measured, their THCO2 values largely overlap (Fig. 6B). Type 2 inclusions show homogenization temperatures from 72 to 388°C (Table 2 and Fig. 5B), and type 3 inclusions have homogenization temperatures from 116 to >310°C (Table 2 and Fig. 5C). TmH2O of type 2 inclusions ranges from -25.6 to -0.2°C, and Tmclath of type 3 inclusions ranges from 2.0 to 8.9°C (Table 2). Type 4 fluid inclusions have TmCO2 ranging from -71.8 to -67.2°C and THCO2 from -69.8 to -48.2°C (Table 2), with estimated XH2O between 0.6 and 0.8 based on the diagrams of Thiery et al. (1994).

**ORIENTATION OF FIPS**

**FIPs in pre-ore ankerite**

The FIPs in pre-ore ankerite are generally intracrystal and short (<0.1 mm) and are typically lined by small (mostly <2 μm), rhombic, elliptic, and irregularly shaped fluid
inclusions. The attitudes of the FIPs are variable, but two preferred orientations are recognized: one strikes NNE-SSW and dips steeply to WNW or ESE (group A1), and the other strikes NW-SE and dips steeply to NE or SW (group A2) (Fig. 7, column A). The strike and dip angle of the preferred orientations vary slightly for different localities, with A1 being 031°/88°, 193°/76°, 017°/74°, and 194°/75° for Red Lake mine, Campbell mine, Cochenour outcrop, and Redcon outcrop, respectively, and A2 being 330°/75°, 326°/70°, 311°/75°, and 136°/84° for these localities (Fig. 7, column B). The two sets of FIPs show mutual crosscutting relationships, suggesting that they are broadly contemporaneous.

**FIPs in syn-ore quartz**

The FIPs in the syn-ore quartz are generally longer and thicker, with fluid inclusions of larger size than those in the pre-ore ankerite. The rounded, irregular, or negative-crystal-shape fluid inclusions that characterize these FIPs are mostly CO₂ dominated and are monophase at room temperature. Crosscutting relationships indicate that the NNE-SSW-trending group is older than the NW-SE-trending one. Locally, both sets are crosscut by some NE-striking and SE-dipping FIPs.

**FIPs in postore quartz**

The postore quartz is less deformed than the pre-ore ankerite and syn-ore quartz, and the FIPs tend to display simple distribution patterns. In contrast to the FIPs in earlier-formed minerals, the FIPs in postore quartz commonly consist of carbonic-aqueous or aqueous inclusions, which are composed of two phases at room temperature. Four groups of FIPs are recognized (Fig. 7, column C). The first and oldest group (C1), consisting of mainly monophase CO₂-dominated inclusions, strikes NW-SE and dips steeply to the NE or SW. The strike and dip angle of group C1 are 318°/83° for the Campbell mine, 311°/76° for the Cochenour outcrop, and 137°/76° for the Sandy Bay–Woodland outcrop (Fig. 7, column C). The second statistically more abundant group (C2) strikes WNW-ESE and dips steeply to the NNE or SSW, and is generally composed of two-phase aqueous or carbonic-aqueous fluid inclusions. The strike and dip angle of group C2 are 280°/76° for the Red Lake mine, 274°/73° for the Campbell mine, and 104°/84° for the Sandy Bay–Woodland outcrop (Fig. 7, column C). The third group of FIPs (C3), also consisting of two-phase aqueous inclusions, strikes NNE and dips steeply to the ESE, with the strike and dip angle being 321°/86°, 330°/78°, 298°/89°, and 146°/87° for Red Lake mine, Campbell mine, Redcon outcrop, and Sandy Bay–Woodland outcrop, respectively (Fig. 7, column C). The two sets of FIPs show mutual crosscutting relationships, suggesting that they are broadly contemporaneous.

angle being 006°/71° for the Campbell mine, 001°/73° for the Cochenour outcrop, and 011°/72° for the Redcon outcrop (Fig. 7, column C). No mutual crosscutting relationships have been observed between C2 and C3. In the Redcon outcrop, group C3 FIPs are cut by FIPs that strike NNW and dip ENE (C4, 341°/71°, Fig. 7).

Implications for the stress field

Both experimental tests (Brace & Bombolakis 1963; Tapponnier & Brace 1976; Krantz 1979) and structural studies dealing with microcracks in relation to regional stress field (Tuttle 1949; Lespinasse & Pecher 1986) have revealed that FIPs tend to develop in a preferred orientation: parallel to the maximum principal compressive stress (σ1) and perpendicular to the least principal stress (σ3). Thus, the FIPs are interpreted as mainly Mode I (extensional) cracks and can be used as structural markers of paleostress fields (Lespinasse & Pecher 1986; Boullier & Robert 1992; Lespinasse 1999), although it cannot be ruled out that some microcracks may have formed as shear fractures.
Similarities and differences in FIP orientations between different stages and between different localities can be perceived in Fig. 7. The similarities between different localities may reflect a common general stress field and similar structural environment along the Red Lake mine trend, and the differences are likely caused by variation of local stress fields. Although the similarities between different stages may be interpreted simply as relatively late deformation events overprinting relatively early mineral phases, it appears more likely that most of the FIPs were formed during the increment-deformation history within individual stages. This inference is based in part on the similarities in microthermometric attributes between FIP and non-FIP inclusions from the same stage, in part on the observation that aqueous and aqueous-carbonic inclusions abundant in postore FIPs were not found in pre-ore and syn-ore FIPs, and in part on comparison of FIP orientations at individual localities. For example, at the Red Lake mine and Sandy Bay–Woodland outcrop, C2 is well developed in the postore quartz, but FIPs of the same orientation are not found in the pre-ore ankerite and syn-ore quartz (Fig. 7, first and last rows). At the Redcon outcrop, C3 and C4 are well developed in the postore quartz, but FIPs with similar orientations are absent in the syn-ore quartz (Fig. 7, fourth row).

Based on the relative timing of the host minerals and the crosscutting relationships of the FIPs, the sequence of development of the major sets of FIPs is summarized in Fig. 8: from foliation-subperpendicular (A1) to foliation-subparallel (A2) in the pre-ore stage, through foliation-perpendicular (B1) and foliation-parallel (B2) in the syn-ore stage, to foliation-parallel (C1), then foliation-subparallel

---

**Fig. 7.** Equal-area, lower-hemisphere stereographic projection of poles to fluid inclusion planes (FIPs) in pre-ore ankerite, syn-ore quartz, and postore quartz. A1–A2, B1–B2, and C1–C4 represent the preferential orientations of poles to FIPs at different stages – see text for detailed discussion.

<table>
<thead>
<tr>
<th></th>
<th>Pre-ore ankerite</th>
<th>Syn-ore quartz</th>
<th>Post-ore quartz</th>
</tr>
</thead>
<tbody>
<tr>
<td>Red Lake mine</td>
<td><img src="image1" alt="Pre-ore ankerite" /></td>
<td><img src="image2" alt="Syn-ore quartz" /></td>
<td><img src="image3" alt="Post-ore quartz" /></td>
</tr>
<tr>
<td>n</td>
<td>337</td>
<td>77</td>
<td>22</td>
</tr>
<tr>
<td>Campbell mine</td>
<td><img src="image4" alt="Pre-ore ankerite" /></td>
<td><img src="image5" alt="Syn-ore quartz" /></td>
<td><img src="image6" alt="Post-ore quartz" /></td>
</tr>
<tr>
<td>n</td>
<td>190</td>
<td>92</td>
<td>60</td>
</tr>
<tr>
<td>Cochenour outcrop</td>
<td><img src="image7" alt="Pre-ore ankerite" /></td>
<td><img src="image8" alt="Syn-ore quartz" /></td>
<td><img src="image9" alt="Post-ore quartz" /></td>
</tr>
<tr>
<td>n</td>
<td>31</td>
<td>92</td>
<td>60</td>
</tr>
<tr>
<td>Redcon outcrop</td>
<td><img src="image10" alt="Pre-ore ankerite" /></td>
<td><img src="image11" alt="Syn-ore quartz" /></td>
<td><img src="image12" alt="Post-ore quartz" /></td>
</tr>
<tr>
<td>n</td>
<td>95</td>
<td>148</td>
<td>65</td>
</tr>
<tr>
<td>Sandy Bay–Woodland outcrop</td>
<td><img src="image13" alt="Pre-ore ankerite" /></td>
<td><img src="image14" alt="Syn-ore quartz" /></td>
<td><img src="image15" alt="Post-ore quartz" /></td>
</tr>
<tr>
<td>n</td>
<td>67</td>
<td>78</td>
<td>65</td>
</tr>
</tbody>
</table>

**Legends:**
- Red
- Blue
- Green
- Black

(C2) and foliation-subperpendicular (C3), and finally foliation-oblique (C4) in the postore stage. The orientations of FIPs from the Redcon outcrop, which is outside the Red Lake mine trend, are similar to those within the Red Lake mine trend (Fig. 7).

The observation that most FIPs are subvertical suggests that \( r_3 \) (represented by poles of the FIPs) was subhorizontal during the formation of the FIPs, whereas the \( r_1–r_2 \) plane was subvertical. This contradicts the inference from boudinage and other stretching structures that \( r_3 \) switched between subhorizontal and subvertical orientations (Zhang et al. 1997; Dubé et al. 2002), and was not just subhorizontal. This discrepancy, together with the orientation of \( r_1 \) and \( r_2 \) (i.e. whether \( r_1 \) was subvertical and \( r_2 \) was subhorizontal, or vice versa), which cannot be determined from the FIP orientation data alone, will be further discussed below.

**DISCUSSION**

**Estimation of fluid pressure during veining and microfracturing**

Fluid pressure plays an important role in fracturing and vein formation (Hubbert & Willis 1957; Secor 1965, 1969; Phillips 1972; Etheridge 1983; Sibson et al. 1988; Hodgson 1989; Cox 1995, 2005). Fluid pressure can also be used to estimate the depth of hydrothermal circulation, and depth is one of the important factors controlling the brittle or ductile behavior of rocks. The fluid pressures in the Red lake mine trend during formation of the pre-ore carbonate ± quartz veins, syn-ore silicification and quartz veins, and postore quartz-carbonate veins, as well as during microfracturing in the various stages, may be estimated from fluid inclusion microthermometric data.

The abundant carbonic (type 1) fluid inclusions provide a good opportunity for fluid pressure estimation. This type of fluid inclusion has relatively gentle isochores in the temperature (horizontal axis)–pressure (vertical axis) space, and represents a better geobarometer than aqueous and aqueous-carbonic inclusions (Roedder & Bodnar 1980; Brown 1989). Calculation of fluid pressure from isochores requires knowing fluid temperature, which can be approximated by the total homogenization temperature of aqueous (type 2) or aqueous-carbonic inclusions (type 3) if these inclusions were entrapped simultaneously from a fluid system consisting of two immiscible phases (Roedder & Bodnar 1980). Unfortunately, such conditions do not appear to be satisfied in the Red Lake mine trend, as there is no petrographic evidence to indicate that the aqueous fluid was in equilibrium with a carbonic phase when it was entrapped, even though both aqueous and carbonic inclusions occur in the same minerals, and phase modeling indicates that the carbonic inclusions may have resulted from phase separation of a CO\(_2\)-dominated carbonic-aqueous fluid (Chi et al. 2006, 2009). The wide range of homogenization temperatures of type 2 fluid inclusions (153–344, 162–380, and 72–388°C for pre-ore, syn-ore and postore minerals, respectively; Table 2 and Fig. 5B) can be best explained by fluid pressure fluctuation and physical separation of aqueous and carbonic fluids during entrapment (Robert & Kelly 1987; Chi et al. 2009), which also implies that the fluid temperatures were higher than the fluid inclusion homogenization temperatures.

Therefore, an independent estimate of temperature conditions is required to calculate fluid pressures from the isochores of type 1 inclusions. As discussed earlier, various lines of evidence from structural and petrographic studies indicate that gold mineralization (and syn-ore silicification) took place during peak regional metamorphism (Dubé...
et al. 2001, 2002), which is in the upper greenschist – lower amphibole transition in the study areas (except Sandy Bay – Woodland, which is in the greenschist facies). Therefore, the formation temperature of the syn-ore quartz is inferred to be in the range of 400–550°C. Although carbonate ± quartz veining predated the main phase of gold mineralization, they likely took place in generally similar conditions rather than as two separate events in distinct environments (Dubé et al. 2002). Therefore, the formation temperature of the carbonate ± quartz veins may have been similar to, or slightly lower than, the 400–550°C range. A temperature range of 300–450°C is tentatively adopted in this study. In the postore stage, the fluid temperatures cannot be determined with certainty, but it is likely that they overlap with or are generally <400–550°C peak metamorphism range, or are in the upper part of the 72–388°C range of type 2 inclusion homogenization temperatures. A temperature range of 250–400°C is tentatively used in this study.

Using the temperature ranges of 300–450, 400–550, and 250–400°C for pre-ore, syn-ore, and postore respectively, and the isochores of type 1 fluid inclusions calculated from the equation of state for CO₂ of Bottinga & Richet (1981) as provided in Brown (1989), ranges of fluid pressures were calculated for each stage (Table 3). Fluid pressures calculated from FIP inclusions overlap with those from non-FIP inclusions (Fig. 9). Fluid pressures in the pre-ore stage range from 919 to 3381 bars, 706 to 3114 bars, 1606 to 3462 bars, and 1012 to 3107 bars for Campbell mine, Red Lake mine, Cochenour, and Redcon, respectively. Fluid pressures in the syn-ore stage range from 1481 to 4322 bars, 1321 to 3610 bars, 1704 to 3891 bars, 942 to 5379 bars, and 3563 to 4827 bars for Campbell mine, Red Lake mine, Cochenour, and Redcon, respectively. Fluid pressures in the postore stage range from 733 to 2873 bars, 981 to 1615 bars, and 1009 to 3621 bars for Campbell mine, Red Lake mine, and Cochenour, respectively. These pressure ranges may have been exaggerated because of the uncertainties of temperatures; the actual scale of pressure fluctuation within each stage may be comparable to the difference between the maximum and minimum pressures for a given temperature, for example, 919–2384 bars at 300°C for the pre-ore stage at the Campbell mine (Table 3).

### Vein formation depths and brittle deformation mechanisms

As discussed earlier, the abundance of colloform-crustiform structures and cockade breccias in the carbonate ± quartz veins from the Red Lake mine trend has led to debate regarding whether the veins were formed in an epithermal environment or at great depths during regional metamorphism. Vein formation depths may be inferred from fluid pressures if the pressure regime is known. Generally the fluid pressure is near-hydrostatic at shallow crustal levels and near-lithostatic at greater depths, with the transition zone being typically between 5 and 15 km (Cox 2005). Fluid pressure fluctuation associated with rock rupturing is typically between hydrostatic and subhydrostatic in epithermal environments, and between lithostatic and hydrostatic in mesothermal environments (Sibson et al. 1988).

In the case of the Red Lake mine trend, the fluid pressure values estimated above are incompatible with an epithermal environment for any of the three stages. For example, for a pressure range of 919–3381 bars for the pre-ore stage in the Campbell mine, a hypothesis of a hydrostatic-subhydrostatic system would mean a formation depth of about 34 km. Therefore, it is more reasonable to assume hydrostatic-lithostatic fluid-pressure fluctuation and that the minimum pressures represent hydrostatic pressures. For the pre-ore stage, when most of the ankerite ± quartz veins were formed, the smallest fluid pressure found was 706 bars (at 300°C) in the Red Lake mine (Table 3), which corresponds to a depth of about 7 km under a hydrostatic regime, still much larger than what is defined as an epithermal environment.

### Table 3 Ranges of fluid pressures (bars) calculated from isochores of type 1 fluid inclusions for different temperature ranges.

<table>
<thead>
<tr>
<th>Localities</th>
<th>Fluid pressure (bars)</th>
<th>Pre-ore</th>
<th>Syn-ore</th>
<th>Postore</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>T = 300°C</td>
<td>T = 450°C</td>
<td>T = 400°C</td>
</tr>
<tr>
<td>Campbell mine</td>
<td>Upper limit</td>
<td>2384</td>
<td>3381</td>
<td>3280</td>
</tr>
<tr>
<td></td>
<td>Lower limit</td>
<td>919</td>
<td>1347</td>
<td>1481</td>
</tr>
<tr>
<td>Red Lake mine</td>
<td>Upper limit</td>
<td>2186</td>
<td>3114</td>
<td>2305</td>
</tr>
<tr>
<td></td>
<td>Lower limit</td>
<td>706</td>
<td>1030</td>
<td>1321</td>
</tr>
<tr>
<td>Cochenour</td>
<td>Upper limit</td>
<td>2443</td>
<td>3462</td>
<td>2946</td>
</tr>
<tr>
<td></td>
<td>Lower limit</td>
<td>1606</td>
<td>2326</td>
<td>1704</td>
</tr>
<tr>
<td>Redcon</td>
<td>Upper limit</td>
<td>2181</td>
<td>3107</td>
<td>4112</td>
</tr>
<tr>
<td></td>
<td>Lower limit</td>
<td>1012</td>
<td>1483</td>
<td>942</td>
</tr>
<tr>
<td>Sandy Bay–Woodland</td>
<td>Upper limit</td>
<td></td>
<td>3675</td>
<td>4827</td>
</tr>
<tr>
<td></td>
<td>Lower limit</td>
<td></td>
<td>3563</td>
<td>4684</td>
</tr>
</tbody>
</table>

© 2011 Blackwell Publishing Ltd, Geofluids, 11, 260–279
Fluid pressure can shift a Mohr circle from the region of stability in a Mohr diagram toward lower normal stress values (the effective stress effect) so that the failure envelope is intersected and fractures produced (Hubbert & Willis 1957; Phillips 1972). Therefore, high fluid pressure tends to neutralize load pressure so that brittle deformation can occur at deep crustal levels (Hodgson 1989). Stated in another way, brittle deformation may not develop at great depths unless high-pressure fluids are involved. High fluid pressures recorded by fluid inclusions in the various hydrothermal stages (Table 3) are at least in part responsible for the brittle deformation observed in the Red Lake mine trend.

The argument that high fluid pressure played a critical role in brittle deformation is also supported by our observation that FIPs containing aqueous-bearing fluid inclusions (types 2 and 3), which are common in postrift minerals, were rarely found in the pre-ore and syn-ore minerals (Table 1). Given that type 2, 3, and 4 inclusions occur as primary and secondary inclusions in postrift minerals, i.e. representing fluids circulating after the formation of the pre-ore and syn-ore minerals, it is remarkable that they rarely occur as secondary inclusions in FIPs in these minerals. The pre-ore and syn-ore minerals may have been cut off from the fluid conduits in the postrift stage because of reduced porosity and permeability, so that there was insufficient fluid around the host minerals to trigger formation of microfractures.

Another factor that may have contributed to brittle deformation in the Red Lake mine trend is the ‘dry’ nature of the fluids in the pre- and syn-ore stage, as reflected by the predominance of CO2-dominated fluid inclusions. It is known that CO2-dominated fluids have higher wetting angles than aqueous fluids (Watson & Brenan 1987). This may have prevented interconnection of the fluid phase in the host rocks (leading to build-up of fluid pressure) and thus enhanced fluid pressure build-up and brittle failure.

**Formation of foliation-parallel veins and stress analysis**

Most shear zone-hosted gold deposits are characterized by abundant veins emplaced at various angles to the main foliation (S fabric) and occupying structural elements such as extensional fractures and various shear fractures (R, R', P, P', and C) (Hodgson 1989). The central veins (fault-fill veins), which are located at the center of the shear zones and parallel to shear zone boundaries, are also commonly oblique – in plan or cross-section view depending on the sense of motion, to the main foliation (S fabric) – even though they are parallel to the foliation immediately surrounding them (the C fabric) (Hodgson 1989). However, in the Red Lake mine trend, most veins are parallel or sub-parallel to the foliation, and the mechanism remains poorly understood. While extensional fracturing caused by outer arc stretching during folding may be invoked to explain the occurrence of foliation-parallel carbonate ± quartz veins near fold hinges (Dubé et al. 2001, 2002), it is difficult to appeal to such a mechanism for veins in other structural locations. The formation of foliation-parallel veins is most likely related to the interplay between the stress field and fluid pressure fluctuation, as explained below.

---

**Fig. 9.** Isochores of type 1 fluid inclusions showing the ranges of fluid pressures for temperature from 400 to 550°C in the syn-ore stage from Campb-ell mine (A), Red Lake mine (B), and Redcon (C); note the overlap of fluid pressures between fluid inclusion planes (FIP) and non-FIP fluid inclusions.
Vein formation requires that fractures be opened, and fracture opening depends on a combination of differential stress, fluid pressure, and fracture orientation (Hubbert & Willis 1957; Phillips 1972). Open space may be produced along shear fractures through mechanisms such as uneven rotation of the fractured rocks (Ramsay 1967), movement on nonplanar shear surfaces (Guha et al. 1983), and inhomogeneous deformation (Hodgson 1989). When the effect of fluid pressure is considered, open space may be created along a fracture if the effective normal stress perpendicular to the fracture \( (\sigma_n - \text{P}_{\text{fluid}}) \) is negative (tensile). Such conditions may be met if the differential stress \( (\sigma_1 - \sigma_3) \) is small and if the fluid pressure is high enough that the effective minimum principal stress \( (\sigma_3 - \text{P}_{\text{fluid}}) \) is negative (Cox 2005), as illustrated in Fig. 10A. When the differential stress is \( >5.7 |T_0| \), where \( T_0 \) is the tensile strength, the Mohr circle will intersect the Coulomb failure envelope at a position with positive normal stress, so shear fractures are created; however, they cannot be opened because the effective normal stress is positive (compressive). For \( \sigma_1 - \sigma_3 < 4 |T_0| \) (circle a3, Fig. 10A), the Mohr circle intersects the Griffith failure envelope at the tensile strength \( (T_0) \), and a tensile fracture is formed; the fracture is perpendicular to \( \sigma_3 \), as shown by the \( 2\theta \) of 0°, where \( \theta \) is the angle between fracture plane and \( \sigma_1 \) (Fig. 10A). For \( 4 |T_0| < \sigma_1 - \sigma_3 < 5.7 |T_0| \) (circle a2, Fig. 10A), extensional-shear (or trans-tensional) fractures are formed and they are able to open because the effective normal stress is negative; the fractures are at high angles to \( \sigma_3 \), as shown by 20 values of 0–50° (Fig. 10A).

For the foliation-parallel veins, the creation and opening of fractures needs special conditions because this orientation (perpendicular to \( \sigma_1 \)) is generally unfavorable for fracturing. The key factor is the high mechanical anisotropy of foliated rocks: the tensile strength in the direction of the foliation (labeled \( T_\parallel \) because the tensile fracture thus produced is parallel to the foliation) is much higher than that perpendicular to the foliation (labeled \( T_\perp \) because the tensile fracture thus produced is perpendicular to the foliation) (Kerrich 1989). Under these circumstances, two Coulomb-Griffith failure envelopes can be used to evaluate failure criteria on the Mohr diagram, one for failure normal to foliation and the other for failure parallel to foliation (Fig. 10B). Foliation-parallel fractures will only occur when the maximum effective stress \( (\sigma_1^*) \) intersects the \( T_\parallel \) tensile failure envelope, while the minimum effective stress \( (\sigma_3^*) \) is not small enough to intersect \( T_\perp \) envelope and cause failure in the direction normal to the foliation

---

**Fig. 10.** (A) Mohr diagram showing the relationship between stress and fracturing in mechanically isotropic rocks. Circles a1, a2, and a3 depict shear fracturing, extensional-shear fracturing, and extensional fracturing, respectively (modified from Davis & Reynolds 1996). \( T_0 \) is the tensile strength of the rock, and \( \theta \) is the angle between fracture plane and \( \sigma_1 \); (B) Mohr diagram showing the relationship between stress and fracturing in foliated rocks, which have different tensile strength along foliation \( (T_\parallel) \) and perpendicular to foliation \( (T_\perp) \). Circle b1 indicates that foliation-parallel fracturing can take place when \( \sigma_1 - \sigma_3 < 1 T_\perp = T_\perp \) and \( \sigma_1^* = T_\parallel \); (C) block diagrams showing three possible stress configurations, which may have been developed in the Red Lake mine trend and the attitudes of veins and fluid inclusion planes; for clarity, only foliation-parallel and foliation-perpendicular veins are shown.

vertical and subhorizontal orientations (Zhang et al. 1997; Dubeé et al. 1986; Zhang et al. 2006, 2009). It is postulated that the upward force associated with the overpressured fluid reservoir below the shear zones, commonly assumed to be present underneath shear zone-controlled vein systems (e.g. Sibson et al. 1988), which may have been episodically larger than the horizontal compressive principal stress. The other is that, as they were formed, the veins may have been physically detached from the surrounding rock mass, resulting in vertical principal stress being locally larger than the horizontal ones. A third possible mechanism is the upward force exerted by magmatic intrusions. However, the mutual crosscutting relationships of FIPs suggest that the stress field changed more frequently than can be accounted for by the limited number of episodes of magmatic activity. Based on fluid composition (CO₂-dominated) and noble gas isotope analysis, it has been proposed that the fluids responsible for carbonate veining and gold mineralization in the Red Lake mine trend were derived from devolatilization in granulite facies conditions at deep crustal levels (Chi et al. 2006, 2009). It is postulated that the upward force associated with the development of this deep fluid reservoir is responsible for the subvertical σ₁ inferred from the FIP data. The coupling of a subvertical maximum principal stress with the fluid source represents an ideal condition for the formation of foliation-parallel veins.

In the first situation (Fig. 10c-1), the main foliation and boudinage structures indicating subvertical stretching were developed, and minor subhorizontal veins or FIPs were formed. In the second situation, the main foliation and boudinage structures indicating subhorizontal stretching were developed, and either foliation-perpendicular veins (if the Mohr circle was in contact with T₁ in Fig. 10b) or foliation-parallel veins (if the Mohr circle was in contact with Tᵢ in Fig. 10b), as well as foliation-perpendicular FIPs, were formed (Fig. 10c-2). In the third situation, foliation-parallel veins and FIPs were developed (Fig. 10c-3); the foliation represents a mechanical weakness in the rock and fractures were preferentially opened along it even when σ₂ was not perpendicular to it. The alternating development of the three stress regimes implies that the differential stress was probably relatively small and that the rapid changes in stress orientations were a local rather than orogen-scale phenomenon.

The subvertical-σ₁ scenario may seem unusual in a tectonic environment such as the Red Lake greenstone belt that was presumably dominated by horizontal shortening. However, it has been inferred in other orogenic gold systems elsewhere, such as the gold-quartz vein deposits in the Val d’Or area in the Abitibi greenstone belt (Boullier & Robert 1992) and the Yellowknife greenstone belt (Kerrich & Allison 1978). It is possible that such a stress regime is inherently associated with shear zone-controlled hydrothermal systems, and two mechanisms are proposed here. One is the upward force associated with the overpressured fluid reservoir below the shear zones, commonly assumed to be present underneath shear zone-controlled vein systems (e.g. Sibson et al. 1988), which may have been episodically larger than the horizontal compressive principal stress. The other is that, as they were formed, the veins may have been physically detached from the surrounding rock mass, resulting in vertical principal stress being locally larger than the horizontal ones. A third possible mechanism is the upward force exerted by magmatic intrusions. However, the mutual crosscutting relationships of FIPs suggest that the stress field changed more frequently than can be accounted for by the limited number of episodes of magmatic activity. Based on fluid composition (CO₂-dominated) and noble gas isotope analysis, it has been proposed that the fluids responsible for carbonate veining and gold mineralization in the Red Lake mine trend were derived from devolatilization in granulite facies conditions at deep crustal levels (Chi et al. 2006, 2009). It is postulated that the upward force associated with the development of this deep fluid reservoir is responsible for the subvertical σ₁ inferred from the FIP data. The coupling of a subvertical maximum principal stress with the fluid source represents an ideal condition for the formation of foliation-parallel veins.
CONCLUSIONS

(1) Fluids circulating in the Red Lake mine trend during carbonate veining and subsequent silicification and gold mineralization were dominated by CO$_2$, whereas aqueous-bearing fluids were active after mineralization. The high fluid pressure and ‘dry’ nature (high wetting angles) of the CO$_2$-dominated fluids may have been responsible for the brittle deformation and formation of veins with colloform-crustiform structures and cockade breccias at depths around 7 km or more, under upper greenish- to lower amphibole-facies conditions.

(2) The fluids circulating in microfractures recorded by fluid inclusion planes (FIPs) were the same as those precipitating vein minerals recorded by non-FIP inclusions. Aqueous-rich fluids active in the postore stage were rarely recorded as FIPs in the pre-ore and synore minerals, indicating that microfractures were not readily generated in previously precipitated vein minerals cut off from the fluids in the main conduits. High fluid pressure therefore appears to have played a critical role in the formation of microfractures as well as fractures.

(3) Field structural analysis indicates that the maximum principal stress ($\sigma_1$) was subhorizontal and in the NE-SW direction, whereas $\sigma_2$ and $\sigma_3$ alternated between subhorizontal and subvertical orientations. FIP analysis, on the other hand, indicates that $\sigma_3$ was mostly subhorizontal during microfracturing. It is postulated that the NE-SW compression regime was punctuated by temporary and episodic change of $\sigma_3$ to a subvertical direction, which may have been caused by the upward force associated with the fluid source region and (or) local detachment of the veins from the surrounding rocks.

(4) Foliation-parallel fracturing and veining may have taken place either during the NE-SW compression or when $\sigma_1$ switched to a subvertical direction. The former situation is limited to special conditions when the differential stress ($\sigma_1-\sigma_3$) was smaller than the difference between foliation-parallel and foliation-perpendicular tensile strengths ($|T_-T_\perp|$), and the effective maximum stress was negative (because of high fluid pressure) and equal to the foliation-parallel tensile strength ($\sigma_1* = T_\perp$). The second situation is particularly favorable for the formation of foliation-parallel veins as the origin of the change of maximum principal stress may be linked to the fluid source.

ACKNOWLEDGEMENTS

This study was supported by an NSERC-Discovery grant (to GC). We would like to thank Andreas Lichtblau and Carmen Storey of the Ontario Geological Survey, Kenneth Williamson and Paul Barc of Goldcorp, Farhad Bouzari of the University of British Columbia, and Mary Louise Hill of Lakehead University for their help in fieldwork and for ensuing discussion. Critical reviews by Drs. Yanhua Zhang, Gema Olivo, and Ole Kaven significantly improved the quality of the paper, for which we are grateful.

REFERENCES


Cox SF (2005) Coupling between deformation, fluid pressure, and fluid flow in ore-producing hydrothermal systems at depth in the crust. Economic Geology 100th Anniversary Volume, 39–75.


**SUPPORTING INFORMATION**

Additional Supporting Information may be found in the online version of this article:

**Appendix S1.** Information of individual samples: location, lithology, relationship to structures, and fluid inclusion studies*.

Please note: Wiley-Blackwell are not responsible for the content or functionality of any supporting materials supplied by the authors. Any queries (other than missing materials) should be directed to the corresponding author for this article.