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**Gold Mineralization at the Campbell Mine, Red Lake Greenstone Belt, Uchi
Subprovince, Ontario**

By

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of the requirements for the degree of Doctor of Philosophy
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Abstract

The Campbell mine is located in the eastern part of the Red Lake greenstone belt, Uchi Subprovince. The center of the belt is comprised predominantly of ≈ 2.99 Ga tholeiitic basalt and komatiites with minor felsic volcanic rocks and sedimentary rocks. The margins of the greenstone belt are comprised of ≈ 2.74 Ga. tholeiitic to calc-alkaline basalts and rhyolites. Voluminous ≈ 2.73 to 2.7 Ga granitoid batholiths border the belt. Metamorphic grade increases from greenschist facies at the center, to amphibolite facies at the margins of the belt. Spatial variations of mineral assemblages indicate that metamorphism resulted from the emplacement of marginal granitoid batholiths. The Campbell mine and adjacent Red Lake mine occur approximately at the greenschist / amphibolites facies metamorphic transition, within the ≈ 2.99 Ga volcanic assemblage. Mineral assemblages, garnet-biotite thermometry and GASP barometry suggest that peak metamorphism at the Campbell mine occurred under conditions of 465°C to 545°C and ≤ 3 kbars. This low P / high T metamorphic event may have postdated greenschist facies metamorphism (M_1), and is termed M_3 .

The eastern part of the Red Lake greenstone belt has experienced a minimum of three deformation events. The earliest event (D_1) involved the development of NE-trending F_1 folds and deformation zones, during NW-SE shortening. D_1 compression occurred prior to 2720 Ma and may correspond to the initial stage of the Kenoran orogeny (Stott, 1997). This deformation may have occurred during emplacement of the Little Vermillion Lake batholith at 2731 ± 3 Ma, and may have been coeval with the earliest metamorphic event (M_1). Chlorite and rare actinolite were developed along S_1 during M_1 . Emplacement of the most volumetrically significant granitoid intrusive units within the belt followed at 2720 to 2718 Ma and they are accompanied by narrow zones of strain at their margins, and narrow M_2 amphibolite facies metamorphic aureoles. D_2 NE-SW shortening and NW trending F_2 folds were contemporaneous with a late marginal phase of the Trout Lake batholith at ≈ 2700 Ma, and a broad zone (≥ 2 km) of M_3 contact metamorphism, in the eastern part of the belt. The coeval nature of M_3 and F_2 is supported by the relationship between M_3 garnet and amphibole extension lineation and F_2 folds. The D_2 extension lineations lie within S_2 foliations, which are axial planar to F_2 folds. These extension lineations form clusters on stereonet across F_2 axial planes, indicating they are not folded by the F_2 event. Northeast F_1 trending folds were re-folded by NW trending D_2 folds. The

Campbell/Goldcorp Red Lake mines lie within the southwest limb of the regional F_2 Balmer synform.

Auriferous veins are parallel to S_2 , and are termed foliation parallel auriferous veins (NW striking, SW dipping). Auriferous veins also occur in shear zones which are oblique to the regional NW trending foliation (foliation-oblique deformation zones). Foliation parallel veins were developed in mafic rock units within zones dominated by flattening strain along attenuated F_2 fold limbs. Foliation-oblique auriferous veins occur at folded mafic / ultramafic rock contacts, within shear zones developed by F_2 flexural slip folding. These auriferous veins occur proximal to the F_2 fold hinges. The deformation zones are typically meters to 10's of meters wide, and are characterized by more intense foliation development, and abundant carbonate veins and veinlets. Progressive D_2 shortening resulted in a late D_2 decollement along the southwestern limb of the Balmer synform, cutting and offsetting auriferous mineralization. The highest Au grade ore bodies at the Campbell mine occur within foliation-oblique auriferous veins. A post-2700 Ma deformation event (D_3) re-activated the East Bay deformation zone as an oblique-slip shear system. Shear zones hosting mineralization are occur within large areas of relatively unstrained rock, and are not restricted to a NW trending orientation, and therefore they do not conform to the description of Andrews et al. (1986) for the D_2 Cochenour – Gullrock Lake deformation zone.

Several stages of hydrothermal activity are recognized in the Campbell mine area. The earliest event, aluminous alteration, produced patchy bleaching containing white mica with minor pyrite in mafic rocks. It is probably synvolcanic and does not display a spatial relation to the ore zones. Later M_3 metamorphism of this alteration produced an assemblage of andalusite + chloritoid + white mica + garnet in amphibolite facies domains. Finally, early M_3 carbonate alteration forms broad pervasive disseminated zones, with carbonate veinlets spatially related to extensive, quartz + carbonate veins, and hosted by D_2 shear zones. Subsequent Au mineralization occurs in quartz veins and quartz + arsenopyrite hydrothermal breccias within quartz + carbonate veins; thus most of the Au mineralization overprints pre-existing quartz + carbonates within the D_2 shear zones.

Wallrock alteration of mafic rocks adjacent to auriferous veins at the Campbell mine is characterized by chlorite + white mica + carbonate in greenschist facies metamorphic domains, and biotite + amphibole + garnet + anorthitic plagioclase in amphibolite facies domains. Pyrite,

pyrrhotite, arsenopyrite, and carbonate occur in all domains, with pyrite and carbonate decreasing in modal abundance with increasing metamorphic grade. AFM diagrams show a progressive change of bulk composition during alteration under peak metamorphic conditions.

Gahnite (Zn spinel) occurs in high Au grade ore from amphibolite facies domains, and textural relations indicate co-deposition with auriferous arsenopyrite. Austrobitite (AuSb_2) has only been observed in greenschist facies domain, compatible with its relatively low thermal stability. Native gold from amphibolite facies domains contains significant Ag, precluding that it formed by the thermal decomposition of austrobitite. Therefore, the spatial variations in occurrence of Au phases reflect the deposition of Au during the peak metamorphism.

Metamorphic mineral assemblages and garnet-biotite thermometry show a steep temperature gradient associated with subvertical isotherms in the mine area. Higher metamorphic grade rocks are extended westward along the auriferous shear zones, suggesting heat transfer along shear zones by hydrothermal fluids. This suggests a magmatic heat source to the east for auriferous hydrothermal activity. Hence, Au mineralization likely took place during the emplacement of the marginal phase of the Trout Lake batholith.

Primary and pseudosecondary Type I, $\text{CO}_2 + \text{CH}_4$ inclusions predominate in the hydrothermal breccias, and are interpreted as a product of phase separation of the auriferous hydrothermal fluids during brecciation and Au deposition. Type II $\text{H}_2\text{O} - \text{CO}_2$ inclusions are characterized by low salinity (1.4 to 7 wt.% NaCl equivalent) and are interpreted to represent the auriferous fluids. Type I inclusions from low Au grade veinlets contain lower CH_4 contents than from high grade ore. The lack of Type II inclusions in high Au grade breccias is interpreted as a result of the selective loss of H_2O following phase separation.

$\delta^{13}\text{C}$ values for carbonates from the eastern Red Lake belt range from +0.06 to 4.24 ‰, which are high compared to those from other Archean Au deposits. Carbonate associated with Au is enriched in $\delta^{13}\text{C}$ relative to pre- and post-Au carbonates. The enrichment of $\delta^{13}\text{C}$ is likely due to discharge of CH_4 during phase separation of auriferous fluids. Phase separation is probably the most important mechanism for the mineralization of high Au grades (>500 g/t Au) at the mine.

$^{87}\text{Sr}/^{86}\text{Sr}$ values for scheelite from auriferous quartz + carbonate veins range from 0.7012 to 0.7017. Basalt and ultramafic rocks of ≈ 2.99 Ga would have had high $^{87}\text{Sr}/^{86}\text{Sr}$ values (> 0.7020) at 2.7 Ga and felsic rocks and sedimentary rocks would have even higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios

at 2.7 Ga. Low $^{87}\text{Sr}/^{86}\text{Sr}$ values for scheelite suggest that the auriferous hydrothermal fluids did not interact with the 3.0 Ga. supracrustal rocks. This indicates the fluids were not derived from the supracrustal rocks through metamorphism, or meteoritic waters. Instead, the low $^{87}\text{Sr}/^{86}\text{Sr}$ values suggest that the fluids were derived from contemporaneous igneous rocks or magmas.

The gold-related wallrock alteration at the Campbell Mine is similar to that in other Archean Au deposits hosted by lower amphibolite facies mafic rocks, such as the Coolgardie goldfields (Knight et al., 2000) and the Norseman district deposits (Mueller, 1992). The P-T conditions for the auriferous hydrothermal activity at many Archean lode-type deposits define a low P and high T path. The path indicates much higher temperatures than the Archean geotherm, suggesting either a contribution of heat from contemporaneous magmas to auriferous fluids, or rapid transfer of heat from deep levels. For syn-metamorphic Au deposits, the data support a magmatic heat source. This is in accord with the low P/high T conditions of the metamorphism accompanying the Au deposits. For post-peak metamorphic deposits, the data suggest transport of heat from a source external to the greenstone sequence.

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1.0 Introduction

1.1 Discovery history of the Campbell/Goldcorp Red Lake deposit

The ground hosting the Campbell mine was staked in 1944 by C.W. Campbell and A.K. McLeod, after visible gold was discovered in exploration trenches. Brewis and White Ltd. funded further exploration, and the Campbell mine portion of the property was eventually acquired by Dome Exploration Ltd. in 1945. A mine was constructed and began production in the spring of 1949. The mine is currently owned by Placer Dome Canada Ltd.

The adjacent Goldcorp Red Lake mine (previously known as A.W. White Mine, and Dickenson Mine, and hereafter referred to as the Goldcorp Mine) was put into production in 1948 by Brewis and White. The Goldcorp mine exploits the eastern portion of the Campbell/Goldcorp Au system.

1.2 Objectives of the project

The Campbell gold mine is located in the Red Lake greenstone belt of the Uchi Subprovince (Superior Province), approximately 160 km NNE of the town of Kenora, at Balmertown, Ontario. The Red Lake gold camp is one of the most significant producers in the Superior Province, having yielded in excess of 543 tonnes of Au. The two most important deposits in the district are the Campbell and adjacent Goldcorp Red Lake mine, responsible for approximately 57 % of the production from the belt. Despite the economic significance of the camp, the structural development of the belt, significance of the regional deformation zones, and timing, controls, and origin of gold mineralization at the Campbell and Goldcorp mines are poorly understood.

The purpose of this study was to evaluate the timing and controls on gold mineralization at the Campbell mine. The relations between Au grade, kinematics of the shear zones hosting gold mineralization, and the timing of deformation, metamorphism, and alteration events were investigated from mine grade plans, level maps and sections, and underground and surface outcrop mapping and sampling. Field work was conducted over the summers of 1993 to 1996, while the writer working as an underground mine geologist. Field work and field visits outside the Campbell mine were conducted in 1993,

1994, 1997, 1998, and 2000, while working as an underground mine geologist for Campbell mine, and later a Research/Exploration geologist, for Placer Dome Exploration.

Ore zones represent the rocks that can be extracted at a profit. Gold grade varies significantly in a vein. As a result, low grade or barren rocks, together with high-grade rocks are mined. At the Campbell mine, high gold grade ore occurs in low gold grade to barren vein structures. Therefore, the term “ore zone” here refers to quartz + carbonate vein structures that are exploited for mining, and may include auriferous and non-auriferous veins, excluding apparently barren late veins.

All rocks have been metamorphosed, so the prefix “meta” is omitted here.

1.3 Previous studies

Dowling (1894) first mapped the geology of the Red Lake belt at a reconnaissance level for the Geological Survey of Canada. More detailed evaluations by Bruce (1924), and Bruce and Hawley (1927) on behalf of the Ontario Department of Mines followed this work. Mineral occurrences in the belt were documented by Hurst (1935), and a later detailed report of mineral occurrences was produced by Horwood (1940), after the initial boom in gold exploration and mine development in the area.

Several phases of Township-scale mapping have been completed in the belt, primarily focussing on the eastern and southern areas (i.e., Chisolm, 1951; Ferguson, 1966; Pirie and Grant, 1978a, Pirie and Grant 1978b; Atkinson, 1993). The eastern part of the Red Lake belt has been the focus of numerous structural (Durocher and Hugon, 1983; Berger and Summers, 1983; Berger and Summers, 1984; Berger and Helmstaedt, 1984; Wilson et al., 1984; Hugon, 1984; Zhang et al., 1997), and stratigraphic and geochronological studies (Pirie, 1981; MacGeehan and Hodgson, 1981; Pirie, 1982; MacGeehan and Hodgson, 1982; Corfu and Wallace, 1986; Corfu and Andrews, 1987; Stott and Corfu, 1991, Corfu et al., 1998; Sanborn-Barrie et al., 2000). All of these workers have contributed to the understanding of alteration and mineralization in the belt. The tectonic setting of volcanic rocks in the greenstone belt has been investigated by the workers mentioned above, and more specifically by Skulski et al. (2000), Hollings et al. (1999), Tomlinson et al. (1998), Tomlinson and Rogers (1999), and Rodger et al. (2000). These workers have also shown that the oldest ultramafic rocks in the Balmer Assemblage were likely derived from partial melting of a Fe-rich mantle, possibly related

to plume activity. By 2.94 to 2.925 Ga, ultramafic volcanic rocks may have been derived from subduction related, refractory, Fe-poor mantle. The Confederation Assemblage volcanic rocks were generated from a younger episode of arc-magmatism, evolving from tholeiitic to calc-alkaline volcanism, reflecting increasing arc maturity.

Geology, alteration and mineralization at the Campbell mine was documented by Rigg (1980), MacGeehan and Hodgson (1982), Christie (1986), Penczak and Mason (1997), Tarnocai and Hattori (1996), Zhang et al. (1997), Tarnocai et al. (1998), and Penczak and Mason (1999). The adjacent Goldcorp Red Lake mine has been described by Rigg (1980), Rigg and Helmstaedt (1981), Mathieson (1982), Mathieson and Hodgson (1984), and Rodgers (1992). Recently, the Madsen deposit in the south-central part of the belt has been re-examined by Dube et al. (2000).

Prior to the mid' 1980's, the gold deposits in the eastern part of the Red Lake belt were regarded as products of syngenetic processes (i.e., Crocket et al., 1981; Kerrich et al., 1981). Rigg and Helmstaedt (1981), and MacGeehan and Hodgson (1982) recognized epigenetic aspects to gold mineralization at the Campbell and Goldcorp mines, and later studies (i.e., Andrews et al., 1986; Corfu and Wallace, 1986; Corfu and Andrews, 1987) showed that gold mineralization significantly post-dated volcanism. Andrews et al. (1986) suggested that Au mineralization at the Campbell and Goldcorp mines was related to belt-scale conjugate simple shear zones, broadly contemporaneous with late phase intrusion of the marginal batholiths and peak metamorphism. In contrast, Penczak and Mason (1997), and Penczak and Mason (1999) proposed that the Campbell/Goldcorp Red Lake deposit is a low sulphidation epithermal vein system, that was subsequently overprinted by later metamorphism and brittle-ductile deformation.

2.0 Regional Geology

2.1 Overview of the Geology of the Red Lake Greenstone belt

The Red Lake greenstone belt is located in the Uchi Subprovince in northwestern Ontario (Fig. 2.0). In this paper, the author divides the supracrustal rocks of the greenstone belt into two sequences (Fig. 2.0). The Lower (MesoArchean) Sequence (2990 ± 3 to 2890 ± 2 Ma; Corfu and Andrews, 1987) contains the Balmer, Ball, Bruce Channel and Slate Bay Assemblages. This is equivalent to the Lower sequence of Corfu and Wallace (1986) Corfu and Andrews (1987), and Corfu et al., (1998). The Balmer Assemblage (2.99 to 2.96 Ga) is composed of tholeiitic basalt and basaltic komatiites, with minor felsic flows and interflow iron formation in the center of the belt. The ultramafic rocks are both flows, and sill-like intrusions that locally cut lithologic boundaries. The Ball Assemblage (2.94 to 2.92 Ga) occurs in the western part of the greenstone belt. The Bruce Channel and Slate Bay Assemblages are unconformable with the Balmer Assemblage. The Bruce Channel Assemblage comprises volcanic fragmental units, fine to coarse-grained clastic sedimentary rocks, and iron formation. The Slate Bay Assemblage comprises clastic sedimentary rocks in the northwest part of the belt. The Upper Sequence is NeoArchean in age, and comprises the Confederation assemblage. It is composed of two phases; an early tholeiitic phase (2760 to 2740 Ma; Corfu and Andrews, 1987) with intercalated sedimentary rocks, and a later calc-alkaline phase (2740 to 2730 Ma; Corfu and Andrews, 1987). The Upper Sequence occurs along the north and south margins of the greenstone belt and at McKenzie Island in the centre of the belt, and the calc-alkaline phases at the contacts with the external batholiths (Fig. 2.0). The boundaries between Upper and Lower sequences may be marked by an unconformity and a thin conglomerate to fine-grained sedimentary unit (Huston Assemblage).

The supracrustal rocks of the Red Lake belt are intruded by subvolcanic sills, granodiorite to diorite stocks and dykes of 2745 ± 15 to 2699 ± 4 Ma age (Corfu and Andrews, 1987). Younger lamprophyre and diabase dykes also intrude the belt. The most volumetrically significant internal intrusive units are the Dome (2718 ± 1 Ma; Corfu and Andrews, 1987) and McKenzie (2720 ± 2 Ma; Corfu and Andrews, 1987) diorite to granodiorite stocks, located in the center of the belt (Fig. 2.0). Composite external

batholiths bound the belt, with latest intrusive phases of approximately 2731 to 2699 Ma (Corfu and Andrews, 1987). Metamorphism, and significant hydrothermal activity may have accompanied emplacement of the latest phases of the marginal batholiths (Andrews et al., 1986; Tarnocai and Hattori, 1996; Tarnocai et al., 1997). Pirie (1978, 1982), MacGeehan and Hodgson (1981), Andrews et al. (1986), and Sanborn-Barrie et al. (2000) have previously described the detailed regional geology and some aspects of the deformation. The timing of volcanism and the late intrusive events are provided on Fig. 2.5.

2.2 Regional metamorphism

Metamorphism in the belt is of low pressure/high temperature type and is associated with late granitoid (≈ 2.73 to 2.7 Ga) intrusions at the margins of the belt (Hugon and Schwerdtner, 1985; Andrews et al., 1986; Menard et al., 1999). The preliminary study by Menard et al. (1999) identified three metamorphic events; each is associated with the intrusion of external granitic rocks and accompanied by deformation. Corfu and Stone (1998a) showed that the belt did not experience a significant thermal event after ≈ 2690 Ma, supporting earlier Ar-Ar studies (McMaster, 1987; Wright et al., 1991). In contrast, Penczak and Mason (1997), and Penczak and Mason (1999) suggested a young age (≈ 2650 Ma) for peak metamorphism that is unrelated to any known intrusive events in the belt.

The metamorphic grade in the belt varies from upper amphibolite facies at the margins of the belt to lower greenschist facies in the center of the belt. Amphibolite facies metamorphism may extend 1 to 3 km's from the margins of the external batholiths into the Red Lake belt (Andrews et al., 1986). The distribution of high metamorphic grade rocks along the margins of the belt is similar to many other greenstone belts in the Superior Province (i.e., Stott and Corfu, 1991; Thurston et al., 1992). This pattern suggests that the intrusion of external batholiths had an important role in the thermal history of the greenstone belt. Internal stocks and dykes have narrow metamorphic aureoles, typically extending less than 10's of meters from the intrusive contacts. Greenschist facies metamorphism was overprinted by contact metamorphism related to the emplacement of the Dome stock, suggesting a widespread metamorphic event prior to

intrusion of the Dome stock at 2718 ± 1 Ma (Corfu and Andrews, 1987). This earlier metamorphic event is termed M_1 and metamorphism related to the emplacement of the Dome and McKenzie stocks is M_2 (Table 2.0). The Trout Lake batholith (2699 ± 1 Ma for the late external phase, Corfu and Andrews, 1987) displays a significant contact metamorphic aureole (Andrews et al., 1986) and this metamorphism is termed M_3 (Table 2.0). Andrews et al. (1986) have shown that the Campbell mine lies approximately at the greenschist – amphibolite grade metamorphic transition of the contact metamorphism (M_3). The study by Mathieson and Hodgson (1984) indicated peak metamorphic temperatures of 520°C to 540°C at 3.8 to 4.2 kbar based on mineral assemblages at the Goldcorp mine, immediately east of the Campbell mine (Fig. 2.0).

2.3 Regional deformation and deformation zones

Deformation zones in the belt have been previously described as NW and NE trending high-strain corridors (Fig. 2.0), which cut the supracrustal rocks and some of the internal intrusive suite (Hugon and Schwerdtner, 1985; Andrews et al., 1986). High strain zones observed in this study are characterized by 2 to 10 m wide discrete zones, separated by weak to moderately deformed rocks (Wilson et al., 1984), rather than the broad (100's meters wide) high strain zones presented by Andrews et al. (1986). The NW-trending deformation zones are the Pipestone-St. Pauls Bay, Chukuni River, and Cochenour–Gullrock Lake deformation zones (Fig. 2.0). The NE-trending deformation zones are the East Bay, Post Narrows, and Flat Lake-Howey Bay deformation zones (Fig. 2.0). Regional NW and NE trending deformation zones in the belt (Fig. 2.0) have been interpreted as conjugate shear zones developed during N-S directed, Superior Province wide Kenorean-age compression (Hugon and Schwerdtner, 1984). This is similar to the interpretation proposed for many shear zones in the Superior Province (e.g. Colvine et al., 1988).

Recent structural work conducted by Zhang et al. (1997) has shown that NW trending shear zones are dominated by strong flattening fabrics and do not display rotations of planar fabrics into the shear zones. Zhang et al. (1997) have suggested that strain along the NW trending Cochenour-Gullrock Lake deformation zone is predominately a result of pure shear, and is inconsistent with the conjugate shear model.

Pure shear deformation also dominated in other deformation zones in the belt (Zhang et al., 1997). Mapping by the author shows that weak, moderately plunging extension lineations occur within the foliation in the Campbell mine area. Extension lineations are defined by preferred long axis orientation of amphibole and garnet trails, chlorite and white mica. Along the Flat Lake – Howey Bay and East Bay deformation zones, extension lineations have shallow plunges within the foliation (Tarnocai, 1996a; Tarnocai, 1998). Extension lineations along NW shear zones in the Pipestone – St. Pauls Bay deformation zone also have shallow to moderate plunges (Tarnocai, 1998). These observations by the author (Tarnocai, 1998) do not support the previously proposed conjugate shear model.

Shear zones within the belt occur as foliation parallel zones (Zhang et al., 1997), and also as lithologic contact-controlled shear zones in various orientations (Tarnocai, 1996a; Tarnocai, 1996b; Tarnocai, 1998). These observations by the author are confirmed by Sanborn-Barrie et al. (2000) that shear zones within the “high strain domains” outlined by Andrews et al. (1986) are not necessarily parallel to the boundaries of the high strain zones. The author’s work, and that of Sanborn-Barrie et al. (2000), also supports the notion that shear zones occur outside the high strain domains of Andrews et al. (1986).

The eastern part of the belt possesses three orientations of foliations (Rigg, 1980; Berger and Summers, 1983; Christie, 1986). A weakly developed 90° to 110° trending cleavage predominates throughout the eastern part of the belt. The NW and NE trending deformation zones of Andrews et al., (1986) also contain foliations parallel to their boundaries, and are interpreted as synchronous with, or post-dating the E-W cleavage (Berger and Summers, 1983). Hugon and Schwerdtner (1985) showed foliation trajectories that are broadly parallel to the margins of the belt, and undergo rotation into some regional deformation zones outlined by Andrews et al. (1986). This supports the interpretation that strain along the regional shear zones occurred during or after the development of regional foliations. Hugon and Schwerdtner (1985) also indicated that strain inferred from foliation intensity increases with metamorphic grade towards the margins of the belt. The occurrence of regional foliation trends broadly parallel to the contacts of the external batholiths (Hugon and Schwerdtner, 1985) does not support other

Table 2.0. Timing of deformation and metamorphism in the eastern Red Lake belt

Deformation and folding	Metamorphism	Age*
D₁/F₁ NE trending folds across the eastern part of the belt. NE trending deformation zones.	M₁ Greenschist facies. May be unrelated to D ₁ in timing.	Pre- 2720±2 Ma
	M₂ Narrow contact metamorphic aureoles adjacent to the Dome and McKenzie stocks.	2718±1 to 2720±2 Ma
D₂/F₂ NW trending, moderate SW plunging folds, developed proximal to the Trout Lake batholith. Variable, but generally NW to WNW trending, SW dipping foliations. NW deformation zones on attenuated F ₂ limbs.	M₃ Wide (km-scale) amphibolite facies contact metamorphic aureole, associated with late felsic phases of the Trout Lake batholith.	2699±1 Ma
D₃/F₃ Sinistral strike-slip shear along the East Bay and Post Narrows D.Z.'s. Shear zone hosted asymmetric folds with steep dipping axial planes		Syn- to post-2699±1 Ma The East Bay D.Z. cuts the Dome stock, and F ₂ fold axes.

*= Age data by U-Pb zircon, from Corfu and Wallace (1986), and Corfu and Andrews (1987)
D.Z.=deformation zone

work that shows regional foliations that are consistently E-W to WNW trending throughout the supracrustal rocks, outside the deformation zones. The orientations of F_1 and F_2 folds vary throughout the belt, and are broadly parallel to its margins (Tarnocai, 1998; Sanborn-Barrie et al. 2000). This suggests that folding and related foliations are not solely a product of regional shortening. It may be a product of, or modified by, local strain variations imposed by the marginal batholiths during or after its emplacement.

2.4 Gold deposits in the Red Lake belt

Numerous past producers and prospects, which yielded in excess of 543 tonnes of Au, occur throughout the belt, principally hosted by Lower Sequence rocks. Some deposits occur within the Dome and McKenzie stocks. They are spatially associated with the regional deformation zones (Hugon and Schwerdtner, 1985; Andrews et al., 1986). The Goldcorp Red Lake mine is immediately east of the Campbell mine and exploits the same ore body. As of the end of 1998, the Campbell mine had produced 9,655,157 ounces Au (300,308 kg Au) at an average grade of approximately 0.58 ounces/ton (19.89 g/t). Current reserves are 1,889,000 ounces (58,754 kg Au) at 0.46 opt. Au (15.77 g/t) (Placer Dome, 1999 Annual Report). The deepest production level at present (summer, 2000) is $\approx$ 4500 feet below surface ($\approx$ 1372 m below surface, 30th level), with shaft access to the 37th level. The Goldcorp mine has produced 3,148,203 ounces Au (97,920 kg Au) at an average grade of 0.379 ounces per ton (12.99 g/t) (Blackburn et al., 1998). Reserves are currently 2,490,000 ounces Au (77,447 kg Au), at an average grade of 0.88 ounces per ton Au (30.17 g/t) (Goldcorp Annual Report, 1998). The Madsen mine, currently owned by Claude Resources, is in the southern part of the belt. The Madsen mine re-opened in 1996, and closed in early 1998. It is presently being dewatered to access the deeper levels for drilling and development. The Madsen mine has produced 2,424,959 ounces of Au (75,424 kg Au), and has proven and probable resources of a further 94,639 ounces of Au (2,944 kg Au) (Claude Resources, 1999).

Auriferous mineralization is hosted in quartz veins in shear zones in all deposits. In addition, auriferous mineralization is hosted by disseminated sulphide bodies at the Campbell, Goldcorp, Cochenour, and Madsen mines, and brecciated quartz + carbonate veins, within matrix quartz at the Campbell, Goldcorp, Cochenour, and Wilmar mines.

Disseminated sulphide mineralization is the principal ore at the Madsen mine, whereas Au is hosted primarily by quartz+carbonate veins at other deposits.

2.5 Geology of Balmer and Dome Townships

Approximately 70% to 80% of the eastern part of the belt is covered by swamp, glacial overburden, or lakes, making mapping and correlation of rock units difficult. Outcrop checks were made by the author in the summer of 1993 and 1994, 1998, 2000, independently and with B. Atkinson and Carmen Storey (OGS Resident District Geologists) and Goldcorp mine geologists. Additional interpretations were made from drill hole data at the Campbell mine and Resident Geologist office, mine geology plans, and regional airborne magnetic survey maps (Ontario Geological Survey, 1978; Placer Dome Canada Ltd. Exploration, 1994). Field data were collected and the locations were recorded using the GPS. Government and mine geology maps, and field data were then compiled in GIS. A geological map of the mine area covering parts of Balmer and Dome Townships was produced by the author from these data. The map (Fig. 2.1) differs significantly from previous geological maps by Ferguson (1962), Pirie and Grant (1978a, 1978b), MacGeehan and Hodgson (1982), and Atkinson and Stone (1993). The locations of the regional deformation zones (Hugon and Schwerdtner, 1985; Andrews et al., 1986) are provided on figure 2.2, together with foliation data.

A voluminous felsic intrusion is shown on geology maps NE of Balmer Lake, and NW of McDougal Lake (Pirie and Grant, 1978a; Pirie and Grant, 1978b; MacGeehan and Hodgson, 1982). As the unit does not cut iron formations on aeromagnetic maps, the author interpreted it as a felsic volcanic unit (felsic volcanic rock unit in the NE portion of the Fig. 2.1). This is confirmed by the author's examination of the unit in the field and thin section examination. The unit shows bedding and crude crystal grading and is termed as a felsic to intermediate crystal-lithic tuff.

Township geology maps (Ferguson, 1962; Pirie and Grant, 1978a; Pirie and Grant, 1978b; Atkinson and Stone, 1993) and mapping by MacGeehan and Hodgson (1982) show regional-scale (>2 km wavelength) gentle to tight NW trending folds in eastern Balmer township. Stratigraphy in western Balmer Township is ENE to E-W (Fig. 2.1) trending, and terminates against NE trending ultramafic units along the East Bay

deformation zone (Fig. 2.2). North-east trending folds occur in the Cochenour area. The distribution of highly magnetic iron formation and ultramafic units allowed the author to re-examine the structure and stratigraphy of the eastern part of the belt, using aeromagnetic maps (Placer Dome Canada Ltd. Exploration, 1994; Ontario Geological Survey, 1978). The highly magnetic lithologies clearly outline NW trending folds, and refolded NE folds, in Balmer and Dome Township. NW trending fold axial traces are linear, and NE trending axial traces are folded about NW axial traces (Fig. 2.3). This indicates NE folding (F_1) pre-dates the NW folding, which is named F_2 . Re-folding has produced a Type 3 fold interference pattern (Ramsay, 1967) in the eastern part of the belt. Two stages of folding in the eastern part of the belt were suggested by Atkinson and Stone (1993), Zhang et al. (1997), and Menard et al. (1999), and are confirmed from outcrop examination (Fig. 2.4). F_1 folds occur throughout the eastern part of the belt, and F_2 folds appear to be restricted to western Balmer Township, being poorly represented in Dome Township and eastern Balmer Township on magnetic maps. Apparent strain related to F_2 folding increases eastward, coincident with metamorphic grade, compatible with observations by Hugon and Schwerdtner (1985). This suggests this strain related to the F_2 folding event increases with proximity to the external batholiths. F_1 and F_2 folds in the eastern part of the belt are interpreted here to be related to D_1 and D_2 deformation events, respectively (Fig. 2.5, Table 2.0). The Bruce Channel assemblage lies south and east of the Campbell and Goldcorp mines, and along the limbs of the Balmer synform.

2.5.1 F_1 Folding and D_1 Faulting

Early NE trending folds in the eastern part of the Red Lake belt (east of the Campbell mine) have not been described by past workers. The Dome stock cuts F_1 folds in the Cochenour area (Atkinson and Stone, 1993), and the composite McKenzie stock cuts a NE trending fold on McKenzie Island (Fig. 2.3). The evidence suggests intrusion post-dated F_1 folding. NE trending deformation zones were developed in the East Bay and Post Narrows areas, along attenuated F_1 limbs.

2.5.2 F₂ Folding and D₂ Faulting

F₂ folds are identified on figure 2.3. On the geology maps by MacGeehan and Hodgson (1982), the most prominent F₂ fold in the eastern part of the Red Lake belt is located immediately east of the Campbell mine and has an axial trace running through Balmer Lake. This structure has been interpreted as a synform (Pirie and Grant, 1978a; Pirie and Grant, 1978b), and is termed the Balmer synform in this study (Fig. 2.3). Antiforms occur to the NE (termed the McDougal Lake Antiform in this study, figure 2.3) and in the Balmertown area. In contrast to the work by MacGeehan and Hodgson (1982), the author suggests that the southwestern limb of the Balmer synform is attenuated and sheared. Stratigraphic boundaries are clearly rotated, and lithological units are thinned into the attenuated limb. Bedding in the sedimentary rocks dips moderate to steeply S to SW, similar to the volcanic rocks (e.g. Pirie and Grant, 1978; MacGeehan and Hodgson, 1982). This suggests the contact between volcanic and sedimentary rocks is a fault, or a faulted unconformity, rather than simply an unconformity as indicated by Sanborn et al. (2000). The attenuated limb is interpreted here as a limb-parallel fault zone (Fig. 2.6). The fault is apparent along the sharp contact between folded volcanic units and sedimentary rocks (Fig. 2.6, 2.7). The attenuated limb corresponds to the “intensely deformed domain of MacGeehan and Hodgson (1981, their Fig. 2). The Goldcorp and H.G. Young mines (Fig. 2.1) occur along the sheared fold limb, and the Campbell mine occurs within folded rocks to the west. The author has interpreted a fault zone along the attenuated limb, merging into bedding and trending WNW in the H.G. Young mine area (Fig. 2.6).

North of the mine area, the extension lineations defined by amphibole grains and trails of garnet plunge consistently WSW, across the F₂ McDougal Lake Antiform (Fig. 2.8). The strike of the foliation planes are parallel to the axial trace of the F₂ folds (Fig. 2.8), suggesting they are axial planar to F₂ folds. Extension lineations lie on the foliation plane, and form a cluster (Fig. 2.8). This indicates that F₂ folding occurred before, or during regional metamorphism. The strike of S₂ changes progressively from WNW to NE across the Post Narrows deformation zone, suggesting this D₁ shear zone operated during D₂, detaching deformation in the west part of the study area.

At the margins of the Trout Lake Batholith, granitic dykes intruded along the foliation planes, and also cut the foliation. Dykes cutting the foliation vary from planar, to tightly folded (Fig. 2.9). Brecciated country rock pillow basalt in the matrix of felsic intrusive rocks occurs at the batholith margin. Country rock fragments are strongly foliated and have long axes parallel to the foliations in the supracrustal rocks (Fig. 2.10). These observations suggest that marginal phases of the Trout Lake batholith in this area were emplaced during F_2 deformation as stated previously by Hugon and Schwerdtner (1985), Andrews et al. (1986), and Zhang et al. (1997).

2.5.3 Post- F_2 deformation

Asymmetric folds in ultramafic rocks along the East Bay deformation zone north of Cochenour (Fig. 2.2) are obvious on aeromagnetic maps (Placer Dome Canada Ltd. Exploration, 1994). These folds are also observed in outcrop at the north end of East Bay. The axial planes of asymmetric folds are NE striking, steeply dipping, and extension lineations in axial planar foliations are subhorizontal. These observations suggest a component of sinistral shear along the East Bay deformation zone. The findings are compatible with the interpretation of Hugon and Schwerdtner (1985). Alternatively, these folds may be parasitic folds within relatively incompetent iron formation, on the eastern limb of a NE closing fold. However, zones of Z shaped folding have not been observed in the area, suggesting the former interpretation may be more appropriate.

F_2 folds are cut by highly strained, and asymmetrically folded ultramafic rocks and iron formation along the East Bay deformation zone (Fig. 2.3), suggesting sinistral shear along this zone post-dated F_2 folding. NW trending foliations, and E-W to WNW trending lithologic units, rotate counter-clockwise westward into the NE trending East Bay and Post Narrows deformation zone (Fig. 2.2). This also suggests that sinistral shear motion along these two deformation zones was syn- to post F_2 folding and NE-SW flattening (Fig. 2.2, 2.4).

2.6 Summary of regional geology

Volcanism and sedimentation in the Red Lake greenstone belt spanned from 2990 Ma to 2730 Ma. The emplacement of the final phases of voluminous, composite batholiths

at the margins of the belt occurred at $2730 \pm 2699 \pm 1$ Ma (Corfu and Andrews, 1987; Fig. 2.5). Predominantly mafic to ultramafic tholeiitic volcanic rocks (2989 ± 3 to 2894 ± 2 Ma) form the centre of the belt, with younger tholeiitic to calc-alkaline mafic to intermediate rocks, and sedimentary units at the north and south margins of the supracrustal sequence (Fig. 2.0). Volumetrically minor ultramafic to felsic intrusive rocks occur in the supracrustal sequence. The older tholeiitic sequence hosts most of the Au deposits and occurrences in the Red Lake belt. Some gold deposits are hosted by felsic stocks of 2720 ± 2 to 2718 ± 1 Ma age.

Supracrustal rocks in the Red Lake belt have been metamorphosed to greenschist and amphibolite grade. Metamorphic grade increases from the centre of the belt to the contacts with the external batholiths. This strongly suggests metamorphism was related to the emplacement of the external batholiths. Foliation planes throughout the belt are broadly parallel to the margins of the external plutons, and strain increases with metamorphic grade (Hugon and Schwerdtner, 1985), suggesting that deformation may have accompanied intrusion (Hugon and Schwerdtner, 1985; Andrews et al., 1986; Tarnocai et al., 1997). In the eastern part of the belt, the distribution of metamorphic facies is parallel to the contact with the Trout Lake batholith, on the eastern margin of the Red Lake belt. This suggests the marginal phase of the Trout Lake batholith was responsible for contact metamorphism in the eastern part of the belt, at 2699 ± 1 Ma (Corfu and Andrews, 1987).

Two episodes of folding are identified in the eastern Red Lake belt (Fig. 2.3, Table 2.0). F_1 folds with NE trending axial traces occurred prior to 2720 ± 2 Ma, and are present throughout the eastern part of the belt. F_1 is interpreted to have occurred during D_1 , and may be related to the emplacement of the Little Vermillion Lake batholith, or the initial stages of regional, Kenorean age NW shortening (Fig. 2.5). F_2 fold axial traces are NW trending, and on aeromagnetic maps are most evident proximal to the Trout Lake batholith. NW trending foliations are axial planar to F_2 folds, and field observations presented by the author support F_2 folding during metamorphism (Table 2.0). D_2 is interpreted here to have occurred with F_2 folding and emplacement of a late, marginal phase of the Trout Lake batholith, at ≈ 2700 Ma (Table 2.0, Fig. 2.5). This supports the proposed relative timing of metamorphism, deformation, and intrusion (Hugon and Schwerdtner, 1985; Andrews et al., 1986; Tarnocai et al., 1997). Corfu and Andrews (1987), and Menard et al. (1999) also concluded that the

metamorphism occurred during deformation, but suggested that the ≈ 2700 Ma late-intrusive age for the Trout Lake batholith (Noble, 1989) may not reflect the intrusive phase accompanying deformation and metamorphism. The data do not support the interpretation of a Superior Province-wide 2650 Ma timing for peak metamorphism (Penczak and Mason, 1997; Penczak and Mason, 1999) affecting the Red Lake belt, as shown by Corfu and Stone (1998).

The author's field work confirmed sinistral strike-slip shear along the East Bay and Post Narrows deformation zones, which was previously proposed by Hugon and Schwerdtner (1985). Motion along these zones occurred during, and possibly later than D_2 , as F_2 fold axial traces, and axial planar foliations are rotated into these zones. The East Bay deformation zone cuts the F_2 folds, and cuts and rotates F_2 related cleavages and foliations. Therefore, the East Bay deformation zone is here termed a D_3 shear zone (Table 2.0, Fig. 2.5), although it may have been active during D_1 and D_2 .

Figure. 2.0

Geology and location of the Red Lake greenstone belt, showing the regional deformation zones, and the location of the Campbell mine. The Upper sequence consists of the NeoArchean Confederation assemblage volcanic rocks. The Lower Sequence contains the MesoArchean Balmer, Ball, Slate Bay and Bruce Channel assemblages. Sedimentary rocks predominate the Slate Bay and Bruce Channel assemblages. Geology modified after Wallace et al. (1986), Andrews et al. (1986) and Corfu et al. (1998).

Figure 2.1

Geology of the eastern part of the Red Lake greenstone belt. The inset map shows the distribution of supracrustal assemblages in the study area, modified from Stott and Corfu (1991) and Corfu et al. (1998). Compilation is based on geological maps (Ferguson, 1962; Pirie and Grant, 1978a; Pirie and Grant, 1978b; Atkinson and Stone, 1993), aeromagnetic maps (Placer Dome Canada Ltd. Exploration, 1994; Ontario Geological Survey, 1978), drill logs at the Campbell mine and Assessment Reports at the Resident Geologist office, and the author's field examination in 1994 and 1995. A felsic intrusive unit shown on maps by previous workers (Ferguson, 1962; Pirie and Grant, 1978a; Pirie and Grant, 1978b) is interpreted here as a felsic volcanic rock unit (shown in the NE portion of the map area). Mine locations are shown as black open circles. Auriferous shear zones are shown as red lines. Boundary coordinates are UTM NAD27, Zone 15. Abbreviation, int=intermediate.

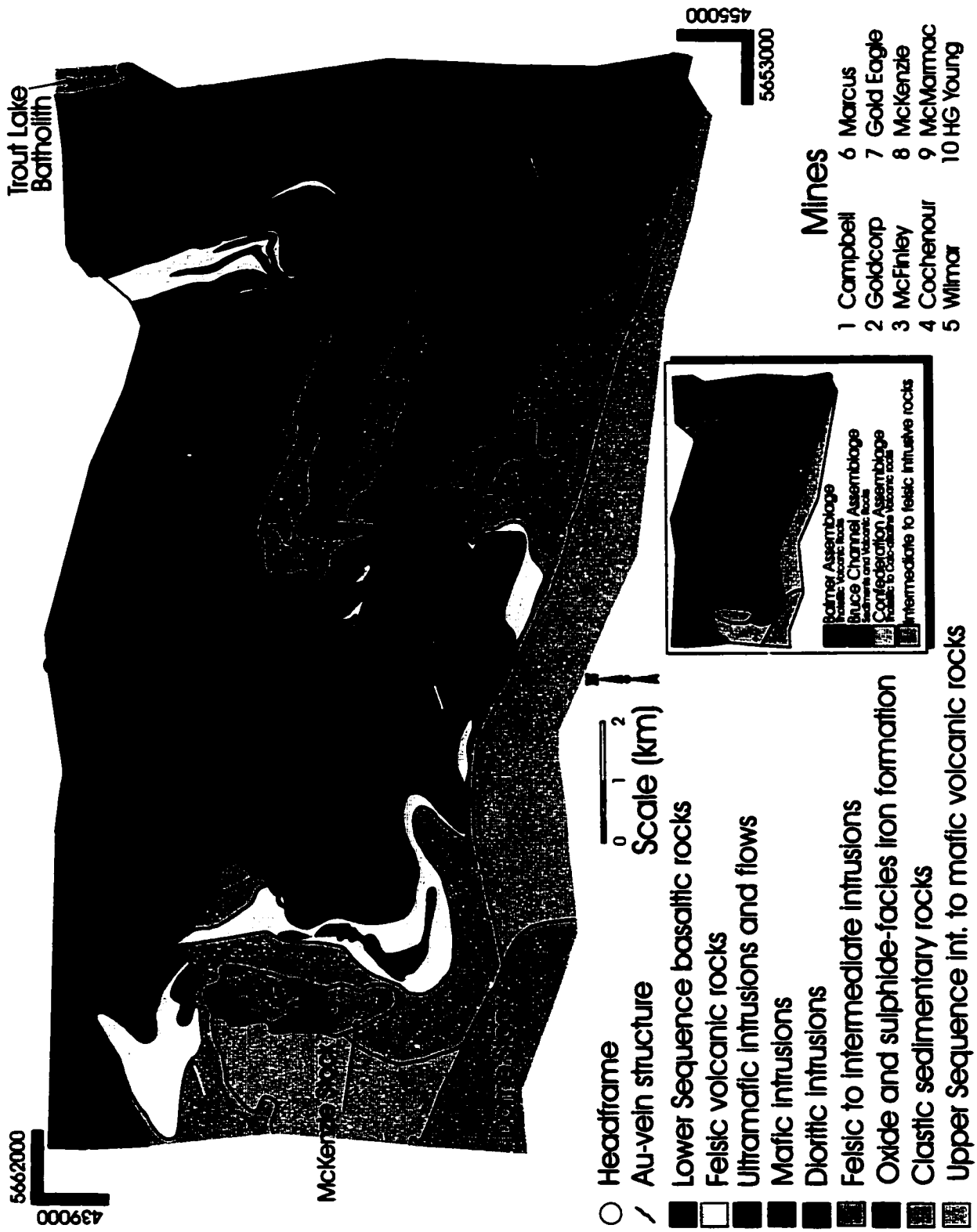


Figure 2.2

Geology of the eastern part of Red Lake greenstone belt, showing the deformation zones (grey shaded bands) after Hugon and Schwerdtner (1984), Andrews et al. (1986), and S₂ foliation planes from the author's field visits in the summers of 1993 and 1994, and from government township maps (Ferguson, 1962; Pirie and Grant, 1978a; Pirie and Grant, 1978b; Atkinson and Stone, 1993). Note that the S₂ foliations rotate into the Post Narrows deformation zone. Stratigraphic boundaries west of the Campbell mine show a similar rotation into the East Bay deformation zone. Foliation planes are WNW to NW trending throughout the central and western area, and subparallel to the contacts of the Trout Lake batholith in the east. Legend for lithologies and mine names provided in Fig. 2.1. Abbreviation, D.Z.=deformation zone.

5662000
439000

Post



Scale (km)
0 1 2

Foliation
Deformation zone

5663000
455000

Figure 2.3

Geology of the eastern Red Lake greenstone belt. NW trending and NE trending fold axial traces (solid black and dashed lines, respectively), are shown. NE fold axial traces are re-folded about NW axial traces in the eastern part of the area, forming a Type 3 refold pattern (Ramsay, 1967). This suggests that NE folding (F_1) pre-dated NW folding (F_2). The southwestern limb of the F_2 Balmer synform is strongly attenuated, and the axial trace terminates against the East Bay deformation zone (Fig. 2.2). The McDougal Lake Antiform occurs to the northeast of the Balmer synform. F_2 with NW striking axial planes appear to be well developed proximal to the margin of the Trout Lake batholith, and are not recognized east of the East Bay deformation zone. F_1 folding is identified throughout the area. Legend for lithologies and mine names provided in Fig. 2.1.

562000
439000



-- F₁ fold axial traces
— F₂ fold axial traces

0 1 2
Scale (km)

N

5653000
455000

Figure 2.4

Folded iron formation NE of Cochenour. A F_1 fold axis is re-folded by NW trending F_2 fold axes.



Figure 2.5

Timing of volcanism and intrusion, D₁ through D₃ deformation, and hydrothermal alteration in the eastern part of the belt.

Timing of volcanism and intrusion are interpreted from U-Pb geochronology for intrusive rocks in the Red Lake belt (Corfu and Wallace, 1986; Corfu and Andrews, 1987; Noble, 1989; Corfu et al., 1998). U-Pb error ranges for the intrusive rocks are indicated by vertical numbered rectangles, and the mean age is indicated by the fine black horizontal line in the rectangle. Post-gold intrusive rocks (pink) in mines are 1 Rowan dyke (Rowan mine, Pipestone – St. Paul Bay deformation zone), 2 Madsen diorite dyke (Madsen mine, Flat Lake – Howey Bay deformation zone). Dates for the external batholiths (purple) are; 3 Trout Lake batholith (Walsh Lake phase), 4 Killala-Baird batholith, 5 Hammel Lake batholith, 6 Little Vermillion Lake batholith. Pre-gold intrusive rocks (orange) are; 7 Dome and McKenzie stocks, 8 Red Crest diorite (Pipestone – St. Paul Bay deformation zone). Minimum ages for the, 9 Wilmar granodiorite (Cochénour – Gullrock Lake deformation zone), 10 Abino dyke (East Bay deformation zone) are indicated by orange arrows. The Goldcorp quartz feldspar porphyry dyke (11, Goldcorp Red Lake mine) is interpreted here to have an uncertain time relation to ore, as discussed in the text. A pre- to syn-ore timing for the dyke is suggested here, based on textural evidence and relations with auriferous shear zones.

The timing for the deformation events interpreted from this study are shown as heavy black lines, with uncertainty indicated by dashed lines. D₁ predates the Dome and McKenzie stocks, and may be equivalent in age to the Little Vermillion Lake batholith. D₂ occurred with the emplacement of the late phase of the Trout Lake batholith, and during or later than the emplacement of the Goldcorp porphyry dyke. D₃ is interpreted to overprint D₂ structures, and is therefore assigned an uncertain age, after the D₂ event. A summary of the deformation events relative to metamorphism are supplied in Table 2.0.

The timing for auriferous hydrothermal activity along the Cochénour – Gullrock Lake deformation zone is shown by the blue bar. The red star indicates the timing of hydrothermal alteration inferred from the secondary age from the altered and mineralized Wilmar granodiorite.

The inset shows the timing of volcanism (green ellipses) for Lower Sequence rocks in the eastern part of the belt (Balmer assemblages). Volcanic units intercalated within the Slate Bay and Bruce Channel assemblage were developed during the MesoArchean. The Upper sequence consists of Confederation assemblage rocks. Clastic sedimentary rocks were deposited between these volcanic periods (Sanborn et al., 2000). In the eastern part of the belt, auriferous hydrothermal activity occurred significantly later than the end of volcanism (blue bar A), and an earlier hydrothermal alteration event occurred during Balmer assemblage volcanism (blue bar B). Syn-volcanic hydrothermal activity during Confederation assemblage volcanism, and associated with base metal deposition, is not represented in the study area.

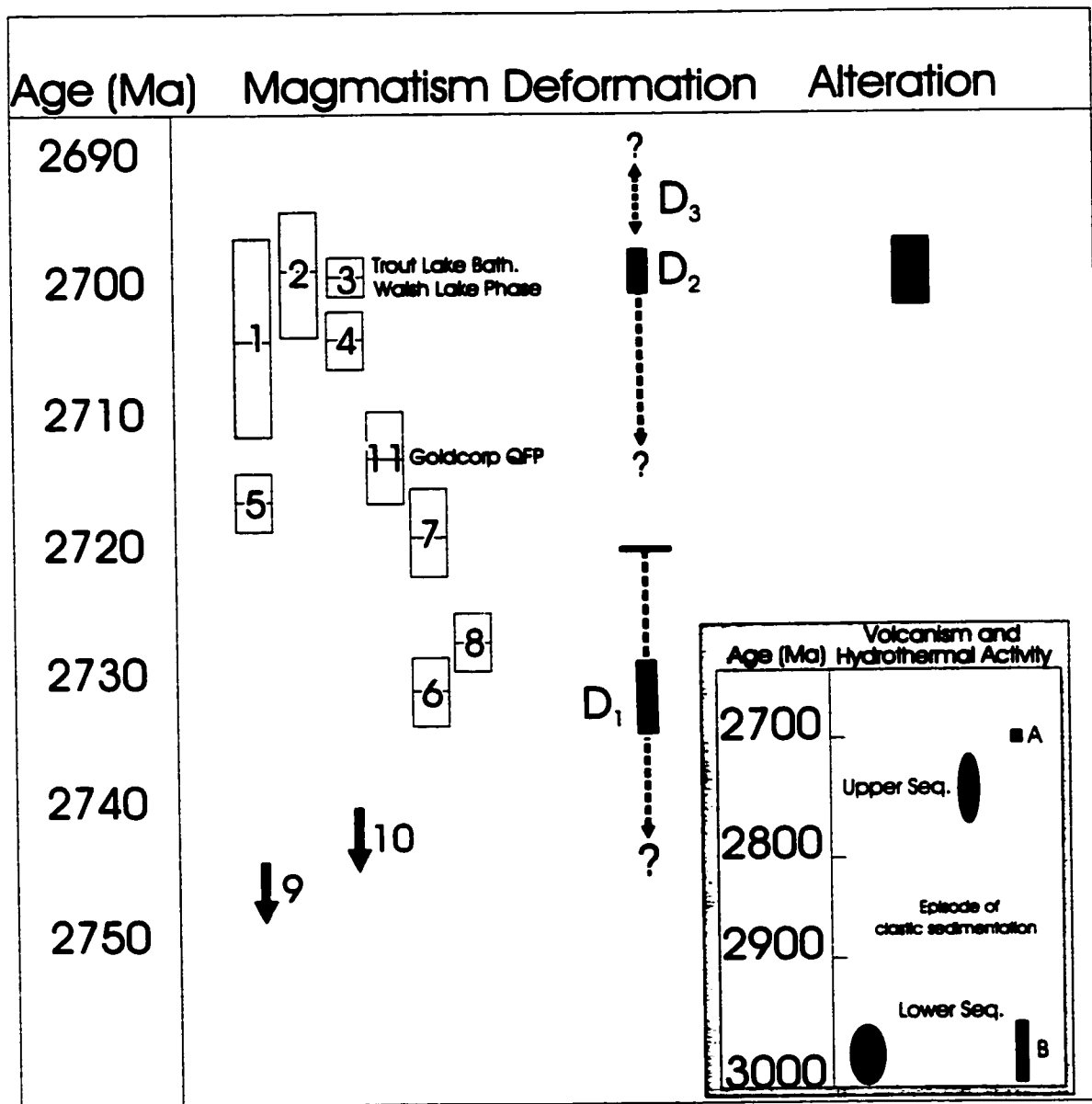


Figure 2.6

Geology of the eastern Red Lake greenstone belt and the East Bay deformation zone (D.Z.). The East Bay D.Z. is interpreted as a sinistral strike-slip shear zone by Hugon and Schwerdtner (1984) and Andrews et al., (1986). This is confirmed by the author's examination of outcrops and aeromagnetic maps (Placer Dome Canada Ltd., 1994; Ontario Geological Survey, 1978). The aeromagnetic maps show asymmetric folds in ultramafic units, with NE trending axial traces. Similar folds are observed in iron formation on outcrops. Extension lineations in sheared iron formations are sub-horizontal and bedding rotates into the East Bay D.Z., compatible with sinistral shear. Strongly attenuated F_2 fold limbs in the Campbell mine area are shown with a component of sinistral shear along limb-parallel faults, which strike NW trending between the Campbell and H.G. Young mines. Southeast of the Goldcorp mine, the fault zone merges into the Upper / Lower sequence contact. The faults terminate at the East Bay deformation, and probably developed during late F_2 folding. This suggests motion along the East Bay deformation zone was synchronous or post-dated F_2 folding and limb shearing. Foliation-parallel auriferous zones at the Campbell, Goldcorp, and Marcus properties correspond spatially to the fault zone interpreted here.

Legend for lithologies and mine names provided in Fig. 2.1.

439000
5662000



455000
5653000

0 1 2
Scale (km)

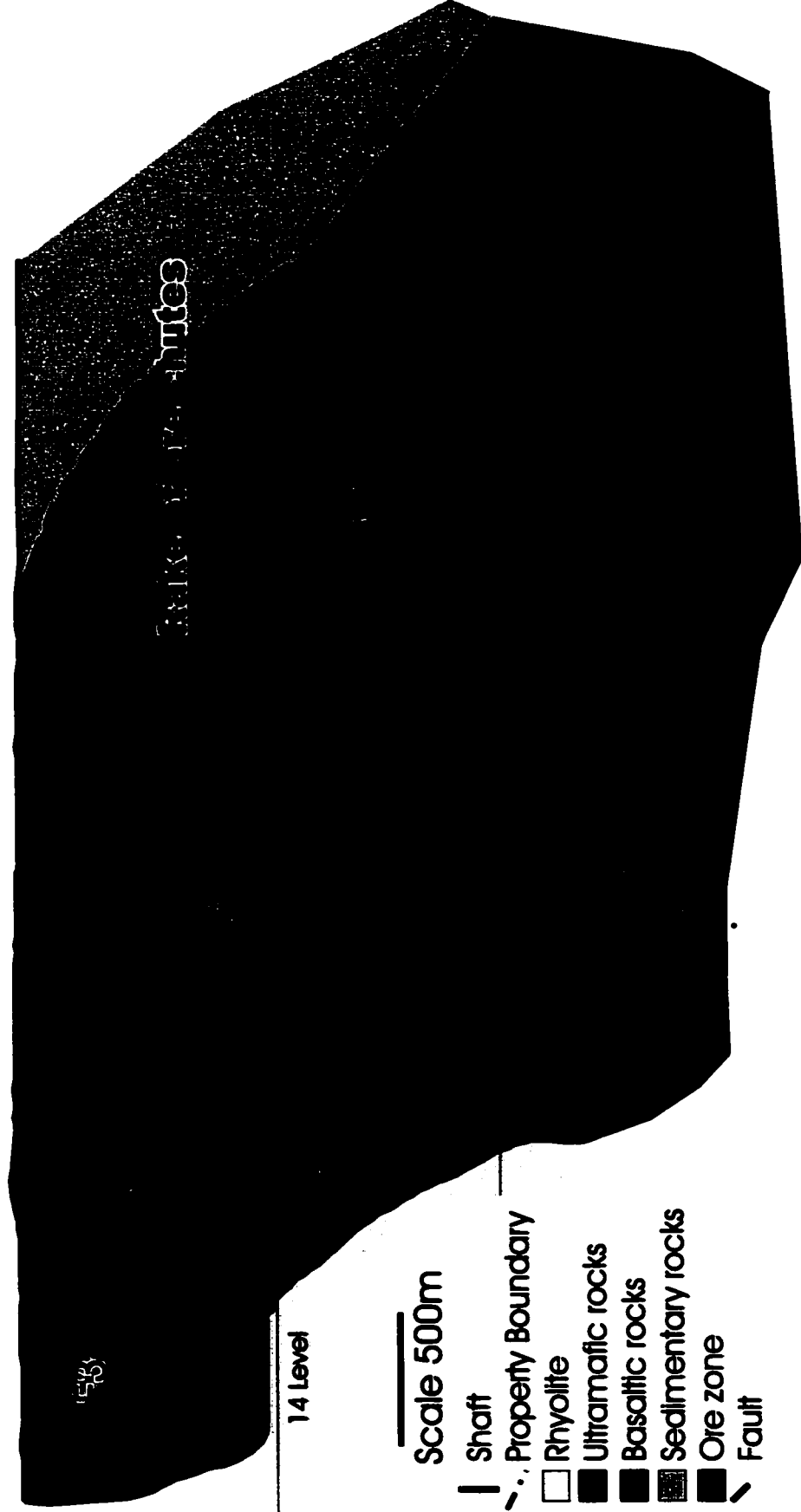
 Fault, with shear sense

Figure 2.7

Longitudinal section of the Campbell (left) and Goldcorp Red Lake mine (right), viewing NE across foliation-parallel (NW trending) auriferous veins. The 14th and 27th level of the Campbell mine are shown as fine horizontal lines. The Reid shaft at the Campbell mine is not shown. Note the folding of ultramafic units. The contact between the sedimentary rocks (Bruce Channel assemblage, east) and volcanic rocks (dash line) is not folded, and truncates folds in the volcanic rocks. The metasedimentary rocks are strongly deformed and folded, as indicated on the diagram. The bedding in volcanic and sedimentary rocks have similar strikes and dips (e.g., Pirie and Grant, 1978). This suggests a fault occurs at the sedimentary / volcanic rock contact (dashed line). The traces of F_2 fold axes in the volcanic sequence are shown as fine dashed lines. Ore chutes rake consistently moderately to steeply SW across folded stratigraphy, suggesting that ore chutes have not been folded. Data was compiled from grade and longitudinal sections of the Campbell mine, and those of the Goldcorp mine (Rogers, 1992). Property boundary between Campbell and Goldcorp mines is shown as a dash-dot line. Ore bodies at the Campbell and Goldcorp Red Lake mine are indicated by SC, F, A, F2, and F3.

NW

SE



14 Level

Scale 500m

- Shaft
- Property Boundary
- Rhyolite
- Ultramafic rocks
- Basaltic rocks
- Sedimentary rocks
- Ore zone
- Fault

Figure 2.8

Geology of the eastern Red Lake greenstone belt, showing structural data along the traverse (black line on the map, with red points showing data collection sites) across the axial trace of the F₂ McDougal Lake Antiform. The stereonet shows the poles to foliations (black squares, □), and the trend and plunge of garnet and amphibole extension lineations (●) on the foliation planes. The mean foliation plane determined from contoured pole data is shown as the black great circle. The minerals defining the extension lineations are part of the contact metamorphic assemblage related to the late phase of the Trout Lake batholith. Lineations do not undergo rotation across the axial trace, indicating folding occurred during, or before metamorphism. Note that the mean foliation plane contains the extension lineations, and is parallel to the axial trace of the McDougal Lake Antiform. This suggests F₂ NE-SW shortening, SW extension, and metamorphism were coeval. Stereonet is a lower hemisphere, equal area projection. Legend for lithologies and mine names provided in Fig. 2.1.

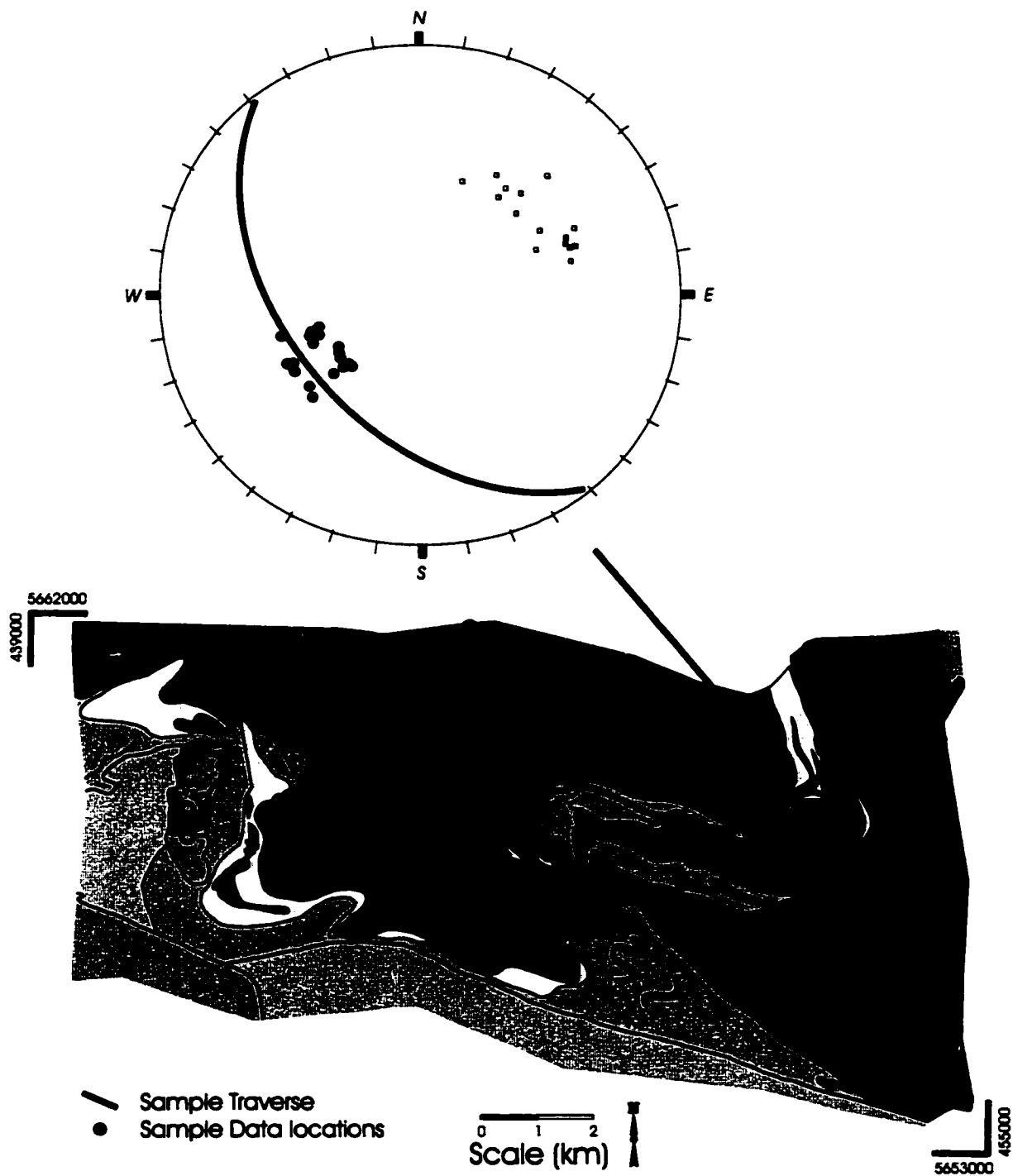


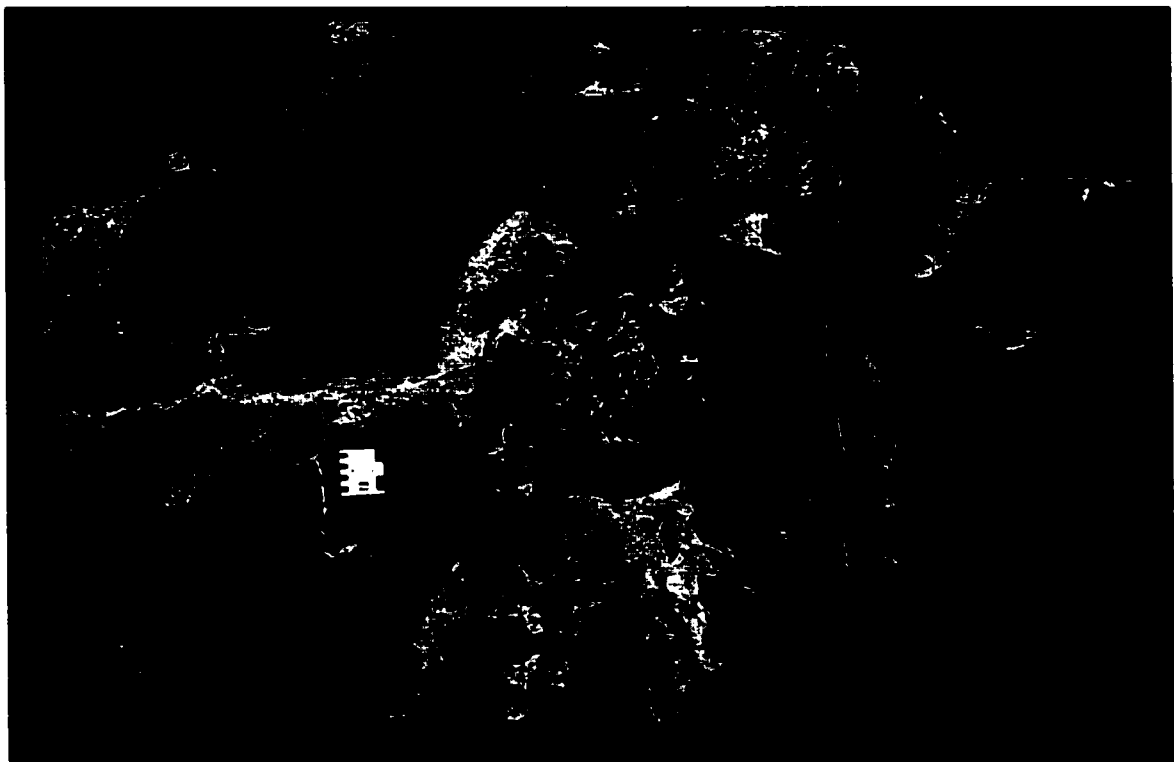
Figure 2.9

Granitic dykes and sills in strongly foliated basalt at the contact with the Trout Lake batholith. Note that calc-silicate (garnet-diopside) bands occur sub-parallel and parallel to well developed foliation planes. Quartz veinlets inclined to the left are boudinaged. Granitic dykes occur along the foliation, and those cutting the foliation are both folded, and relatively planar. The evidence indicates the intrusion of the Trout Lake batholith during and after shearing and foliation in the basalt. Location: 456117E, 5659899N, UTM NAD27.



Figure 2.10

Brecciated pillow basalt with granodiorite near the Trout Lake batholith. Brecciation extends > 500m from the margin of the Trout Lake batholith, suggesting stoping of the batholith. NW trending foliations (parallel to the short side of the photograph) in the country rock fragments are cut by intrusive rocks. This suggests that intrusion occurred after the initial development of foliation in the country rocks. Location 454615E, 5659927N, UTM NAD27.



3.0 Geology of the Campbell mine

The host rocks at the mine are E-W to NW trending mafic-ultramafic rocks and rhyolite of the Lower Sequence (Fig. 3.0). Two komatiite units occur within the predominantly mafic tholeiitic basalt sequence. Basalt comprises ≈ 55 vol.% of the mine rocks, ultramafic rock ≈ 30 vol.%, and rhyolite ≈ 15 vol.% (Fig. 3.0). Numerous NW and $\approx$ N-S trending diabase and lamprophyre dykes are present in the mine area. Interflow sedimentary rocks are volumetrically minor in the mine. All rocks in the mine are deformed. Fe-carbonate alteration occurs as disseminated blebs, veins and veinlets in the volcanic sequence throughout the property.

3.1 Rock Types

3.1.1 Ultramafic rocks

Two ultramafic units (basaltic komatiites to komatiites, Pirie, 1982, and Fig. 3.1a) occur in the mine and are conformable with the mafic and felsic volcanic units. Representative compositions are provided in Appendix 1. The margins of the ultramafic units are dark green, fine-grained, strongly foliated, and commonly brecciated (Fig. 3.2). Breccia fragments are semi-rounded to angular, and 10 to 100 cm in size. The mine term for the brecciated ultramafic rocks is “mottled fine diorite chert”. The breccias are composed of ≈ 30 vol.% chlorite along the foliation, with quartz and minor talc, chromite, feldspar, and magnetite. Carbonate (20 to 40 vol.%) occurs as coarse-grained clots up to 2 to 4 mm, wrapped by chlorite, and also as fine anhedral grains in the matrix. The breccia matrix is composed of chlorite, chlorite + carbonate with minor quartz, and carbonate + quartz with biotite and fuchsite. The matrix compositions vary in the proximity of auriferous veins. The occurrence of abundant chromite and sporadic fuchsite support the interpretation that these rocks were originally ultramafic rocks.

The mine term for non-brecciated, fine-grained chloritic and carbonate porphyroclastic ultramafic rocks is “chloritic rock”. These rocks are composed predominantly of fine-grained dark green chlorite (< 35 vol.%), fine-grained anhedral, to coarse-grained (< 7.5 mm) carbonate (30 to 50 vol.%), quartz, and < 2 vol.% fine chromite. Chlorite wraps around carbonate grains. Biotite, and locally pyrite, overprints carbonate and chlorite. Christie (1986) recognized relic spinifex textures in this rock

type, supporting a komatiite origin. Chloritic ultramafic rocks plot in the basaltic komatiite field on a Jensen cation plot (Fig. 3.1b). Representative “chloritic” ultramafic rock compositions are provided in Appendix 2.

The interiors of ultramafic units vary from green, fine-grained rocks containing chlorite + talc + serpentine, with minor magnetite and carbonate, to coarse-grained light grey rocks, with talc + magnetite + serpentine pseudomorphs after olivine, in a serpentine + magnetite matrix. Patches of carbonate are intergrown with talc and serpentine. The variations from chlorite + carbonate + quartz assemblage in ultramafic rocks, at the margins of ultramafic units, to talc + serpentine bearing assemblages in the cores of ultramafic units suggests decreasing carbonate alteration of the ultramafic units, inwards.

3.1.2 Basaltic rocks

Basaltic rocks are dark green, ranging from massive to variolitic flows and occasional pillowed units. Representative compositions are given in Table 3.0, and generally lie within Fe to Mg tholeiite fields on a Jensen (1976) classification diagram (Fig. 3.1c). Garnetiferous basalt occurs throughout the eastern and deeper levels of the mine, and garnet is particularly abundant in basalt adjacent to some auriferous veins, chloritic pillow selvages, and bleached, buff-brown color basaltic rocks. With increasing proximity to the auriferous veins, garnetiferous basalt commonly grades to green-grey or grey-brown silicified basalt containing less MgO and CO₂ than basalt distal to the auriferous veins (Table 3.0). Garnetiferous basalts are somewhat MnO-rich and K₂O and Na₂O-poorer than silicified basalt, and basalt distal to the auriferous veins, although there is significant overlap in the concentrations of the major oxide elements in all varieties of basaltic rocks (Table 3.0). Increasing SiO₂ concentration, and decreasing CaO is noted from relatively unaltered basalt to garnetiferous basalt, to silicified basalt compositions. The mine term for basaltic rocks is “andesite”.

3.1.3 Rhyolitic rocks

Rhyolite is cream-yellow to white, fine-grained, with sporadic quartz phenocrysts. The unit is commonly cleaved and brecciated, with flattened angular breccias (75 to 130 cm) cut by anastomosing dark brown to green cleavage seams (Fig. 3.3) composed of

chlorite + biotite + tourmaline, and subordinate anorthitic plagioclase, pyrite, and carbonate. Rhyolites have been pervasively altered to an assemblage of quartz and sericite with minor pyrite, flecks of chromian muscovite and biotite, and occasional andalusite.

3.1.4 Interflow sedimentary Rocks

Sedimentary rocks occur as narrow (< 1 to 2 m) units, at mafic volcanic - felsic volcanic contacts, and in basalt units. Interflow sedimentary rocks are fine-grained, grey-brown to black pelitic units, and commonly contain sulphide-facies iron formations. They contain bands and nodules of Fe sulphides, a mixture of pyrite and pyrrhotite, up to 30 vol. %. In the eastern and deeper levels of the mine, these rocks contain garnet, white mica, syn-tectonic andalusite porphyroblasts, and minor chlorite.

3.1.5 Intrusive rocks

3.1.5.1 Lamprophyre

Dykes are generally ≤ 1 to 3 meters wide. Lamprophyre dykes cut all rock types including quartz+feldspar porphyry dykes (Fig. 3.4), and are therefore the youngest dykes in the area. Lamprophyre dykes also cut all auriferous veins underground and on the surface outcrops (Figure 3.4). Three sets of grey to black lamprophyre dykes are present at the mine: foliation-parallel dykes; N-S trending and steeply west dipping dykes, and N- to NE-trending dykes with shallow to moderate NW dips.

Chlorite pseudomorphs after hornblende suggest foliation-parallel dykes and some N-S dykes are spessartite lamprophyres, following the IUGS classification (Le Maitre, 1989). This is consistent with the occurrence of less-altered spessartite dykes north and west of the mine. Lamprophyre dykes are boudinaged with sub-vertical and sub-horizontal boudin axes, weakly foliated and hydrothermally altered. Quartz veins with coarse-grained carbonate and black tourmaline cut lamprophyre dykes.

3.1.5.2 Diabase dykes

North to NE and N-S trending dykes (≤ 4 m wide) also include minor grey to black diabase dykes, containing up to 40 vol.% zoned plagioclase phenocrysts, in a

Table 3.0. Bulk chemical compositions of basalt at the Campbell mine

	Average* Silicified 6#	Average Garnetiferous 3#	Average Basalt 15#	CRL02*	CRL09*	CRL19*	CRL49a*
SiO ₂	70.23±5.09	58.87±3.91	49.49±5.04	68.35	50.12	51.51	52.16
Al ₂ O ₃	14.85±0.90	15.03±3.31	13.34±1.01	12.41	13.87	16.52	14.91
TiO ₂	0.96±0.06	0.92±0.01	0.84±0.06	0.77	0.88	1.52	0.83
Fe ₂ O ₃ *	11.67±5.37	16.03±2.39	12.81±2.01	8.14	10.16	11.15	7.26
MgO	2.40±1.00	3.44±0.48	6.03±1.11	2.61	5.33	4.81	3.66
MnO	0.26±0.12	0.33±0.06	0.24±0.05	0.28	0.17	0.17	0.12
CaO	0.99±0.99	1.47±1.46	7.95±1.47	2.32	7.72	5.33	6.68
Na ₂ O	0.24±0.16	0.09±0.03	0.35±0.20	0.17	0.23	0.16	3.35
K ₂ O	0.34±0.32	0.07±0.07	0.32±0.27	0.27	1.03	0.15	0.50
Cr ₂ O ₃	0.05±0.01	0.04±0.01	0.05±0.02	0.06	0.07	0.03	0.00
P ₂ O ₅	0.11±0.02	0.10±0.02	0.10±0.01	0.07	0.06	0.13	0.32
CO ₂	0.18±0.22	0.18±0.08	5.90±2.49	NA	NA	NA	NA
S	0.32±0.15	0.19±0.06	0.24±0.16	0.50	5.41	0.19	0.41
LOI	2.65±0.54	3.17±0.15	7.78±2.56	3.64	4.38	7.86	9.54
Total	99.32	99.57	99.30	99.59	99.44	99.53	99.74
Trace elements (ppm)							
Ni	NA	NA	NA	160	137	239	28
Cu	NA	NA	NA	81	169	BD	165
Zn	NA	NA	NA	90	157	61	83
Rb	NA	NA	NA	6	18	8	14
Sr	NA	NA	NA	30	137	70	1370
Y	NA	NA	NA	BD	8	9	6
Zr	NA	NA	NA	51	53	88	201
Nb	NA	NA	NA	5	BD	6	BD
Pb	NA	NA	NA	BD	BD	9	BD

#=number of samples, data obtained by Placer Dome Canada, see Appendix 3 for method.

*=Total Fe reported as Fe₂O₃, determined by XRF.

NA= not analyzed.

BD=below detection limits

Ba below detection limit.

CRL02 Garnetiferous, silicified basalt with chloritoid. Garnets are subhedral, up to 2 mm in size and 2 vol.%. 1652A stope, L zone.

CRL09 Basalt adjacent to high Au grade ore mineralization. 2351-5 stope, L zone.

CRL19 Relatively unaltered dark green to grey green basalt. DDH 14-461.

CRL49a Relatively unaltered dark green basalt. DDH 14-470.

CO₂ determined by Leco furnace

S determined by Leco furnace

matrix composed of fine plagioclase, green anhedral to subhedral biotite, and minor anhedral fine-grained epidote. Pyrite and carbonate also occur in the matrix.

3.1.5.3 Quartz + feldspar porphyry dykes

Felsic dykes are rare at the mine, and are recorded on upper-level mine plans. The author did not recognize them in accessible underground areas during the course of this study. The mine geology maps shows that quartz + feldspar porphyry dykes are NW to WNW trending.

A narrow NW-trending porphyritic dyke occurs on a stripped outcrop immediately north of the G zone ore body (Fig. 3.4). It contains dark grey quartz (10 to 20 vol.%, ≤ 1 to 3 mm) and feldspar (≤ 10 vol.%) phenocrysts with biotite (≤ 5 vol.%) in the matrix. This dyke has clearly experienced hydrothermal alteration, containing quartz veins (Fig. 3.5a), quartz+carbonate veins (Fig. 3.5b), and fine disseminated carbonate and pyrite (Fig. 3.6). Foliation in the dyke (Fig. 3.7) is parallel to those in the wallrocks. The dyke cuts across the shear zone hosting the G ore body. A direct cross cutting relationships between the dyke and ore are not observed, but the dyke is thinned (Fig. 3.8) and offset with a dextral sense (Fig. 3.4) by the deformation zone hosting the G ore zone (Fig. 3.4). Strain in the dyke (inferred from the intensity of foliation development) increases towards the shear zone. The relatively unstrained part of the dyke is exposed east of the shear zone (Fig. 3.4), and is approximately 1 to 1.5 m wide, while the strongly foliated dyke within the shear zone is < 15 cm wide. Clearly, the dyke is not post-tectonic, as indicated by Sanborn-Barrie et al. (2000). These observations suggest that the intrusion predated the mineralization, and the dyke was introduced prior to, or early during D₂ shortening and carbonate alteration.

Felsic dykes with similar NW to WNW trends are observed in the adjacent Goldcorp mine (MacGeehan and Hodgson, 1982). A U-Pb zircon age (2714 ± 4 Ma) was obtained by Corfu and Andrews (1987) from a moderately altered, and moderately to intensely deformed dyke. The intrusion of the dyke has been interpreted to postdate the gold mineralization at the Goldcorp mine (Corfu and Andrews, 1987; Andrews et al., 1986, Menard et al., 1999). This has been used to constrain the youngest age for mineralization (Corfu and Andrews, 1987; Penczak and Mason, 1997; Menard et al.,

1999), but the geological relationship between ore veins and the dyke is not well established.

3.1.5.4 Dioritic intrusions

Grey to grey-blue medium grained equigranular to porphyritic “Campbell diorite” occurs at intermediate to upper levels in the mine (Fig. 3.0). Rounded quartz phenocrysts are locally present in the groundmass of quartz and plagioclase. Biotite and chlorite occur in the groundmass, but are most abundant in irregular patches and wisps, which together with trace garnet give the rocks a mottled appearance.

3.2 Deformation at the Campbell mine

3.2.1 Previous work

The style of deformation related to auriferous mineralization at the Campbell mine has been in debate. Andrews et al. (1986) noted that extension lineations on foliation planes rake moderate to steeply SW, and auriferous chutes in veins plunge parallel to extension lineations at the Campbell and Goldcorp mines. They suggested that foliation-oblique crustiform quartz + carbonate veins formed during simple shear, characterized by normal to oblique (dextral-normal) motion. They interpreted that the foliation-parallel veins post-dated the foliation-oblique zones. Wilson et al. (1984) also suggested that auriferous mineralization is hosted by strike-slip shear zones, but the shear zones developed under a regime of bulk subhorizontal extension and subhorizontal NE-SW shortening after earlier penetrative foliations were developed. They considered that the earlier penetrative foliations developed during the same progressive NE-SW shortening event, but with a steeply dipping minimum stress direction. Zhang et al. (1997) proposed that the deformation at the mine is characterized by pure shear during the auriferous mineralization. Foliation-parallel auriferous veins formed during a predominantly flattening strain regime. Foliation-oblique auriferous veins are hosted by second-order conjugate shear zones, post-dating foliation-parallel veins (Zhang et al., 1997). The maximum principal stress direction was NE-SW trending and subhorizontal throughout the ductile to brittle-ductile deformation. The intermediate principle stress direction switched from subhorizontal, NW-SE trending during the generation of

foliation-parallel veins, to sub-vertical during the development of NNW and E-W conjugate arrays, and auriferous mineralization. The proposed interpretation of Zhang et al. (1997) is that some auriferous mineralization occurred within local, conjugate strike-slip shear zones, under a regime of NE-SW compression and NW-SE subhorizontal extension, is compatible with the interpretation of Wilson et al. (1984). Zhang et al. (1997) also showed that two stages of deformation continued after auriferous mineralization, during the progressive NE-SW shortening.

3.2.2 Folding

Lithological boundaries in the Campbell and adjacent Goldcorp mine display numerous folds on geology plan maps, with NW to NNW trending fold axial traces (Fig. 3.0). These folds may represent tightened and attenuated F_1 folds (Fig. 3.9.1), overprinted by F_2 folds. The F_2 folds in the mine are parasitic Z-shaped F_2 folds developed on the SW limb of the Balmer Synform (Fig. 2.3). Eastward, from the Campbell through Goldcorp mines, NW trending folds progress from tight to isoclinal or rootless (Fig. 3.0). This suggests that strain during F_2 folding increased eastward. The attenuation of NW trending fold limbs may have occurred during progressive NE-SW shortening, coeval with attenuation and shearing of the eastern limb of the Balmer Synform.

The orientation of F_2 fold axes were determined from bedding data measured from the mafic – ultramafic rock contact in the L and NL ore zone areas, at intermediate levels in the mine. The poles to bedding planes lie on a great circle distribution on a stereonet (Fig. 3.9). The distribution defines a strong girdle on an eigenvalue ratio plot (Woodcock, 1977). The pole to the great circle plane, representing the fold axis, plunges moderately SW (Fig. 3.9). The distribution of bedding poles corresponds to either an open NE trending fold closing to the SW, or a tight, reclined NW trending fold. Mapping supports that the fold is tight and reclined, with a NW striking axial surface (Fig. 3.9, inset).

3.2.3 Foliations and lineations

Foliations are defined by the alignment of platy minerals (chlorite, biotite, and white mica), oblate carbonate grains, wispy sulphides, and a preferred shape fabric of

quartz grains. Throughout the mine, foliation planes are weakly developed in mafic rocks, and more prominently developed in chloritic ultramafic rocks, which contain more carbonate and chlorite. Rhyolite rocks contain an anastomosing cleavage. Foliation intensity in basaltic and ultramafic rocks increases over a short (typically < 10 m) distance towards auriferous veins, suggesting increasing strain, as shown by Zhang et al. (1997). Eastward and with depth, foliations are more prominently developed outside the shear zones, through the Campbell and Goldcorp Red Lake mine properties. This suggests apparent strain increases eastward and with depth, compatible with increasingly tight folding, and fold limb attenuation. Rigg (1980) reported a localized second foliation, subparallel to the first foliation, although the relations between these foliations are reported as uncertain. Penczak (1996) reported a rare crenulation cleavage. This writer suggests the crenulation cleavage represents a folded S_1 fabric, which is overprinted at a high angle by S_2 in the hinges of F_2 folds. Along F_2 limbs, S_1 and S_2 are subparallel, similar to the relationship observed by Rigg (1980). A prominent L fabric is not well developed on the foliation planes, supporting observations by Rigg (1980) and Zhang et al. (1997). Extension lineations are rare at the mine, and generally plunge moderately SW on the foliation planes. Rigg (1980) and Zhang et al. (1997) report lineations on foliation planes are commonly intersection lineations between foliation, and bedding. One prominent foliation was observed by the author. Progressive changes in the orientation of foliations occur with increasing foliation intensity into NNW and E-W shear zones, as described below. The author has observed that extension lineations on foliations also undergo rotation into the shear zones.

Foliation planes measured by the author throughout the mine trend NW, and dip moderately to steeply SW (Fig. 3.10), having a mean orientation similar to foliations observed outside the mine area. The poles to foliation planes lie on a partial great circle distribution. The great circle distribution of poles to foliation planes (Fig. 3.10) is similar to the great circle corresponding to the poles to bedding planes (Fig. 3.9). This suggests that the foliations are parallel to bedding and also folded. However, S_2 foliations are at a high angle to bedding in F_2 fold hinges (Fig. 3.9.1). S_2 foliations rotate progressively with increasing apparent strain, into shear zones on F_2 limbs (Fig. 3.4, 3.13). These relationships support that the foliation shown in figure 3.10 developed during D_2 . The

distribution of poles to foliation planes shown in this figure are a result of the kinematics of discrete D_2 shear zones on F_2 limbs near F_2 hinges, as discussed below. The fold axis determined from the bedding data lies very close to the mean foliation plane (Fig. 3.10) which bisects the fold limbs (Fig. 3.9), suggesting NW trending foliations are axial planar to moderate SW plunging, NW trending F_2 folds.

3.2.4 Shear zones and relations to folds

MacGeehan and Hodgson (1982), Wilson et al. (1984), and Zhang et al. (1997) have shown that dextral shear occurred along NNW trending auriferous veins. The evidence for dextral shear includes anticlockwise rotation of foliation planes and NW veins into NNW trending veins (Fig. 3.11) on stereonet and maps (Fig. 3.4, 3.13), rotation and offsets of foliation-parallel veins (Fig. 3.4), and asymmetric folding of NNW veins (Zhang et al., 1997). Zhang et al. (1997) also proposed that auriferous E-W veins occur in sinistral shear zones, from rotations of foliations and NW veins into E-W shear zones, and asymmetric folding of E-W veins.

The author's mapping of surface outcrops and underground exposures support progressive rotation of veins and foliations into E-W and NNW shear zones, and the strike-slip motion proposed by Zhang et al. (1997). East-west trending veins display asymmetric folding (Fig. 3.12a) and clockwise rotation of foliation planes and NW trending veins (Fig. 3.11) consistent with sinistral shear. Rotation of foliation planes, and rotation and offset of NW trending quartz + carbonate veins into NNW trending zones is supported here from underground (Fig. 3.13) and surface mapping (Fig. 3.4, 3.14). Subvertical boudin axes are parallel to the intersection of NNW and E-W veins (Fig. 3.11), and together with the shear senses along NNW and E-W zones, indicate that they are a conjugate array, as proposed by Zhang et al. (1997). The plane of NW veins and foliations approximately bisects the NNW and E-W vein sets (Fig. 3.11), suggesting they formed during the same deformation episode (Zhang et al., 1997).

The author shows that poles to bedding planes lie between E-W and NNW shear zones (compare figures 3.9 and 3.11). E-W shear zones, and NNW shear zones, are spatially related to E-W and NNW fold limbs at NNW and E-W mafic-ultramafic contacts, respectively. The fold axis determined from the bedding data corresponds

closely to subvertical boudin axes in E-W, NNW, and NW veins, and to the mean NW foliation plane. The author has shown above that NW trending foliations are axial planar to F_2 folds. NW foliations undergo progressive rotation into E-W and NNW shear zones, therefore the shear zones were developed during F_2 NE-SW shortening.

The kinematics of E-W and NNW shear zones, and their spatial and temporal relation to F_2 folds, suggest these shear zones may have been developed at lithologic contacts during flexural slip folding accompanying D_2 shortening.

3.2.5 Folding, shearing, and controls on high Au grade ore chutes in foliation-oblique zones

High Au grade mineralization occurs as podiform to pipe – like hydrothermal breccia bodies of quartz and arsenopyrite, within earlier quartz + carbonate veins. The trend and plunge of these pipe-like quartz + arsenopyrite breccia bodies was determined from stope and grade maps. The pipe-like high gold grade ore bodies plunge $\approx 243^\circ/65^\circ$ in the L ore zone (51 ore body, 14, 17, and 18th levels), and $\approx 249^\circ/52^\circ$ in the P ore zone (51-6 ore body, 22, 23, and 24th levels). This orientation coincides with the intersection of NNW and E-W shear zones, subvertical boudins, and axes of NW folds estimated from bedding data (Fig. 3.11).

In addition, the quartz + arsenopyrite breccias are spatially related to the hinges of NW trending folds (Fig. 3.15), and the intersection of E-W and NNW shear zones on F_2 limbs near hinges (Fig. 3.16). Dilation at fold hinges, at the mafic – ultramafic rock contacts, may have occurred during F_2 flexural-slip folding, and resulted in preferentially focused fluid flow.

3.2.6 Deformation related to foliation-parallel zones

Previous workers have interpreted the A zone vein system at the Campbell mine (Fig. 3.0) as a sinistral shear zone (MacGeehan and Hodgson, 1981), dextral shear zone (Penczak and Mason, 1997), and a deformation zone predominated by pure shear (Zhang et al., 1997). Penczak and Mason (1997) and Penczak and Mason (1999) have also proposed that the planar veins formed by “strong flattening” of earlier auriferous quartz + carbonate veins.

Auriferous foliation-parallel quartz + carbonate veins dip steeply, and are boudinaged. Vertical longitudinal sections along foliation-parallel zones for the Goldcorp and Campbell mines show that ore chutes plunge consistently SW (Fig. 2.7). It was shown above (section 3.2.5) that high gold grade breccias also plunge moderately SW. This suggests that the plunge of auriferous mineralization is similar in both NNW auriferous veins, and the foliation-parallel auriferous veins. Folded ultramafic units are clearly evident on both plan (Fig. 3.0) and section maps (Fig. 2.7). The vertical longitudinal section (Fig. 2.7) is not an axial planar section to F_2 folds, and therefore shows F_2 axial traces at a low angle to the horizontal, on the section. The consistent plunge of the ore chutes across folded strata suggest the ore chutes have not been folded, and must have developed during or after the folding. It is also unlikely that the NW veins formed initially at a high angle to S_2 , and were rotated into the flattening plane. These vein systems are planar over large strike and dip extents (> 1 km), and do not display strong internal folding. On plan maps, foliation-parallel zones are linear, in contrast to folded strata. Figure 3.17 shows that foliation-parallel auriferous veins are spatially related to highly attenuated limbs of F_2 folds at the Campbell and Goldcorp mines. These observations suggest that auriferous mineralization occurred during, or after F_2 folding. It was suggested above that a temporal relationship exists between F_2 folding, and the development of high gold grade ore chutes in foliation-oblique auriferous veins. The geometric similarity between the plunge of ore chutes along NNW auriferous veins and NW auriferous veins, and the spatial relations of foliation-parallel auriferous veins to attenuated F_2 fold limbs, suggests foliation-parallel auriferous veins may also be temporally related to F_2 folding. These observations are compatible with the interpretation of Zhang et al. (1997), which indicates that shear zones hosting foliation-parallel auriferous veins accommodated significant NE-SW shortening.

The foliation-parallel auriferous veins form left-stepping en-echelon arrays. Many workers interpret en-echelon vein arrays to have formed under a simple shear regime (i.e. Hodgson, 1989). Foliation planes do not undergo rotation into the foliation-parallel auriferous veins, and the mean foliation plane and mean NW vein plane are not significantly different (Fig. 3.11). Earlier formed veins are symmetrically shortened about the foliation. These observations support the interpretations of Zhang et al. (1997).

The foliation-parallel veins are a tabular system with over > 1 km of strike and down-dip extents. These veins are not significantly folded. These observations suggest that the veins were not initially developed at a high angle to the S_2 foliation, and rotated into the S_2 fabric.

3.2.7 Post-ore deformation

Auriferous mineralization is cut by shallow-dipping veins of milky quartz $\pm$ tourmaline (Fig. 3.18), pyrite, pyrrhotite, and carbonate. Quartz $\pm$ tourmaline and carbonate tension veins occur in rocks throughout the mine area. Ore bodies, and some lamprophyre dykes, show minor offsets (typically < 3 m) across narrow (< 2 cm wide) faults (Fig. 3.19), termed “black line faults” at the mine. The “black line faults” are planar to gently folded, and composed of chlorite, quartz, tourmaline, and occasional coarse-grained scheelite. Black line faults cut and displace auriferous veins and lamprophyre dykes, and therefore post-date these events. Larger offsets (.10 to 100 m) of auriferous zones may occur across reverse faults (51 zone). In these areas, black line faults and auriferous mineralization are strongly deformed.

Lamprophyre dykes cut auriferous veins, but are boudinaged, and cut by shallow-dipping quartz $\pm$ tourmaline veinlets. Therefore, these barren veins post-dated the auriferous mineralization. In E-W shear zones, shallow-dipping quartz veinlets are rotated and deformed into S-shaped sigmoids, and offset across E-W shear planes (Fig. 3.20), indicating sinistral strike-slip shear continued along some E-W auriferous veins, after auriferous mineralization. The E-W shear zones observed in this study occur on the northerly limbs of F_2 folds. Leaf-texture and coarse-grained visible gold locally occurs in these veinlets where they cut high-grade gold mineralization, suggesting local remobilization of gold, after the main gold mineralization event. The shallow to moderate dips of post-ore veinlets is compatible with the interpretation of Zhang et al. (1997) that lamprophyre dykes were deformed by NE-SW shortening, with a moderate-plunging minimum stress direction.

Black line faults and ore bodies are strongly deformed and offset by WNW to NW striking, moderate to steep SW dipping brittle-ductile faults, at and below the 27th level, and on the 14th and 15th levels, in the mine. The faults at and below the 27th level were

observed during brief underground visits to the 27th and 30th level, and were not studied in detail. Slickenlines on fault surfaces rake steeply. The rotation of foliation planes, occurrence of cylindrical boudinage of earlier carbonate veins, with shallow plunging boudin axes, and shallow dipping tension veinlets (Fig. 3.21), suggest that reverse-motion operated across discrete (several m's wide) brittle-ductile fault zones subparallel to the foliation. A component of left lateral motion is evident from offset veins, asymmetric folding of veinlets (Fig. 3.22), and rotated carbonate porphyroclasts. This is compatible with the final phase of NE-SW shortening, post-dating initial deformation of the lamprophyre dykes, suggested by Zhang et al. (1997). During this event, foliation planes and planar, foliation-parallel anisotropies may have been reactivated as reverse shear zones. These post-ore shear zones are spatially related to attenuated F_2 limbs (Fig. 3.17), and correspond to some earlier NW shear zones containing auriferous veins. In the B ore zone at the Campbell mine (Fig. 3.17) a post-ore shear zones cuts auriferous mineralization. The "01" fault below the 27th level of the Campbell mine cuts ore in the 51 zone. The development of these shear zones is suggested to be related to late D_2 shortening and attenuation of F_2 limbs.

3.2.8 Summary of deformation at the Campbell mine

D_1 developed NE trending F_1 folds (Fig. 3.9.1) in the Campbell mine area, and a weak S_1 foliation. D_2 NE-SW shortening refolded F_1 folds and S_1 foliations about tight, reclined, F_2 folds with moderate to steep SW dipping, NW striking axial surfaces (Fig. 3.9.1). Along attenuated F_2 limbs, S_1 is subparallel to S_2 . Rare crenulation of S_1 by S_2 is observed in F_2 hinge areas. The relationship between tight folds with axial surfaces subparallel to the F_2 limbs suggests some moderate SW plunging lineations in the foliation may represent intersection lineations.

Banded to crustiform quartz + carbonate veins were developed in the shear zones, prior to auriferous mineralization. S_2 foliation-parallel and oblique (E-W and NNW) shear zones. Auriferous mineralization overprints the quartz+carbonate veins, within the D_2 shear zones foliation-parallel auriferous veins are planar, and are spatially related to attenuated F_2 fold limbs. The foliation-oblique shear zones are developed at mafic / ultramafic rock contacts, along F_2 limbs near the hinge area. East-west auriferous veins

are characterized by sheeted veinlets, and NNW auriferous veins are characterized by large crustiform to banded quartz + carbonate veins. The E-W and NNW shear zones also cut and offset F_2 folded mafic – ultramafic rock contacts (Fig. 3.16). Gold occurs in quartz + arsenopyrite breccias in NW, E-W, and NNW veins, suggesting mineralization was broadly coeval in these vein systems. However, the occurrence of NNW (wide, crustiform banded vein) and E-W (sheeted veinlets) carbonate veins hosting auriferous mineralization differ. Therefore, NNW and E-W foliation-oblique auriferous vein systems do not represent the folded equivalents of the same, pre- F_2 , bedding controlled auriferous veins. Podiform ore bodies consistently plunge moderately SW across folded strata in the Campbell and Goldcorp mines. If the linear ore chutes pre-dated the folding, they should also be folded, and would therefore not be expected to have a consistent trend and plunge across the Campbell and Goldcorp mines. Therefore, the author considers it unlikely that the auriferous veins were developed prior to F_2 folding.

Foliation-parallel auriferous veins (A ore zone, parts of NL ore zone) are characterized by increasing foliation intensity and folding of earlier formed carbonate veinlets, suggesting an increase in strain towards the auriferous veins. Relatively early carbonate veinlets display symmetrical folding about the foliation (Zhang et al., 1997). This suggests that a large component of flattening was accommodated across the foliation-parallel auriferous veins, supporting the interpretation of Zhang et al. (1997), who proposed that deformation across foliation-parallel auriferous veins was predominantly pure shear. Development of these veins must have occurred under conditions of relatively high fluid pressure, allowing for dilation of the foliation during D_2 .

Foliation-oblique auriferous veins occur in E-W and NNW trending, sinistral and dextral simple shear zones (Zhang et al., 1997), respectively. These represent conjugate shear zones developed during NE-SW shortening (Zhang et al., 1997). This interpretation is supported here. The author has shown that NNW and E-W zones are spatially related to NNW and E-W trending limbs of F_2 folds near hinges. Progressive rotation of S_2 foliations into these zones indicates shear occurred during D_2 related NE-SW shortening. Ore-stage silicification overprints S_2 foliations, indicating auriferous hydrothermal alteration occurred during, or after D_2 related shortening. The author has

also shown that high gold grade quartz + arsenopyrite hydrothermal breccias are spatially related to the intersections of E-W and NNW zones and F_2 fold hinges. Mutual cross cutting relations between E-W and NNW shear zones suggest coeval strike-slip motion. Shear zones developed at mafic / ultramafic rock contacts during F_2 , possibly during folding accommodated by flexural slip. Development of conjugate shear zones during flexural slip F_2 (Fig. 3.9.1) folding is supported by the opposing shear senses along NNW and E-W zones, the spatial relations of the auriferous veins to F_2 fold limbs, and the occurrence of shear zone intersections parallel to F_2 fold axes.

High gold grade podiform breccia mineralization plunges parallel to the orientation of F_2 fold axes in both foliation-parallel and oblique auriferous veins. This suggests hydrothermal brecciation of early D_2 quartz + carbonate veins occurred during the same F_2 related event in foliation-parallel and oblique auriferous veins. Auriferous hydrothermal activity may have been focussed along the σ_2 direction in foliation-parallel zones, parallel to the orientation of F_2 fold axes and intersections of E-W and NNW auriferous veins.

Auriferous veins were deformed during a protracted NE-SW shortening event that resulted in minor offset of ore bodies across thin black line faults, and later significant left lateral-reverse offsets across discrete foliation subparallel fault zones. These fault zones are interpreted as late F_2 decollements along the attenuated western limb of the Balmer synform (i.e. Fig. 2.6), and has produced hook-shaped F_2 folds at the Campbell and Goldcorp mines (Fig. 3.17).

The data do not support the interpretation of Penczak and Mason (1997) and Penczak and Mason (1999) that the entire NE-SW shortening post-dated auriferous mineralization. Instead, the data support the interpretation of Wilson et al. (1984), that auriferous mineralization in NNW shear zones occurred during NE-SW shortening, after the initial stages of shortening and development of conjugate strike-slip shear zones. It supports the interpretation of Zhang et al. (1997) that the deformation associated with foliation-parallel auriferous veins is characterized by strong flattening, and that foliation-oblique auriferous veins developed within higher-order simple shear zones under predominantly pure-shear conditions. It also supports the proposal by Zhang et al. (1997)

that the rocks in the mine experienced a NE-SW shortening that continued after auriferous mineralization.

3.3 Metamorphism at the Campbell mine

Andrews et al. (1986) indicated the Campbell and Goldcorp mines occur approximately at the greenschist – amphibolite facies metamorphic transition. Mathieson and Hodgson (1984) confirmed that rocks at the Goldcorp mine experienced lower amphibolite-facies metamorphism under relatively low pressure conditions. The spatial distribution of metamorphic facies in the eastern part of belt suggest that emplacement of the marginal phases of the Trout Lake batholith were responsible for the metamorphism. As shown above, the D₂ event related to F₂ folding (Table 2.0) was coeval with contact metamorphism, and rocks show increasing strain eastward, towards the Trout Lake batholith. Rigg (1980) has shown that some porphyroblasts (muscovite, garnet, and amphibole) preserve an earlier foliation. This writer suggests the first foliation was developed during D₁.

The author determined the P-T conditions of metamorphism and metamorphic temperature gradients from relatively unaltered, and variably altered basaltic and interflow sedimentary rocks using mineral assemblages, garnet – biotite thermometry, and GASP (garnet + aluminosilicate + quartz + plagioclase) geobarometry. Data presented in Mathieson and Hodgson (1984) was re-evaluated using the petrogenetic grids presented here, to maintain consistency. Mineral compositions were determined by SEM-EDS, and the method is provided in Appendix 4.

3.3.1 Mineral assemblages

Basalt does not retain primary magmatic minerals. Fine-grained chlorite, white mica, and plagioclase (Assemblage 1, Table 3.1) occur in basalt in the upper and western parts of the mine, and locally in the western portion of the study area. In the western portion of the study area, basalt contains fine-grained actinolite, with minor fine-grained subhedral epidote (Assemblage 2, Table 3.1). The first appearance of garnet (Assemblage 3, Table 3.1) in basalt is closely associated with the first appearance of garnet in interflow sedimentary rocks, pelites, and greywackes. Fibrous cummingtonite

occurs in some garnet + chlorite bearing basaltic rocks (Assemblage 4, Table 3.1), as both inclusions in garnet and in the matrix. In the vicinity of the NL and L ore zones, north and east of the mine shafts (Fig. 3.0), basalt contains anthophyllite and gedrite (Assemblage 5, Table 3.1). Chloritoid (Assemblage 12, Table 3.1) appears in basalt in the mine shaft area, and distal to the auriferous veins. Adjacent to the auriferous veins at intermediate levels in the Campbell mine, chloritoid-bearing and chloritoid-free staurolitic assemblages (8, and 9 and 10, respectively; Table 3.1) appear in sedimentary rocks and altered basalt. The first appearance of staurolite is farther east, and at deeper levels than the first appearance of garnet, and chloritoid, in the Campbell mine. Chlorite and staurolite occur in the necks of boudinaged garnet bands (Assemblage 9, Table 3.1) in some rocks containing Assemblage 8. To the north and east of the mine shaft, basalt contains chlorite, ferro- and magnesio-tschermakite, and plagioclase (Assemblage 6, Table 3.1). Intermediate tuffs contain plagioclase, anthophyllite, gedrite, garnet, and biotite (Assemblage 7, Table 3.1). Assemblage 11 (Table 3.1), which includes garnet and andalusite, was observed in a single sample from a drill core below the 27th level of the Campbell mine. Andalusite, and rare cordierite are present in interflow sedimentary rocks and altered basalt from the eastern portions of the Campbell mine property, and to the north and east. Rhyolite in the underground workings of the mine locally contains andalusite porphyroblasts. Kyanite or sillimanite have not been observed in the study area. White mica is common in basalt, rhyolite, and sedimentary rocks in the mine. In basaltic rocks it exceeds 2 vol.% only in rocks immediately adjacent to auriferous veins. Ankerite, dolomite, and occasional calcite is present in rocks throughout the central and western portions of the study area, decreasing in modal abundance to the northeast and east of the mine, as recognized by Rodgers (1992) for the Goldcorp mine.

The spatial variation of minerals indicates an increasing metamorphic temperature to the east under low pressure conditions, compatible with previous work (Mathieson and Hodgson, 1984; Andrews et al., 1986). The evidence also supports one or more phases of hydrothermal alteration prior to, or during metamorphism (MacGeehan and Hodgson, 1982; Mathieson and Hodgson, 1984; Christie, 1986; Andrews et al., 1986).

Table 3.1. Metamorphic mineral assemblages

Rock Type	Assemblage #	Mineralogy*																	
		Bt	Chl	And	St	Cld	Oam*	Hbl	Grt	Msc	Act	Ep	Cum	Crd	Ilm	Mag	Po/Py	Tur	
Basalt	1		X							X					X	X		X	
	2		X								X	X			X				
	3		X						X							X	X		
		X	X						X							X	X		
	4		X						X				X		X				
	5		X				X	X	X							X	X		
Intermed. tuff		X	X				X	X	X						X	X	X	X	
		X	X				X	X	X						X	X	X		
	6		X					X							X	X			
	7	X	X				X		X							X	X		
	Interflow Sediments and Altered basalt	8	X	X		X	X			X	X						X	X	
		9	X	X		X					X								X
10		X			X				X	X			X			X	X	X	
		X	X		X				X	X			X			X	X	X	
11		X	X	X	X			X						X			X	X	
12			X			X		X	X							X	X		

* All assemblages contain quartz, plagioclase, and carbonate.

* Either anthophyllite or gedrite

Abbreviations: Act=actinolite, And=andalusite, Bt=biotite, Chl=chlorite, Cld=chloritoid, Cum=cummingtonite, Crd=cordierite, Ep=epidote, Grt=garnet, Ilm=ilmenite, Mag=magnetite, Msc=muscovite, Oam=orthoamphibole, Po/Py=pyrrhotite/pyrite, St=staurolite, Tur=tourmaline.

3.3.2 Mineral textures and mineral chemistry

Garnet: Almandine-rich garnet occurs as isolated crystals, and as layers and lenses in basalt and interflow sedimentary rocks in the eastern and deeper levels of the Campbell mine (> 700 m below surface), and to the north and east of the deposit. Coarse-grained garnet grains display zoning, with inclusion-rich (quartz, chlorite, cummingtonite), pink-orange to pink cores, to inclusion-poor, pinkish rims. The boundary between core and rim is sharp, marked by a change in color and abundance of inclusions. The X_{Alm} (mole% almandine) and X_{Prp} (mole% pyrope) increase towards the margins, with decreasing X_{Grs} (mole% grossular) and X_{Sps} (mole% spessartine), and atomic ratios of $\text{Fe}/(\text{Fe}+\text{Mg})$ (Fig. 3.23). Large garnet crystals (> 0.75 cm) at the mine are anhedral to subhedral, commonly fragmented and locally display S-shaped inclusion trails of quartz, magnetite, and pyrrhotite continuous with the foliation in the matrix (Fig. 3.24). Euhedral crystal faces are developed in pressure shadows and necks of boudinaged garnet bands (Fig. 3.25). Tabular garnet porphyroblasts occur within the foliation.

The values of X_{Alm} range from 57.9% to 86.9%, and X_{Prp} from 2.9% to 10.8%, with most values between 3.0% and 5.0%. The X_{Grs} varies between 0.0% and 15.3%, and is generally > 4.0%, and X_{Sps} varies between 0.5% and 22.5%. Representative garnet compositions are provided in Table 3.2.

Cummingtonite occurs as randomly oriented grains in the cores of compositionally zoned garnets, within foliated rocks. This suggests that metamorphism preceded deformation. The occurrence syntectonic garnet, and euhedral growth faces of garnet into boudin necks supports that metamorphism continued during deformation. The occurrence of zoned garnet may suggest that either the rocks at the mine experienced two metamorphic events, or zoning reflects a change in the garnet forming reaction during progressive metamorphism. The lack of cummingtonite in garnet rims may reflect consumption of cummingtonite during metamorphic reactions, or a change in bulk rock composition such that cummingtonite was not stable. This later hypothesis would imply that hydrothermal alteration occurred during or after amphibolite grade metamorphism, and cummingtonite within garnet was armored from alteration reactions in the surrounding rock.

Andalusite: Andalusite (100 μm to 1 mm) is subhedral to anhedral, and occurs in rhyolite, interflow sedimentary rocks, and some altered basaltic rocks from the eastern portions of the Campbell mine. Andalusite occasionally forms lens-shaped syntectonic crystals (Fig. 3.26).

Plagioclase: Fine-grained (< 40 to 800 μm) un-twinned andesine to labradorite (An_{45-65}) occurs in chloritic and garnetiferous basalt. Plagioclase is a matrix component in chloritic basalt, and occurs as lensoidal porphyroblasts in strongly deformed, garnetiferous basalt.

Chloritoid: Subhedral to anhedral twinned chloritoid laths (100 to 600 μm) occur in altered basalt and interflow sedimentary rocks near the Campbell mine shaft. Chloritoid is rarely in contact with garnet, and has ragged contacts with staurolite. In rocks lacking biotite or staurolite, chloritoid forms subhedral to euhedral laths overprinting chlorite (Assemblage 12).

Biotite: Red-brown to brown and green to brown biotite occurs in interflow sedimentary rocks, basalt, and in clastic sedimentary rocks to the east of the mine shaft. Biotite grains, together with chlorite, define the foliation. Biotite in interflow sedimentary rocks (60 to 400 μm and < 10 vol. %) replaces Fe-oxides and tschermakitic amphiboles. Green-brown biotites are locally red - brown where they replace Fe-oxide minerals.

The composition of representative biotite grains is given in Table 3.3. Biotites contain vacancies in octahedral and interlayer sites, and plot in an intermediate position close to the siderophyllite - eastonite join, on the Mg-Fe_{tot}-Al ternary diagram (Mitchell 1989)(Fig. 3.27).

Chlorite: Chlorite is light green to green-grey pleochroic, subhedral, and is generally aligned along the foliation. Randomly oriented chlorite locally replaces biotite. Chlorites are predominantly ripidolite in composition (Table 3.4), following the classification of Hey (1954) (Fig. 3.28). Chlorite comprises 30 to 40 vol.% of basaltic rocks, and decreases in abundance to < 10 vol.% in rocks north-east of the Campbell mine property. In general, chlorite becomes Mg-rich with increasing depth and to the east.

Amphibole: Fine-grained (< 80 to 100 μm) actinolite is common in basaltic rocks to the west and northwest of the mine shaft, and is locally overprinted by carbonate. Northeast

Table 3.2. Representative garnet compositions

Sample	22	16	84	11	10
SiO ₂	36.60	36.42	37.31	36.49	37.14
Al ₂ O ₃	20.52	20.59	20.75	20.21	20.38
TiO ₂	0.00	0.00	0.00	0.11	0.00
Fe ₂ O ₃	0.54	1.34	0.00	0.61	0.05
FeO	31.08	26.74	34.98	33.98	30.88
MgO	1.85	2.33	2.59	0.98	0.96
MnO	5.57	9.69	1.74	4.51	6.86
CaO	2.90	2.24	2.45	2.67	3.79
TOTAL	99.06	99.35	99.82	99.56	100.06
Formula (Basis 24 oxygens)					
Si ⁺⁴	5.989	5.938	6.029	5.993	6.043
Al ⁺³ 0.011	0.062	0.000	0.007	0.000	
Σ T site	6.000	6.000	6.029	6.000	6.043
Al ⁺³ 3.947	3.895	3.952	3.906	3.909	
Ti ⁺⁴	0.000	0.000	0.000	0.014	0.000
Fe ⁺³ *	0.066	0.164	0.000	0.075	0.006
Σ Oct site	4.013	4.059	3.952	3.995	3.915
Fe ⁺² 4.254	3.646	4.728	4.667	4.202	
Mg ⁺²	0.452	0.566	0.624	0.241	0.233
Mn ⁺²	0.772	1.339	0.238	0.627	0.945
Ca ⁺²	0.508	0.391	0.424	0.470	0.662
Σ DC site*	5.986	5.942	6.014	6.005	6.042
Endmembers (mole%)					
Alm	71.07	61.36	78.62	77.72	69.55
FeAdr	1.65	4.14	0.00	1.87	0.15
TiAdr	0.00	0.00	0.00	0.35	0.00
Grs	6.83	2.44	7.05	5.60	10.81
Prp	7.55	9.53	10.38	4.01	3.86
Sps	12.90	22.53	3.96	10.44	15.64

* Fe⁺³ calculated by charge balance.

* DC=distorted cubic

Alm = almandine, FeAdr = Fe andradite, TiAdr = Ti andradite, Grs = grossular, Prp = pyrope, Sps = spessartine

22 Subhedral garnet from Assemblage 8. Drill hole 27-269.

16 Anhedral garnet from Assemblage 11. Drill hole 27-267

84 Euhedral to subhedral garnet from Assemblage 7, from a surface outcrop southwest of Walsh Lake.

11 Anhedral garnet with abundant anthophyllite and gedrite inclusions in Assemblage 5. 1851 stope L zone area.

10 Subhedral to anhedral garnet, in Assemblage 7.

Sample locations shown in Fig. 3.31.

and east of the mine shaft, basalt contains subhedral to euhedral ferro-tschermakite and magnesio-hornblende, following the classification of Hawthorne (1983). Hornblende crystals in moderately foliated rocks are rotated. Carbonate occurs in pressure shadows adjacent to hornblende, and cutting hornblende. This indicates hornblende growth occurred prior to, or during deformation. The first occurrence of cummingtonite (60 to 600 μm) roughly coincides with the first appearance of garnet in basaltic rocks. Randomly oriented fibrolitic cummingtonite occurs as inclusions in garnet from the deeper and eastern portions of the Campbell mine (Fig. 3.25, 3.29). Gedrite and anthophyllite (60 μm to 2 mm) occur in both hornblende-bearing and hornblende-free rocks at intermediate levels in the mine, in the area of the mine shaft, and to the north and east. Gedrite also occurs as flattened, fibrous radial aggregates, with the radial axis oriented perpendicular to the foliation. Representative amphibole compositions are provided in Table 3.5.

Staurolite: Staurolite occurs in almandine-bearing altered basalts and interflow sedimentary rocks from deeper levels (> 1140 m below ground level) in the Campbell mine, and in sedimentary rocks northwest of the mine shaft. Subhedral staurolite occurs within the rims of coarse-grained garnets, and in the necks of boudinaged garnet bands. Staurolite is moderately well aligned along the foliation.

Cordierite: Rare cordierite is identified in altered basalts of Assemblage 11. Cordierite is deformed and cut by fractures containing quartz and chlorite. Cordierite also contains fine-grained inclusions of quartz, and Fe-sulphides (Fig. 3.30). Inclusions within cordierite are finer-grained than the minerals in the matrix of the rock, and do not show a preferred mineral orientation (Fig. 3.30). This suggests cordierite growth occurred prior to deformation, and that it retains the pre-deformation fine-grained rock fabric. This observation is compatible with the time relation between growth of Mn-rich garnet cores and the occurrence of randomly oriented cummingtonite inclusions described above. Cordierite is retrograded to chlorite on its margins, and along fractures.

3.3.3 Temperature estimates from garnet-biotite thermometry

Metamorphic temperatures from the study area are evaluated by garnet-biotite thermometry. Garnet and biotite compositions were determined by SEM-EDS, and

Table 3.3. Representative biotite compositions

	12	10a	10b	7	84
SiO ₂	34.34	34.03	33.84	34.61	36.07
Al ₂ O ₃	18.19	17.71	18.33	18.44	17.40
TiO ₂	0.52	1.87	1.96	1.70	1.55
FeO	27.12	26.68	25.88	21.96	19.96
MnO	0.00	0.20	0.11	0.16	0.00
MgO	6.34	6.27	5.75	8.38	10.53
CaO	0.00	0.00	0.13	0.00	0.00
K ₂ O	9.16	9.00	9.35	9.23	9.41
TOTAL	95.67	95.76	95.35	94.48	94.92
Formula (Basis 22 oxygens)					
Si ⁺⁴	5.412	5.357	5.339	5.390	5.525
Al ⁺³	2.588	2.643	2.661	2.610	2.475
Σ T site	8.000	8.000	8.000	8.000	8.000
Al ⁺³	0.791	0.643	0.748	0.775	0.666
Ti ⁺⁴	0.061	0.222	0.232	0.199	0.179
Fe ⁺²	3.574	3.512	3.414	2.860	2.557
Mn ⁺²	0.000	0.027	0.015	0.021	0.000
Mg ⁺²	1.490	1.472	1.352	1.945	2.405
Σ Oct site	5.916	5.876	5.761	5.800	5.807
Ca ⁺²	0.000	0.000	0.022	0.000	0.000
K ⁺	1.842	1.807	1.883	1.835	1.839
Σ A site	1.842	1.807	1.905	1.835	1.839

Sample locations are shown in Fig. 3.31.

- 12 Biotite from a Assemblage 10. Drill hole S172
- 10 Brown biotite in contact with garnet, from Assemblage 7. Drill hole 14-928.
- 7 Fine-grained biotite from Assemblage 3. P zone area, 2451 stope.
- 84 Euhedral to subhedral garnet from Assemblage 7, from a surface outcrop southwest of Walsh Lake.

Table 3.4. Representative chlorite compositions

	22#1	22#2	22#3	22#4	16	11	8
SiO ₂	22.55	23.36	24.46	23.52	24.12	22.87	22.58
Al ₂ O ₃	22.55	22.90	21.86	22.26	22.58	22.30	22.24
FeO	30.79	29.02	23.29	28.01	22.80	30.29	37.17
MgO	8.99	10.86	15.17	11.63	15.81	10.01	5.85
MnO	0.00	0.20	0.28	0.38	0.20	0.15	0.00
CaO	0.18	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	85.06	86.34	85.06	85.8	85.51	85.62	87.84
Formula Basis (28 Oxygens)							
Si ⁺⁴	5.079	5.116	5.273	5.166	5.157	5.101	5.086
Al ⁺³	2.921	2.884	2.727	2.834	2.843	2.899	2.914
Σ T site	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al ⁺³	3.065	3.028	2.826	2.928	2.846	2.962	2.992
Fe ⁺²	5.800	5.316	4.198	5.146	4.077	5.650	7.004
Mg ⁺²	3.018	3.546	4.875	3.807	5.039	3.328	1.965
Mn ⁺²	0.000	0.038	0.051	0.071	0.036	0.028	0.000
Σ Oct site	11.927	11.928	11.950	11.952	11.998	11.968	11.961

Na₂O, TiO₂, and Cr₂O₃ are below detection limits, = 0.0 wt. %

Sample locations are shown in Fig. 3.31.

22#1-4 Blue birefringent chlorite from Assemblage 8. DDH 27-264

16 Chlorite from Assemblage 11.

11 Chlorite in Assemblage 5, from the north end of 1851 stope, L zone.

8 Chlorite from Assemblage 3 garnetiferous basalt, with minor biotite. 1902 E drift.

Refers to the analysis number for a separate grain in a given sample.

garnet rim compositions were used with co-existing biotite. Garnet from the study area contains significant Ca and Mn, which expand the almandine stability field and affects the Fe-Mg exchange coefficients between garnet and biotite ($K_D^{\text{Gr-Bt}}$, hereafter referred to as K_D). This effect has been evaluated by numerous workers (e.g., Hodges and Spear, 1982; Koziol, 1990; Berman and Aranovich, 1996). Mixing of Ca in garnet has been shown to be nearly ideal for grossular-almandine binary mixtures (Koziol, 1990; Berman and Aranovich, 1996), particularly for almandine-rich garnet. Grossular-spessartine mixing is ideal to nearly ideal (Berman 1990; Gavrieli et al., 1996), and mixing of grossular-pyrope is non-ideal (Ganguly and Saxena 1984, Hoinkes 1986, Berman 1990). It is generally agreed that Mn mixes ideally within garnet (Hodges and Spear, 1982; Kretz, 1990; Koziol, 1990, Berman, 1990), although some workers have suggested non-ideal interaction (Ganguly and Saxena, 1984; Williams and Grambling, 1990). Recent work by Ganguly et al. (1996) indicates non-ideal mixing between spessartine and pyrope.

Using empirical data, Alcock (1996) concluded that temperature correction for Ca in garnet is unnecessary for $\leq 14.0\%$ X_{Grs} . Garnets examined by Alcock (1996) also contain 0.05 X_{SpS} , suggesting that a correction for small X_{SpS} is not required. This is consistent with the conclusion of Kretz (1990), that K_D is essentially unaffected by X_{SpS} up to 0.12%, and is further supported by Williams and Grambling (1990, their Fig. 2.6) for $X_{\text{SpS}} \leq 0.10$, and by Hoinkes (1986, and Table 4 therein) for $X_{\text{SpS}} \leq 0.4$. These data suggest that a K_D correction for Mn and Ca may be small for garnets which contain $\leq 0.10 X_{\text{SpS}}$ and $\leq 0.14\% X_{\text{Grs}}$. All garnet grains used for temperature calculations here contain $\leq 0.16 X_{\text{Grs}}$, with only two analyses $> 0.10 X_{\text{Grs}}$. This suggests that the Fe-Mg thermometers may provide reasonable temperature estimates for garnet-biotite pairs in this study.

Metamorphic temperatures are calculated with the experimental Fe-Mg exchange thermometers of Ferry and Spear (1978), and Perchuk and Laverent'eva (1981), with results presented in Table 3.6. Kretz (1990) has reconciled significant differences in these calibrations, and results are presented in Table 3.6. The thermometer of Ganguly and Saxena (1984) is applied to account for the grossular and spessartine content in

Table 3.5. Representative metamorphic amphibole compositions

	12#14 *	49#1	49#2	95a#9	22#9	22#25
SiO ₂	38.05	46.23	45.85	40.09	53.67	52.51
Al ₂ O ₃	18.57	9.85	11.00	18.89	0.34	1.14
TiO ₂	0.00	0.28	0.28	0.15	0.00	0.00
Fe ₂ O ₃	10.00	5.56	6.79	16.99	28.30	28.63
FeO	15.26	10.01	9.44	9.54	0.00	0.00
MnO	0.14	0.39	0.44	0.61	0.98	1.36
MgO	2.88	12.44	11.82	3.12	16.23	14.75
CaO	10.41	11.90	11.84	9.41	0.00	0.31
Na ₂ O	1.03	1.64	1.34	0.00	0.00	0.00
K ₂ O	0.33	0.22	0.25	0.27	0.00	0.00
TOTAL	96.67	98.52	99.05	99.07	99.52	98.70
Formula Basis (23 Oxygens)						
Si ⁺⁴	5.843	6.717	6.622	5.896	7.485	7.418
Al ⁺³	2.157	1.283	1.378	2.104	0.056	0.189
Σ T site	8.000	8.000	8.000	8.000	7.541	7.607
Al ⁺³	1.205	0.404	0.494	1.171	0.000	0.000
Ti ⁺⁴	0.000	0.031	0.030	0.017	0.000	0.000
Fe ⁺³	1.156	0.608	0.738	1.880	2.970	3.043
Fe ⁺²	1.960	1.216	1.141	1.173	0.000	0.000
Mn ⁺²	0.018	0.048	0.054	0.076	0.116	0.163
Mg ⁺²	0.660	2.695	2.545	0.683	3.375	3.105
Σ C site	4.999	5.002	5.002	5.000	6.461	6.311
Ca ⁺²	1.713	1.853	1.832	1.483	0.000	0.048
Na ⁺	0.287	0.147	0.168	0.000	0.000	0.000
Σ B site	2.000	2.000	2.000	1.483	0.000	0.048
Na ⁺	0.020	0.315	0.207	0.000	0.000	0.000
K ⁺	0.066	0.041	0.047	0.051	0.000	0.000
Σ A site	0.086	0.356	0.254	0.051	0.000	0.000

Cr₂O₃ is below detection limit, = 0.0 wt. %.

* Fe⁺³ by charge balance.

Sample locations are shown in Fig. 3.31.

12#14 Ferro-tschermakite in Assemblage 10. Amphibole is blue-green, anhedral, and comprises < 3 vol. % of the rock, with minor retrogression to chlorite. Drill core S172.

49#1, 2 Magnesio-hornblende in fine-grained basalt of Assemblage 6.

95a#9 Ferro-tschermakite in Assemblage 5.

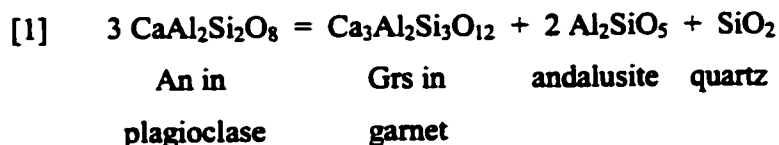
22#9, 25 Cumingtonite included in garnet in assemblage 8. From drill core below the 27th level

Refers to the analysis number of an individual grain in a given sample.

garnet (Table 3.6). The thermometer of Williams and Grambling (1990), which considers non-ideal Mn mixing in garnet, and the Ca correction of Hoinkes (1986) applied to the Ferry and Spear (1978) thermometer, are provided for comparison (Table 3.6). All thermometers provide comparable temperature estimates, except that of Williams and Grambling (1990). The Williams and Grambling (1990) thermometer yielded significantly higher temperatures, probably due to their proposed compensation for non-ideal mixing of Mn in Fe-Mg garnet. It is assumed here that the effect of Mn and Ca on K_D , for the samples used in this study, is minor. This paper uses metamorphic temperatures obtained from the thermometer of Kretz (1990), provided in column 3 of Table 3.6. The results show variability in temperature up to 25 °C (Sample 12a and b, Table 3.6), which is within the calculated error range of the thermometer ($\approx \pm 30$ °C). Sample locations for the temperature estimates are provided on Fig. 3.31.

3.3.4 Metamorphic pressure estimate from GASP barometry

The presence of andalusite and the lack of sillimanite and kyanite indicate that metamorphic pressures did not exceed 3.7 kilobars. The presence of cordierite also indicates low pressure metamorphic conditions. Andalusite was commonly observed, but was only observed in a single sample with garnet and plagioclase. This assemblage indicates the reaction;



which is the basis for GASP (garnet + aluminosilicate + quartz + plagioclase) barometry. The univariant curve for reaction [1] is calculated using the TWQ 1.02 program of Berman (1991), incorporating the thermodynamic data of Berman (1988), the quaternary garnet model of Berman (1990), and the ternary plagioclase mixing model of Furhman and Lindsey (1988). Reaction [1] is plotted on Fig. 3.32 with the Fe-Mg exchange reaction between garnet and biotite, and pressure estimate of 3 kilobars at 550 °C is obtained (Fig. 3.32). This determination has an error of $\approx \pm 1$ kilobar (Berman, 1991).

Table 3.6. Temperatures (°C) obtained from garnet - biotite thermometry

Sample*	LnK _D	1 [†]	2	3	4	5	6
84	1.67	589	581	601	596	687	600
12a	1.98	492	526	524	559	580	494
12b	1.86	526	546	551	589	614	533
10	1.97	496	528	527	581	657	469
7a	2.17	444	497	483	514	611	357
7b	2.20	436	492	477	506	602	351
22	1.87	525	546	551	565	661	524
16	1.91	512	538	540	527	663	524

1 Ferry and Spear (1978).

2 Perchuk and Laverent'eva (1981).

3 Kretz (1990), calculated using Fe⁺² determined by stoichiometry.

4 Ganguly and Saxena (1984), calculated using the GPT spreadsheet of Reche and Martinez (1996).

5 Williams and Grambling (1990), with corrections in American Mineralogist 1992. v. 77 p. 188.

6 Hoinkes (1986), applied to the Ferry and Spear (1978) thermometer.

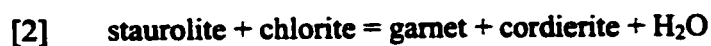
a and b refer to separate pairs from a single hand specimen. Sample locations are provided on Fig. 3.31.

Sillimanite has not been observed in rocks of the northeastern part of the study area. The lack of sillimanite suggests either an over-estimate of temperature or pressure, or that the single pressure estimate may not truly represent pressure conditions for the area.

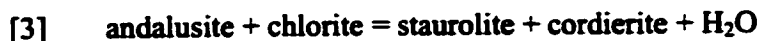
Nevertheless, the low metamorphic pressure determined with this barometer is consistent with the occurrence of andalusite and cordierite in the study area. The author considers the estimate by GASP as a maximum pressure for the study area, given the absence of sillimanite in rocks at the Goldcorp mine, and to the NE.

3.3.5 Temperature and pressure estimates from mineral assemblages

Pressure and temperature conditions were obtained for prograde mineral assemblages in Table 3.1, with two petrogenetic grids; the Spear and Cheney (1989) grid (Fig. 3.33) for rocks with a pelitic composition, and the grid of Spear (1993) for rocks containing quartz, hornblende, and plagioclase, and lacking a potassic phase (Fig. 3.34). The pelitic grid was calculated using the thermodynamic data of Berman (1988), and Fe/Mg exchange between minerals. This grid is appropriate for rocks in this study, as it predicts chloritoid stability within the andalusite field. It also uses an aluminosilicate triple point similar in position to that determined by Holdaway (1971), and more recently confirmed by Holdaway and Mukhopadhyay (1993). The Spear (1993) grid is largely empirical, and constructed for cordierite - anthophyllite rocks. The reactions on the grid for rocks with low Al bulk compositions are applicable to some rocks in the study area. Because the grid of Spear (1993) is empirical, the position of the univariant curves are subject to uncertainty. For this reason, the reaction;



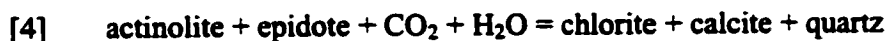
is calculated using the TWQ 1.02 program of Berman (1991). Based on the result, the low pressure and temperature portion of the grid is modified, with a position for the univariant curves intermediate to that presented in Spear and Rumble (1986) and Spear (1993). It also incorporates a change to the position of the univariant curve for the reaction;



as suggested by F.S. Spear (written communication, May 8, 1997).

Carbon dioxide is a significant component in auriferous fluids from Archean gold deposits. As petrogenetic grids are commonly projected from a pure H₂O fluid, its effect on metamorphic mineral equilibria is considered here by two methods. Qualitatively, low fluid XCO₂ is indicated if there is reasonable agreement between temperature determined from the petrogenetic grid, and from fluid absent exchange reactions, such as the Fe-Mg garnet-biotite exchange reaction. An assemblage from relatively unaltered rocks is also evaluated with a T-XCO₂ diagram, calculated using the TWQ 1.02 program of Berman (1991).

Assemblage 2 rocks contain fine-grained actinolite, indicating greenschist facies metamorphism in the western portions of the study area. Basaltic rocks containing Assemblage 1 indicate the reaction;



has proceeded, and is confirmed by the presence of carbonate overgrowing actinolite needles in some Assemblage 2 rocks. Assemblage 1 is also indicative of the greenschist facies, and represents relatively high fluid XCO₂.

Garnet first appears in basalts and interflow sedimentary rocks in a chloritic assemblage (Assemblage 3) in the eastern portion of the mine property, and contain variable X_{Gr}s (0.00-10.81 mole%) and X_{Sp}s (11.28-17.06 mole%). Garnet was likely formed under upper-greenschist facies or lowermost-amphibolite facies conditions, facilitated *locally* by rock compositions with high Mn and Ca contents. The X_{Sp}s content of garnets in Assemblage 3 suggests temperatures ≥ 460 °C to 475 °C, as illustrated in Fig. 3.33. Cumingtonite is associated with the first appearance of garnet in the P zone area, in garnet and chlorite bearing, muscovite- and biotite-free rocks (Assemblage 4). The grid of Spear (1993) yields a metamorphic temperature of < 475 °C to 480 °C (Fig. 3.34). This is consistent with a metamorphic temperature of ≈ 480 °C (Table 3.6,

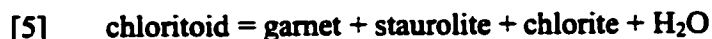
samples 7#2a and 7#2b) determined from garnet-biotite thermometry, and is compatible with the approximate position of the greenschist / amphibolite transition.

Basalts east of the mine shaft in the underground workings contain Assemblages 5 and 6. The occurrence of tschermakitic amphibole and magnesio-hornblende clearly indicate these rocks have experienced amphibolite facies metamorphism. Garnets from Assemblage 5 contain lower X_{SpS} (≤ 11.28 mole%) than garnets from Assemblage 3. Assemblage 6 rocks occur at intermediate levels in the mine, and contain gedrite, anthophyllite, garnet, chlorite, and ferro-tschermakite with sporadic biotite. This assemblage indicates temperatures of > 475 °C to 480 °C, and between 1.5 and 3.7 kilobars pressure (Fig. 3.33). Further, for rocks of Assemblage 5, lacking hornblende, a temperature of ≈ 495 °C at 3 kilobars pressure, and X_{CO_2} for the metamorphic fluid of ≈ 0.025 is determined (Fig. 3.35). This indicates that rocks at intermediate levels in the Campbell mine workings, east and north of the shaft, have been subjected to metamorphic temperatures of ≥ 475 °C. These rocks are relatively unaltered compared to rocks adjacent to the ore bodies, which may explain the low fluid X_{CO_2} value. The P-T estimates are comparable to those obtained from the central part of the mine.

Interflow sedimentary rocks and moderately altered basalts from underground drill cores (> 1240 m below surface) at the Campbell mine, contain Assemblages 8 and 9. Chloritoid forms subhedral to anhedral laths along the foliation in Assemblage 8 rocks, staurolite is developed as euhedral crystals, and biotite and chlorite occur with plagioclase in the matrix. Subhedral to anhedral garnet porphyroblasts are abundant, commonly comprising 5 vol.% of the rock.

Randomly oriented cummingtonite inclusions occur within garnet cores, in a rock with a strong S_2 foliation. This indicates garnet core growth occurred prior to development of the foliation. Assemblage 9 occurs within boudin necks of garnet bands (Fig. 3.25), and was developed during the S_2 parallel extension. The margins of the garnet, outside of the boudin necks, are deformed. Euhedral growth faces occur on garnet within boudin necks. This indicates Fe + Mg rich garnet rim growth occurred during shortening. It could not have pre-dated shortening, as boudin necks were developed during D_2 shortening which produced the foliation. It is also unlikely that Fe + Mg-rich garnet growth post-dated the shortening, as euhedral growth faces would not

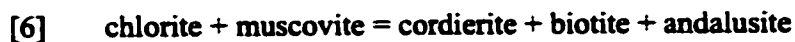
be preferentially developed in boudin necks. The presence of euhedral crystal faces on garnet in pressure shadows which contain chlorite and staurolite, but not chloritoid, and the occurrence of staurolite inclusions in garnet rims, suggest chloritoid decomposition to form garnet and staurolite by the reaction:



The presence of staurolite as inclusions in almandine rich garnet rims indicates staurolite formation during garnet growth, and the reaction proceeded with the growth of almandine and pyrope rich garnet during the development of Assemblage 9. Reaction 5 occurs at < 3 kilobars pressure, and between approximately 520 °C and 540 °C (Fig. 3.33).

The occurrence of cummingtonite in garnet cores, and its absence in the matrix, indicates the loss of cummingtonite either during metamorphic reaction, or hydrothermal alteration.

A representative sample containing Assemblage 9 yields a temperature of 550 °C, from garnet-biotite thermometry (sample 22, Table 3.6), consistent with another sample in the area which provided a temperature estimate of 540 °C (sample 16, Table 3.6). This estimate is consistent with the textural relationship between Assemblages 8 and 9 which indicates a prograde metamorphic path in the area. An increase in X_{Alm} and X_{Pyp} , with decreasing X_{Sps} and X_{Grs} , and $\text{Fe}/(\text{Fe}+\text{Mg})$ towards garnet margins (Fig. 3.23) also indicates prograde garnet growth. Assemblage 11 rocks, containing rare cordierite, suggest the reaction;



which takes place at ≈ 530 °C to 565 °C, and ≤ 2.7 kilobars pressure (Fig. 3.33). Results from GASP barometry (Fig. 3.32) from this assemblage also indicate low metamorphic pressure (≈ 3 kilobars), and the metamorphic temperature is also consistent with that determined for other rocks in the area.

Sedimentary rocks (Assemblage 10) and altered basalt (Assemblage 6) east and northeast of the Campbell mine shaft contain staurolite and biotite, and euhedral garnet.

Ferro-tschermakite in Assemblage 6 is subhedral, included within garnet, and weakly altered to biotite along grain margins. Assemblage 10 rocks contain cummingtonite and chlorite. A temperature of ≈ 530 °C to 540 °C was obtained for these rocks (Table 3.6, samples 10 and 12a, 12b for Assemblage 10 and 6, respectively) using garnet-biotite thermometry.

Mathieson and Hodgson (1984) have estimated metamorphic pressure and temperature at 520 °C to 540 °C, and 3.8 to 4.2 kilobars, for rocks in the Goldcorp Red Lake mine. Mineral assemblages provided in Table 2 of Mathieson and Hodgson (1984) are applied to the petrogenetic grids used in this study. The assemblages of chlorite + biotite + andalusite + staurolite + cordierite + plagioclase + quartz (sample 157-03) indicates reaction [6], corresponding to ≤ 2.7 kilobars, and 530 °C to 565 °C (Fig. 3.33). The assemblage, andalusite + biotite + chlorite + plagioclase + quartz (Sample 67-06) indicates the reaction;



and suggests that metamorphic pressure was ≤ 2.5 kilobars at ≥ 530 °C to 565 °C (Fig. 3.33). The assemblage, chlorite + andalusite + staurolite + cordierite + plagioclase + quartz + hornblende (Sample 157-05) in muscovite and biotite free rocks corresponds to the univariant curve for reaction [3]. This reaction occurs at ≤ 500 °C (Fig. 3.34), although the position of the curve is not well constrained. The occurrence of chloritoid in many assemblages indicates that the terminal chloritoid reaction (reaction [5], Fig. 3.33) has not occurred. Sillimanite was not observed Mathieson and Hodgson (1984) during their study of the Goldcorp mine. The silicate assemblages at the Goldcorp mine indicate metamorphic conditions of < 540 °C to ≥ 565 °C, and ≤ 3 kilobars pressure.

Assemblage 7 occurs in intermediate tuffs and basaltic rocks north and east of the Campbell mine. Chlorite is a minor component in these rocks (< 10 vol.%), and amphiboles are anthophyllite to gedrite in composition. Garnet is more Fe + Mg-rich than garnets from the Campbell mine. The loss of chlorite, and occurrence of almandine and pyrope-rich garnet compositions (Table 3.2) reflect higher metamorphic temperatures, consistent with results from garnet-biotite thermometry (≈ 600 °C, sample

84, Table 3.6). Sillimanite was not observed in clastic sedimentary rocks in the area of sample 84 (Fig. 3.31). This may suggest that the metamorphic pressure determined by GASP barometry may slightly over-estimate the true pressure.

3.3.6 Metamorphic temperature gradient

The mineral assemblages and results from garnet-biotite thermometry show a significant thermal gradient at low (≤ 3 kilobars) pressure across the study area. Temperature ranges from $< 460^\circ\text{C}$ in greenschist domains west of the Campbell mine shafts, 460°C to 480°C in greenschist to amphibolite facies transitional domains in the Campbell mine shafts area, to approximately 475°C to 565°C (Campbell mine west boundary and Goldcorp mine), and $\approx 600^\circ\text{C}$ in the eastern area proximal to the Trout Lake batholith (Fig. 3.33).

3.3.7 The garnet isograd

The occurrence of garnet was selected to map the thermal gradient in the area, as it is coarse-grained and easily identified in hand-specimen, unlike other fine-grained metamorphic minerals. The occurrence of garnet was mapped from surface and underground exposures in the Campbell mine, and from drill log data of surface and underground diamond drill cores (Figs. 3.31 and 3.36).

The occurrence of garnet is a function of bulk rock composition and temperature. The occurrence of garnet must be in part controlled by heterogeneous composition of the rocks, as it is recognized in thin chloritic pillow selvages. However, the bulk rock composition of relatively fresh basalt in the mine area is quite uniform (Table 3.0), aluminous altered basalt occurs sporadically in the basalt sequence throughout the Campbell and Goldcorp mine areas, and pillow basalt is ubiquitous at the mine. The temperature gradient within the mine area is steep, suggesting that the first occurrence of garnet provides a reasonable approximation for an isograd.

The isograd is parallel to the margins of the Trout Lake batholith in the north, and is extended westward along the Cochenour-Gullrock Lake deformation zone in the south (Fig. 3.31). Broadly, the garnet isograd appears to be parallel to the southwestern limb of the Balmer synform. However, the axial trace of the McDougal Lake Antiform, and

folded strata in the northeast portion of the study area are cross cut by the garnet isograd (Fig. 3.31).

The garnet isograd on Campbell mine level maps clearly cuts litho-stratigraphic boundaries (Fig. 3.36). Garnetiferous rocks are associated with the main northwest striking A, F, and NL auriferous veins (Fig. 3.36), which also cross cuts lithologic units. Basaltic rocks containing abundant, large garnet (> 3 vol.%, up to 1 cm in size) are spatially associated with the L ore zone.

The position of the garnet isograd on the 14, 15, 17, 20, and 23 levels of the mine ($\approx$ 600 to 1200 m below surface) are summarized in Fig. 3.37. Sporadic garnet is observed on the 14th level near the northeast limits of the underground workings, and there is a strong spatial association between garnet occurrences and abundance, and the auriferous veins. Between the 15th and 17th levels, the isograd dips moderately westward, and is extended along the foliation-parallel A and F auriferous veins. Garnetiferous basalt occurs northeast and east of the mine shaft on the mine property, at and below the 15th level. Between the 17th and 23rd levels (> 300 m below the surface), garnet-bearing basalts are strongly attenuated along the A and F auriferous veins, and the boundary between garnet-bearing and garnet-free basalt is near vertical (Fig. 3.37). Sporadic garnet is also present in interflow sedimentary rocks adjacent to the G ore zone. At, and below the 26th level, garnet is ubiquitous throughout the mine workings, which extend slightly west of the P ore zone on the 23rd level (Fig. 3.37).

The absence of substantial later deformation and tilting (Zhang et al., 1997) suggests that the orientation of the isograd has not been significantly modified since the peak of metamorphism. The vertical garnet isograd, therefore, suggests a horizontal thermal gradient, with temperature increasing eastward.

Figure 3.0

Geological plan map of the Campbell (14th level) and Goldcorp (15th level) mines on the grid of the Campbell mine (in feet), showing the locations of auriferous vein systems (labeled by ore zone). The volcanic rocks are clearly folded about NW trending axial traces. Modified after MacGeehan and Hodgson (1982), with additional data compiled by the author from Campbell mine geology plans and drill logs. Serpentinite rocks at the Goldcorp mine are intrusive rocks.

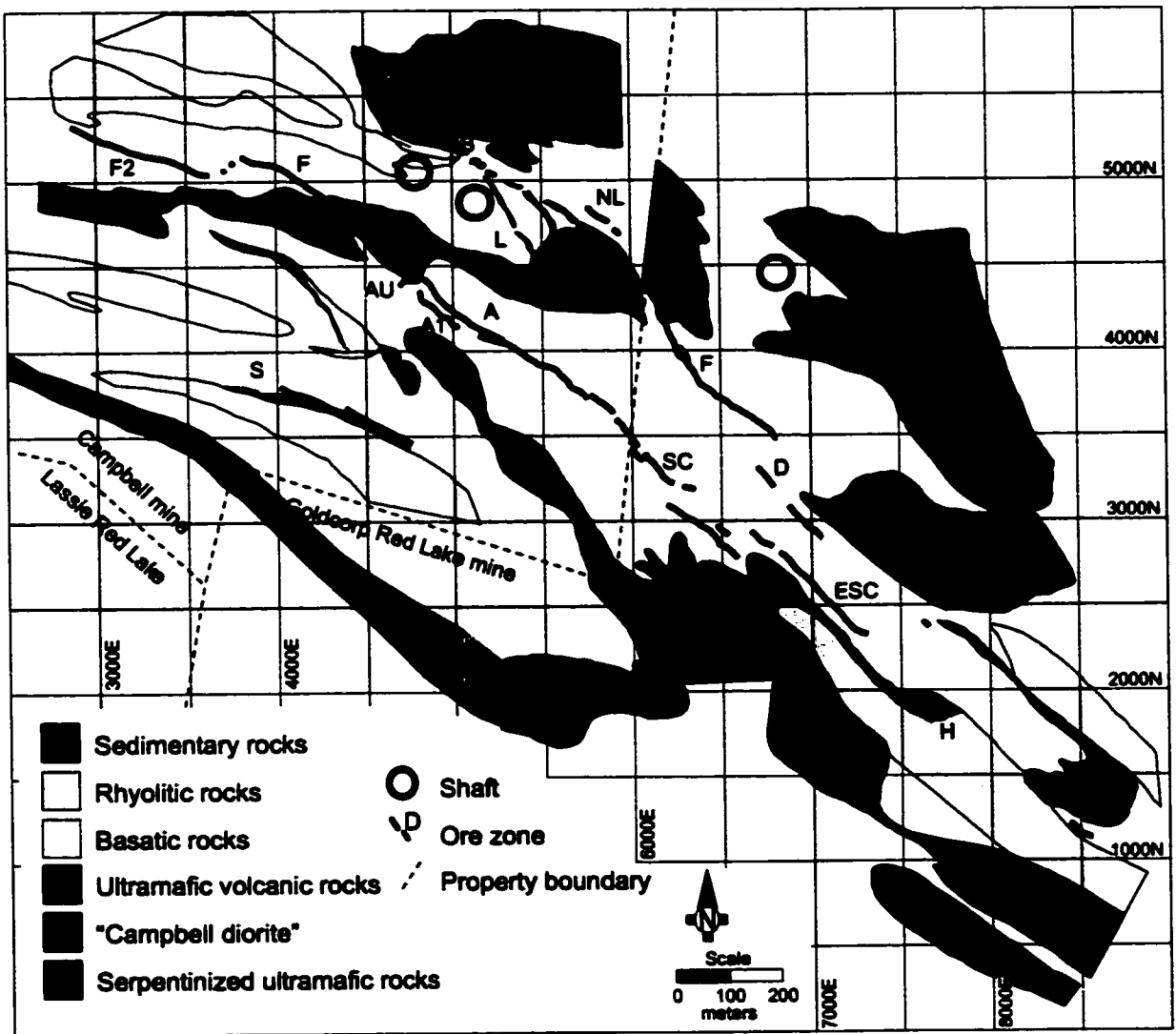


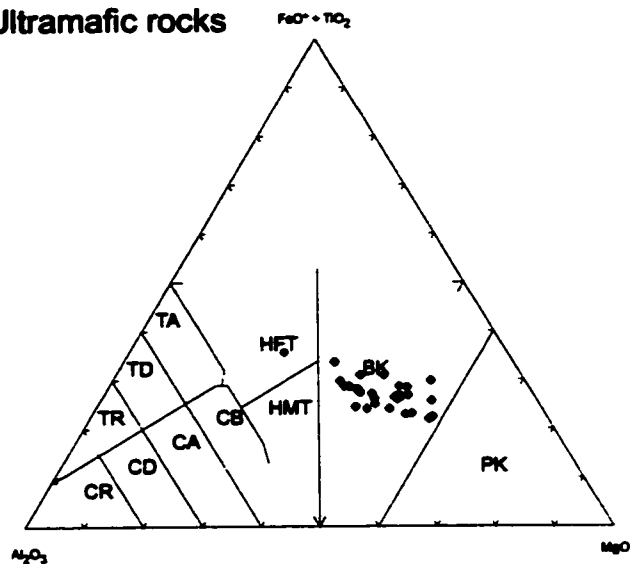
Figure 3.1

Ternary cation diagram of ultramafic rocks (A), “chloritic” ultramafic rocks (B), and basalt (C). Ultramafic rocks and “chloritic” ultramafic rocks, including brecciated units, plot in the field of basaltic komatiites (BK field). Basalt compositions lie predominantly in the field of high Fe and Mg tholeiites (HFT and MFT). Increasingly altered basalt compositions plot towards the Al_2O_3 apex, resulting in apparent andesitic to tholeiitic rhyolite compositions. Diagram modified from Jensen (1976).

Abbreviations: BK=basaltic komatiite, CA= calc-alkaline andesite, CB=calc-alkaline basalt, CD= calc-alkaline dacite, CR= calc-alkaline rhyolite, HFT=high iron tholeiite, MFT=high magnesian tholeiite, PK=peridotitic komatiite, TA= tholeiite andesite, TD= tholeiite dacite, TR= tholeiite rhyolite.

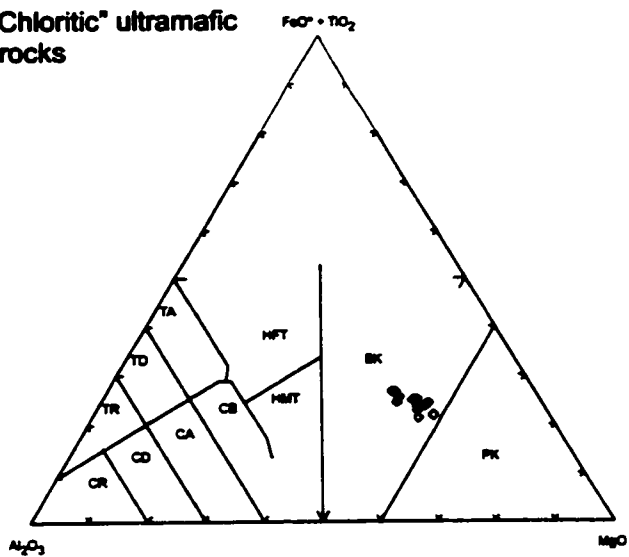
A

Ultramafic rocks



B

"Chloritic" ultramafic rocks



C

Basalt

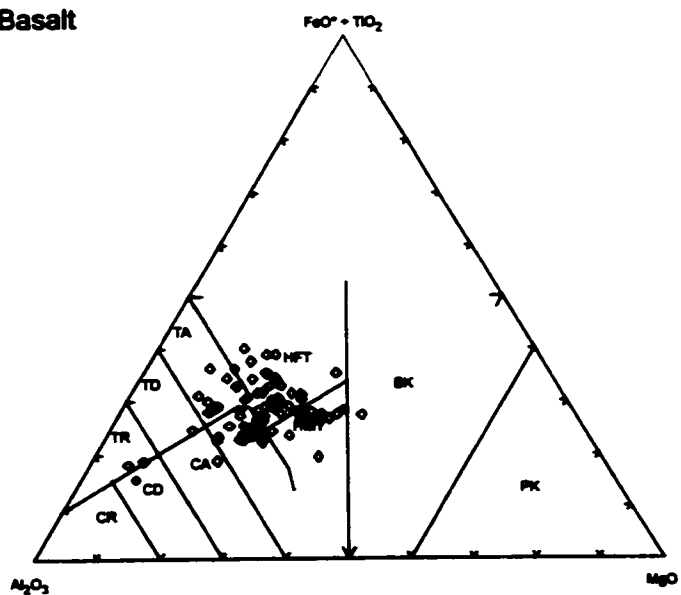


Figure 3.2

Photograph of brecciated ultramafic rock, with a matrix of grey cherty quartz. The breccia is cut by a series of sub-parallel “sheeted” carbonate veinlets (inclined to the right in the photograph). Location: 1854 stope. Screen is approximately 15 cm square.

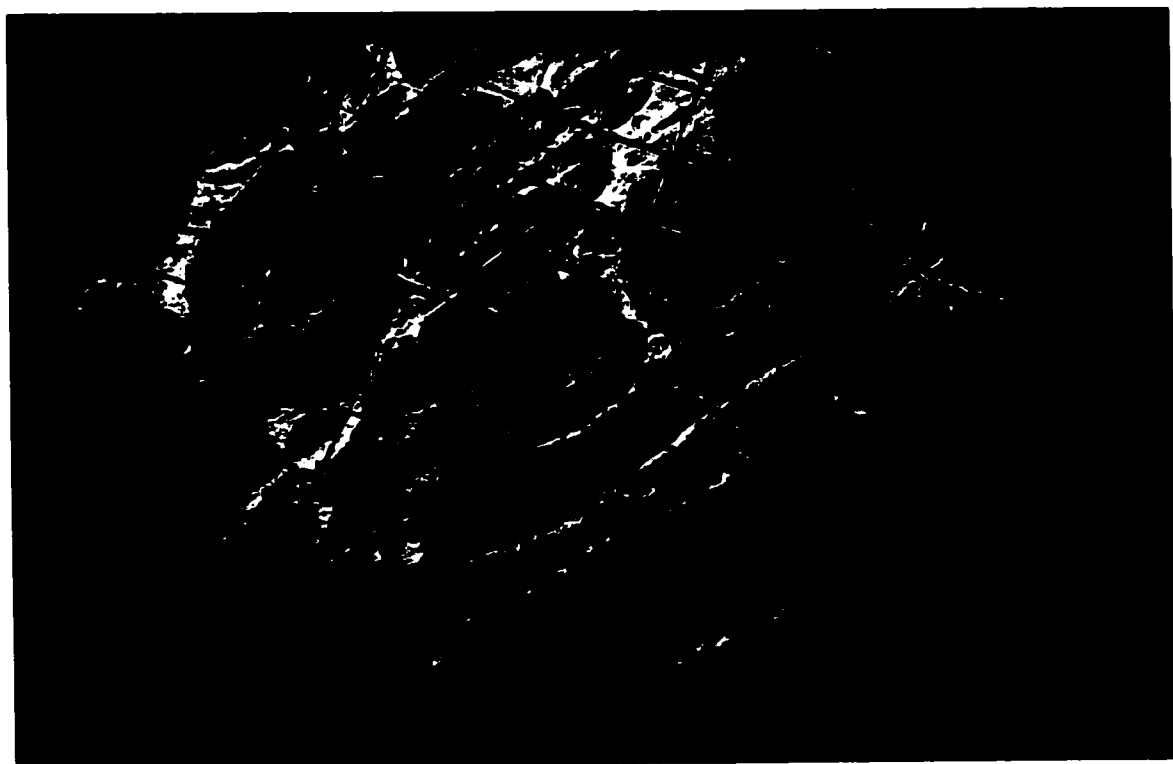


Figure 3.3

Cream-yellow rhyolite cut by dark anastomosing cleavage seams composed of chlorite and tourmaline, with minor anorthite, pyrite, and carbonate. The cleavage seams are subparallel to the foliation in the mafic and ultramafic rocks, suggesting they formed during NE-SW shortening. The occurrence of alteration minerals along the cleavage suggests hydrothermal activity during or after cleavage development. Photo of drift wall, screen is approximately 15 cm square.

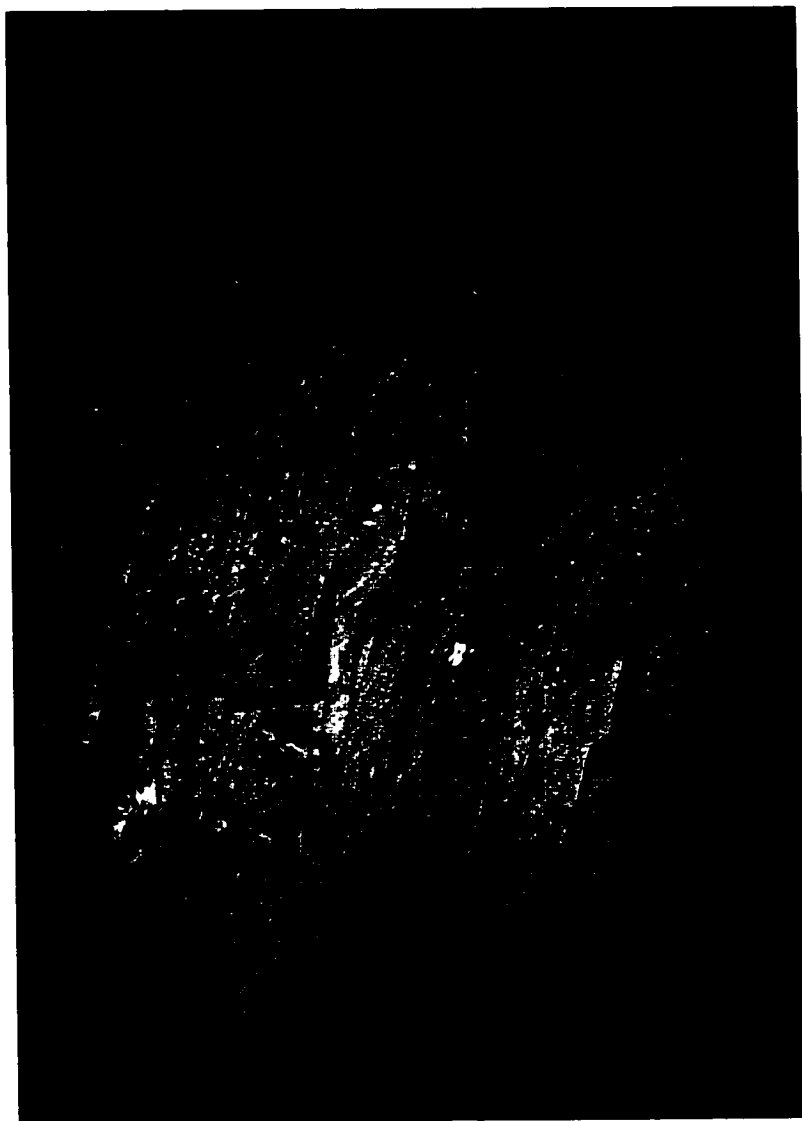
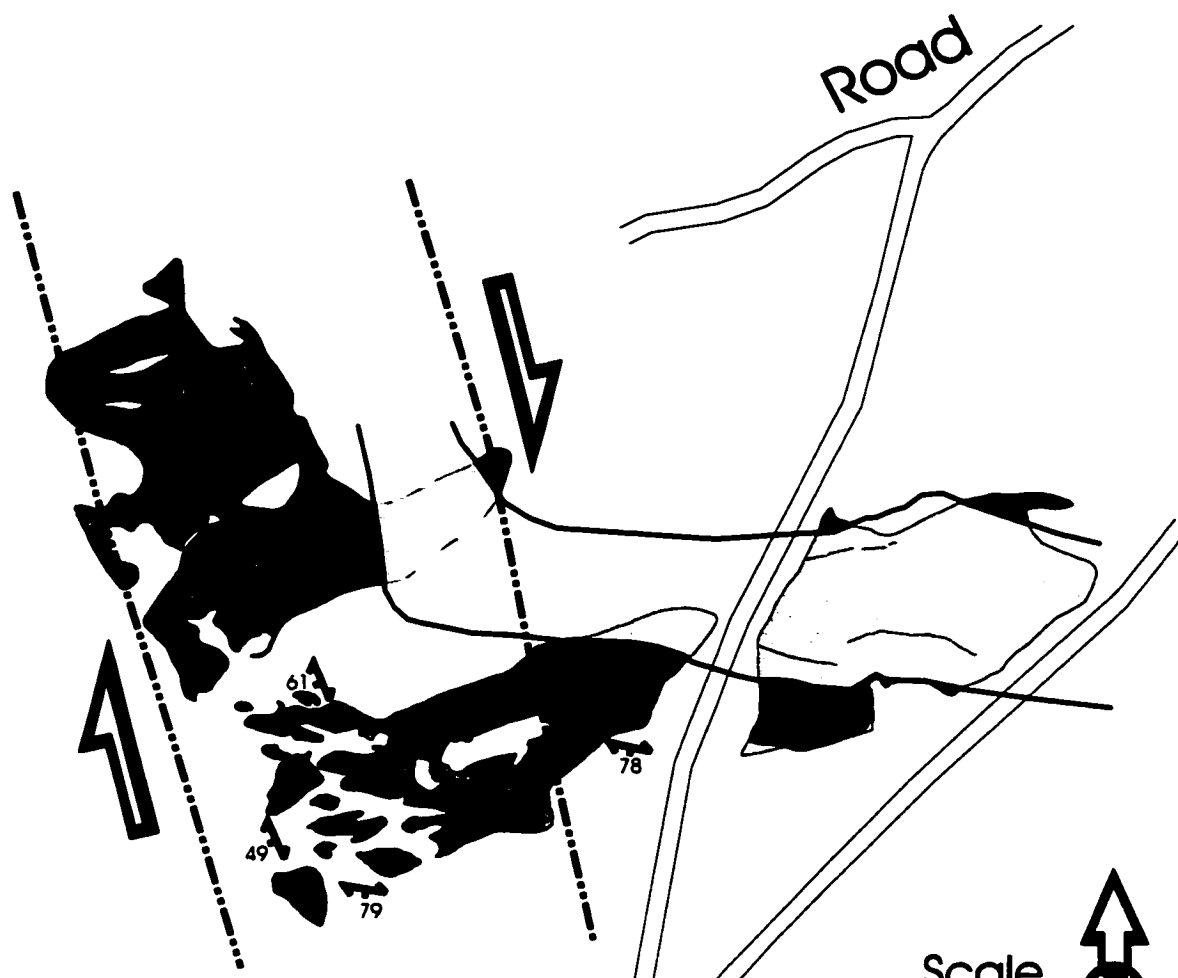


Figure 3.4

Surface plan map of the area immediately north of the G zone ("G zone extension outcrop"), Campbell mine. The up-dip projection of the G ore zone is approximately 30 m to the south of the map area. Foliation planes, quartz + feldspar porphyry dykes, and quartz + carbonate veins undergo rotation into the shear zone which hosts the G orebody. Strain intensity, inferred from the intensity of foliation development, also increases into the shear zone. The sense of rotation of veins and foliations indicates dextral shear operated along the shear zone hosting the G ore body. The lamprophyre dykes cut the quartz + feldspar porphyry dyke.



-  Mafic/Ultramafic volcanic rocks
-  Rhyolitic rocks
-  Quartz-feldspar porphyry dyke
-  Lamprophyre dyke
-  Quartz-carbonate veins
-  Geologic contact
-  Shear zone boundary (approx.)

Scale
0 10
Meters



Figure 3.5

Time relations between vein and quartz + feldspar porphyry dyke emplacement. A) Photograph of coarse-grained quartz veinlets in a deformed quartz + feldspar porphyry dyke adjacent to the G zone shear structure. B) Photograph of quartz+carbonate vein cutting a quartz + feldspar porphyry dyke.

These observations indicate that quartz and quartz + carbonate veining occurred late or after quartz + feldspar porphyry dyke emplacement.

Location: G zone extension outcrop.

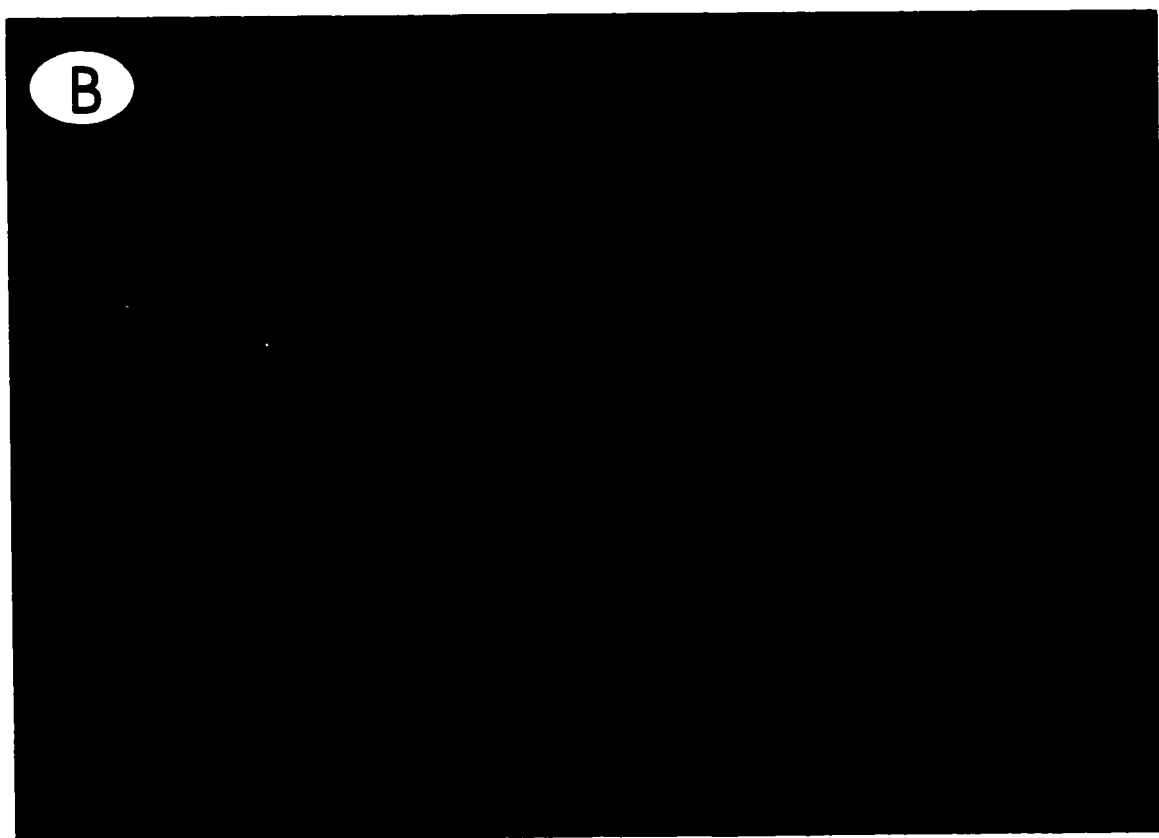


Figure 3.6

Photograph showing weak carbonate alteration (brown streaks) in a deformed quartz + feldspar porphyry dyke adjacent to the G zone shear structure, G zone extension outcrop. Viewing subvertical face.

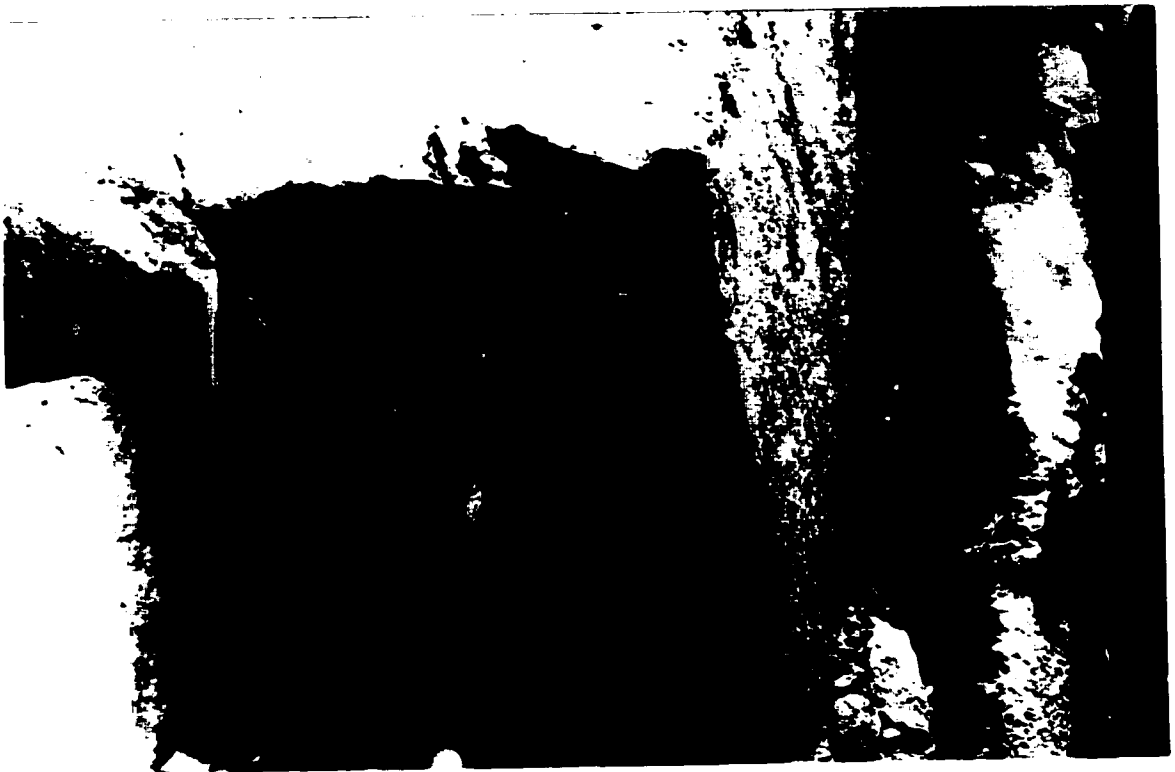


Figure 3.7

Thinly spaced foliation composed of biotite and white mica, in a quartz + feldspar porphyry dyke, adjacent to the shear zone which hosts the G ore body, G zone extension outcrop.



Figure 3.8

Strongly sheared and tectonically thinned quartz + feldspar porphyry dyke (left of the compass) within the G zone shear structure at the G zone extension outcrop. The dyke outside the shear zone is approximately 1 m wide. Where the dyke transects the shear zone which hosts the G ore body, it is rotated, thinned (< 15 cm wide), and contains a well-developed foliation.



Figure 3.9

Stereonet showing the poles (●) to the contact between the mafic and ultramafic rocks in the L and NL ore zone areas. Fold limbs are shown as dashed great circle planes, whose poles correspond to the extreme ends of the bedding pole data distribution. The pole data correspond to a strong girdle (best-fit great circle shown as heavy black line) on an eigenvalue plot (shape parameter 0.50, strength parameter 4.35). The pole to the best-fit great circle (■) is $218^{\circ}/58^{\circ}$. The data correspond to a tight reclined fold with a NW striking axial plane, as discussed in the text. Data obtained from the 14th through 18th levels in the mine, from stope and level mapping by the author and Campbell mine geology staff. Statistics calculated with the stereonet program GEOorient 7.2 (Halcombe, R.J., 1998). Lower hemisphere, equal area projection. N = 29 poles.

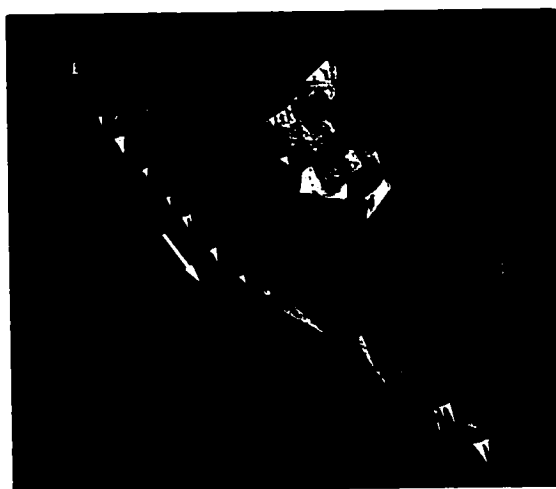
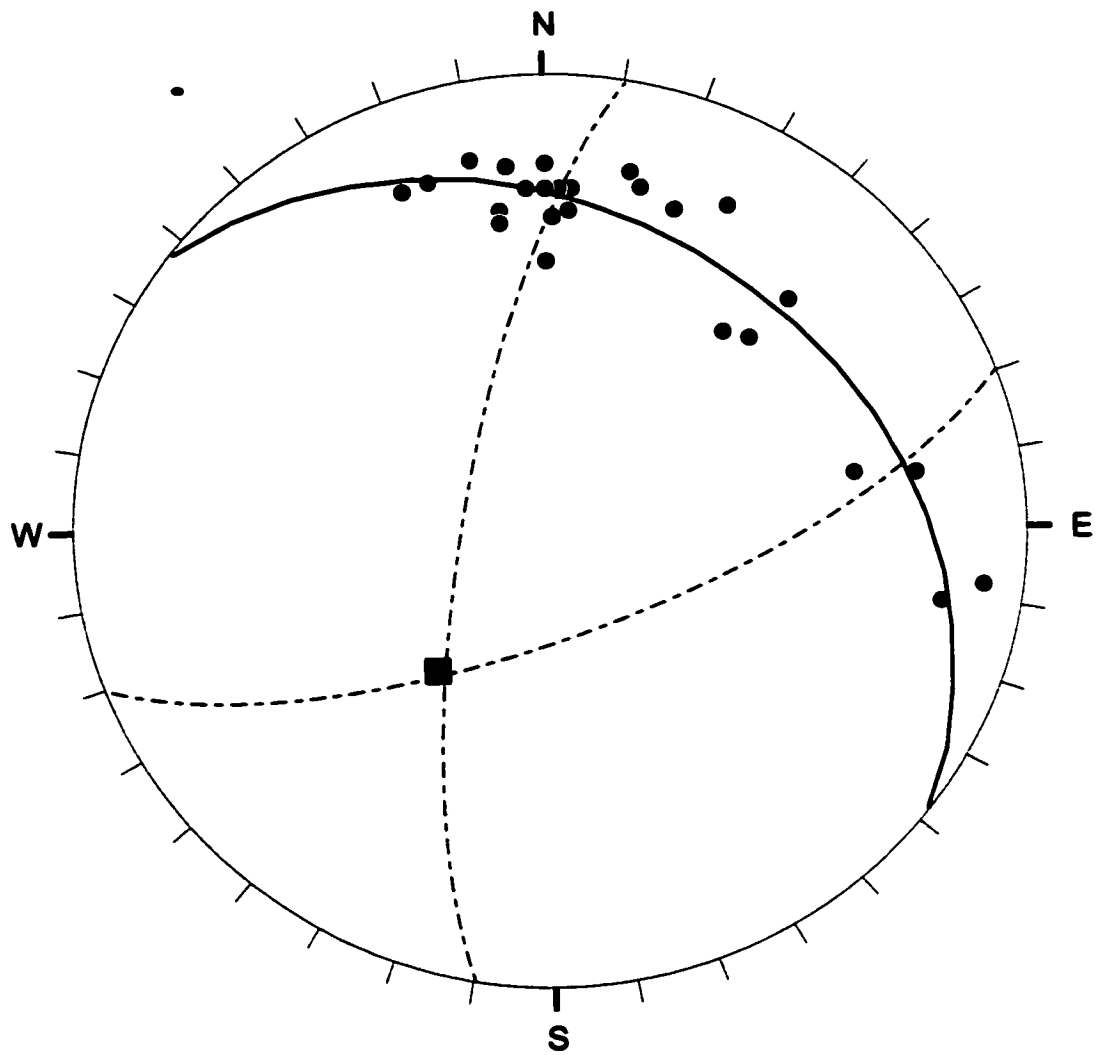


Figure 3.9.1

Schematic diagram showing the deformation history at the Campbell mine. The location of the schematic diagram is shown as the light box on the geology map, lower right. D_1 (A) developed NE trending folds and weak S_1 foliations. During D_2 (B), F_1 and S_1 were refolded about F_2 folds with NW trending axial planes (axial trace shown as heavy solid line). NW trending shear zones hosting auriferous veins (ESC, A, H, and ore zones shown in red on the diagram) were developed along attenuated F_2 limbs. Conjugate shear zones at mafic / ultramafic rock contacts, near F_2 hinges, host auriferous veins (G and L ore zones shown in red on the diagram). A schematic diagram (lower right) shows the relations between folded lithologic contacts, the shear zones and the auriferous veins near the F_2 hinges. These shear zones may have developed by flexural slip along lithologic contacts during F_2 folding. During late D_2 (C) shortening, shear zones (fine dashed lines) transpose highly attenuated fold limbs, forming hook-shaped folds. Auriferous veins were also deformed (B ore zone, shown in red) during late D_2 shortening.

Figure 3.10

Stereonet showing the poles to foliation planes (■) from throughout the Campbell mine, and the mean foliation plane (black great circle). The best-fit great circle to the poles is shown as the dashed great circle, and the pole to the plane is the grey square. The axis of NW trending F_2 folds in the mine, determined from bedding data (Fig. 3.9), is shown as the open square (□). Note the fold axis lies close to the mean foliation plane, and the mean foliation plane approximately bisects the fold limbs (see Fig. 3.9). These observations suggest the NW trending foliations are axial planar to the F_2 folds.

The inset shows an oblique view (Z, up; N, north; E, east) of a folded ultramafic rock contact (purple). The top of the solid model corresponds to the 22nd level of the Campbell mine, and the intersection of the ultramafic rock contact with the 23rd, 24th, and 25th levels are also shown. The axis of the tight, reclined F_2 fold plunges moderately to the SW. P zone, Campbell mine. The length of the reference axes is approximately 30 m. The solid model constructed from stope and level plans.

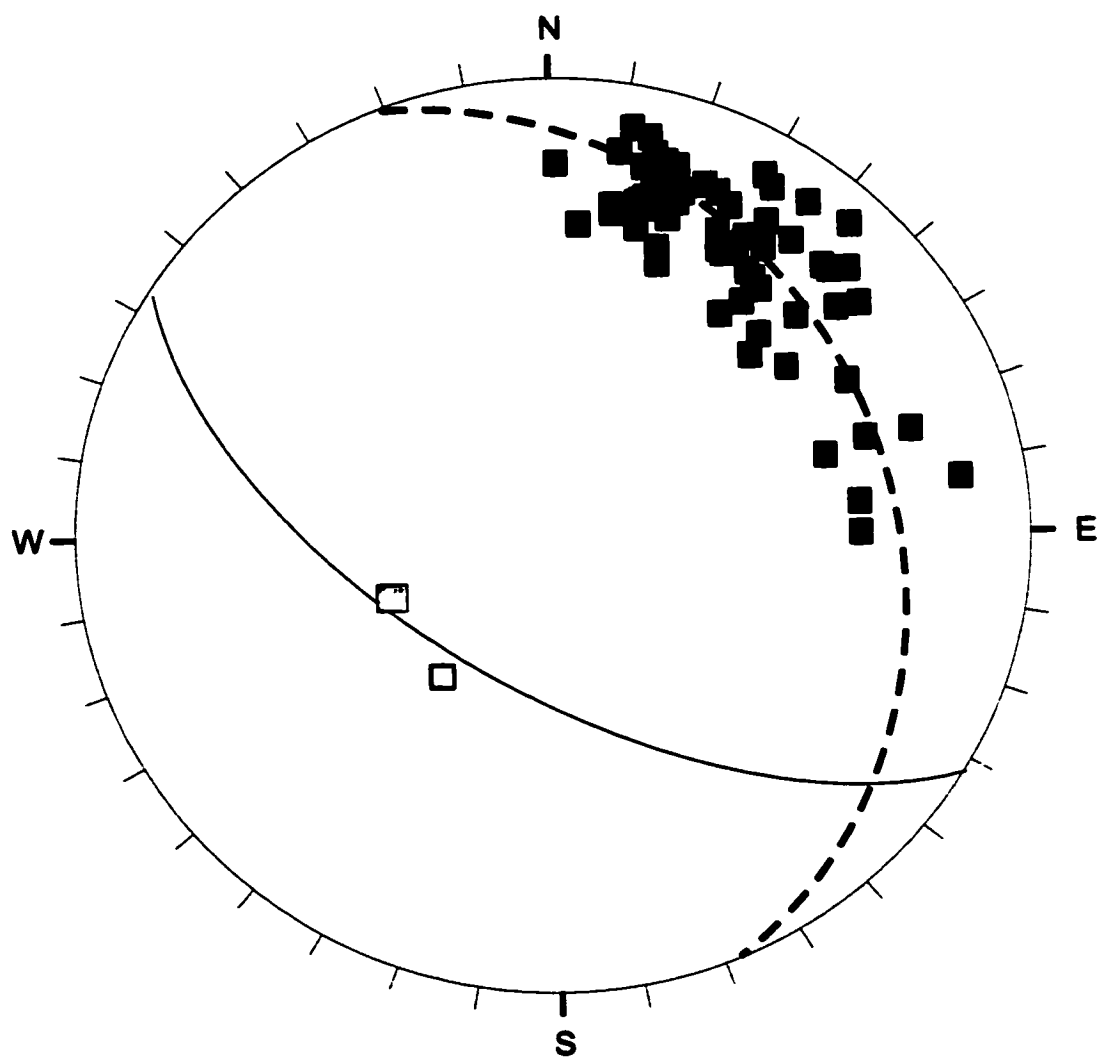


Figure 3.11

Stereonet showing the mean plane to NNW and E-W auriferous veins and shear zones (heavy black great circles, with shear sense), NW veins (dash-dot great circle), and subvertical boudins (●) after Zhang et al. (1997). The mean foliation plane (fine black great circle) from the authors mapping (Fig. 3.10) bisects E-W and NNW shear zones, and the intersection contains the subvertical boudins, suggesting these shear zones are a conjugate array related to NE-SW shortening and NW-SE extension (Zhang et al., 1997).

The mean foliation plane lies close to the mean NW auriferous vein plane and the axis of F_2 folds (□). The mean foliation plane also contains the subvertical boudin axes. Note also the poles to bedding lie between the mean planes of NNW and E-W shear zones (compare to Fig. 3.9), and the fold axis corresponds closely to the shear zone intersections. This indicates that conjugate vein arrays are related to the deformation which produced F_2 folding. The trend and plunge of high-Au grade quartz + arsenopyrite breccia mineralization in the P and L zones (⊕) lies at the intersection of NNW and E-W shear zones, and NW veins and the mean foliation plane. The plunge of the zones also corresponds closely to the orientation of F_2 fold axis. This suggests hydrothermal breccias were developed along the intermediate finite strain direction. Lower hemisphere, equal area projection.

The inset shows an oblique view (Z, up; N, north; E, east) of a folded ultramafic rock contact (purple), presented in figure 3.10. A NNW trending auriferous vein (red) occurs at the ultramafic rock contact, and is shown between the 22 and 24th levels. High Au grade ore chutes occur within the vein, proximal to the F_2 hinge. P zone, Campbell mine. The length of the reference axes is approximately 30 m.

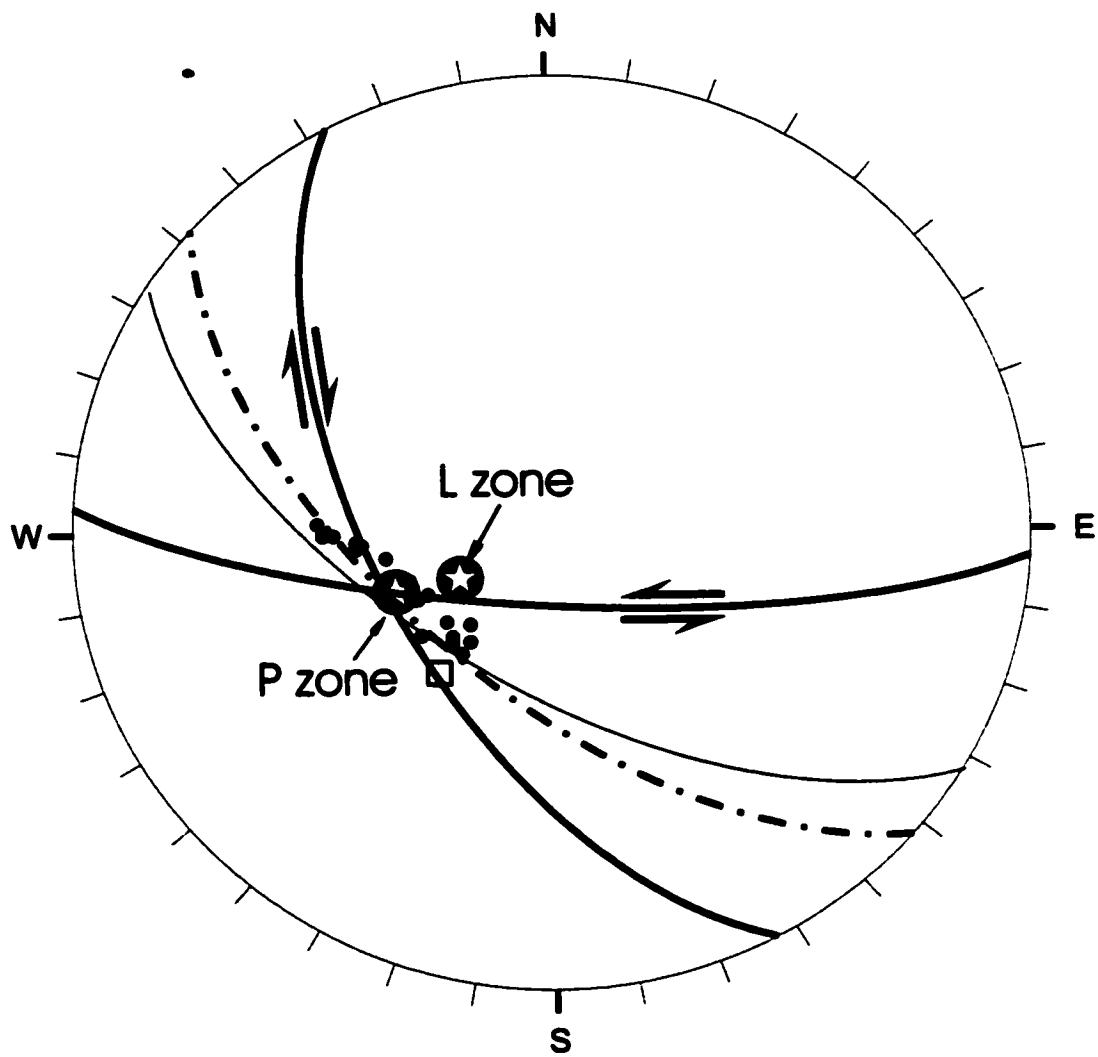


Figure 3.12

Timing of Fe-carbonate alteration and foliation development. A) Pervasive alteration comprising Fe-carbonate and sheeted veinlets in foliated ultramafic rock. Foliation planes are over-printed by pervasive Fe-carbonate alteration, and fibrous-texture sheeted carbonate veins. Note that the sheeted carbonate veins are gently asymmetrically folded, suggesting a component of sinistral shear across the veins. Photo viewing downward, outcrop immediately north of the up-dip projection of the G ore body. B) Carbonate veinlets along axial planar cleavage planes in an F_2 fold in iron formation. Outcrop at the Cochenour airport, compass for scale.

These observations indicate that pervasive Fe-carbonate alteration and veining occurred during and after F_2 folding and the development of NW trending foliations and cleavages.



Figure 3.13

Geology plan map of the G zone, 1861 stope. The quartz + carbonate veining is variably overprinted by auriferous silicification and brecciation. Foliation trend lines are shown as black lines. Wallrocks adjacent to NNW trending vein segments are characterized by progressive rotation of foliation planes into auriferous veins, indicating dextral shear along these veins. N-S trending veins are oriented at a high angle to the foliation and are symmetrically folded. Strain increases towards the NNW zones, inferred from the intensity of folding of N-S veins and development of foliation. Campbell mine grid, feet.

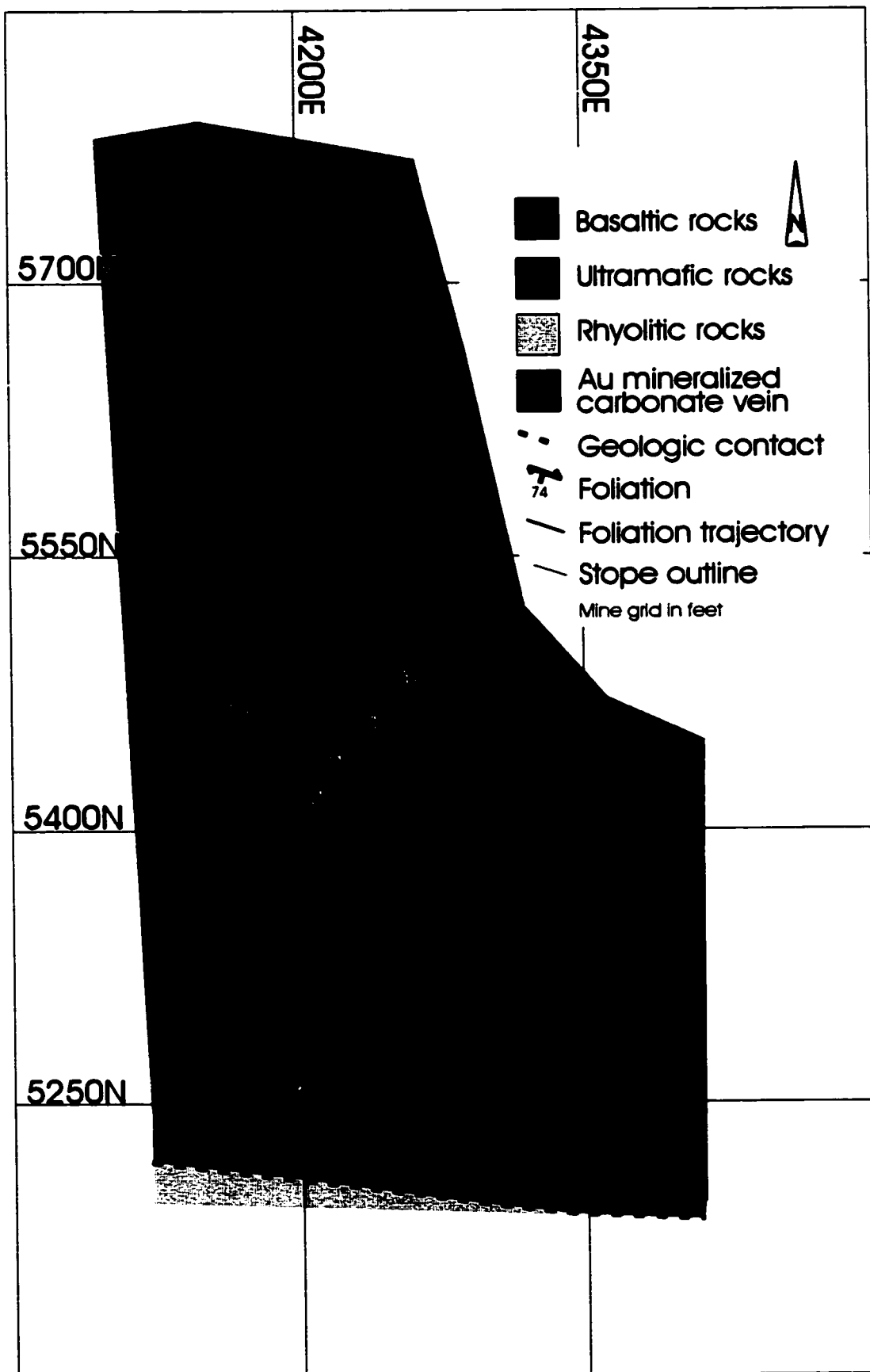


Figure 3.14

Photo of the brittle-ductile shear zone hosting the G ore body (center of photo, trending right-left). NW-trending Fe-carbonate veins are offset with an apparent dextral sense across the NNW trending shear zone, indicating dextral shear occurred during, or continued after NW carbonate veining. Carbonate veins have undergone rotation into the brittle-ductile shear zone. Pillow basalt in foreground, ultramafic rocks with strong pervasive Fe-carbonate alteration in the background. Compass for scale, foreground left side of photo. Photo viewing east.

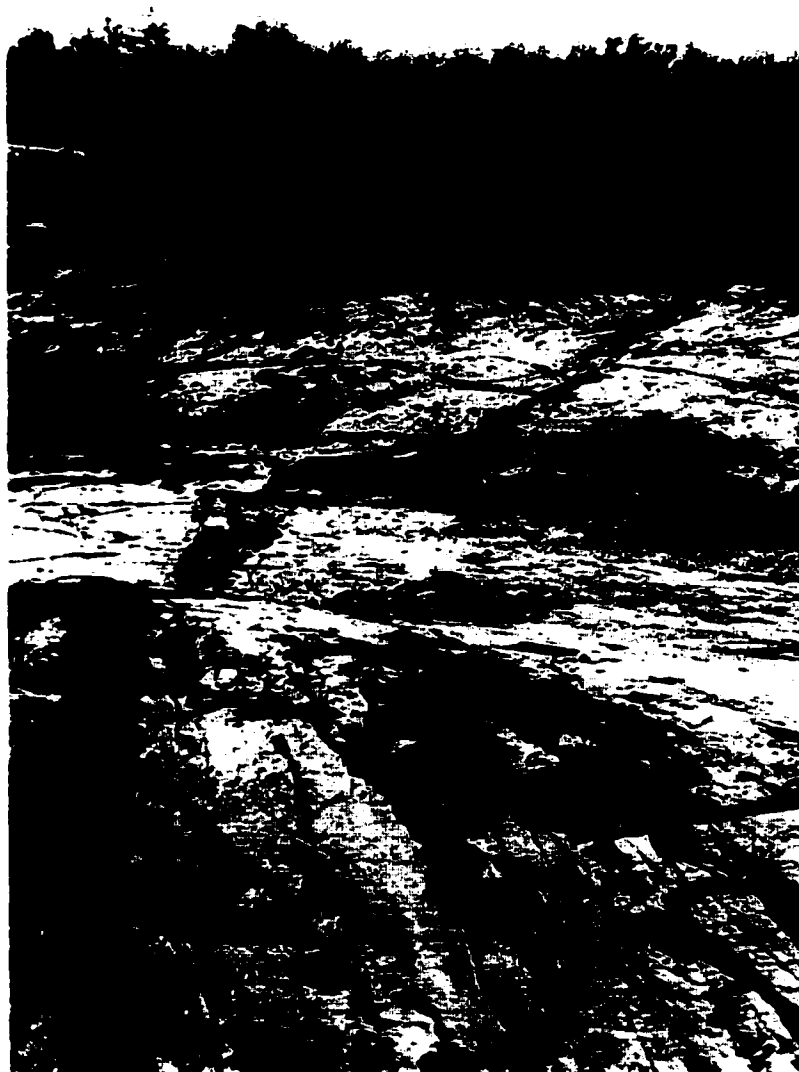


Figure 3.15

Geology plan map of the P zone, 2451-6 stope and -5 sub. Auriferous brecciated crustiform quartz + carbonate veins are shown in orange, and high Au grade quartz + arsenopyrite breccia bodies are shown in red. A WNW auriferous quartz + carbonate vein ("sheeted veinlet zone") occurs at the WNW trending mafic – ultramafic rock contact. High gold grade mineralization occurs at the intersections of auriferous NNW and E-W veins, and at the fold hinges in the mafic – ultramafic rock contact. Note that E-W veins undergo rotation into the NNW vein at the high gold grade breccia in the left portion of the diagram, and the NNW vein is offset to the left approximately 3 to 5 m, in the area of intersection with the E-W vein zone. Auriferous NNW veins on the right side of the diagram offset the ultramafic rock contact with an apparent dextral sense of shear, and show mutual crosscutting relations with E-W veinlets. These observations suggest that motion along NNW and E-W veins were coeval.

- Lamprophyre dykes
- Basaltic rocks
- Ultramafic rocks
- Banded to crustiform quartz-carbonate vein
- \\ Sheeted carbonate veinlets, $\pm$ biotite + pyrite alteration
- Carbonate vein breccia, quartz-Fe sulphide, Au mineralized
- High-Au grade quartz-arsenopyrite breccia
- Magnetite + pyrite + pyrrhotite breccia in quartz + carbonate vein

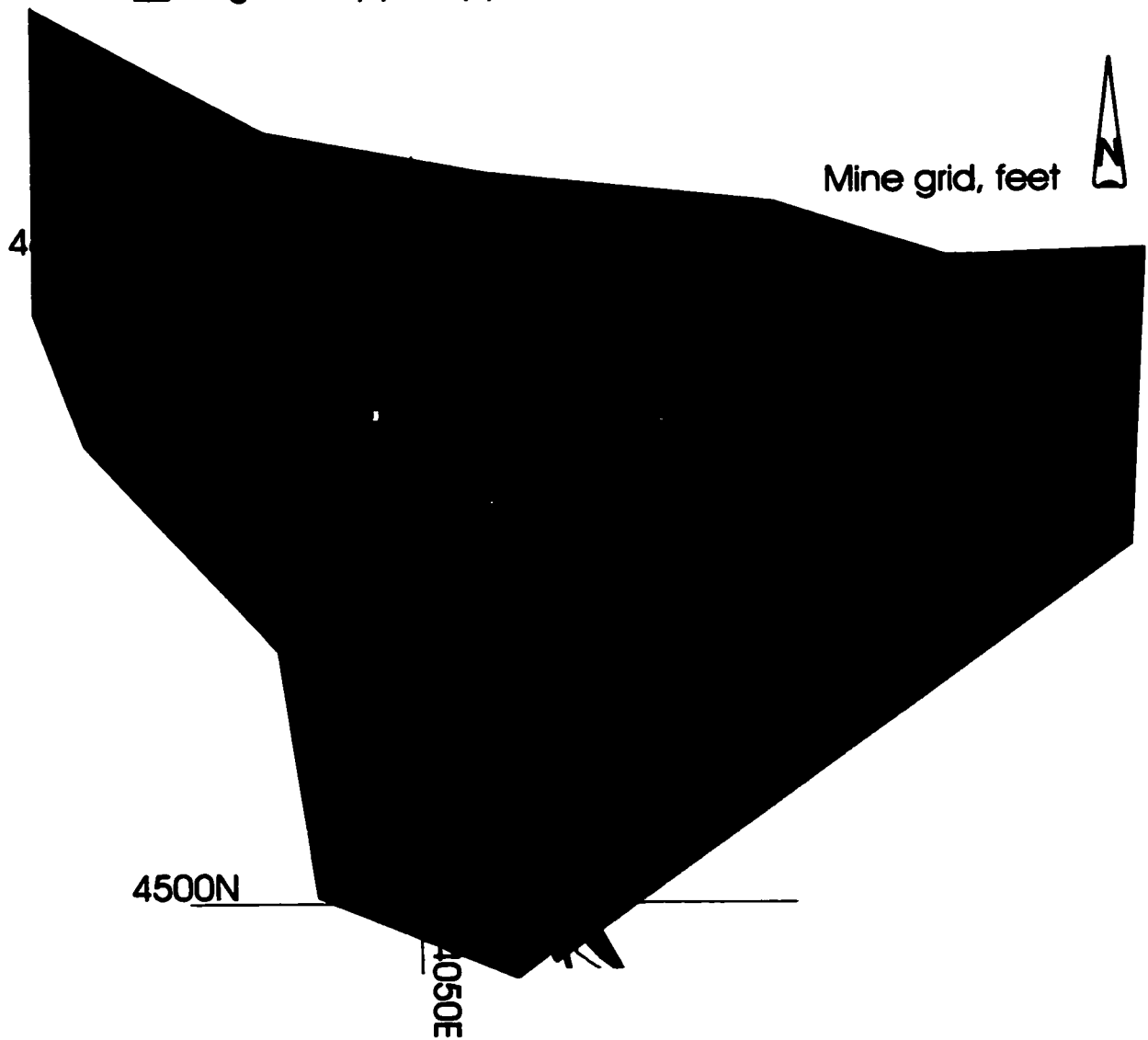


Figure 3.16

Geology plan map of the P zone, 2451-6 stope and –5 sub, showing shear zones (shaded bands, with shear sense). An E-W trending shear zone occurs along the E-W fold limb, with an apparent left-lateral shear sense and 3 to 5m of horizontal displacement. NNW shear zones are related to NNW trending veins, and have an apparent right lateral shear sense, from lithologic offsets and rotations of E-W veins. Shear zones are spatially related to fold limbs, and high gold grade breccia mineralization (red) is spatially related to the intersection of E-W and NNW zones, and the fold hinge in the mafic – ultramafic rock contact.

- Lamprophyre dykes
- Basaltic rocks
- Ultramafic rocks
- Banded to crustiform quartz-carbonate vein
- Sheeted carbonate veinlets, ± biotite + pyrite alteration
- Carbonate vein breccia, quartz-Fe sulphide, Au mineralized
- High-Au grade quartz-arsenopyrite breccia
- Magnetite + pyrite + pyrrhotite breccia in quartz + carbonate vein

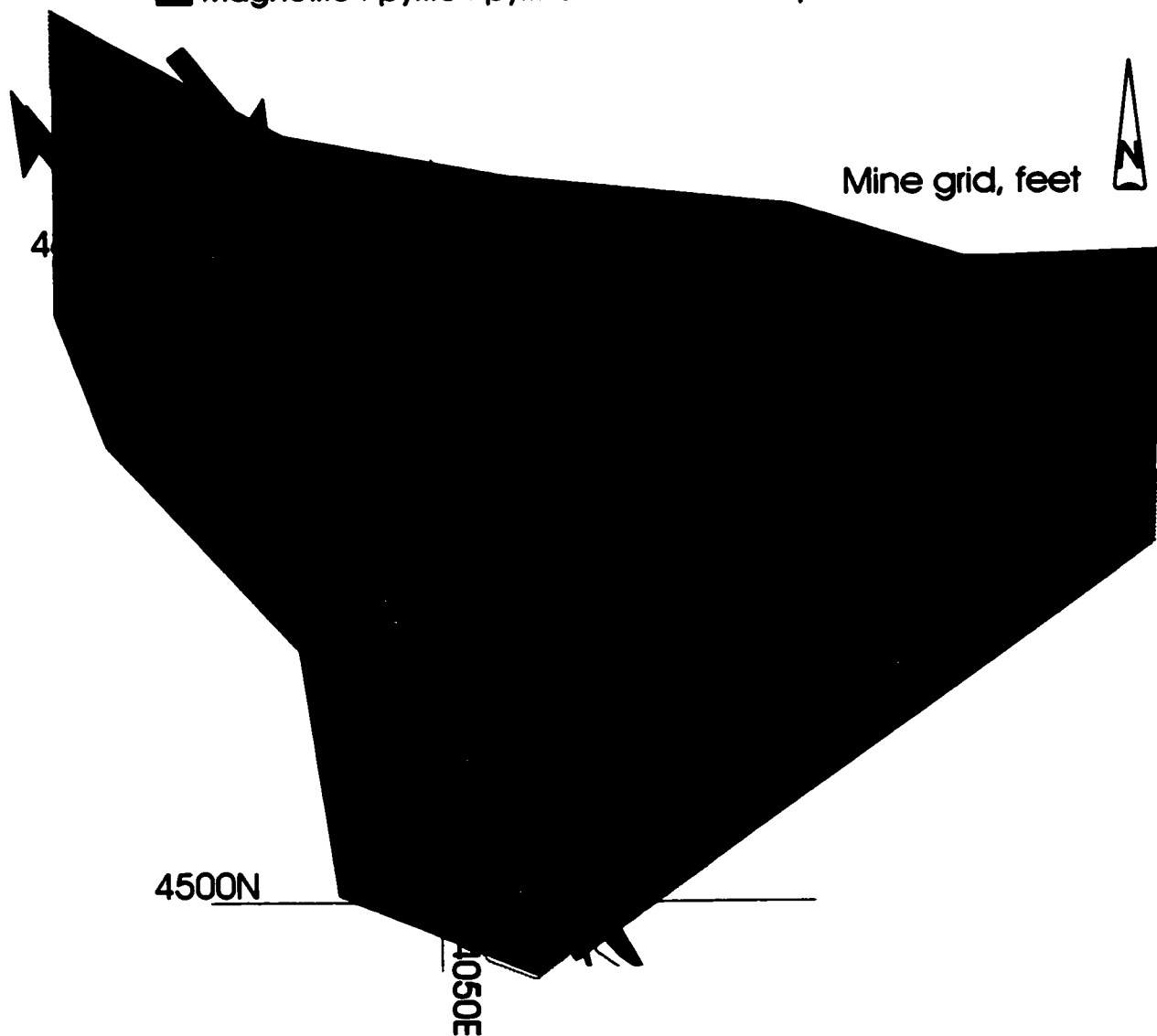


Figure 3.17

Geological plan map of the Campbell (14th level) and Goldcorp (15th level), showing the locations of auriferous veins. The volcanic units are clearly folded, with NW trending axial traces. Foliation-parallel auriferous veins are spatially associated with the highly attenuated limbs of hook-shaped NW folds. Note that folding is tight in the west, becoming isoclinal and rootless in the east. This suggests strain related to NW folding increased eastward. Attenuation of NW fold limbs at the mine may be coeval with attenuation of the eastern limb of the NW trending Balmer Synform. The grey shaded bands show the approximate position of the late D₂ shear zones, some of which operate on earlier, D₂ shear zones which host the auriferous veins. The shear sense for the late D₂ shear zones is shown by the arrows. Modified after MacGeehan and Hodgson (1982), with additional data from Campbell mine geology plans and drill logs. Campbell mine grid, feet.

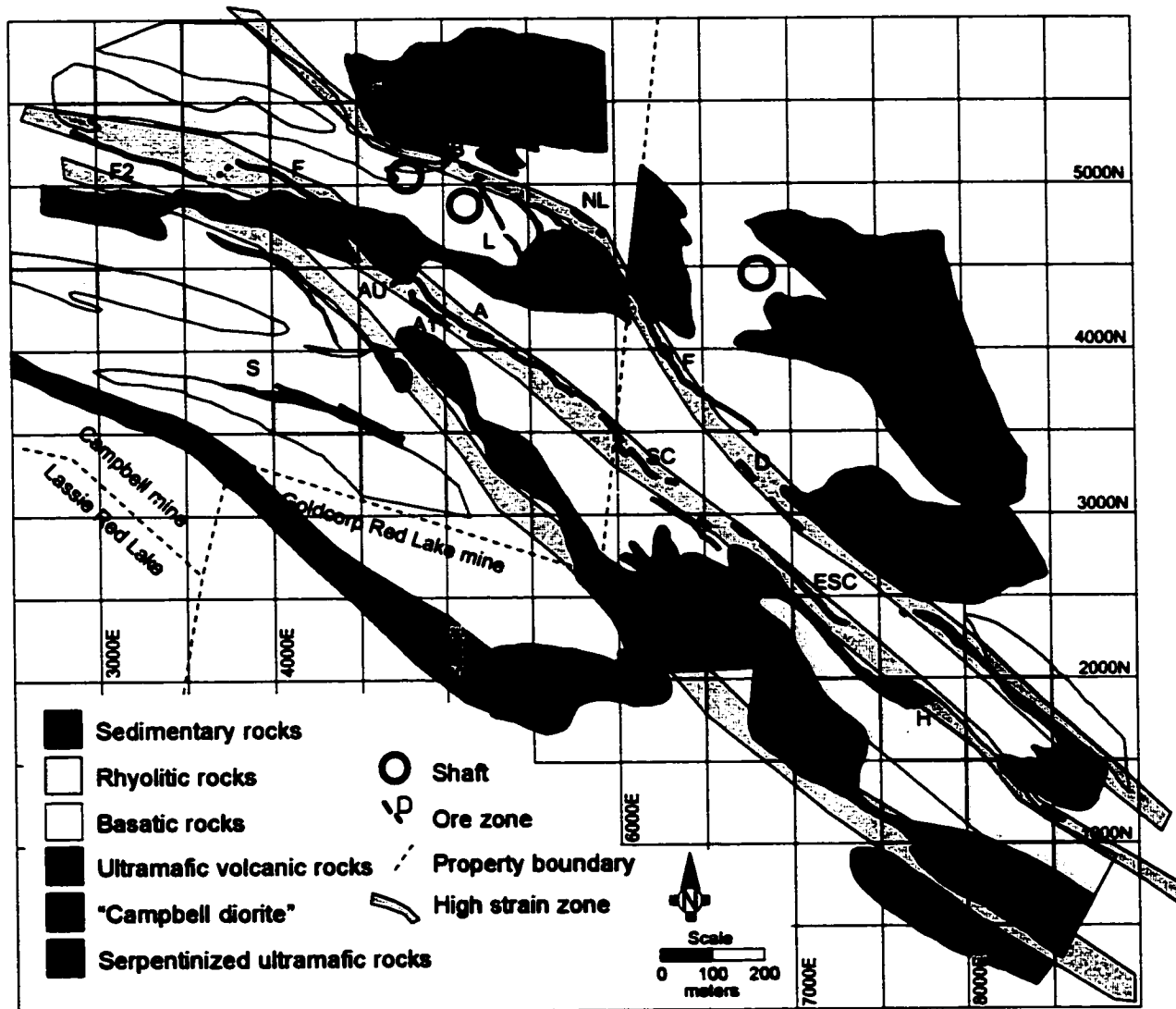


Figure 3.18

Auriferous foliation-parallel quartz + carbonate vein in the A ore zone. The vein dips steeply, is boudinaged, and contains white quartz veinlets in the boudin necks.

Auriferous quartz + arsenopyrite (+ pyrite and pyrrhotite) mineralization (grey, left side of vein) overprints earlier crustiform quartz + carbonate banding (white fragments, predominantly on right side of vein), and are cut by late D₂ white quartz veins (subhorizontal) in the boudin necks. Photo of face, 1302 stope.



Figure 3.19

Pillow basalt cut by a brittle to brittle-ductile “black line fault” (trending left-right, below coin). The fault is composed of quartz, tourmaline, and minor Fe-carbonate. Campbell mine, G zone extension outcrop.

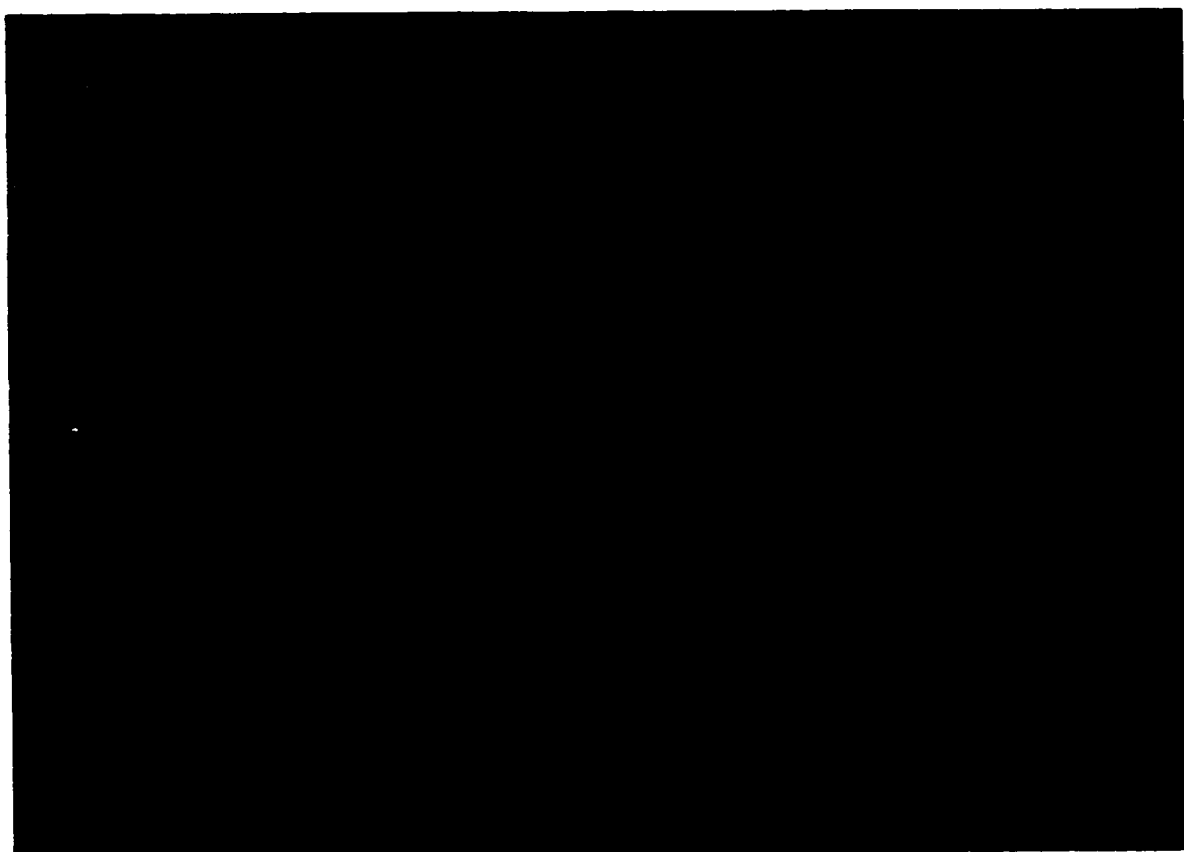


Figure 3.20

Photograph showing pervasive Fe-carbonate alteration and bleaching of ultramafic rocks. Note that the E-W trending foliation overprints pervasive carbonate alteration. Late north trending, shallow dipping quartz veinlets displaying weak symmetrically folded, and show left-lateral offsets across foliation planes. Time relations between pervasive carbonate alteration and foliations suggest carbonate alteration was coeval with the development of E-W shear zones. Fibrous coarse-grained sheeted veinlets cutting the quartz veins are not intensely sheared. This suggests the later phase of carbonate deposition and sheeted veinlet development occurred during, or after sinistral shear. Location: G zone extension outcrop immediately north of the up-dip projection of the G ore body.



Figure 3.21

A photograph of quartz and carbonate veins in highly foliated basalt. Milky white quartz tension veins (grey in the photo, white in the line drawing) cutting sheared foliation-parallel quartz + carbonate vein (blue in the line drawing), in basalt (green). Black fine lines show the foliation planes in the basalt. The inclination of the quartz veins with respect to the foliation suggests reverse shear along the foliation planes and NW trending carbonate vein. This is supported by the geometry of relatively early, sigmoidal, rotated quartz veins (grey). Location: 2701 south cross cut, vertical face, viewing WNW.

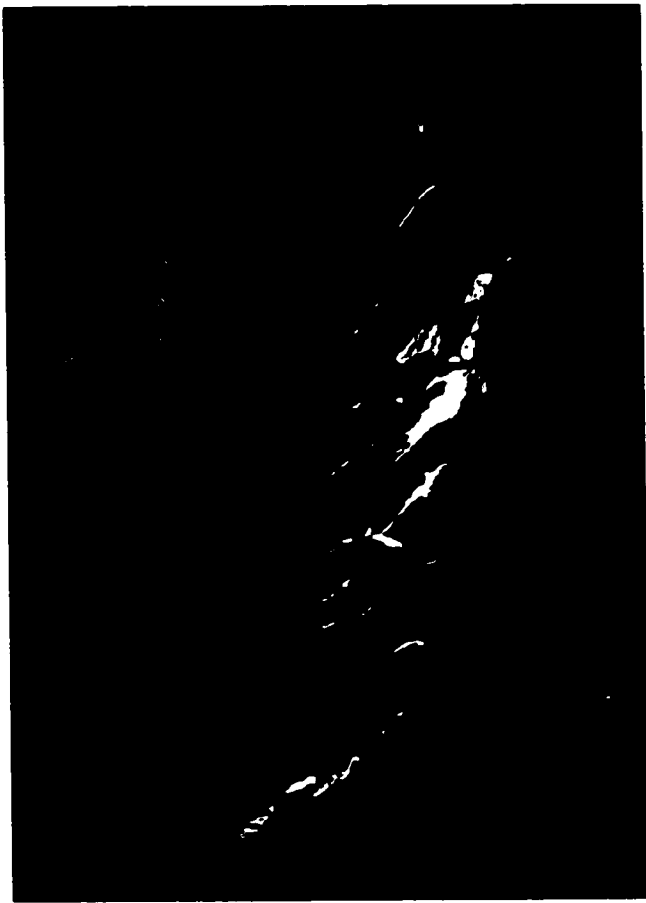


Figure 3.22

Photo of a post-ore fault zone. The inset shows carbonate veins (light grey) with asymmetric folds indicating left-lateral shear. Photo of back, West drift fault area, 30th level Campbell mine. Note that shear senses are opposite of the apparent shear sense when viewing the back.

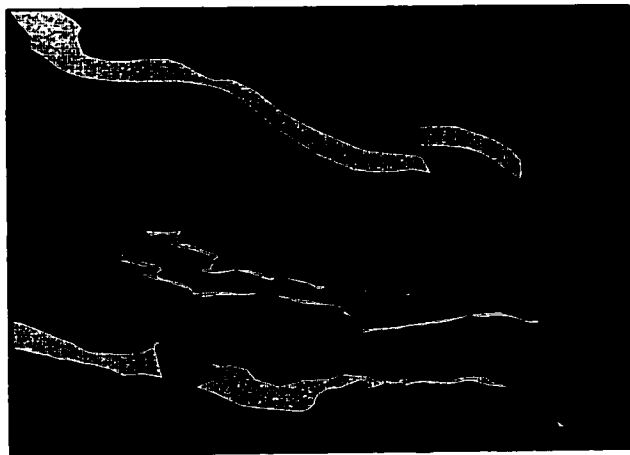
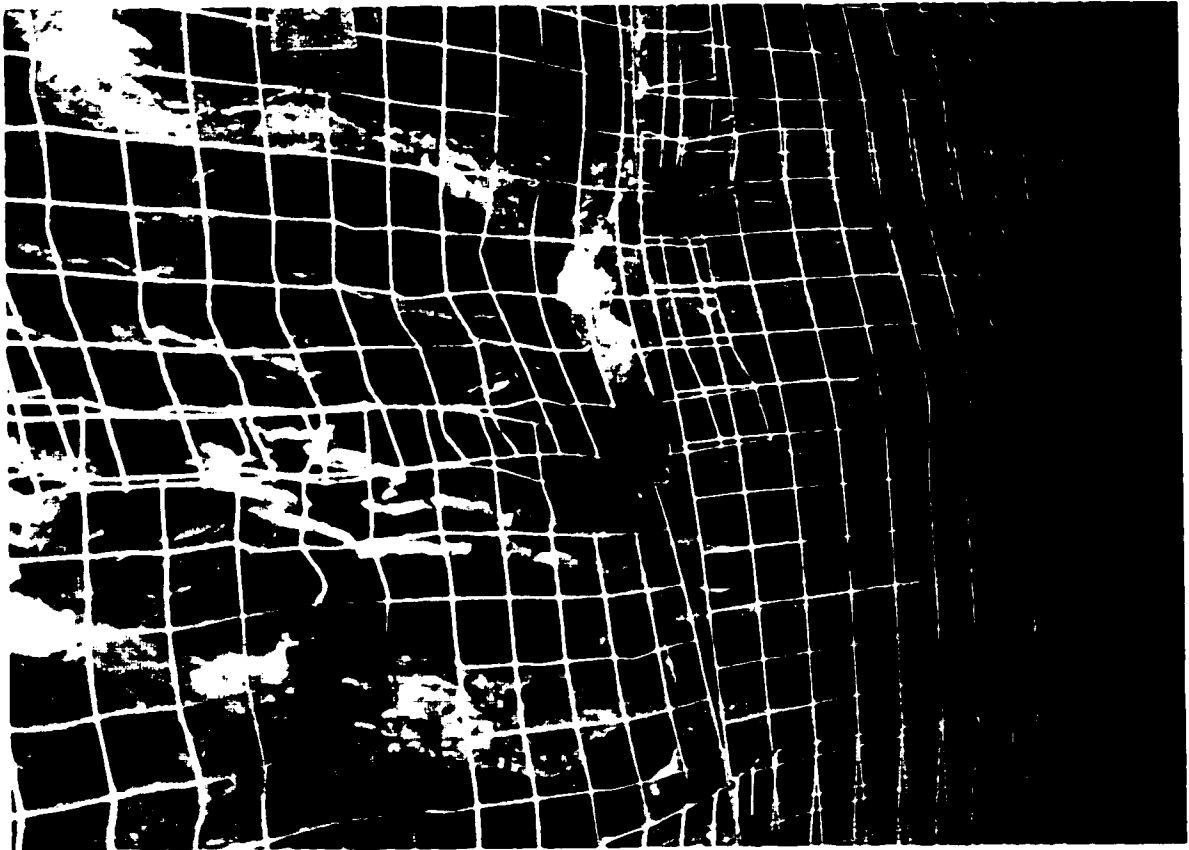


Figure 3.23

Zoning of a coarse-grained garnet from the Campbell mine. Abundant mineral inclusions (quartz, cummingtonite, chlorite, sulphide and oxide minerals) in the garnet made it difficult to obtain a continuous compositional zoning profile. Analysis points are represented by the solid circles. Cores are X_{Sps} and X_{Grs} rich, rims are X_{Alm} and X_{Prp} rich, and the atomic ratio of $\text{Fe}/\text{Fe}+\text{Mg}$ decreases towards the rim. Dotted lines show the estimated compositional profile, and its jog is based on the sharp color change between the core and rim. The abbreviations X_{Sps} , X_{Grs} , X_{Alm} , and X_{Prp} refer to the mole fractions of spessartine, grossular, almandine and pyrope, respectively.

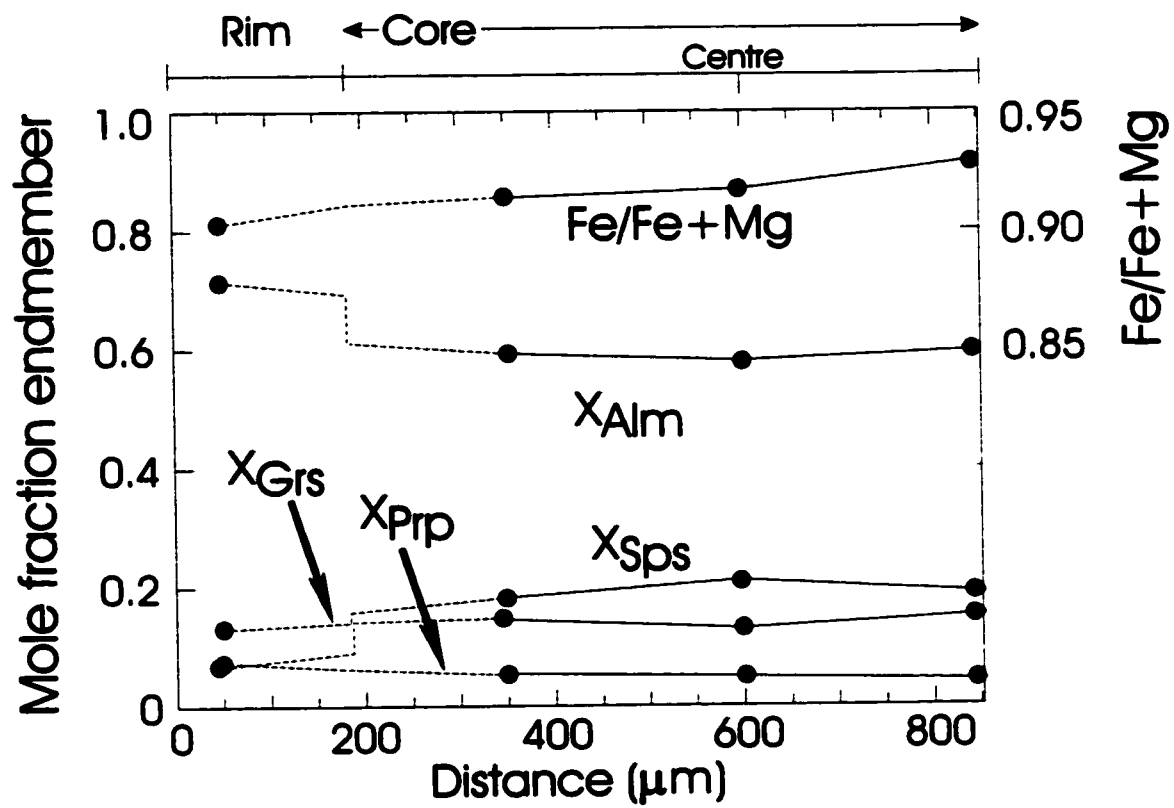


Figure 3.24

Reflected light photomicrograph of a syn-tectonic garnet (Grt) porphyroblast displaying S shaped inclusion trails of quartz (Qtz), continuous with the foliation in the matrix. Garnet also contains inclusions of pyrrhotite (po). The matrix is predominately composed of chlorite, quartz, plagioclase, and biotite. Sample 22, from drill core beneath the 27th level. Abbreviations; Po = Pyrrhotite

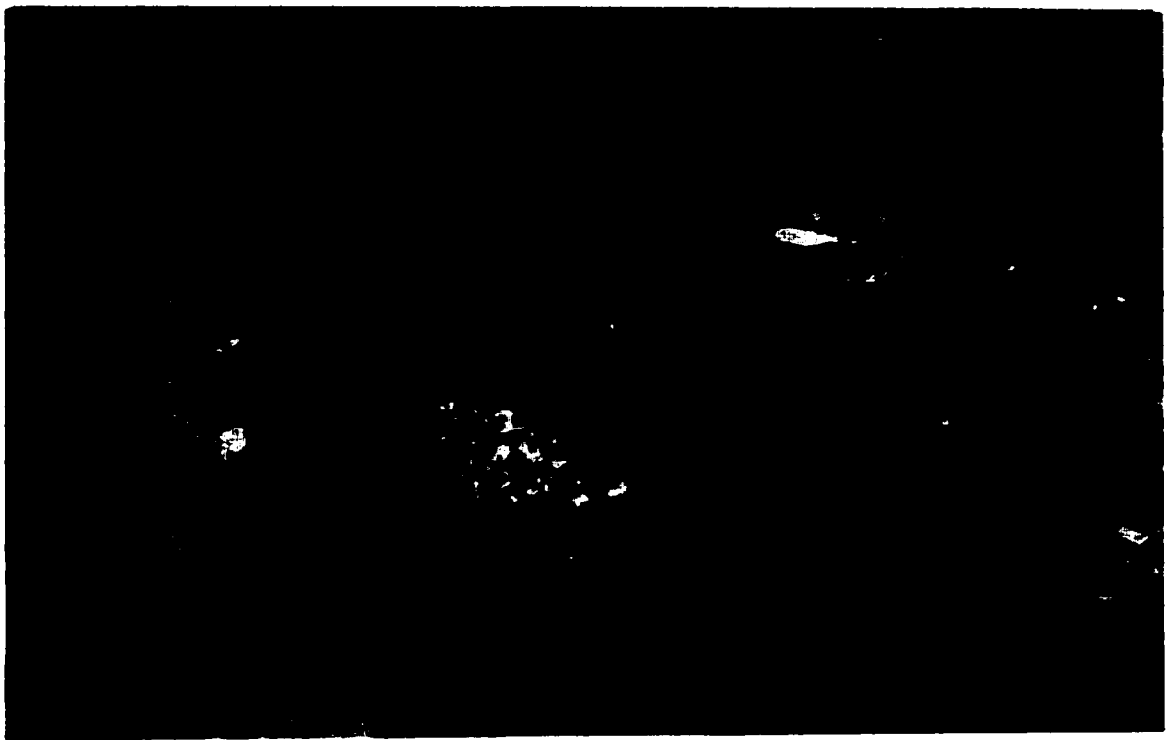


Figure 3.25

Photomicrograph showing Assemblage 8, with Assemblage 9 in the garnet boudin neck (Table 3.1). Garnet (Grt) has developed euhedral growth faces adjacent to the boudin neck. The foliation planes, shown as dashed lines, are defined by chlorite, biotite and chloritoid (Cld) in Assemblage 8, and chlorite and biotite in Assemblage 9. Both Assemblage 8 and 9 contain biotite. Hatched area in garnet contains abundant quartz and cummingtonite inclusions. The plane of the section is perpendicular to the foliation, and boudin axes. The reaction, chloritoid = garnet + biotite + chlorite (Fig. 3.33) is interpreted to have occurred during shortening. Assemblage descriptions are provided in Table 3.1. Sample from drill core 27-264, Campbell mine. Abbreviation; St = staurolite

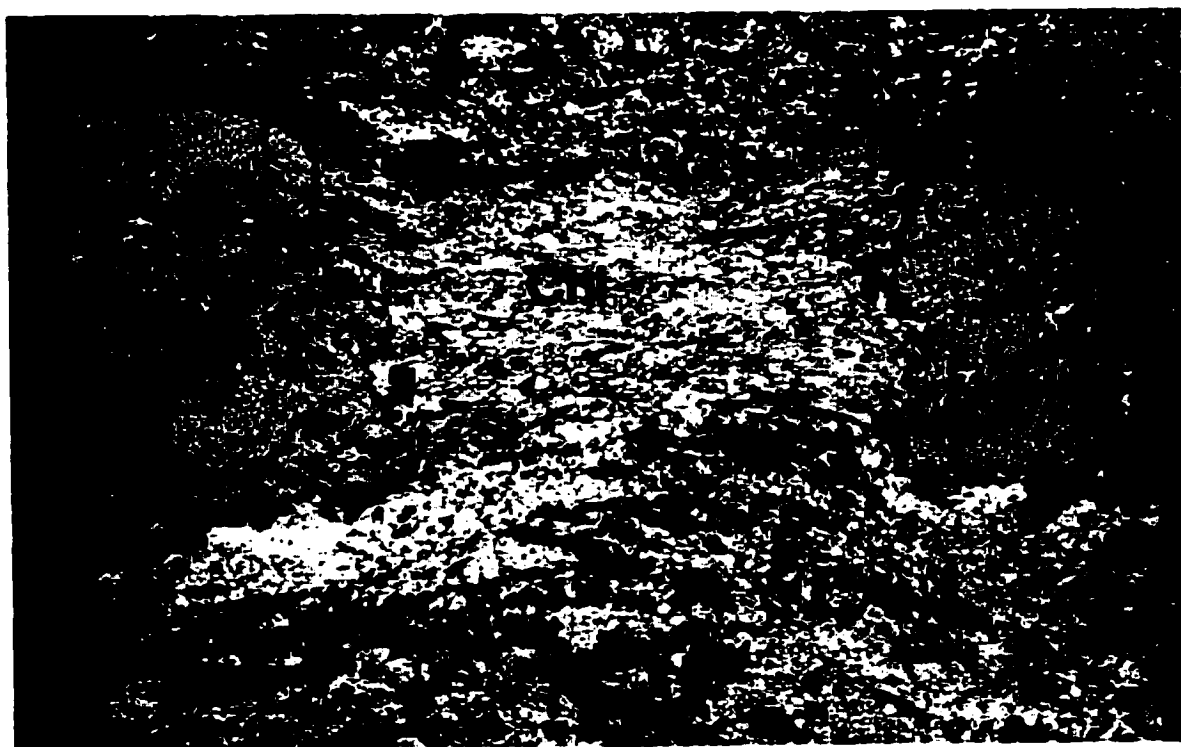
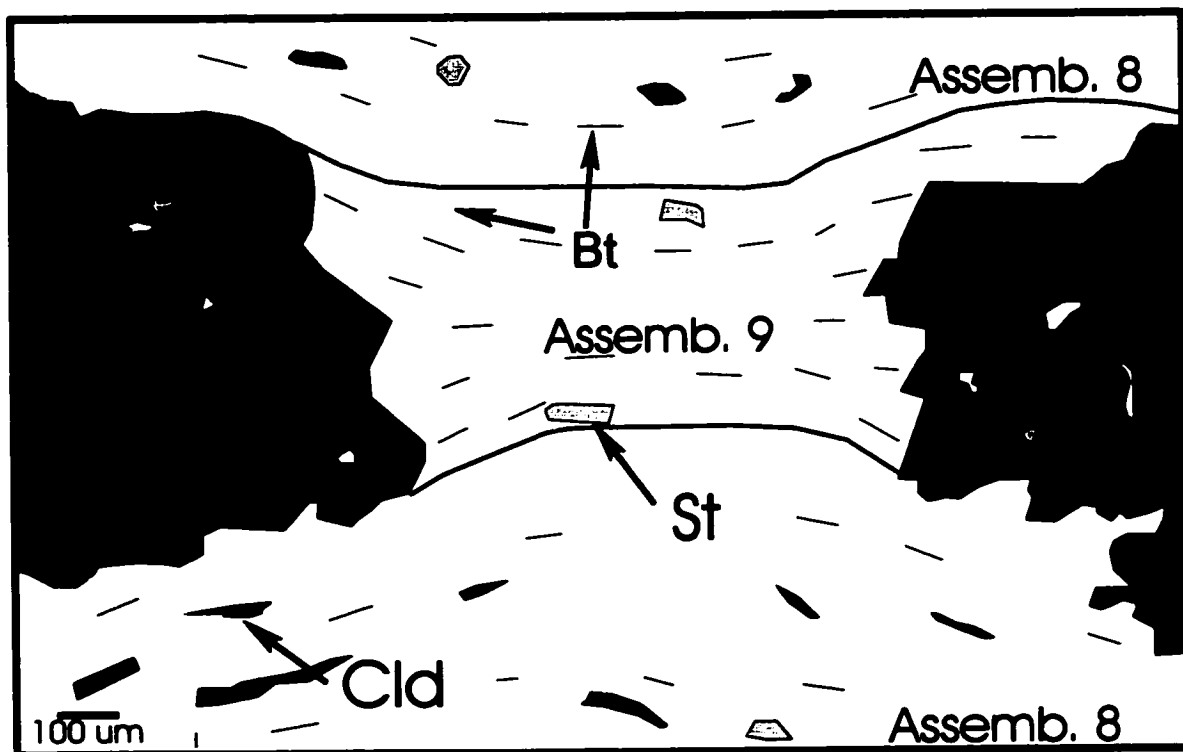


Figure 3.26

Photomicrograph showing a syn-kinematic, rotated andalusite (And, grey) porphyroblasts. Quartz and pyrite inclusions in the porphyroblasts form trails continuous with foliation in the matrix, shown as dashed lines in the drawing. The matrix is composed of very fine-grained quartz (Qtz), white mica, biotite, and abundant pyrite (Py). From interflow sedimentary rock from drill core beneath the 27th level of the Campbell mine.

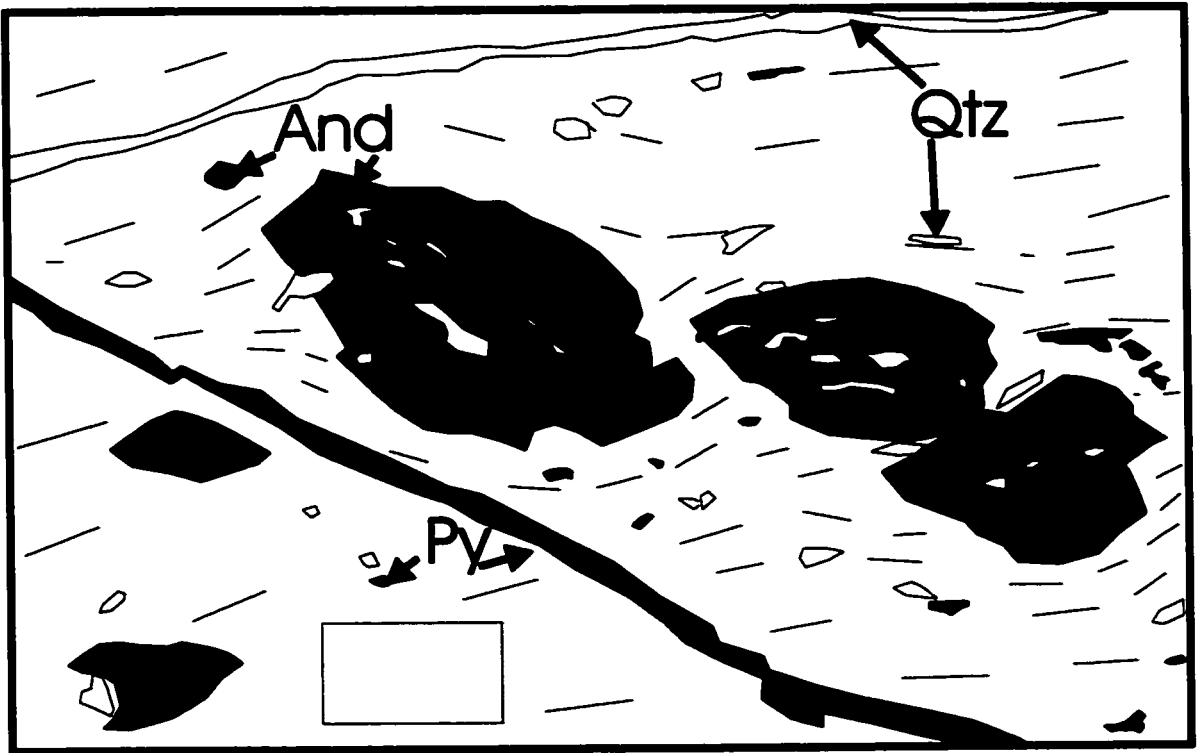
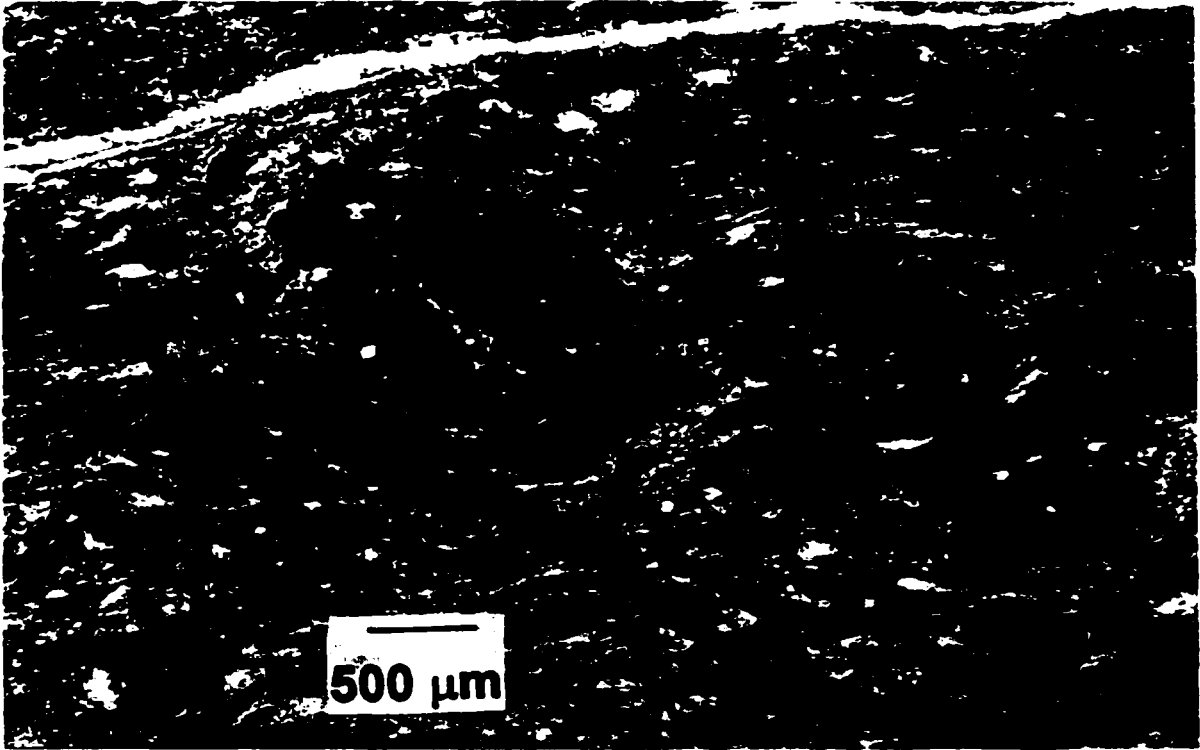


Figure 3.27

Ternary diagram showing biotite compositions. The boundary between biotite (Bt) and phlogopite (Phl) is from Deer et al. (1962). Biotite from the study area plot close to the siderophyllite (Sid) - eastonite (Eas) join.

Biotite from the relatively unaltered rocks (●) are Fe-rich and Al-poor compared to biotite from the auriferous veins, and from altered rocks adjacent to auriferous veins (◆). Arrow indicates the trend of increasing Al and Mg in biotite with decreasing distance to auriferous veins. The star represents the average composition of biotite used in the construction of figures 10.9 and 10.10. Diagram modified after Mitchell (1989). Abbreviations; Ann = annite, Sid = siderophyllite.

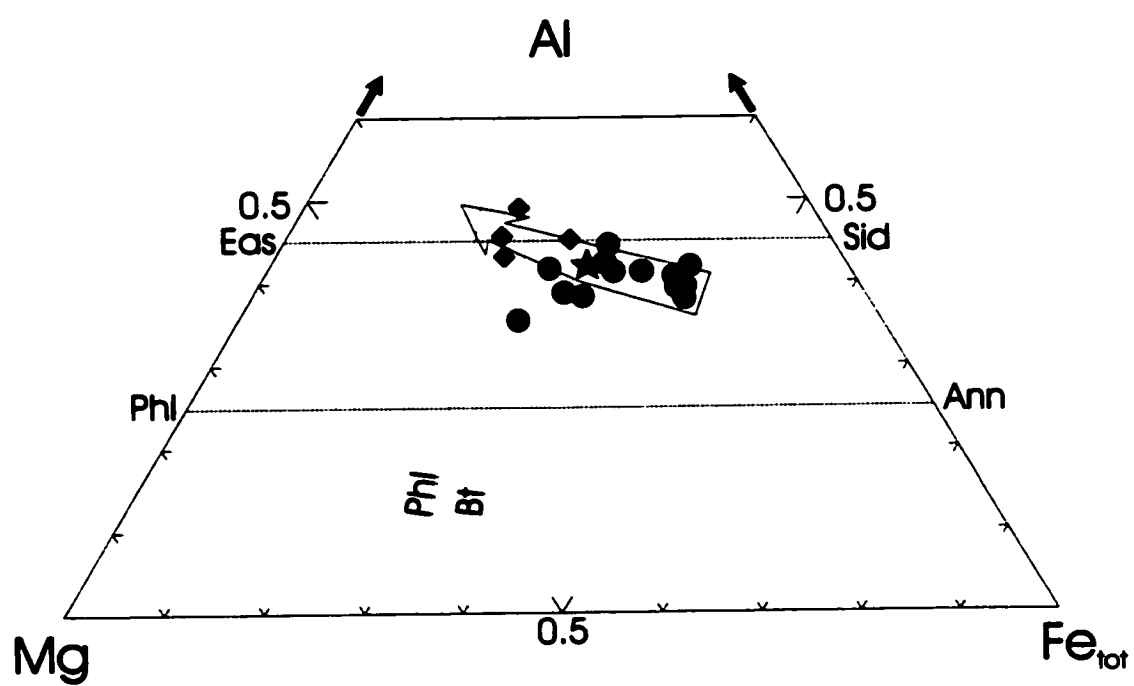


Figure 3.28

Composition of chlorites from the Campbell mine (solid circles).

They plot in the field of ripidolites (Rp). Diagram modified after Hey (1954).

Abbreviations: Br = brunsvigite, Cc = clinochlore, Cp = corunophilite, Db = diabantite, Dn = daphnite, Pc = pycnochlorite, Pn = pennine, Pt = pseudothuringite, Sh = sheridanite.

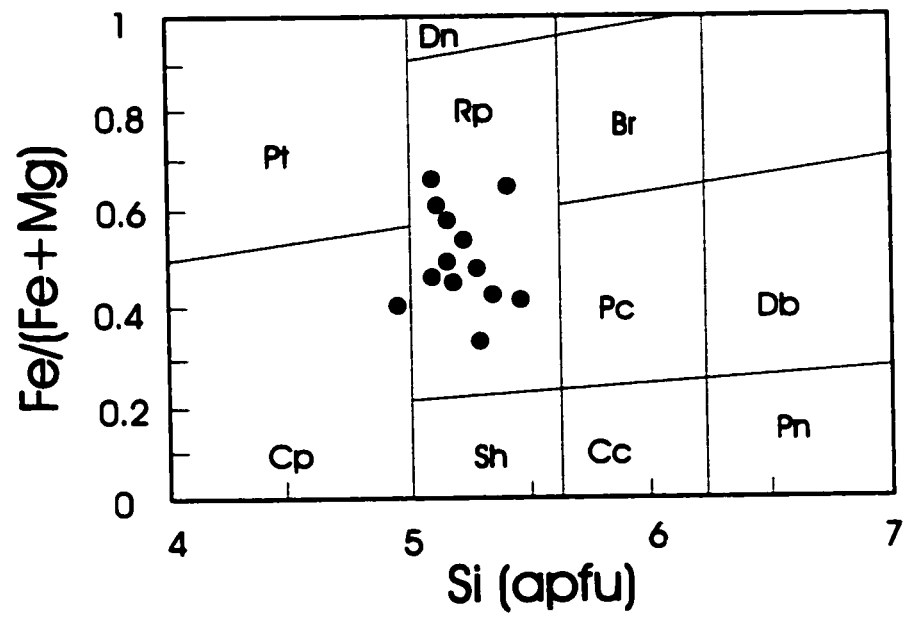


Figure 3.29

Photomicrograph and line drawing showing cummingtonite (Cmt) inclusions in garnet (Grt). Cummingtonite occurs as randomly oriented inclusions in the interior of coarse-grained garnet, and is not observed in foliated rock outside the garnet grains. This suggests that cummingtonite growth occurred prior to deformation. The loss of cummingtonite from the matrix suggests either its consumption during metamorphic mineral reactions, or a change in bulk rock composition leading to cummingtonite instability. Sample 22, from drill core beneath the 27th level of Campbell mine.

Abbreviations: Mag = Magnetite, Qtz = quartz.

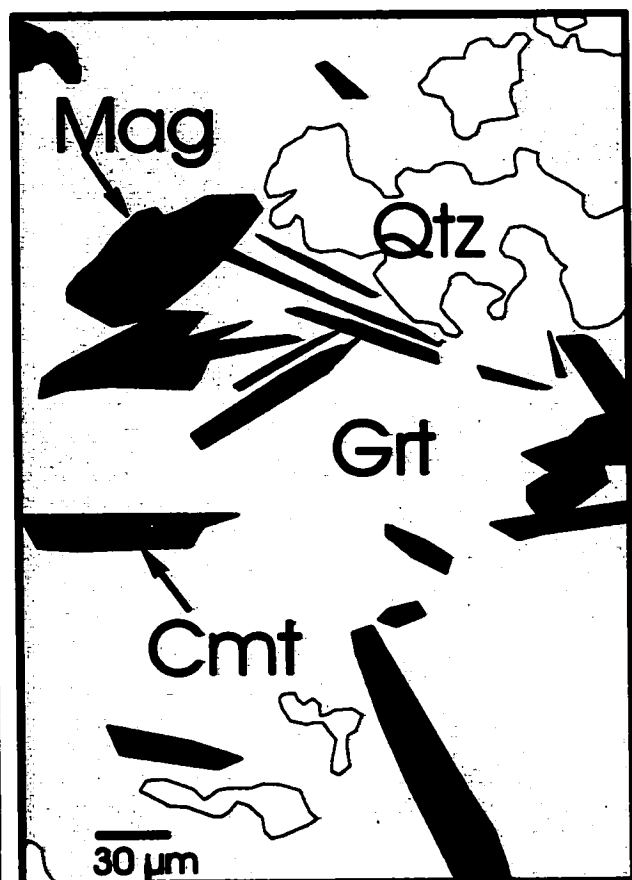


Figure 3.30

Photomicrograph of lens-shaped, deformed cordierite porphyroblast in aluminous altered basalt. Cordierite encloses fine-grained quartz, and minor fine-grained opaque phases, and occurs in a coarser-grained matrix of quartz, biotite, chlorite, and opaque phases. This suggests that inclusions preserve the initial, fine-grained pre-metamorphic rock texture. Cordierite is altered to chlorite (light green) and coarse-grained biotite (brown) along its margins, and along fractures. These observations suggest flattening and shearing occurred after the onset of metamorphism, and was coeval with hydrothermal activity resulting in retrograde alteration of cordierite. The field of view is approximately 0.75 millimeters.

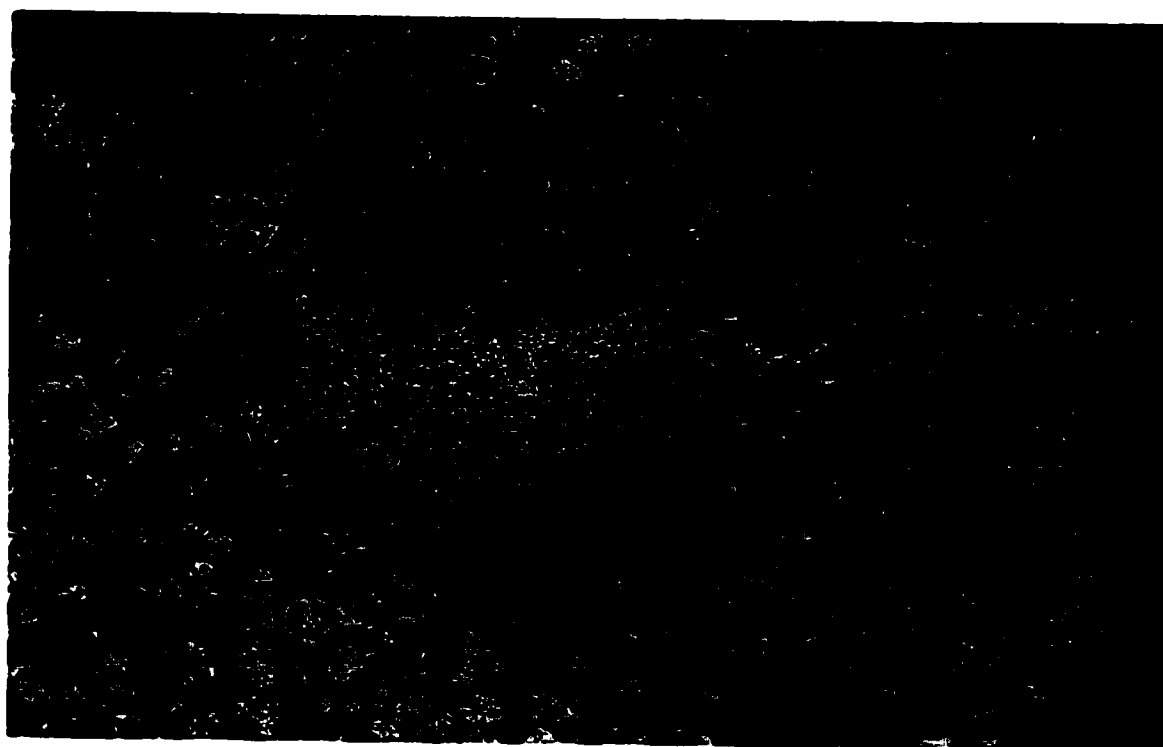


Figure 3.31

Geology map of the eastern Red Lake belt. Sample locations for garnet-biotite thermometry are shown as circles. Sample numbers (Table 3.6) with the depth-below-surface (in meters) are provided for each sample location. Temperature estimates use the thermometer of Kretz (1990), Table 3.6. Heavy dashed line is the garnet isograd at surface (determined in this study), in basalt and sedimentary rock, with garnet bearing rocks on the east-side of the line. The isograd position was determined from surface outcrops, and data from > 100 surface diamond drill holes. The location of representative traces of drill holes collared at surface, and outcrop areas, are also shown. The heavy solid line marks the Campbell mine property boundary. Major auriferous veins at the Campbell and Goldcorp mines are labeled, and indicated as heavy lines. Inset map (upper left) shows F_2 fold axial traces and stratigraphy, with the position of the garnet isograd for reference. Note the isograd cross-cuts both folded stratigraphy and F_2 fold axial traces in the north-central part of the study area. Map modified from MacGeehan and Hodgson (1982), and Pirie and Grant (1978a).

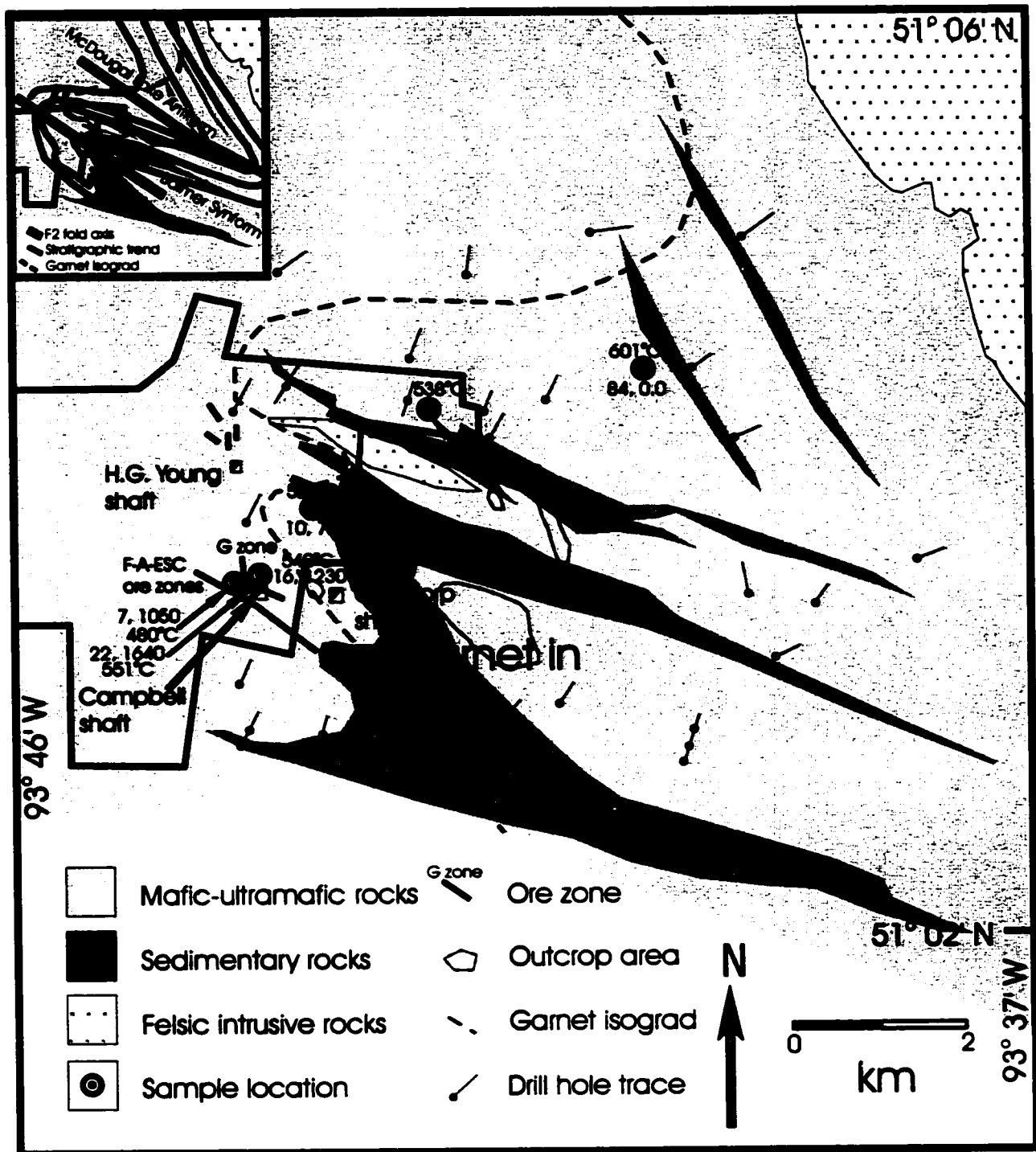


Figure 3.32

Evaluation of metamorphic pressure by GASP (garnet + andalusite + quartz + plagioclase) barometry for Assemblage 11, at a temperature determined from garnet-biotite thermometry. Results indicate a metamorphic pressure of ≈ 3 kilobars at ≈ 550 °C, close to the andalusite-sillimanite boundary. Abbreviations; And = andalusite, Alm = almandine, Ann = annite, An = anorthite, Als = aluminosilicate, Grs = grossular, Ky = kyanite, Prl = pyrophyllite, Prp = pyrope, Sil = sillimanite.

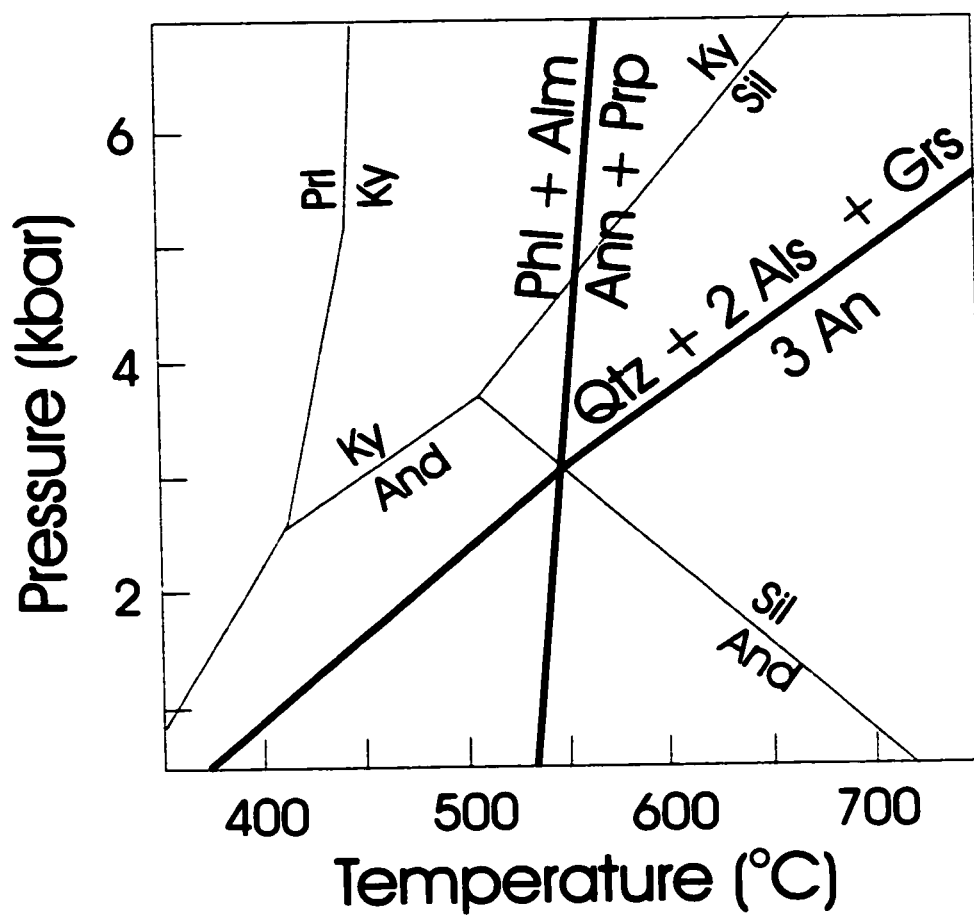


Figure 3.33

Petrogenetic grid of Spear and Cheney (1989), in the KFMASH system. The reaction $\text{Fe Cld} + \text{Ann} = \text{Alm}$ is a KFASH reaction. The dash lines labeled 10, 20, and 30 refer to X_{Sps} (or X_{Grs}) in garnet, showing the expansion of the garnet + chlorite stability field to lower temperatures with increasing X_{Sps} (Spear and Cheney, 1989). Thick solid lines plot the stability fields for the aluminosilicate minerals. The absence of chloritoid in Assemblage 9 (Fig. 3.25, Table 3.1) and lack of chloritoid in contact with garnet in Assemblage 8 suggests the reaction chloritoid = garnet + biotite + chlorite (red line). Solid circles with sample numbers refer to temperatures determined with the garnet-biotite thermometer of Kretz (1990) (Table 3.6). Temperature for the samples are plotted at 3 Kbars pressure, which is considered a maximum pressure estimate, as discussed in the text. Abbreviations: Alm = almandine, Als = aluminosilicate, And = andalusite, Ann = annite, Bt = biotite, Chl = chlorite, Cld = chloritoid, Crd = cordierite, Fe Cld = Fe chloritoid, Grt = garnet, Ky = kyanite, Prl = pyrophyllite, St = staurolite.

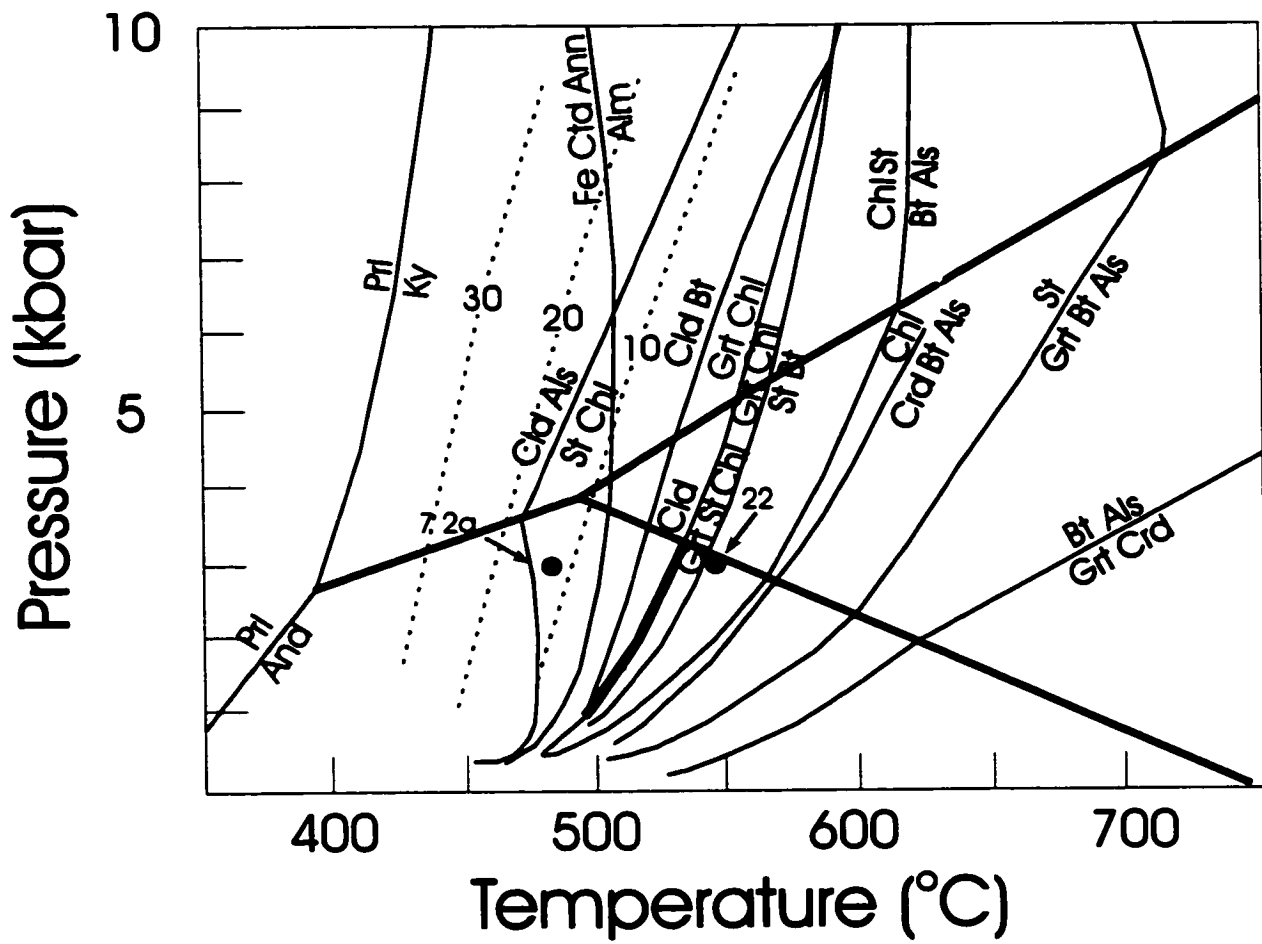


Figure 3.34

Petrogenetic grid for cordierite + anthophyllite rocks, modified after Spear (1993).

Reactions on this grid for Al-poor bulk compositions are appropriate for some rocks in the study area. The heavy black line represents the reaction; garnet + chlorite + cummingtonite = gedrite. The assemblage of gedrite and anthophyllite occur on the high-temperature side of the curve, and those containing the assemblage of garnet + chlorite + cummingtonite on the low-temperature side. The medium black lines outline the labeled aluminosilicate stability fields, and that of pyrophyllite. Red arrows represent approximate P-T regions for rocks of the central to western portion (P ore zone area) of the Campbell mine, and the eastern portion of the Campbell mine (L ore zone area), and Goldcorps' Red Lake mine. The reactions are; 1 Als+Ged = Crd+St, 2 St+Ged=Crd+Grt, 3 Ath=Crd+Ged+Cum, 4 Chl+Ged=Crd+Cum, 5 Grt+Chl=Crd+Ged, 6 Ged+Cum=Grt+Ath, 7 St+Chl=Ky+Ged, 8 Grt+Chl=St+Ged, 9 Ged=Crd+Grt+Cum, 10 Oam=Crd+Grt+Cum, 11 Sil+Ged=Crd+St, 12 Sil+Oam=Crd+Grt. Abbreviations; Als = aluminosilicate, Ath = anthophyllite, Chl = chlorite, Cum = cummingtonite, Crd = cordierite, Ged = gedrite, Grt = garnet, Ky = kyanite, Oam = orthoamphibole, Prl = pyrophyllite, St = staurolite.

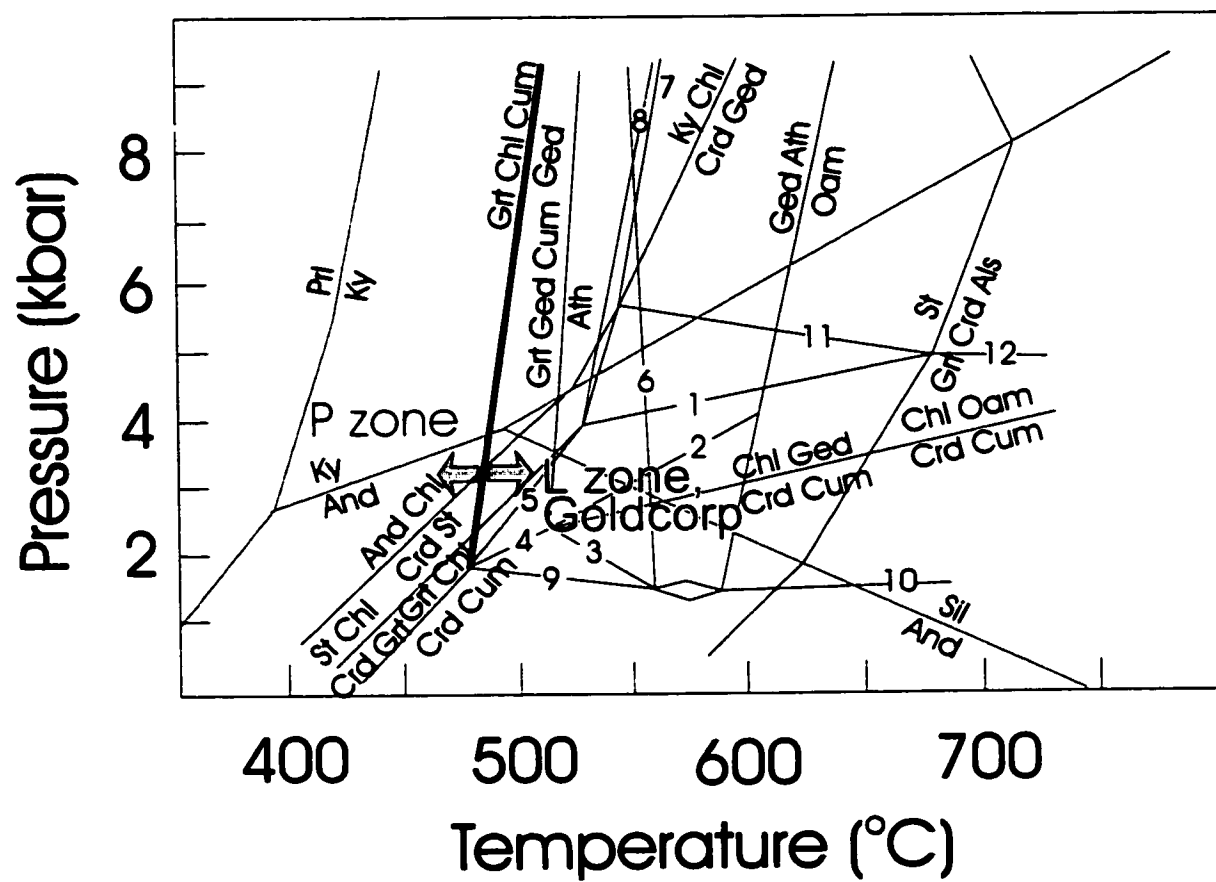


Figure 3.35

Mineral stability diagram of temperature vs. X_{CO_2} at 3 kbar total pressure, for basaltic rocks northeast of the Campbell mine shaft which contain chlorite, gedrite, anthophyllite, garnet, carbonate, anorthitic plagioclase, quartz, and biotite. The diagram is calculated using the TWQ 1.02 program of Berman (1991), with thermodynamic data of Berman (1988), the garnet activity model of Berman (1990), the ternary plagioclase model of Furman and Lindsey (1988), and the biotite activity model of McMullin et al. (1991). The compositions of plagioclase, garnet, and biotite from the L zone area of the Campbell mine were used in the calculation. The assemblage yields a temperature of $\approx 495^\circ\text{C}$, and an X_{CO_2} of ≈ 0.025 . Note that with increasing X_{CO_2} , carbonate, biotite, and garnet will replace anthophyllite, whereas increasing temperature at $X_{\text{CO}_2} \leq 0.025$ will not result in replacement of anthophyllite by carbonate. The numbered reactions refer to; 1 $7 \text{ Ann} + 7 \text{ An} + 4 \text{ Ath} + 14 \text{ CO}_2 = 25 \text{ aQz} + 7 \text{ Phl} + 7 \text{ Do} + 7 \text{ Alm} + 4 \text{ W}$, 2 $3 \text{ Alm} + \text{Do} + 3 \text{ Phl} + 8 \text{ W} = \text{aQz} + 2 \text{ Chl} + \text{An} + 3 \text{ Ann} + 2 \text{ CO}_2$, 4 $6 \text{ Ath} + 7 \text{ An} + 22 \text{ W} + 14 \text{ CO}_2 = 7 \text{ Chl} + 7 \text{ Do} + 41 \text{ aQz}$, 5 $41 \text{ Alm} + 16 \text{ Do} + 41 \text{ Phl} + 102 \text{ W} = \text{Chl} + 2 \text{ Ath} + 16 \text{ An} + 41 \text{ Ann} + 32 \text{ CO}_2$. Abbreviations; aQz = alpha quartz, Alm = almandine, An = anorthite, Ann = annite, Ath = anthophyllite, Chl = chlorite, Do = dolomite, Phl = phlogopite, W = H_2O .

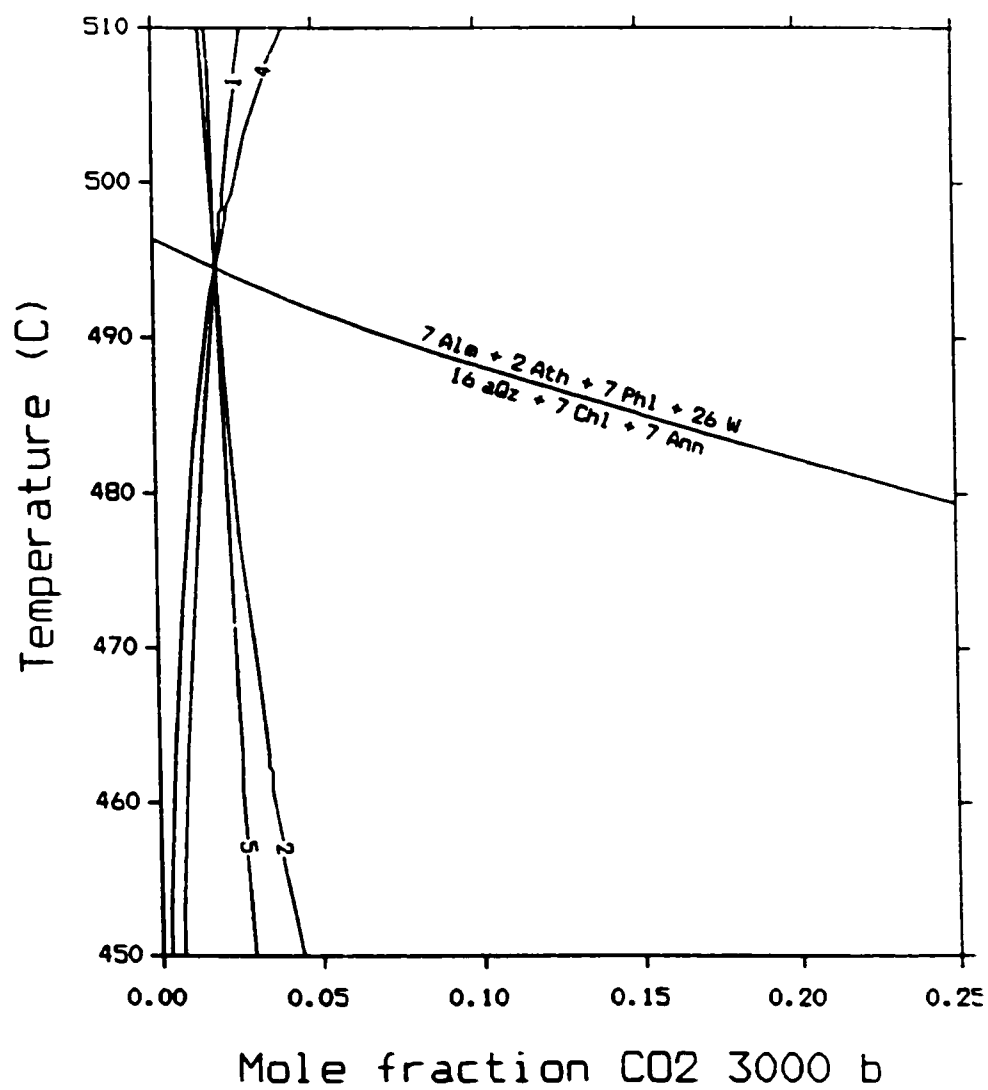


Figure 3.36

Geology plan map of the 17th level of the Campbell mine, showing geology, auriferous veins (A, F, G, L, NL), and the position of the garnet isograd. Note that the garnet isograd extends along the northwest trending A and F auriferous veins, and crosscuts lithologic units. Occurrences of abundant, large garnets are associated with the L ore zone. The Goldcorp Red Lake mine occurs to the right of the Campbell mine boundary.

Legend

- Basaltic rocks
- Rhyolitic rocks
- Ultramafic rocks

- Drill hole trace
- Garnet absent
- Garnet bearing
- Fine garnets in pillow rims

- Drifts and stopes
- Mine shaft
- Garnet isograd
- Inferred isograd
- A Ore zone

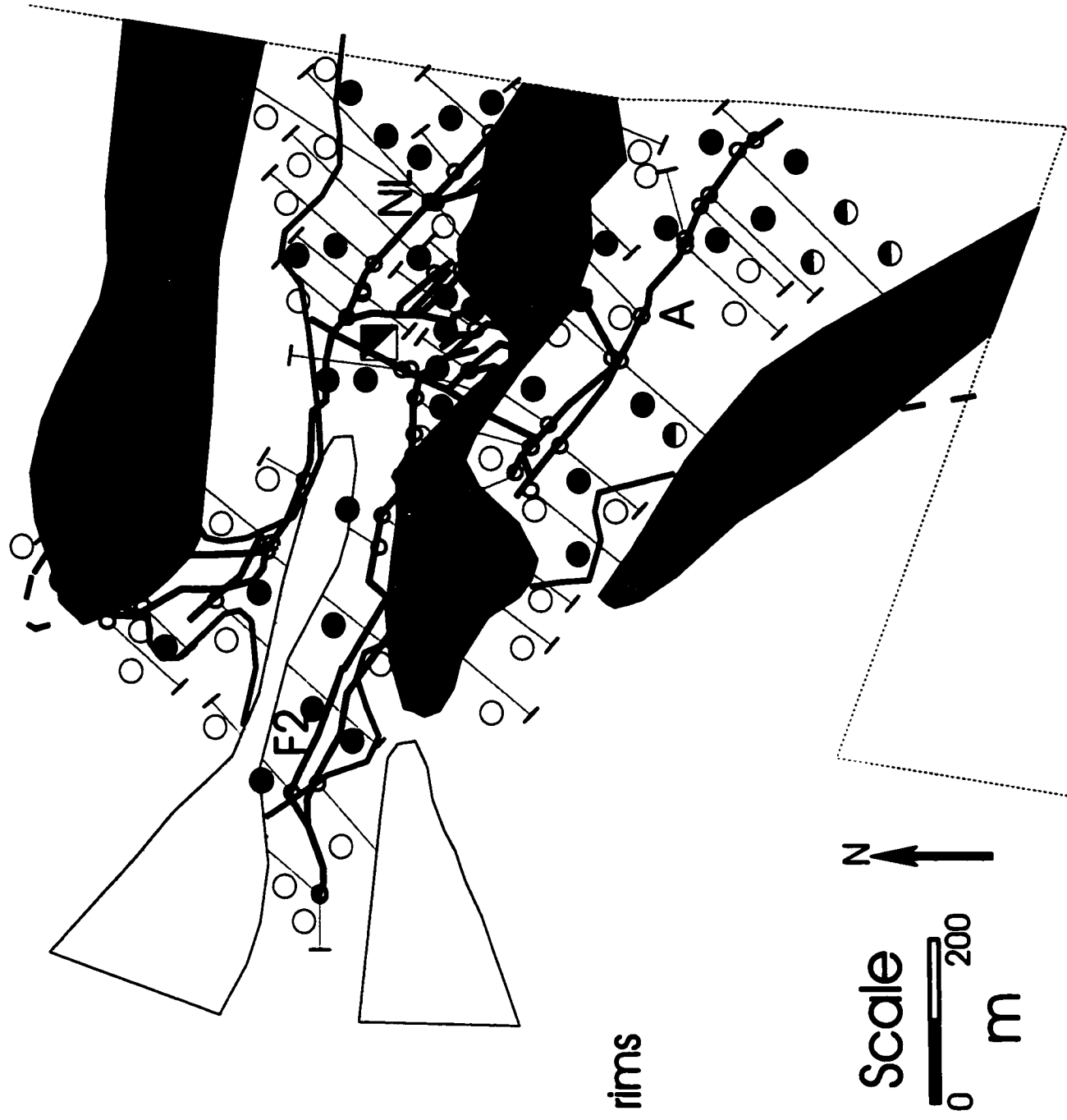
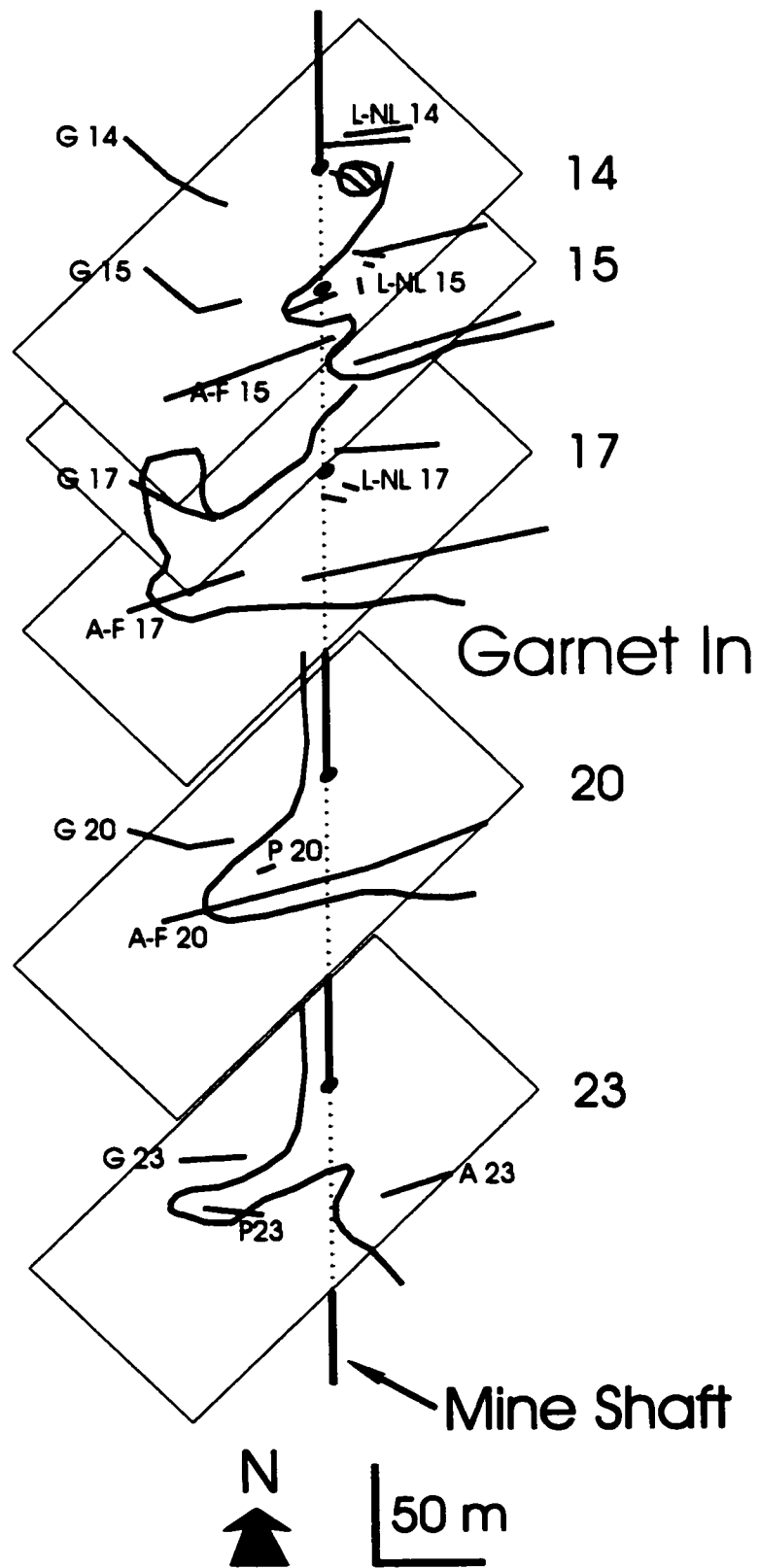


Figure 3.37

Three-dimensional view of the garnet isograd (red curves) at the Campbell mine. Garnet occurs on the east (right) side of the isograd.

The mine shaft (vertical line) intersections on mine levels (rectangles) are shown as solid circles. Heavy gray lines are auriferous veins and the numbers depicted correspond their levels: P 23 stands for P ore zone on 23 level. For clarity, the A-F ore zone has been excluded from the 14th level. Basaltic rocks of Assemblages 1 and 2 (Table 3.1) occur to the left of the isograd. The first appearance of garnet in basalt is marked by Assemblage 3 (Table 3.1). Note that the isograd is near vertical and attenuated along the A and F ore zone structure. Garnet occurs in interflow sediments and basalt throughout the underground workings on the 26th level of the mine (not shown).



4.0 Veining and Mineralization

4.1 Veining and auriferous structures

The auriferous veins at the Campbell mine are distinguished on the basis of their relation to the property wide NW trending foliation, as either oblique, or sub-parallel (“parallel”), following the nomenclature of MacGeehan and Hodgson (1982). These auriferous veins show differences in grade, tonnage, geologic location, and lateral and vertical extent. Foliation-parallel veins (A, F, F2, parts of NL ore zones) are NW trending, and hosted by basalt. Foliation-oblique veins (P, L, G, AU, and parts of NL ore zones) are NNW and E-W trending, and occur at basalt/ultramafic rock contacts. Auriferous veins at the mine are hosted by shear zones, characterized by ductile to brittle-ductile shear fabrics. The locations of auriferous veins at the Campbell mine (and adjacent Goldcorp mine) are provided in Figure 3.0. Carbonate veins and veinlets occur in mafic and ultramafic rocks throughout the mine, and are barren. Thick banded to crustiform quartz + carbonate veins are also barren. Ore is hosted by brecciated portions of the quartz + carbonate veins, and therefore post-dates most of the quartz+carbonate veins. Non-ore gold mineralization also occurs in basalt-hosted disseminated pyritic zones.

4.1.1 Quartz + Carbonate Veins

The proportions of quartz and carbonate in quartz + carbonate veins varies from 1:1, to 1:10. Narrow veins and veinlets (< 10 cm) are typically coarse-grained, composed of interlocking carbonate grains in bands of several mm’s in width. Large (> 1 m wide) banded and crustiform texture quartz + carbonate veins are called “snowbank” veins at the Campbell mine. Quartz + carbonate vein textures in the thicker veins vary from fine-grained crustiform banding (mm to cm scale) of quartz and carbonate (Fig. 4.0), and coarse-grained subhedral to euhedral carbonate rhombs with crustiform textures surrounding angular wallrock inclusions (“carbonate vein wallrock breccias”, Fig. 4.1). Angular wallrock inclusions display a cleavage or foliation, and commonly have a “jig-saw” (after Sibson, 1986) arrangement. The angularity of the fragments indicates minimal fragment transport, and the occurrence of foliations in wallrock fragments in carbonate veins preserving delicate carbonate crystal textures, indicate brecciation during

or after shortening. These breccias are interpreted as implosion breccias after the nomenclature of Sibson (1986), and formed during vein development. Wallrock breccias are most common in NNW trending quartz + carbonate veins at mafic-ultramafic rock contacts. Carbonate and quartz occur in the matrix of brecciated, banded carbonate veins, and grade into crustiform layers surrounding angular wallrock fragments. Quartz + carbonate veinlets in the wallrocks, are typically coarse-grained, locally with a medial band of fine-grained white to cream-yellow quartz. These quartz + carbonate veins, snowbank veins, and implosion breccias, are barren of gold mineralization. All quartz+carbonate veins at the mine display weak to moderate folding and boudinage. These veins also display epithermal-like textures, as described by Penczak and Mason (1997).

4.1.2 Auriferous Breccias

Several textural types of gold mineralization are hosted in breccias overprinting quartz + carbonate veins. Auriferous breccias host most of the gold at the Campbell mine, and although they occur principally in earlier quartz + carbonate veins, the orientation is not the same as the host vein. Breccia bodies are podiform to cylindrical, and show consistent moderate plunge to the SW, within the earlier carbonate vein. Ore chutes represent higher gold grade portions of the vein structure, rather than thicker portions of the vein structure. These breccia types show differences in texture and mineralogy, and an understanding of these breccias is critical to understanding the nature of auriferous mineralization. Some breccias show a low ratio of matrix to fragment, and are termed crackle breccias. The crackle breccia matrix is composed of a network of narrow quartz + pyrite + pyrrhotite veinlets, with minor sphalerite, magnetite, and arsenopyrite. Crackle breccias often occur adjacent to quartz + arsenopyrite breccias, described below.

The breccia variety with high matrix/fragment ratios, and a matrix composed predominantly of grey translucent quartz and arsenopyrite with pyrrhotite, and minor sphalerite, is termed here quartz + arsenopyrite breccia. Carbonate vein and wallrock fragments are sub-angular, and occur in a fine-grained matrix of quartz, arsenopyrite and minor carbonate, suggesting that they are hydrothermal breccias characterized by minor

fragment transport. The highest gold grades at the Campbell mine are associated with this style of mineralization. Arsenopyrite occurs as fine disseminated grains, bands and wisps within quartz. Sphalerite occurs as deep red wisps. Pyrrhotite and pyrite occur as disseminated grains, and patchy grain aggregates, constituting up to 30 vol.%. Auriferous quartz + arsenopyrite breccia is cut by fractures and veinlets of pyrite, pyrrhotite, chlorite and sericite, as well as carbonate and milky quartz veins with occasional tourmaline. Individual quartz + arsenopyrite breccia bodies have down-plunge lengths typically < 100 m. The widths of the auriferous breccias are 1 to 15 m. The ore zones are typically a composite of several auriferous hydrothermal breccia bodies.

4.1.3 Basalt-hosted foliation-parallel auriferous veins

Auriferous veins of this type occur in basalt-hosted crustiform to banded quartz + carbonate veins (25 cm to ≥ 100 cm thick), form NW-trending (125° to 135°) planar vein systems (Fig. 3.18), and have dip (65° to 75° SW) and strike continuity of ≥ 1 km. They are the A, F, F2, and parts of the NL vein systems (Fig. 3.0). Vein systems are composed of en-echelon vein segments on both plan and vertical section maps. Vein segments are offset into the hanging-wall to the NW (left-stepping), and down-dip. Some offsets in foliation parallel auriferous veins occur on the strike extensions of E-W shear zones hosting E-W foliation-oblique auriferous veins. White quartz veins occur in boudin necks (Fig. 3.18) in the vein, and narrow (1 to 5 cm) foliation-parallel quartz + carbonate veins occur between some vein segments. Quartz + carbonate veins are more commonly folded and boudinaged with increasing depth in the mine.

4.1.4 Foliation-oblique mafic-ultramafic contact-related auriferous veins

These auriferous veins occur close to, or along the contacts between mafic and ultramafic rocks, and are the most important for gold grade and tonnage at the Campbell mine. They include the G, L, AU, P, and parts of the NL ore zones (Fig. 3.0). Thick (> 1 m) colloform to crustiform quartz + carbonate veins hosting auriferous breccia mineralization in the G, L, and P ore zones are developed along the western, NNW trending margins of ultramafic units. Along E-W trending margins of ultramafic units, auriferous mineralization is hosted by E-W trending “sheeted” veinlet zones in ultramafic

rocks (Fig. 4.2). Mineralization of this style occurs in the AU ore zone, the SW parts of the NL ore zone, and the southern part of the L ore zone (Fig. 3.0). Gold mineralization forms podiform to pipe-like bodies which are most vertically extensive in NNW trending zones (≤ 100 m). Auriferous foliation-oblique veins have dip extents typically < 500 m. Auriferous mineralization also occurs in NW, E-W, and NNW veinlets in basalts (Fig. 4.3) adjacent to the foliation-oblique veins.

4.1.4.1 Foliation-oblique NNW-trending auriferous veins

North-northwest trending auriferous veins occur in thick (> 1 m) crustiform banded to banded texture quartz + carbonate veins, with rare carbonate vein wallrock breccia zones, at NNW trending mafic – ultramafic rock contacts. Quartz and carbonate layers in the veins are typically parallel to the vein margins, and veins are moderately boudinaged. Smaller (< 1 m wide) NW trending quartz + carbonate veins rotate into NNW trending veins, suggesting that some NW veins predated NNW veins. North-south trending vein splays are commonly folded (Fig. 3.13). Auriferous mineralization occurs in brecciated portions of the quartz + carbonate veins, and overprints folds. In ultramafic rocks adjacent to NNW trending auriferous veins, auriferous E-W to NW trending veinlet zones are developed, containing carbonate overprinted by fine-grained grey quartz, fuchsite, and arsenopyrite with subordinate pyrite, sphalerite, and pyrrhotite. In basalt, narrow (0.5 to 5 cm), spaced (decimeter to meter) NNW, E-W, and NW trending veinlets of quartz $\pm$ coarse-grained Fe-carbonate, with sphalerite, arsenopyrite, pyrite and pyrrhotite, and rare stibnite occur adjacent to the auriferous veins.

High-grade (> 150 g/t Au, locally > 500 g/t Au) quartz + arsenopyrite breccias form podiform and pipe-like bodies in NNW-trending contact veins, and are spatially associated with fold hinges along the NNW trending contacts (Fig. 3.15). These quartz + arsenopyrite breccias commonly have widths of ≤ 5 to 10 m, and down-plunge extents locally > 100 m. Quartz + arsenopyrite breccias are composed primarily of quartz with ≤ 30 vol.% arsenopyrite. Pyrrhotite (≤ 10 to 15 vol.%) and sphalerite (< 10 vol.%) occur as coarse-grained aggregates, blebs and wisps in quartz + arsenopyrite breccias. Banded quartz + carbonate vein fragments are locally preserved in the ore. Coarse-grained carbonate and white quartz veinlets cut some podiform ore bodies.

North-northwest trending banded to crustiform quartz + carbonate veins are also cut by NNW and E-W replacement zones (20 to 100 cm wide) of grey quartz with fine-grained arsenopyrite and minor biotite, pyrite, pyrrhotite, and tourmaline (Fig. 4.4). The zones have gradational margins, and are commonly continuous with auriferous quartz veins in the wallrocks. Scheelite and margarite occur in the grey, quartz-rich zones in the central and western parts of the deposit, and orthoamphibole occurs as bands in the eastern parts of the deposit.

Low Au-grade (< 2.5 to 3 g/t) dark grey veins of brecciated carbonate, with pyrite (< 20 vol.%) pyrrhotite (< 20 vol.%) and magnetite (40 to 60 vol.%) in the matrix are developed adjacent to some NNW trending ore bodies (Fig. 4.5). Sphalerite, and patchy silicification overprint magnetite. These breccia veins are < 5 to 10 cm wide, and cut E-W and NW quartz + carbonate veins (Fig. 3.15). They are generally present in basaltic wallrocks within 10 m of the ore veins, and are a volumetrically minor form of veining.

4.1.4.2 Foliation-oblique, E-W auriferous veins

Auriferous mineralization in ultramafic rocks is hosted by silicified, fuchsitic, arsenopyrite-rich patches overprinting E-W to WNW trending “sheeted” carbonate + quartz + chlorite veinlets, parallel to E-W trending mafic – ultramafic rock contacts. The E-W auriferous veins generally occur within 35 meters of the ultramafic-mafic contact. Veinlets are generally 5 mm to 5 cm wide and spaced 1 to 5 cm apart. Wider veins (10 to 15 cm) are spaced 20 to 40 cm apart. Non-auriferous sheeted veinlet zones are composed of coarse-grained fibrous carbonate and rare chlorite and grey to white quartz, with carbonate crystals oriented perpendicular to vein walls. Auriferous sheeted veinlets contain grey to green-grey fine-grained quartz, with abundant (10 to 15 vol.%) arsenopyrite, and minor coarse-grained carbonate. Altered wallrocks are steel-grey in color and contain abundant fine-grained arsenopyrite.

4.1.5 Basalt-hosted, auriferous disseminated pyrite

This style of auriferous mineralization (07 zone) is low-Au grade (< 2.5 to 3 g/t Au), with ≤ 15 to 20 vol.% disseminated pyrite. At present, this style of gold mineralization is not exploited. Auriferous mineralization is hosted by biotitic, white

mica $\pm$ fine-grained Fe-carbonate altered light brown to light green basalt with rare pyrrhotite and arsenopyrite. The fine-grained pervasive alteration is cut by cleavages and veinlets containing sphalerite, tourmaline, gahnite, and Fe-sulphides. SEM studies by the author indicate that fine-grained native gold occurs within cleavages and tourmaline veinlets. N-S trending, pygmatically folded quartz veinlets with coarse visible gold cut the disseminated auriferous mineralization. Basalt contains fine-grained carbonate (20 to 30 vol.%), white mica (10 to 15 vol.%), fine-grained quartz ($\leq$ 65 vol.%), biotite ($<$ 10 vol.%), and minor tourmaline and sphalerite. The 07 zone at the Campbell mine occurs in the upper east portion of the Campbell mine, beneath the surface infrastructure near the Campbell / Goldcorp boundary. This style of mineralization is equivalent to the auriferous disseminated sulphide zones at upper levels in the Goldcorp mine (i.e., the ESC-23 ore zone, Andrews et al., 1986)

Figure 4.0

Typical finely banded quartz + carbonate vein. The bands are composed of alternating millimeter scale layers of fine-grained white to grey quartz, and white carbonate (Fe-dolomite to ankerite). Scale in centimeters.

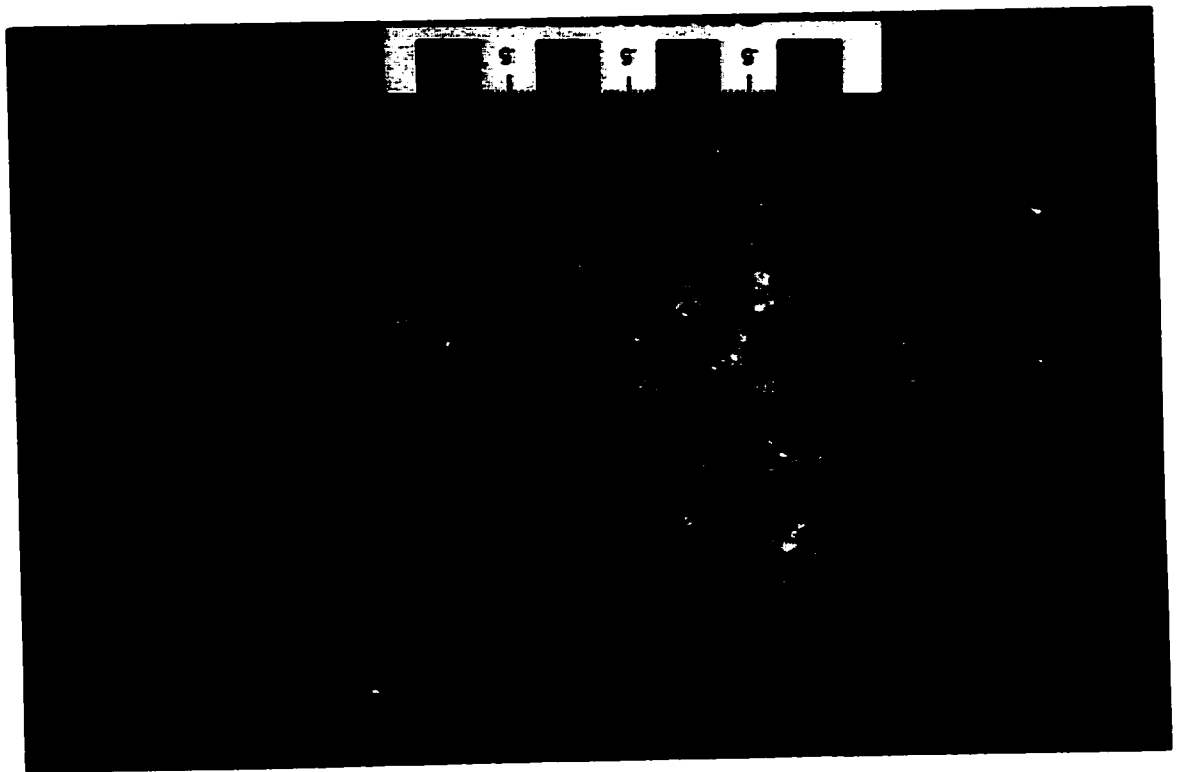


Figure 4.1

Breccias of angular wallrock basalt (dark grey to black) and banded quartz + carbonate vein fragments, in a matrix of banded to crustiform quartz + carbonate. Basalt breccias show a jig-saw arrangement, and the textures suggest minimal transport. The breccia occurs along a NNW trending quartz + carbonate vein, and suggests that carbonate veining occurred by implosive brecciation (Sibson, 1986). Photo of rib, 1854 stope, L ore zone. Screening is approximately 15cm square.

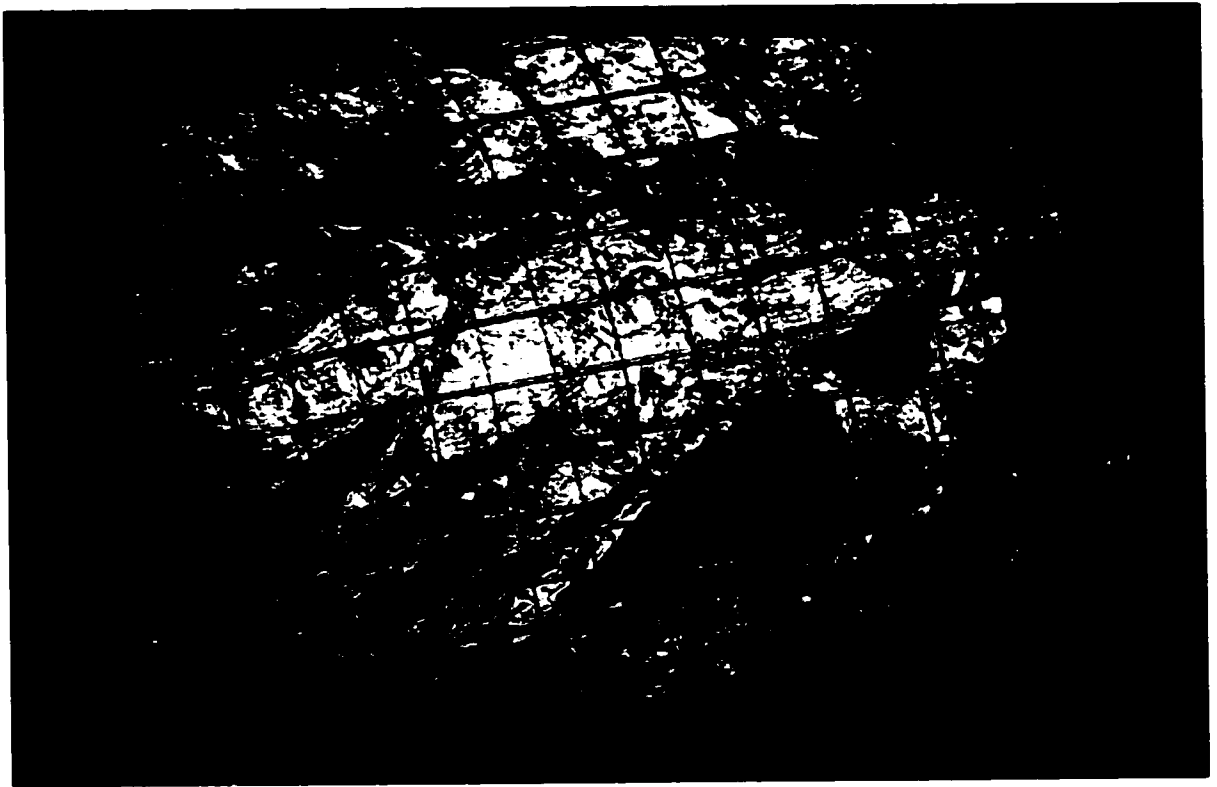


Figure 4.2

Non-auriferous sheared carbonate veinlets within an easterly trending shear zone in foliated ultramafic rocks. Carbonate crystals in the veinlets are coarse-grained and acicular, oriented approximately perpendicular to the vein walls. Photo viewing downwards, surface outcrop immediately north of the up-dip projection of the G ore body.

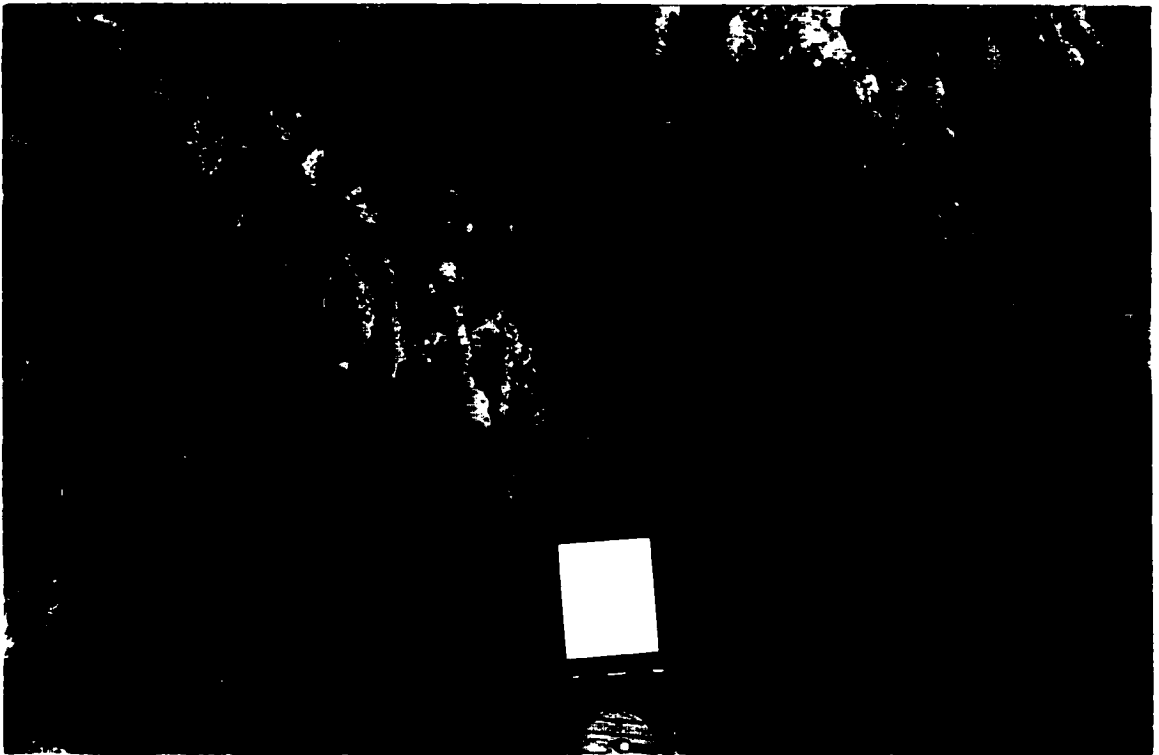


Figure 4.3

Auriferous quartz veinlet (0.5 cm wide, shown in purple) displaying a typical wallrock alteration zonation for the eastern part of the deposit. The veinlet contains grey translucent to transparent quartz with abundant visible gold (yellow). Wallrock alteration grades outwards from a steel-grey zone of fine-grained arsenopyrite and white mica (grey), to white mica, biotite and pyrite altered wallrocks (light yellow). Relatively unaltered wallrock basalt are dark grey to black (shown in green). Note that relatively early white fine-grained carbonate veinlets in the wallrock (blue) are overprinted by silicification in the altered wallrocks. Photo of face, 1961 stope, G zone.

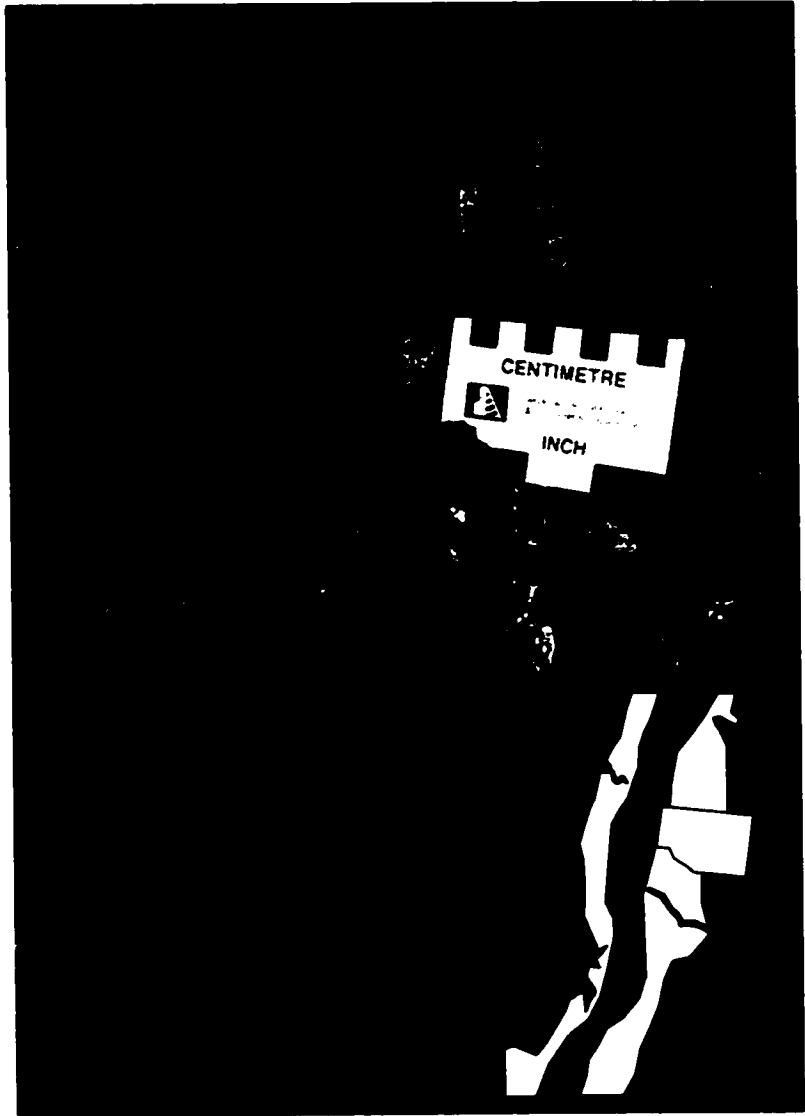


Figure 4.4

Replacement of crustiform banded quartz + carbonate vein by later stage auriferous, fine-grained quartz + arsenopyrite (grey vertical band, center of photo) mineralization, containing disseminated pyrite, pyrrhotite, scheelite, and margarite. Photo of face, 2061B stope, G zone.

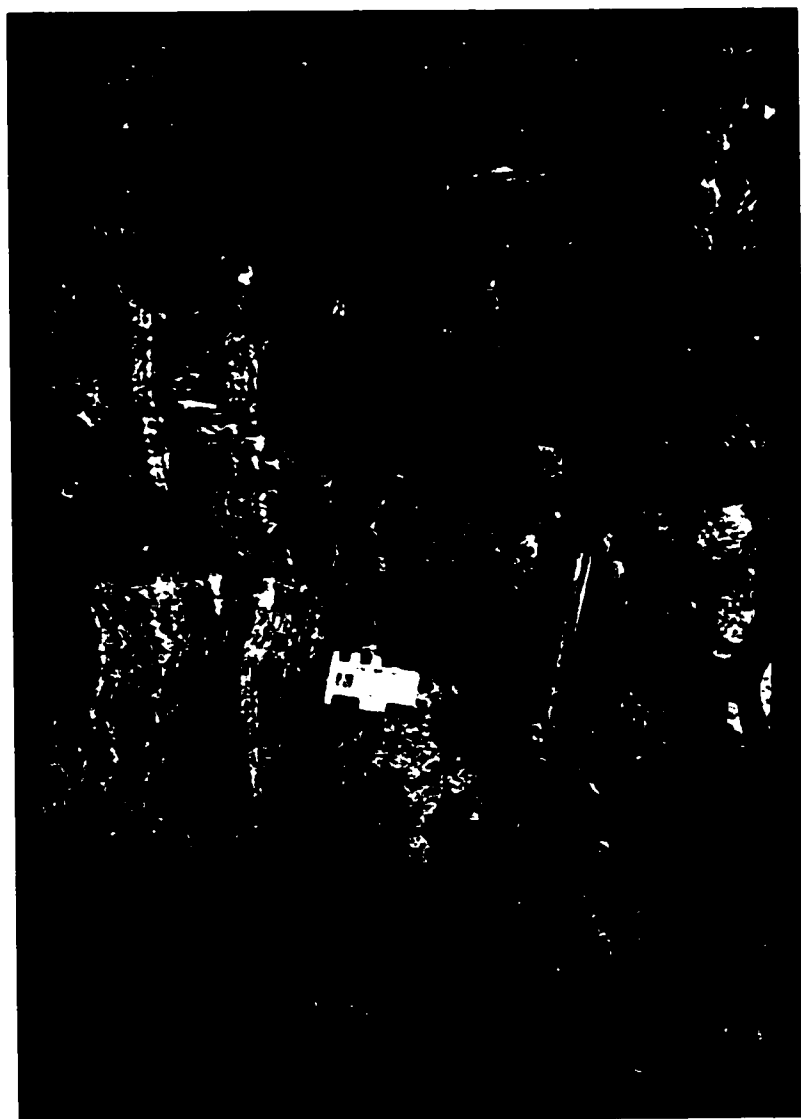
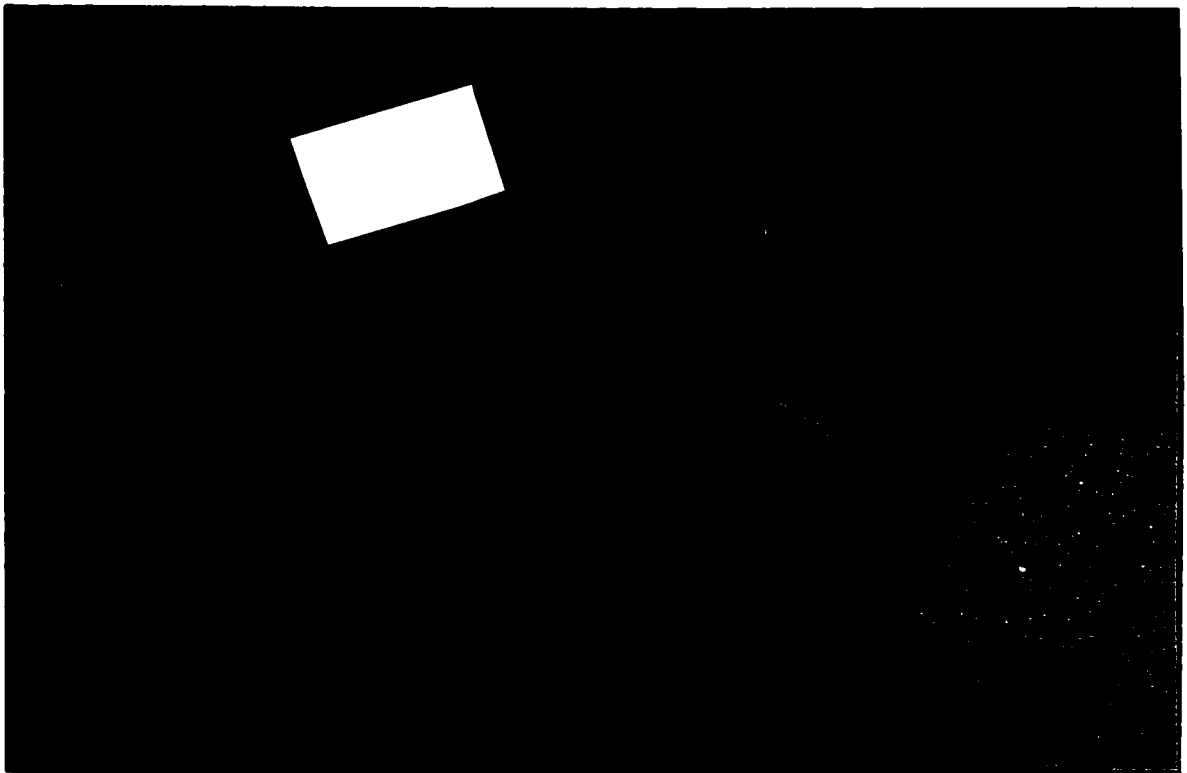
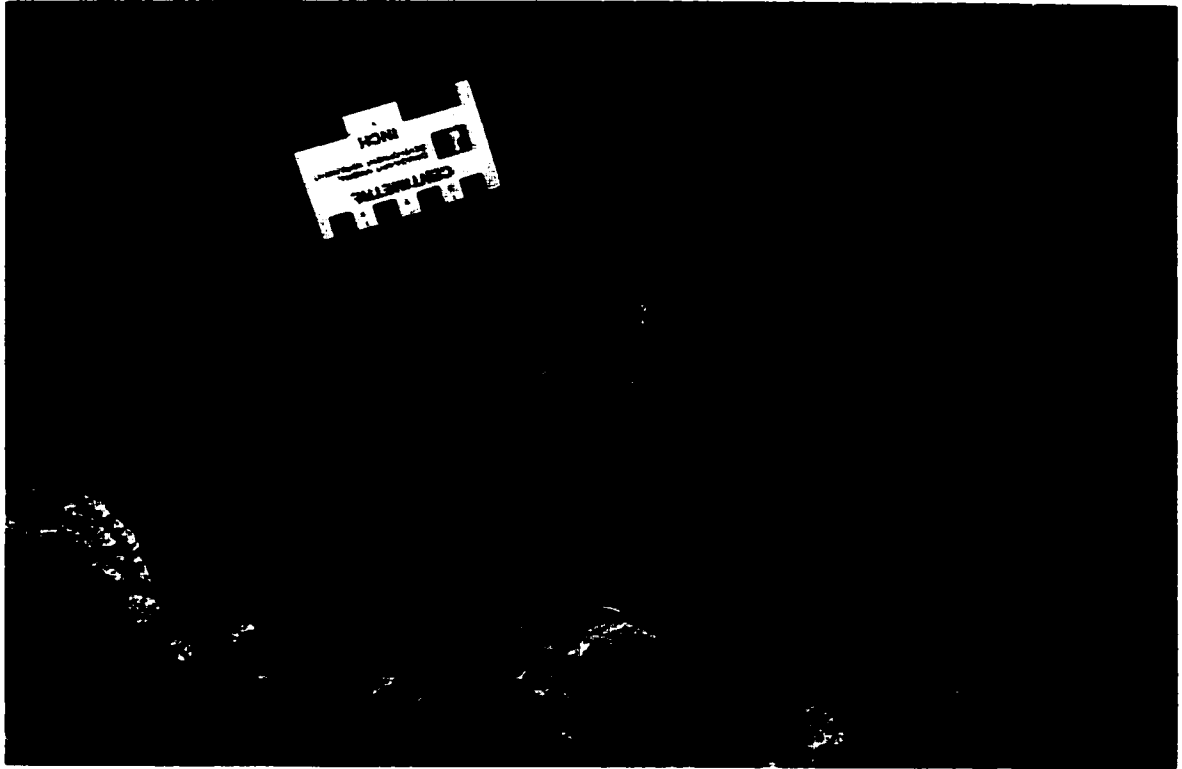


Figure 4.5

Brecciated basalt-hosted banded carbonate veins (blue), with a matrix (grey) containing magnetite and pyrrhotite. Patchy grey silicification (light grey) and sphalerite (red) overprint magnetite mineralization. Basalt wallrocks are shown in green. Photo of wall, 1656 stope, L zone.



5.0 Metallic vein mineralogy

The Au mineralization is associated with quartz, occurring as veinlets, silicified patches, and as the matrix of brecciated quartz + carbonate veins. Gold occurs as native gold and aurostibite (AuSb_2). Pyrite and pyrrhotite occur throughout the deposit, in ore and barren veins. Arsenopyrite shows a close spatial relationship with native Au and aurostibite in the ore veins throughout the deposit. Minor minerals include sphalerite, gersdorffite (NiAsS), Sb-bearing sulfides (ullmanite, $[\text{Ni},\text{Co}]\text{SbS}$; nisbite, chalcostibite), sulphosalts, and gahnite, all of which except sphalerite are fine-grained.

5.1 Arsenopyrite

Arsenopyrite is the most abundant metallic mineral (15 to 25 vol.%) in quartz + arsenopyrite breccias (Fig. 3.0). Arsenopyrite occurs as euhedral to subhedral disseminated grains (typically 50 to 100 μm) in auriferous veins and wallrocks immediately adjacent to auriferous veins. In ore, arsenopyrite occurs as bands, and rosettes mantling pyrite, pyrrhotite, and gahnite, and as inclusions in gahnite and pyrrhotite. In the P ore zone, arsenopyrite encloses magnetite, and is included within magnetite. Arsenopyrite and magnetite have sharp mutual contacts, suggesting equilibrium. Arsenopyrite contains significant Au (≤ 0.56 wt.%, Tarnocai et al., 1997) as a structurally bound component, and displays compositional zoning with Fe and S rich cores, and As and Au rich rims (Tarnocai et al., 1997).

5.2 Gold minerals

Au occurs as native gold and aurostibite. Native gold is present throughout the deposit, and aurostibite is restricted to the G, P, and F2 ore zones in the western portions of the mine. Native gold occurs as very fine blebs disseminated in quartz (Fig. 5.0), pyrite, tourmaline, chalcopyrite, and gersdorffite, and was deposited synchronously with arsenopyrite in quartz + arsenopyrite breccia ore. In the auriferous disseminated pyrite style gold mineralization, native gold occurs in fractures in pyrite, together with sphalerite (Fig. 5.1). Native gold is present as vermicular intergrowths with nisbite (NiSb_2) and ullmanite in pyrite, and in fractures in pyrite with sphalerite, tetrahedrite, and

berthierite. Native gold contains 4 to 30 mole% Ag, tending to be Ag-rich on the E side of the mine.

Leaf gold occurs in anastomosing fractures cutting quartz + arsenopyrite breccias, and coarse gold (≥ 0.25 cm) blebs occur in quartz veins and veinlets which cut auriferous mineralization. Leaf gold also cuts carbonate and quartz veinlets cutting auriferous mineralization. This suggests a second stage of gold, possibly representing remobilization of earlier deposited gold. This second stage of gold mineralization occupies late D₂ veinlets.

Aurostibite grains are small, generally < 20 μm , and rarely up to 1 mm in size. It is intergrown with ullmanite and sphalerite, and with chalcopyrite which encloses native gold. Aurostibite also encloses arsenopyrite, has smooth contacts with native gold grains (Fig. 5.2), and occurs as inclusions within tetrahedrite.

5.2.1 Spatial variations in the occurrence of Au minerals

Native gold occurs in ore throughout the mine. Occurrences of aurostibite are restricted to ore from the W part of the mine. Aurostibite breaks down at 460 °C to gold and native antimony (Hansen, 1958; Nekrasov, 1996) marking its upper temperature limit. Native gold associated with aurostibite at the mine contains ≈ 4 mole% Ag, indicating that the activity of Au is nearly 1, and the temperature conditions for this reaction are appropriate for aurostibite-bearing ore at the mine. The absence of aurostibite from the eastern parts of the mine suggests hydrothermal or metamorphic temperatures exceeded 460 °C, exceeding the thermal stability for aurostibite. Aurostibite does not accommodate Ag in its structure, and metamorphism of aurostibite results in native gold of high fineness (Nekrasov, 1996). However, native gold from aurostibite-free ore is Ag-rich (≈ 30 mole%) compared to aurostibite-bearing ore, suggesting pre-peak metamorphic deposition of aurostibite is unlikely.

5.3 Pyrite and pyrrhotite

Veins and wallrocks throughout the mine contain pyrite and pyrrhotite. Three stages of vein pyrite and pyrrhotite deposition are identified. The earliest stage accompanied, and may have slightly preceded, ore deposition. Pyrite is subhedral to

anhedral, and pyrite and pyrrhotite occur as fine disseminated grains, and grain aggregates in auriferous quartz, and at quartz grain boundaries, to 8 vol.%. Pyrite grains contain structurally bound Au contents of ≤ 168 ppm as patchy zoning (Tarnocai et al., 1997).

A second stage of Fe-sulphide deposition post-dating ore occurs as a matrix to brecciated auriferous quartz in quartz + arsenopyrite breccia bodies, and as veinlets cutting auriferous quartz. This stage of Fe-sulphide may be temporally equivalent to Fe-sulphides in the fractured neck regions of boudinaged, barren quartz + carbonate veins. A final stage of Fe-sulphide deposition occurs as veinlets cutting milky quartz veins and carbonate veinlets which post-date auriferous mineralization.

Pyrrhotite is more abundant than pyrite in ore veins than wallrocks, and somewhat more abundant with depth and in eastern parts of the mine. Pyrrhotite occurs as inclusions in garnet, arsenopyrite, and magnetite grains.

5.4 Sphalerite

Dusty, fine-grained aggregates, patches and wisps of light yellow to dark red sphalerite occurs in auriferous veins and altered wallrocks throughout the mine. Wispy sphalerite aggregates occur with pyrite in deformed, low gold grade (< 2 g/t Au) NNW trending quartz + carbonate veins.

5.5 Stibnite

Stibnite is rare, and occurs as sporadic grains in quartz, and coarse-grained aggregates (< 20 cm, commonly < 5 cm) in auriferous veins. Rare occurrences of stibnite as a matrix to brecciated, banded texture quartz + carbonate veinlets are observed in the G ore zone. Stibnite most commonly occurs in veins proximal to mafic-ultramafic rock contacts. Stibnite encloses arsenopyrite, and has not been observed with aurostibite.

5.6 Sb-bearing sulfides, and sulphosalts

These minerals are berthierite (FeSb_2S_4), jamesonite ($\text{Pb}_4\text{FeSb}_6\text{S}_{14}$), tetrahedrite and friebertite. Berthierite occurs in ore at intermediate levels (16th to 24th levels) throughout the mine, with pyrrhotite and arsenopyrite, and occasional magnetite.

Tetrahedrite, fribergite, and jamesonite are locally replaced by chalcostibite (CuSbS_2) along their margins, and the Sb-bearing sulfides and sulphosalts replace pyrite (Fig. 5.3). Although common in ore, these minerals constitute < 1 vol.% of the vein.

5.7 Magnetite

Magnetite occurs as anhedral grains (< 150 μm) in quartz + arsenopyrite breccia ore, and as the primary matrix component in magnetite breccias which overprint earlier banded quartz + carbonate veins (Fig. 4.5). In quartz + arsenopyrite breccias, magnetite occurs as inclusions in arsenopyrite, and in smooth contact with arsenopyrite. This suggests equilibrium and co-deposition of arsenopyrite and magnetite. Magnetite occurs as inclusions within pyrrhotite, and commonly mantles pyrite and pyrrhotite. Contacts between the Fe-sulphides and magnetite vary from smooth, to irregular, suggesting variable equilibrium, and overprinting of the Fe-sulphides by magnetite. Magnetite in magnetite-rich breccias overprinting earlier quartz + carbonate veins are coarse-grained, and occur with quartz, pyrite and pyrrhotite. The breccias overprint veining in E-W quartz + carbonate veinlet zones. The breccias are weakly auriferous, and cross cutting relations suggest that brecciation occurred approximately coeval with, or after auriferous hydrothermal activity.

5.8 Summary of the metallic minerals

Gold minerals show a strong spatial and temporal relation with arsenopyrite. The sulphosalts, Sb-bearing sulphides, pyrite, pyrrhotite, chalcopyrite, and sphalerite were deposited with Au in auriferous veins. Pyrite and arsenopyrite contain significant structurally bound Au (Tarnocai et al., 1997). Leaf gold cuts post-ore veins, suggesting local remobilization of Au. A paragenetic sequence for the vein metallic and gangue minerals is provided in figure 5.4. The spatial variations in the occurrence of the Au minerals are unlikely to be a product of metamorphism of an earlier Au bearing assemblage. In addition, the preservation of fine textural and compositional zoning, including Au zoning, in pyrite and arsenopyrite (Tarnocai et al., 1997), are unlikely to survive amphibolite facies peak metamorphic conditions. These observations suggest that auriferous mineralization occurred during, or after peak metamorphism.

Figure 5.0

Reflected light photomicrograph of fine disseminated native gold (au) with pyrrhotite (po) in a quartz + arsenopyrite breccia, L ore zone. Light grey matrix is quartz with minor carbonate. The width of view is approximately 0.75 mm.

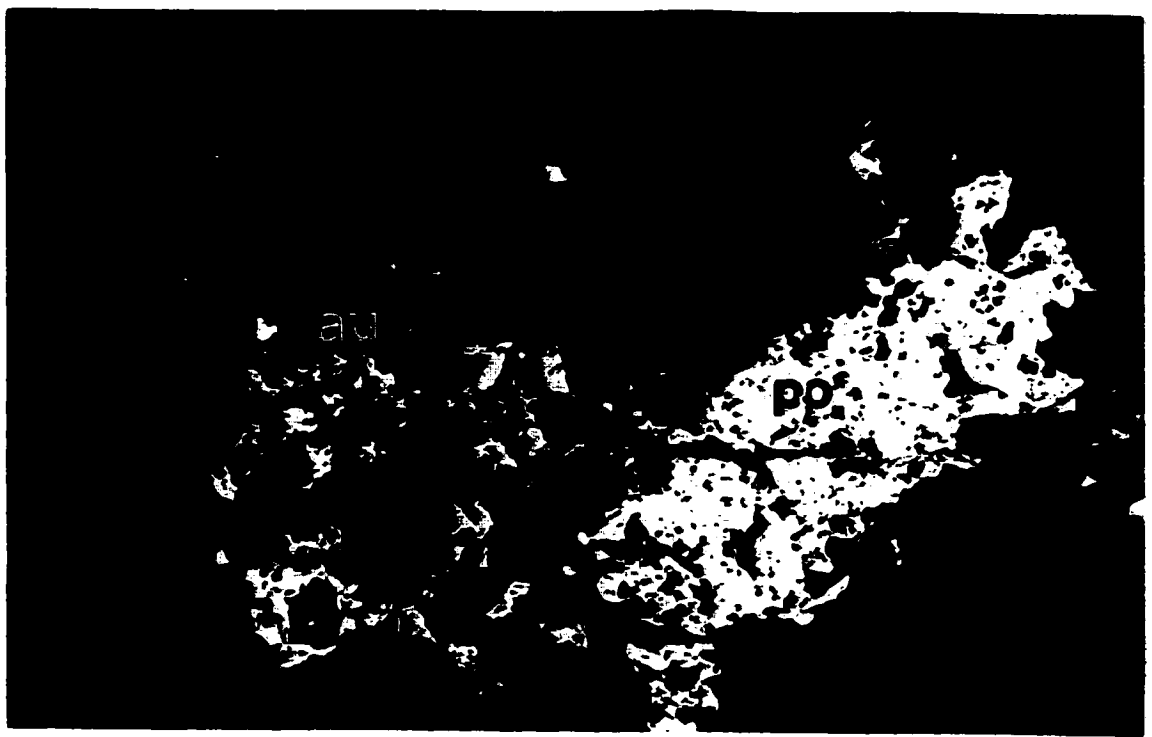


Figure 5.1

Backscattered electron image of pyrite (Py), with fractures containing native gold (Au) and sphalerite (Sph). Sample from the 07 ore zone, Campbell mine.

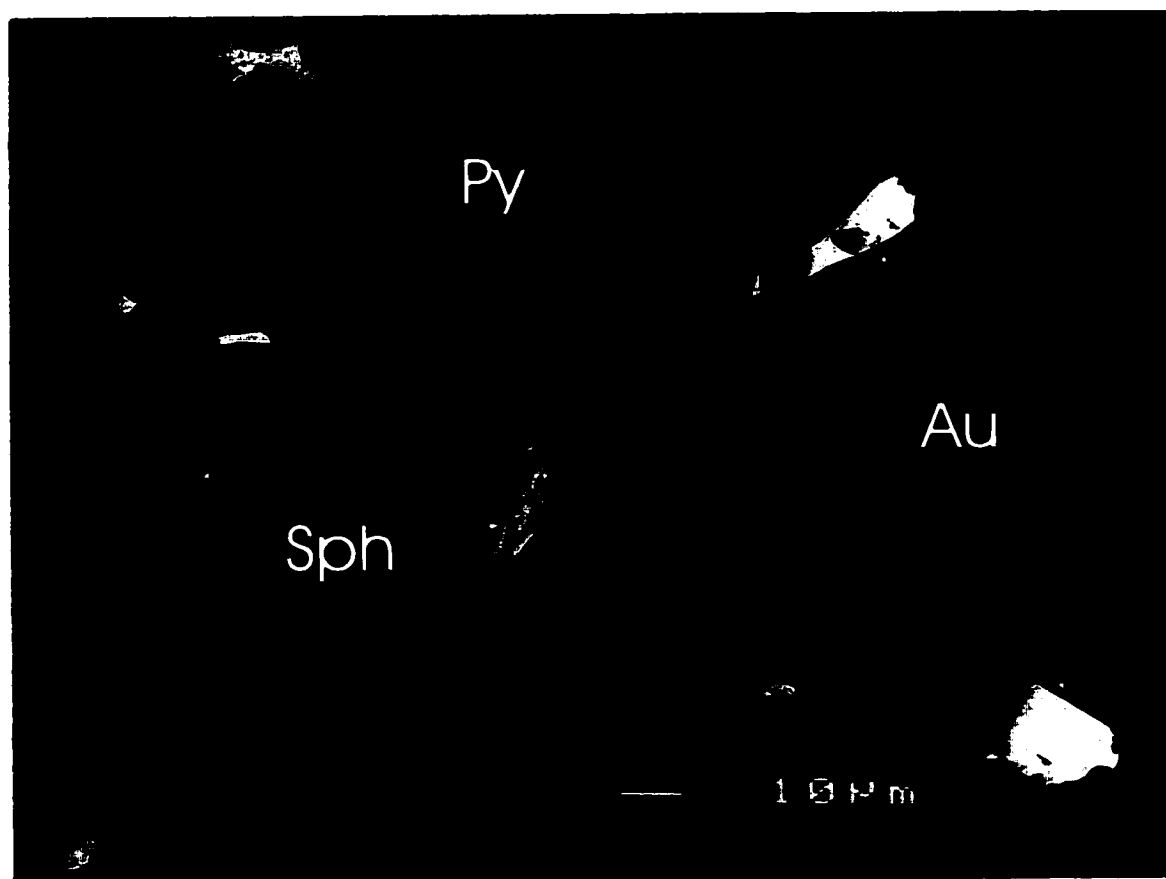


Figure 5.2

Backscattered electron image of aurostibite (ast) in contact with native gold (au). The moderate grey mineral is carbonate, and the black matrix is quartz. Sample from the P ore zone.



Figure 5.3

Backscattered electron image of tetrahedrite (Tet) and jamesonite (Jms) replacing pyrite (Py). Tetrahedrite is partially mantled by chalcostibite (Csb). The black matrix is predominately quartz. Fine-grained aurostibite (Ast) occurs at the bottom of the photo. P ore zone.

Csb

Tet

Py

Ast

Figure 5.4

Paragenetic sequence for the vein metallic and gangue minerals. Red = Au minerals, purple = metallic minerals, green = non-metallic mineral, blue = non-metallic minerals which does not occur throughout the deposit. The volume of the ellipse indicates relative volume of deposition of the mineral over time, and does not imply the modal abundance of the mineral, relative to the other minerals on the diagram. The dashed lines indicate minor deposition of the mineral.

The pre-ore carbonate alteration event (Stage 1 carbonate) involved the development of banded to crustiform texture quartz + carbonate veins, and carbonate veinlets, prior to auriferous mineralization. Auriferous quartz + arsenopyrite hydrothermal breccias (QAP), and auriferous quartz veins and veinlets overprint the Stage 1 carbonate veins. Post-ore veins are quartz $\pm$ tourmaline, pyrite, pyrrhotite, and carbonate veins that cut auriferous veins. Local coarse native gold occurs in the late veins where they cut auriferous veins. This second gold event is interpreted as local remobilization of gold from the auriferous veins. Black line faults are discussed in the text, and cut auriferous veins and the post-ore veinlets. Scheelite occurs locally in auriferous veins and breccias, and banded to crustiform carbonate veins, in the western part of the Campbell mine. Gahnite, plagioclase, and amphibole (anthophyllite – gedrite, and hornblende) occur in auriferous veins from the eastern part of the Campbell mine, hosted by amphibolite facies rocks.

Time



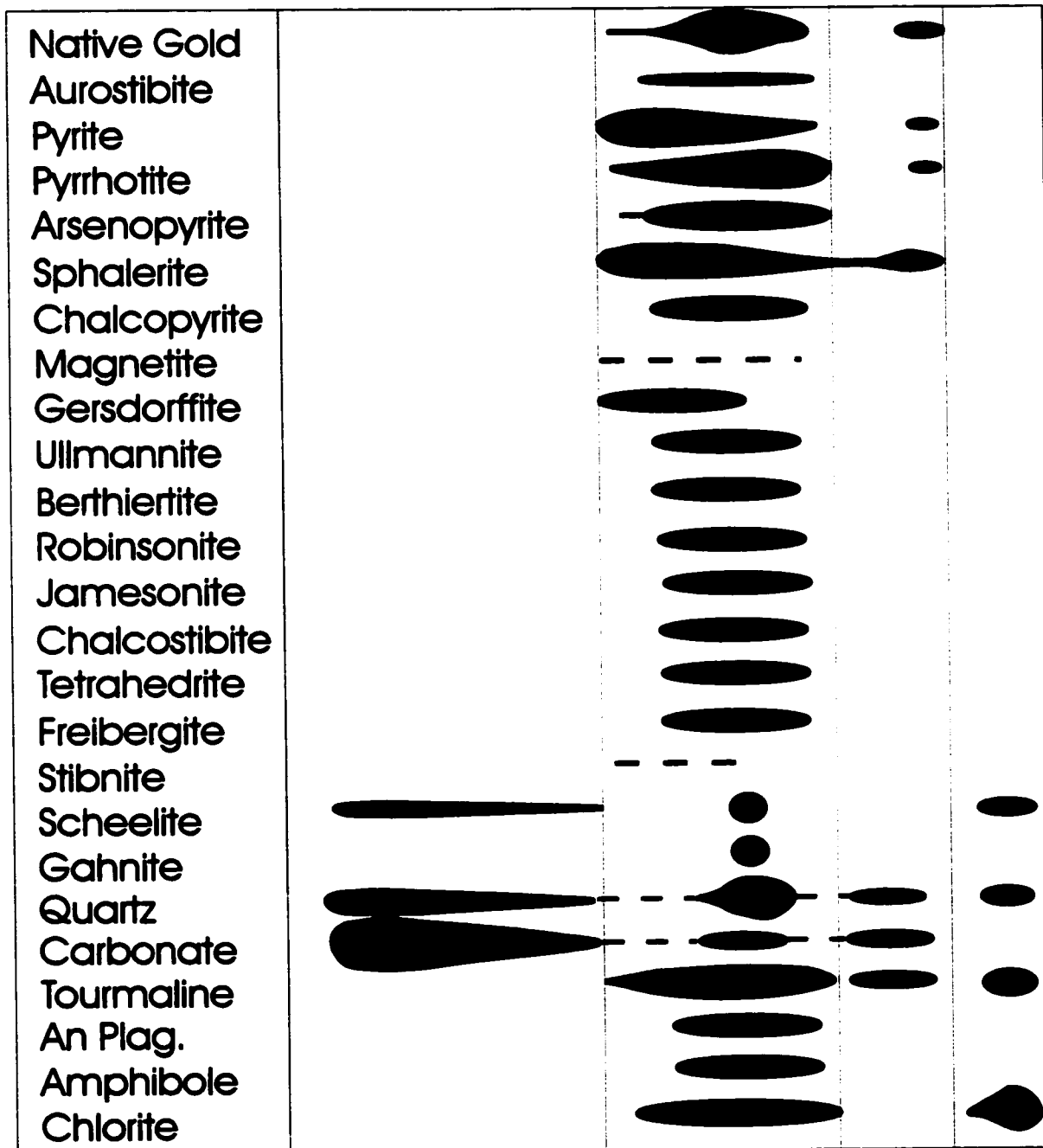
Vein Mineral

Pre-ore carbonate
veining

Auriferous QAP and
quartz veins

Post-ore
veins

Black line
faulting



6.0 Non-metallic vein minerals

6.1 Quartz

Three stages of quartz deposition are identified based on cross-cutting relationships. The earliest stage accompanied deposition of carbonate in banded to crustiform quartz + carbonate veins. In hand sample, this quartz is light grey and translucent to opaque. In thin section, quartz is intergrown with euhedral carbonate.

Fine-grained, second stage translucent to transparent grey quartz is the principal gangue mineral associated with auriferous mineralization, and cuts earlier quartz in quartz + carbonate veins. Quartz occurs as auriferous quartz veinlets, as a matrix in auriferous, quartz + carbonate vein-hosted quartz + arsenopyrite breccia ore, and as patchy replacement of the wallrocks. Quartz from veins displays weak undulose extinction, serrate grain boundaries, and commonly forms tabular grains parallel to the vein margins. Contacts between early carbonate and quartz, and second stage quartz are irregular, suggesting replacement of the early carbonate, and replacement and recrystallization of first stage quartz.

Third stage quartz is milky white, contains abundant fluid inclusions and rare tourmaline. It occurs in veins in boudin necks cutting banded to crustiform quartz + carbonate veins, veins in lamprophyre dykes, and locally cuts quartz + arsenopyrite breccia ore. Quartz grains are subhedral, and display very weak undulose extinction.

6.2 Carbonate

Carbonate in the mine is ankerite to ferroan dolomite, with minor calcite and siderite (Fig. 6.0). Three stages of carbonate deposition are identified, from crosscutting relations with auriferous mineralization, and relations to the foliation. The earliest stage of carbonate (Stage 1) deposition involved pervasive veins and veinlets with variable amounts of quartz, and as finely disseminated grains and patches throughout the mafic and ultramafic wallrocks. This stage of deposition may have been synchronous with the development of banded to crustiform texture quartz + carbonate veins. Carbonate grains contain abundant fluid inclusions, and display fine zonations in composition and fluid inclusion content.

Carbonate occurs as veinlets, and along cleavage planes in rhyolitic rocks, which are subparallel to the foliations in adjacent mafic and ultramafic rocks, and are related to the NE-SW shortening event. This indicates that carbonate deposition occurred during, or after the development of the F_2 -related foliations (Fig. 3.12a,b). Foliation-oblique shear zones overprint pervasive carbonate alteration (Fig. 3.20), and deform quartz + carbonate veins, suggesting shear occurred during, and continued after carbonate deposition.

Northwest, NNW and E-W carbonate veins were formed prior to brecciation which accompanied the introduction of Au and quartz, as auriferous mineralization overprints all quartz + carbonate textures in these veins. Carbonate spatially related to high gold-grade mineralization occurs as fragments of crustiform texture veins, locally rimmed by pyrite, pyrrhotite, and arsenopyrite. This stage of carbonate clearly pre-dates auriferous mineralization. Syn-ore (Stage 2) carbonate occurs as volumetrically minor (< 5 vol. %) fine carbonate blebs and wisps in quartz, and locally as tension fractures cutting auriferous veins. Carbonate is anhedral, and contains abundant fluid inclusions, with zonation in fluid inclusion abundance from cores to rims of grains. Veinlets containing auriferous quartz, biotite, and tourmaline cut Stage 2 carbonate, indicating it is coeval with auriferous veining.

A later stage of carbonate (Stage 3) occurs as fibrous texture sheeted carbonate veinlets cutting pervasive carbonate alteration, and coarse-grained carbonate veinlets cutting high Au-grade quartz + arsenopyrite ore bodies. It is unclear if these two occurrences of late carbonate are temporally related. Carbonate compositions are provided in Table 6.0.

6.3 Scheelite

The distribution of scheelite was investigated with an ultra-violet lamp during field work. It occurs sporadically in the western part of the mine, and is absent from eastern parts of the deposit. Scheelite was observed in auriferous, silicified, arsenopyrite-rich replacement bands in NNW-trending contact-hosted quartz + carbonate veins (Fig. 4.4), "black line faults" which consist of chlorite + carbonate + tourmaline + quartz cutting auriferous

Table 6.0. Representative carbonate compositions.

	1	2	3	4	5
CaO	29.73	30.04	55.35	29.21	29.67
MgO	14.63	13.17	0.00	12.93	18.30
MnO	1.63	1.36	0.65	0.68	0.18
FeO	10.52	11.68	0.00	12.62	3.76
Total	56.51	56.25	56.00	55.44	51.91
Formula (Basis 6 Oxygens)					
Ca ⁺²	2.994	3.078	5.945	3.044	3.058
Mg ⁺²	2.049	1.877	0.000	1.874	2.625
Mn ⁺²	0.130	0.110	0.055	0.056	0.015
Fe ⁺²	0.827	0.935	0.000	1.026	0.303
Total	6.000	6.000	6.000	6.000	6.001
Mol.% Endmembers					
CaCO ₃	49.90	51.30	99.08	50.73	50.96
MgCO ₃	34.15	31.28	0.00	31.23	43.74
MnCO ₃	2.17	1.83	0.92	0.93	0.25
FeCO ₃	13.78	15.58	0.00	17.10	5.05

- 1,2 Fine-grained disseminated ankerite from auriferous disseminated pyrite, 307E drift.
- 3 Calcite from foliation-parallel auriferous quartz veinlet in the wallrocks of the G ore zone, 1961 stope.
- 4 Ankerite from a basalt-hosted foliation-parallel auriferous vein, A zone 1702 W stope.
- 5 Ferroan dolomite from a coarse-grained carbonate veinlet cutting podiform quartz + arsenopyrite breccia. L zone, 1756-2E stope.

veins, and most abundantly (≤ 3 vol.%) in auriferous quartz + arsenopyrite breccia ore (Fig. 6.1). This suggests at least two stages of scheelite deposition, one occurring during ore deposition, and a later, post-ore stage of scheelite mineralization associated with “black line” faulting. In auriferous mineralization, scheelite (< 1 mm, commonly < 30 μm) contains quartz and pyrrhotite inclusions, and is locally cut by fractures containing carbonate, quartz, Fe sulfides, and tourmaline. Scheelite from black line faults is coarse-grained (< 1.75 cm), and intergrown with quartz and chlorite.

6.4 Gahnite

Gahnite (50 to 75 μm) occurs in quartz + arsenopyrite ore and disseminated pyritic zones as subhedral to anhedral grains, and aggregates of gahnite grains form lenses to (locally) 10 vol.%. Within auriferous disseminated pyrite mineralization, gahnite occurs in fractures cutting disseminated pyritic mineralization, with biotite, chlorite, tourmaline, sphalerite and pyrite. Gahnite contains sphalerite inclusions, and also has smooth grain boundary contacts with sphalerite (Fig. 6.2).

In quartz + arsenopyrite breccias, gahnite is associated with quartz, biotite, sphalerite, pyrite, pyrrhotite, and arsenopyrite. Arsenopyrite occurs as both mantles on gahnite, and as inclusions in gahnite (Fig. 6.3). Wallrock basalt fragments in auriferous hydrothermal breccias are extensively replaced by gahnite, arsenopyrite, pyrrhotite, and quartz. Gahnite in quartz + arsenopyrite breccias are pseudomorphed by quartz.

Gahnite from auriferous disseminated pyrite mineralization occurs with (locally) 3 to 5 vol.% sphalerite, and are Zn-poor compared to gahnite from relatively sphalerite-poor, podiform quartz + arsenopyrite ore. Gahnite compositions are provided in Table 6.1.

6.4.1 Spatial variations in the occurrence of gahnite

Gahnite is uncommon in Archean gold deposits, and is typically associated with volcanogenic massive sulfide deposits metamorphosed to the amphibolite facies. Gahnites from the mine have compositions similar to spinel from massive sulfide deposits and aluminous metasediments (Table 6.1, Fig. 6.4). Although gahnite has been

Table 6.1. Representative gahnite compositions

	1	2	3
Al₂O₃	55.41	55.33	54.06
Cr₂O₃	0.00	0.00	0.18
Fe₂O₃	1.65	1.27	1.94
V₂O₃	0.27	0.00	0.00
FeO*	6.09	6.03	5.09
MgO	1.34	1.19	1.25
ZnO	35.61	35.56	35.94
Total	100.37	99.38	98.46
Formula (Basis 32 Oxygens)			
Al⁺³	15.649	15.769	15.607
Cr⁺³	0.000	0.000	0.036
Fe⁺³	0.297	0.231	0.358
V⁺³	0.053	0.000	0.000
Σ B site	15.999	16.000	16.001
Fe⁺²	1.221	1.219	1.042
Mg⁺²	0.480	0.430	0.458
Zn⁺²	6.301	6.350	6.499
Σ A site	8.002	7.999	7.999
Mole% Endmembers**			
Spl	5.89	5.30	5.58
Hc	14.65	15.02	12.71
Ghn	77.28	78.24	79.25
Mgn	0.11	0.08	0.13
Mag	0.28	0.22	0.29
Frk	1.47	1.15	1.82

* Fe⁺² by charge balance

TiO₂ and MnO below detection limits, = 0.01 wt.%

** Abbreviations: Spl=spinel, Hc=hercynite, Ghn=gahnite, Mgn=magnesioferrite, Mag=magnetite, Frk=franklinite.

- 1,2 Subhedral gahnite from basalt-hosted disseminated pyritic ore. Gahnite occurs in fractures with sphalerite, tourmaline, and chlorite. 307E drift.
- 3 Subhedral gahnite from contact-hosted, podiform quartz + arsenopyrite ore. 1756-2E stope, L zone.

recorded in greenschist facies sedimentary rocks, it is rare and extremely fine-grained (Spry and Scott, 1986), unlike the gahnite occurrences at the mine. Gahnite occurrences are restricted to the eastern part of the mine, indicating either these rocks have experienced amphibolite grade metamorphism, or gahnite was formed under syn-amphibolite facies metamorphic hydrothermal conditions. The occurrence of auriferous arsenopyrite as inclusions in gahnite, and also mantling gahnite (Fig. 6.3), indicates that the growth of gahnite was synchronous with auriferous hydrothermal activity (Tarnocai et al., 1997, Appendix 9).

6.5 Tourmaline

Tourmaline occurs in barren and auriferous veins, and in the wallrocks throughout the mine. Green tourmaline occurs along foliation planes in basalt and ultramafic rocks, and in chloritic cleavages in rhyolites. Garnet contains tourmaline inclusions oriented parallel to the foliation, and garnet and andalusite are cut by fractures containing tourmaline. This suggests a phase of tourmaline growth during the development of NW cleavages and foliations and metamorphism.

In auriferous veins, tourmaline is green, or zoned with green rims and greenish-brown, relatively Ti and Fe-rich cores. Tourmaline contains inclusions of quartz, pyrrhotite, and rare native gold, sphalerite, and arsenopyrite. Tourmaline not spatially associated with auriferous veins has intermediate dravite - schorl composition and is relatively Fe-rich. Tourmaline in auriferous veins and the altered wallrocks immediately adjacent to auriferous veins is relatively Al- and Mg-rich (Fig. 6.5, Table 6.2).

6.6 Plagioclase

Plagioclase occurs as anhedral grains (< 1 mm) in veins, intergrown with quartz and cut by carbonate and quartz. Plagioclase are An-rich (Table 6.3), between An₆₃₋₉₂, except for andesine (An₃₀), from garnetiferous basalt.

Table 6.2. Representative tourmaline compositions.

	1	2	3	4	5	6	7
SiO ₂	36.74	36.95	35.69	34.89	35.59	36.00	37.56
Al ₂ O ₃	34.41	36.64	29.85	28.28	32.68	33.12	33.09
Fe ₂ O ₃ *	0.00	0.00	0.00	1.17	0.00	0.87	0.00
FeO	6.77	4.07	8.29	8.49	8.46	6.00	4.38
MgO	6.35	6.53	8.58	8.10	5.89	5.69	7.40
MnO	0.00	0.00	0.00	0.00	0.00	0.16	0.00
TiO ₂	0.00	0.00	0.79	0.83	0.34	0.29	0.20
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.38	0.29
CaO	0.00	0.00	1.42	1.54	0.67	0.64	0.23
Na ₂ O	2.62	3.17	2.68	2.09	1.80	0.00	2.31
Total	86.89	87.36	87.30	85.39	85.43	83.15	85.46
Formula (Basis 49 Oxygens)							
Si ⁺⁴	11.844	11.689	11.724	11.776	11.798	12.021	12.151
Al ⁺³	12.156	12.311	11.554	11.247	12.202	11.979	11.849
Σ Z site	24.000	24.000	23.278	23.023	24.000	24.000	24.000
Al ⁺³	0.920	1.352	0.000	0.000	0.570	1.055	0.770
Fe ⁺³	0.000	0.000	0.000	0.296	0.000	0.218	0.000
Fe ⁺²	1.827	1.076	2.277	2.395	2.347	1.675	1.185
Mg ⁺²	3.052	3.082	4.201	4.076	2.911	2.834	3.567
Mn ⁺²	0.000	0.000	0.000	0.000	0.000	0.046	0.000
Ti ⁺⁴	0.000	0.000	0.195	0.210	0.085	0.073	0.050
Cr ⁺³	0.000	0.000	0.000	0.000	0.000	0.100	0.075
Σ Y Site	5.799	5.510	6.673	6.977	5.913	6.001	5.647
Ca ⁺²	0.000	0.000	0.499	0.558	0.240	0.229	0.081
Na ⁺	1.639	1.942	1.706	1.366	1.156	0.000	1.449
Σ X Site	1.639	1.942	2.205	1.924	1.396	0.229	1.530

K₂O below detection limit, = 0.01 wt. %

* Fe⁺³ determined by charge balance.

- 1,2 Pale green tourmaline with sphalerite, chlorite and gahnite, in fractures from basalt-hosted disseminated pyritic ore. 307E drift.
- 3,4 Green to blue-green tourmaline along the foliation, in basalt. Drill core 14-470
- 5 Fine-grained, brown-green tourmaline in auriferous quartz vein. Drill core 27-267.
- 6 Tourmaline from contact-hosted, podiform quartz + arsenopyrite ore. P zone 2351 stope.
- 7 Brown tourmaline with Au inclusion from carbonate breccia-hosted ore in a foliation-parallel quartz-carbonate vein, L zone 1854 stope.

Table 6.3. Typical plagioclase compositions.

	1	2	3	4	5	6
SiO ₂	46.04	59.78	44.89	46.04	45.65	51.36
Al ₂ O ₃	33.80	24.48	35.37	34.69	34.47	30.55
Fe ₂ O ₃	0.56	0.13	0.29	0.35	0.47	0.31
Na ₂ O	1.35	8.05	0.87	1.59	1.17	4.43
CaO	17.54	6.18	18.94	16.66	17.45	13.39
K ₂ O	0.00	0.08	0.00	0.00	0.00	0.00
Total	99.29	98.70	100.36	99.33	99.21	100.04

Formula (Basis 32 Oxygens)

Si ⁺⁴	8.537	10.780	8.265	8.506	8.463	9.352
Al ⁺³	7.388	5.202	7.678	7.555	7.532	6.557
Fe ⁺³	0.078	0.017	0.041	0.049	0.066	0.043
Σ Z site	16.003	15.999	15.984	16.110	16.061	15.952

Na ⁺	0.485	2.813	0.311	0.568	0.419	1.565
Ca ⁺²	3.486	1.195	3.736	3.298	3.467	2.614
K ⁺	0.000	0.019	0.000	0.000	0.000	0.000
Σ X site	4.027	4.047	3.971	3.866	3.886	4.179

Mol.% Endmembers*

An	87.79	29.67	92.32	85.31	89.22	62.55
Ab	12.21	69.85	7.68	14.69	10.78	37.45
Or	0.00	0.47	0.00	0.00	0.00	0.00

- 1 Anorthitic bytownite from the wallrocks adjacent to an auriferous vein in basalt, 1851 stope, L zone.
- 2 Andesine from a garnetiferous basalt. Drill core S172.
- 3 Anorthite from the matrix of brecciated rhyolite. The matrix is composed of chlorite, biotite, tourmaline, carbonate and pyrite with plagioclase. Drill core 27-263.
- 4,5,6 Plagioclase from altered, staurolite-bearing basalt beneath the 27th level. Analysis 4 and 5 are bytownite from the core of a quartz + chlorite veinlet. Analysis 6 is nodular labradorite from the wallrocks. Drill core 27-264.
- Abbreviation; An = anorthite, Ab = albite, Or = orthoclase

6.7 Summary of the gangue minerals

Significant carbonate and quartz + carbonate veining preceded gold deposition. Auriferous mineralization occurs in the quartz matrix to brecciated parts of quartz + carbonate veins, together with minor ore-stage carbonate (Stage 2). Gahnite and plagioclase occur in some auriferous veins and breccias in amphibolite facies metamorphic domains. Plagioclase is cut by carbonate, indicating carbonate deposition after plagioclase. Gahnite crystallized synchronously with auriferous arsenopyrite (Tarnocai et al., 1997), and suggests auriferous hydrothermal mineralization was coeval with peak metamorphism. Scheelite occurs with Au in the western, greenschist facies domain, and is absent from auriferous veins in the eastern part of the deposit.

Figure 6.0

Ternary diagram showing the compositions of carbonate from auriferous veins.

Carbonates are predominantly ankerite, with minor siderite, calcite, and Fe-dolomite.

The dashed line shows the boundary between Fe-dolomite and ankerite. Star shows the average carbonate composition used in the construction of figure 10.9 and 10.10.

Abbreviations; Ank=ankerite, Dol=dolomite

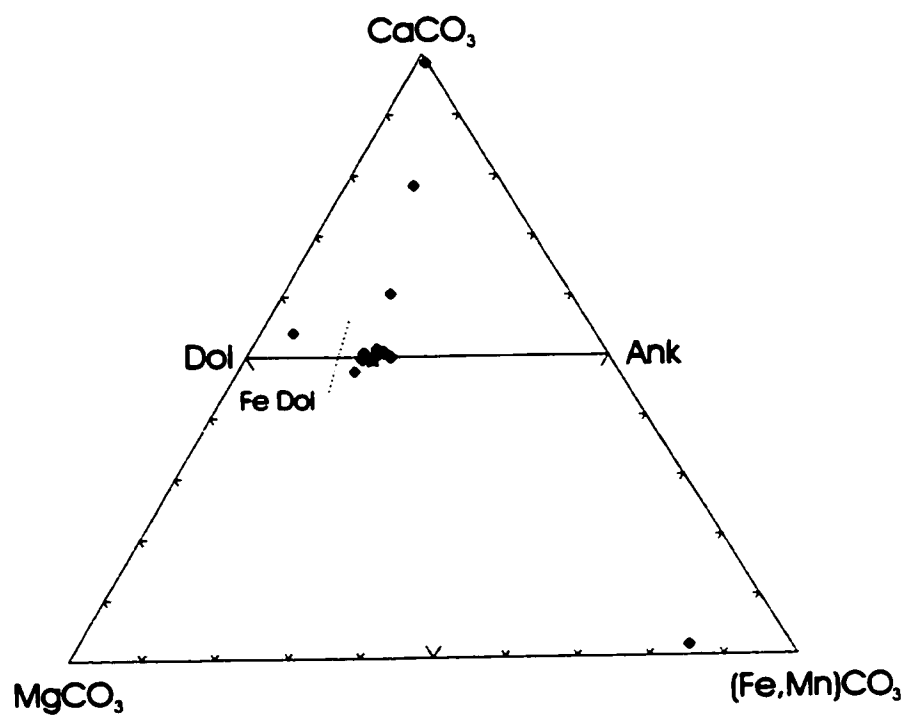


Figure 6.1

Photomicrograph showing scheelite (high relief, center of photo) intergrown with calcite and quartz, in an auriferous quartz + carbonate vein. Sample from the 27th level of the Campbell mine. The width of the photo is approximately 2 mm.



Figure 6.2

Backscattered electron image of subhedral gahnite (Ghn) containing sphalerite (Sph) inclusions in contact with sphalerite. Chlorite (Chl) occurs as inclusions in sphalerite. Sample from disseminated pyritic mineralization zone, 307E drift.

Sph

Ghn

Chi

1911

Figure 6.3

Backscattered electron image of gahnite (Ghn, medium grey) mantled by arsenopyrite (Apy, white acicular needles), and also with arsenopyrite inclusions, in podiform quartz + arsenopyrite ore. The matrix is predominantly quartz. Sample from the 17th level, L zone.

Ghn

Apy

1000

Figure 6.4

Ternary diagram for the spinel group minerals, showing the composition of gahnites (●) from ore at the Campbell mine. Numbered fields correspond to the composition of spinel from; 1, metamorphosed massive sulfide deposits; 2, pegmatite's; 3, marbles; 4, aluminous metasediments. Gahnites from the mine fall in the fields of those from metamorphosed massive sulfide deposits and aluminous metasedimentary rocks. Modified after Spry and Scott (1986).

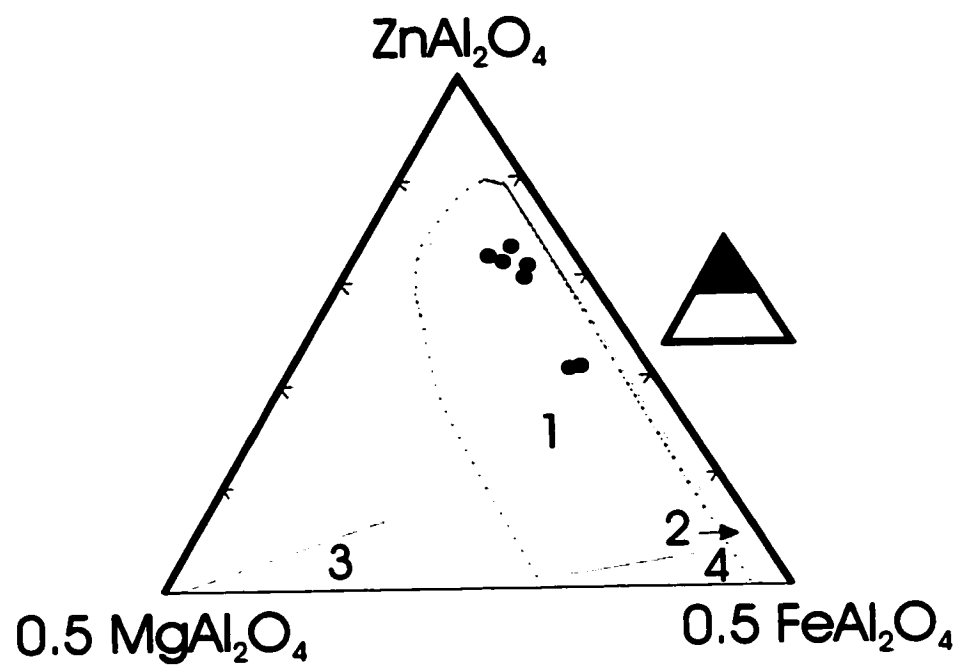
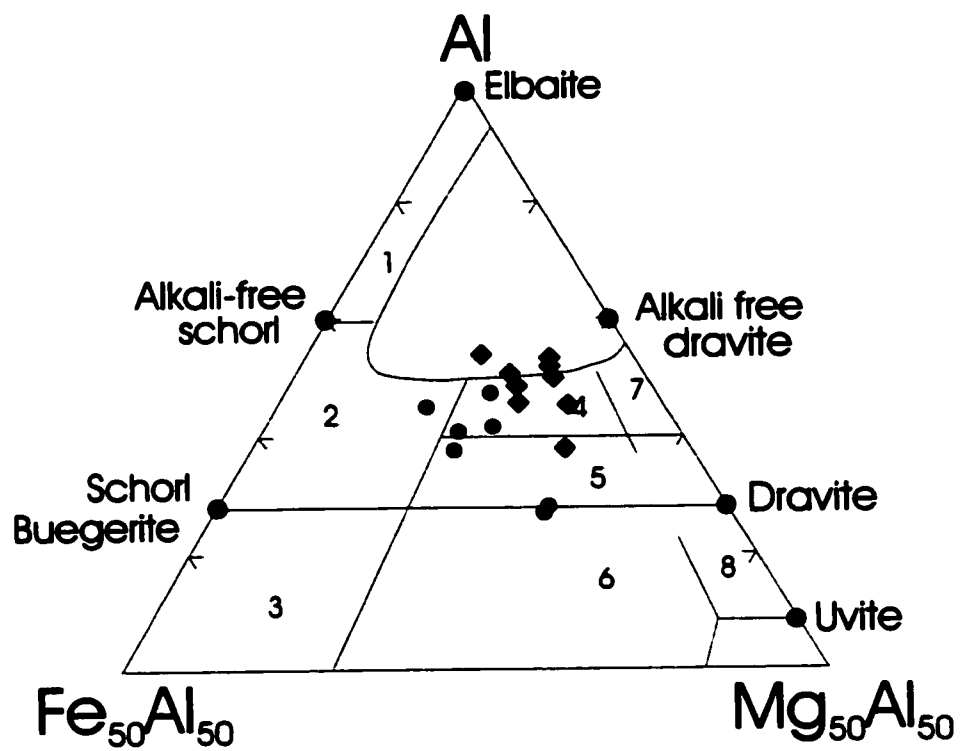


Figure 6.5

Ternary diagram showing tourmaline compositions. Tourmaline from basaltic rocks distal to auriferous veins (●) are intermediate dravite-schorl, whereas tourmaline from ore veins, and wallrocks immediately adjacent to ore veins (◆) are Al- and Mg-rich. Endmember tourmaline compositions are shown as (⊙). Diagram modified after Henry and Guidotti (1985). Labeled fields are 1, Li-rich granitoids and pegmatites, 2, Li-poor granitoids and pegmatites, 3, Fe⁺³ rich quartz-tourmaline rocks, 4, metapelites and metapsammities with an Al-saturating phase, 5, metapelites and metapsammities without an Al-saturating phase, 6, Fe⁺³ rich quartz-tourmaline rocks and calc-silicates, 7, low-Ca ultramafic rocks and Cr, V-rich sediments, 8, metacarbonates and metapyroxenites.



7.0 Stable Isotopic Composition of Carbonates in the Campbell mine and eastern Red Lake belt

7.1 Introduction

Carbon isotopic composition of hydrothermal carbonate is controlled by the isotopic composition of carbon in the hydrothermal fluid, temperature, pH, and fO_2 . As the isotopic fractionation factors between carbon species are large, the chemical conditions that existed during ore deposition, and potentially fluid source, can be examined using carbon isotopes.

7.2 Sample selection

Fifteen samples were analyzed from the Campbell mine (P, G, A and L ore zones), and from three barren crustiform quartz + carbonate vein samples were analyzed from outside the mine, to the W and NW. The samples from the mine represent carbonate deposited with gold (Stage 2 carbonate), crustiform and banded texture Stage 1 carbonate veins, and Stage 3 carbonate veinlets. The stages of carbonate deposition have been described in section 6.0. Powder samples were obtained from hand samples with a small drill. Quartz is the predominant sample impurity, typically < 25% of the sample volume. Weathered rinds were removed from the carbonate samples with a rock saw, prior to drilling. Three crustiform layers were individually sampled from sample CRL-15, to check between-layer compositional variability. Additional details of sampling and analytical method are provided in Appendix 5.

7.3 Carbon and oxygen isotopic compositions of carbonate

The O and C isotopic composition of the Campbell mine and eastern Red Lake belt carbonates are presented on figure 7.0 and Table 7.0. The sample locations for samples collected outside the Campbell mine underground are provided on figure 7.1 and in Table 7.0. The $\delta^{13}C$ value for these carbonate samples ranges from +0.06 to +4.24 ‰. The isotopic composition of carbonates from the mine and mine area, relative other gold deposits from the Slave and Superior Province (Kerrick et al., 1987) are also provided. Clearly, carbonates from the Campbell mine area are isotopically enriched in ^{13}C and $\delta^{18}O$ relative to other gold deposits. Carbonate from other Archean Au systems typically

Table 7.0. Carbonate O and C compositions

Sample	$\delta^{18}\text{O}_{(\text{SMOW})}$	$\delta^{13}\text{C}$	North	East	Stope/Zone	Description
cr3-94	15.73	0.63	N/A	N/A	2451-6, P zone	E-W trending quartz + carbonate veinlet.
cr17-94*	16.31	0.94	N/A	N/A	2451-6, P zone	Wispy coarse-grained carbonate in auriferous, E-W trending veinlet. Occurs with quartz, arsenopyrite, and native gold.
cd14-94*	17.13	1.68	N/A	N/A	1854, L zone	Coarse-grained wispy carbonate in high Au grade quartz + arsenopyrite breccia in NNW trending quartz + carbonate vein.
cr19-94*	18.13	3.75	N/A	N/A	2451-6, P zone	Coarse-grained wispy carbonate in high Au grade quartz + arsenopyrite breccia in NNW trending quartz + carbonate vein.
cr1103	16.82	0.78	N/A	N/A	2061B, G zone	Post-ore coarse-grained carbonate vein cutting auriferous quartz + arsenopyrite mineralization.
cr15-94	14.78	1.63	443863	5656878	N/A	Wilmar, banded to crustiform Fe-carbonate vein.
cr16-94	14.78	1.48	449095	5661483	N/A	Redcon banded to crustiform quartz + carbonate vein.
cr13-94*	18.67	3.43	N/A	N/A	1851, L zone	Crackle brecciated, auriferous quartz + carbonate vein.
cr18-94	16.57	1.82	446416	5661926	N/A	Banded to crustiform texture quartz + carbonate vein at the Campbell west property boundary, stripped outcrop beneath the airport approach.
cr4-94	15.06	0.37	N/A	N/A	2061, G zone	Banded to crustiform quartz + carbonate vein, NNW trending
cr1-94	15.14	1.42	N/A	N/A	1851, L zone	Brecciated crustiform quartz + carbonate vein, NNW trending.
cr20-94	17.53	1.6	N/A	N/A	1902 E, A zone	1902 E dr. A zone
cr2-94	15.16	0.06	N/A	N/A	2451-6, P zone	Banded to crustiform quartz + carbonate vein, NNW trending.
cr5-94	15.16	1.06	N/A	N/A	2451-6, P zone	Banded to crustiform quartz + carbonate vein, NNW trending.
cm1-cr115*	14.17	1.59	N/A	N/A	2651-1W	Banded quartz + carbonate vein.
cm2-cr115*	14.17	1.44	N/A	N/A	2651-1W	Banded quartz + carbonate vein.
cm3-cr115*	17.61	4.24	N/A	N/A	2651-1W	Banded quartz + carbonate vein.

Northing and Easting (meters), UTM NAD27, Zone 15

Stope/Zone = location in Campbell mine

N/A = not applicable

* auriferous, Stage 2 carbonate sample

have negative C isotopic compositions, and generally lie within a range -8.5 and -2.7 ‰ (Colvine et al., 1988). Samples collected from Stage 1 crustiform carbonate veins in the mine typically display lower $\delta^{13}\text{C}$ values ($+0.06$ to $+1.06$ ‰) than Stage 2 carbonate ($+0.94$ to $+3.75$ ‰) (Table 7.0). Stage 3 carbonate displays relatively low values ($+0.78$ ‰), similar to Stage 1 carbonate samples. In the P ore zone, carbonate associated with Au mineralization is ^{13}C enriched, relative to carbonate not closely associated with Au, in the same vein (Fig. 7.2). The highest gold grades in the Campbell mine are associated with quartz + arsenopyrite breccia bodies, which also display the heaviest $\delta^{13}\text{C}$ isotopic values ($+1.68$ to $+3.75$ ‰, Table 7.0). Crustiform carbonate veins from W and NW of the mine display significantly heavier $\delta^{13}\text{C}$ isotopic values ($+1.48$ to $+1.62$ ‰), than similar veins within the Campbell mine. Carbonates from the Campbell mine and the eastern Red Lake belt display some of the highest $\delta^{13}\text{C}$ values analyzed from an Archean gold system.

Carbonates from the eastern Red Lake belt are relatively $\delta^{13}\text{C}$ enriched compared to carbonates from the East Bay, Post Narrows, and Flat Lake-Howey Bay deformation zones (El Bilali and Hattori, 1998). The uniformly high $\delta^{13}\text{C}$ from the eastern Red Lake belt suggests that carbonate deposition may have occurred from a similar fluid throughout the eastern part of the belt. Samples were obtained from rocks that experienced a wide range of metamorphic temperatures, from middle greenschist facies, to lower to middle amphibolite facies, and E to W variations in isotopic composition of carbonates from the region may reflect metamorphic decarbonation. The heavy $\delta^{13}\text{C}$ values for carbonates from the eastern part of the belt, compared to other Archean gold deposits, may indicate the fluid had an anomalously enriched source, inter-acted with an enriched rock prior to carbonate deposition, or acquired its signature from mixing of isotopically different fluids. The C isotope composition of the fluid must have been relatively constant over time, as both Stage 1 and 3 carbonates display similar $\delta^{13}\text{C}$ values. Significant variations in the carbonate isotopic composition occur at the stope-scale, between ore and non-ore related carbonate. This small-scale variability may be due to the processes which operated during gold deposition, such as fluid mixing, changes in $f\text{O}_2$, and isotopic fractionation between C species during phase separation. Carbonate bands from

individual auriferous veins (Table 7.0) show large variations in ^{13}C (up to 2.80 ‰), suggesting episodic changes in fluid composition, temperature, pressure, or $f\text{O}_2$.

7.4 Possible causes for ^{13}C enrichment in carbonate from the Campbell mine and eastern Red Lake belt

7.4.1 Interaction with relatively enriched rocks

The fluid may have acquired its C isotopic composition through dissolution of carbonate rock high in ^{13}C at the source, or during transport. Marine carbonates have $\delta^{13}\text{C}$ values between -2 and +3 ‰ (Hoefs, 1997), however, marine carbonates are uncommon in Archean greenstone sequences, and are absent from the eastern Red Lake belt. Volcanic rocks that were altered during submarine hydrothermal activity or submarine weathering would contain carbonate with isotopic compositions similar to those of marine carbonates. Therefore, the dissolution of carbonates from carbonate altered submarine volcanic rocks would have similar effects as the dissolution of marine carbonates. Pervasive submarine carbonate alteration has not been described from petrographic reports for the eastern Red Lake belt. Although dissolution of marine carbonate cannot be discounted, it is suggested as unlikely that the hydrothermal fluid C isotopic composition was derived from interaction with marine carbonates.

Interaction of auriferous fluids with carbonaceous sedimentary rocks has been shown to result in both positive (Colvine et al., 1988; King and Kerrich, 1987) and negative shifts in the ^{13}C for carbonates. A positive shift may arise from incorporation of methane to solution at constant $^{13}\text{C}\Sigma\text{C}$ (proportion of ^{13}C relative to total C in solution). Isotopic fractionation between methane and oxidized C species may result in the enrichment of ^{13}C in the dissolved carbonate species. The extent of the upward shift is controlled by the degree of reduction of the fluid. A decrease in carbonate ^{13}C following interaction of a fluid with carbonaceous sediments may arise from the incorporation of isotopically light reduced carbon, lowering $^{13}\text{C}\Sigma\text{C}$ of the fluids. The ^{13}C for total dissolved C, and carbonates deposited from the fluid, will therefore depend on the balance of C acquired by oxidation of reduced C, dissolution of enriched carbonate, and the addition of CH_4 to the fluid.

Clastic and pelitic sedimentary rocks with minor graphite occur immediately east of the Campbell mine. Faults interpreted here to be spatially associated with mineralization, and which could have acted as hydrothermal conduits, transect the sedimentary sequence (Fig. 2.6). It is therefore possible that the isotopic composition of carbonates at the mine, and in the eastern part of the belt, may be related to changes in fluid fO_2 as a result of interaction with graphitic sediments. The $\delta^{13}C$ value for carbonates in sedimentary rocks around 0 ‰ (El Bilali and Hattori, 1998) does not support significant interaction of hydrothermal fluids with sedimentary rocks. Interaction with carbonaceous sediments does not adequately explain the positive shift in the isotopic composition of Stage 2 carbonates at the stope scale, where carbonaceous sediments are absent.

7.4.2 Cooling of the hydrothermal fluid

Assuming constant isotopic compositions of total dissolved carbon in hydrothermal fluid ($^{13}C_{\Sigma c}$), cooling of the fluid can result in progressive isotopic enrichment in the deposited carbonates. Cooling between 500 °C and 300 °C (amphibolite to greenschist facies temperatures) may account for at most a +1 to +2 ‰ shift. Cooling may explain the relatively isotopically heavy composition for crustiform carbonate samples east of the mine (enriched by +0.42 to +1.56 ‰). However, cooling alone does not explain the isotopically heavy composition for ore-related carbonates (2.15 to 3.72 ‰) relative to pre- and post-ore carbonates.

7.4.3 Metamorphic decarbonation and dissolution

Dissolution of pre-existing carbonate may result in a small increase in $\delta^{13}C$ of H_2CO_3 in solution (+1 to +2 ‰), within a reasonable hydrothermal temperature range. Decarbonation of carbonates during metamorphism results in coupled depletions of ^{18}O and ^{13}C in the residual carbonate, with isotopically heavy carbon partitioned to the fluid (Faure, 1986; Hoefs, 1997). Carbon dioxide added to the fluid by decarbonation may be enriched by up to 4‰ over the parent carbonate, between 200 °C and 400 °C. Carbonate deposited from this fluid would also be enriched relative to the parent carbonate, but the extent of the enrichment would depend on the proportion of CO_2 added as a product of

decarbonation to the total carbon budget of the fluid, the bulk isotopic composition of carbon in the fluid prior to addition, and the carbonate isotopic composition of the parent carbonate.

There is field and petrographic evidence for decarbonation in rocks between the eastern boundary of the Campbell mine (Fig. 7.3) and the Trout Lake Batholith. If progressive deposition of carbonate occurred from the fluid, westward from the Goldcorp mine, the isotopic composition of carbonates may show an enrichment trend resulting from Rayleigh distillation or cooling, given that no further modifications occurred to the fluid $^{13}\text{C}_{\Sigma\text{c}}$. If carbonate in amphibolite facies rocks at Goldcorp, and to the east of the mine supplied CO_2 to the fluid during decarbonation, then carbonates at Goldcorp may be expected to have a somewhat reduced value compared to samples from greenschist facies rocks. Carbonates at Goldcorp ($\delta^{13}\text{C} +0.7$ to $+2.3\text{‰}$, El Bilali and Hattori, 1998) are somewhat lighter than those at Campbell, and those west of Campbell are heavier (Table 7.0). This suggests that decarbonation to the east may have contributed to increased $^{13}\text{C}_{\Sigma\text{c}}$ of the fluid, and would imply that carbonates at Campbell, and to the west, were deposited during metamorphism. The stope-scale patterns of enrichment in Stage 2 carbonates are difficult to envisage as a product of metamorphic decarbonation.

7.4.4 Isotopic fractionation during phase separation

The differential partition coefficient for phases such as CO_2 and CH_4 , between fluid and vapor can result in isotopic fraction during vapor phase extraction, particularly under open system conditions. The isotopic fractionation between aqueous and vapor phase CO_2 is relatively small, and extraction of CO_2 from the fluid would probably result in minor changes to the fluid $^{13}\text{C}_{\Sigma\text{c}}$. However, if methane is present in the fluid, owing to its relatively large partition coefficient compared to CO_2 , and the large $\delta^{13}\text{C}$ fractionation factor between CO_2 and CH_4 , a large positive shift in $^{13}\text{C}_{\Sigma\text{c}}$ can also occur during vapor phase extraction. Fluid immiscibility and vapor phase extraction may be induced by a drop in temperature or pressure. A pressure drop in the hydrothermal conduit may occur during fault motion (Sibson, 1986, 1988), and result in hydrothermal brecciation. Evidence for hydrothermal brecciation during gold deposition is ubiquitous

at the mine. The magnitude of the shift will depend on the mole % CH_4 initially in solution, temperature, and the degree of vapor phase extraction.

7.5 Summary of carbonate $\delta^{13}\text{C}$ values in the eastern Red Lake belt

The $\delta^{13}\text{C}$ values of carbonate samples from the eastern Red Lake belt are anomalously high compared to carbonate $\delta^{13}\text{C}$ values reported from other Archean gold deposits (Kerrick et al., 1987). The values presented by the author, and El Bilali and Hattori (1998), are uniformly high across the eastern part of the belt (+0.06‰ to +4.24‰), suggesting carbonate deposition from a similar common fluid composition. Although the values are compatible with a source from dissolution of marine carbonate, it is considered unlikely because of the lack of identified marine carbonates or extensive seafloor hydrothermal carbonate alteration. Carbonate from the eastern part of the belt is ^{13}C enriched relative to carbonate from the East Bay and Flat Lake – Howey Bay deformation zones (El Bilali and Hattori, 1998), suggesting that the fluid responsible for carbonate alteration along these deformation zones may be different.

Stage 2 carbonate, deposited with gold, is enriched by up to 2.69‰ over both Stage 1 and 3 carbonate. Stage 1 and 3 carbonate show similar $\delta^{13}\text{C}$ compositions, suggesting the mechanism for ^{13}C enrichment in Stage 2 carbonate is related to the gold deposition process. Phase separation accompanying hydrothermal brecciation, resulting in loss of CH_4 and an increase in fluid ^{13}C , may have resulted in both the deposition of gold, and ^{13}C enrichment in Stage 2 carbonate.

Figure 7.0

Oxygen and carbon isotopic compositions for carbonates (dolomite to ankerite) from veins at the Campbell mine, and from banded quartz + carbonate veins to the W and NW of the mine (blue dots). The field of carbonate compositions from Au deposits in the Slave (dash-dot line) and Superior Province (dashed line) are shown for reference.

Note that the carbonate samples from the Campbell mine are enriched in ^{13}C compared to samples from other Archean Au deposits. Data fields for Slave and Superior Province Au deposits from Kerrich et al. (1987).

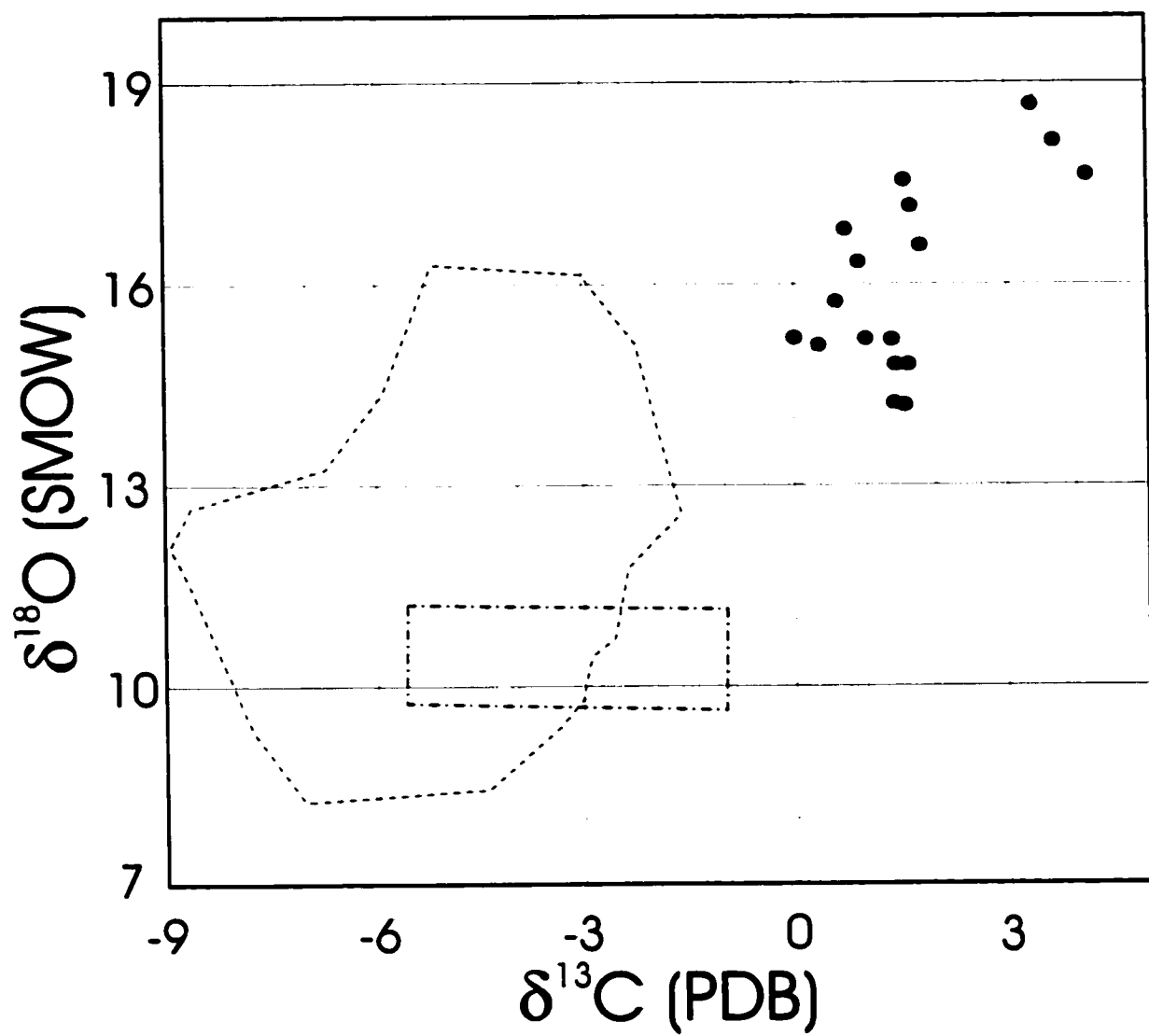


Figure 7.1

Geology map of the eastern Red Lake belt, showing the location of banded to crustiform quartz + carbonate vein samples (red dots) for isotopic analysis. The sample numbers and $\delta^{13}\text{C}$ values are shown beside the sample location.

cr15-94, Wilmar mine; cr18-94, Campbell western property boundary; cr16-94, Redcon property. Geology legend and mine names are provided in figure 2.1.

439000
5662000

● 1.48
● C16-94



455000
5653000

● 1.04
● C14-94

C-O isotope location,
with ^{13}C (PDB) analysis, sample number

Figure 7.2

Geology map of the P zone, 2451-6 stope and -5 sub, showing the $\delta^{13}\text{C}$ values of carbonates collected from banded to crustiform quartz + carbonate veins, brecciated auriferous carbonate veins, and high gold grade quartz + arsenopyrite breccia bodies. Along both NNW and E-W veins, the $\delta^{13}\text{C}$ values of carbonates become greater from non-auriferous, to high gold-grade mineralization.

- Lamprophyre dykes
- Basaltic rocks
- Ultramafic rocks
- Banded to crustiform quartz-carbonate vein
- Sheeted carbonate veinlets, ±biotite+pyrite alteration
- Carbonate vein breccia, quartz-Fe sulphide, Au mineralized
- High-Au grade quartz-arsenopyrite breccia
- Magnetite+pyrite+pyrrhotite breccia in quartz+carbonate vein

● +0.06 Carbonate ^{13}C composition (‰)

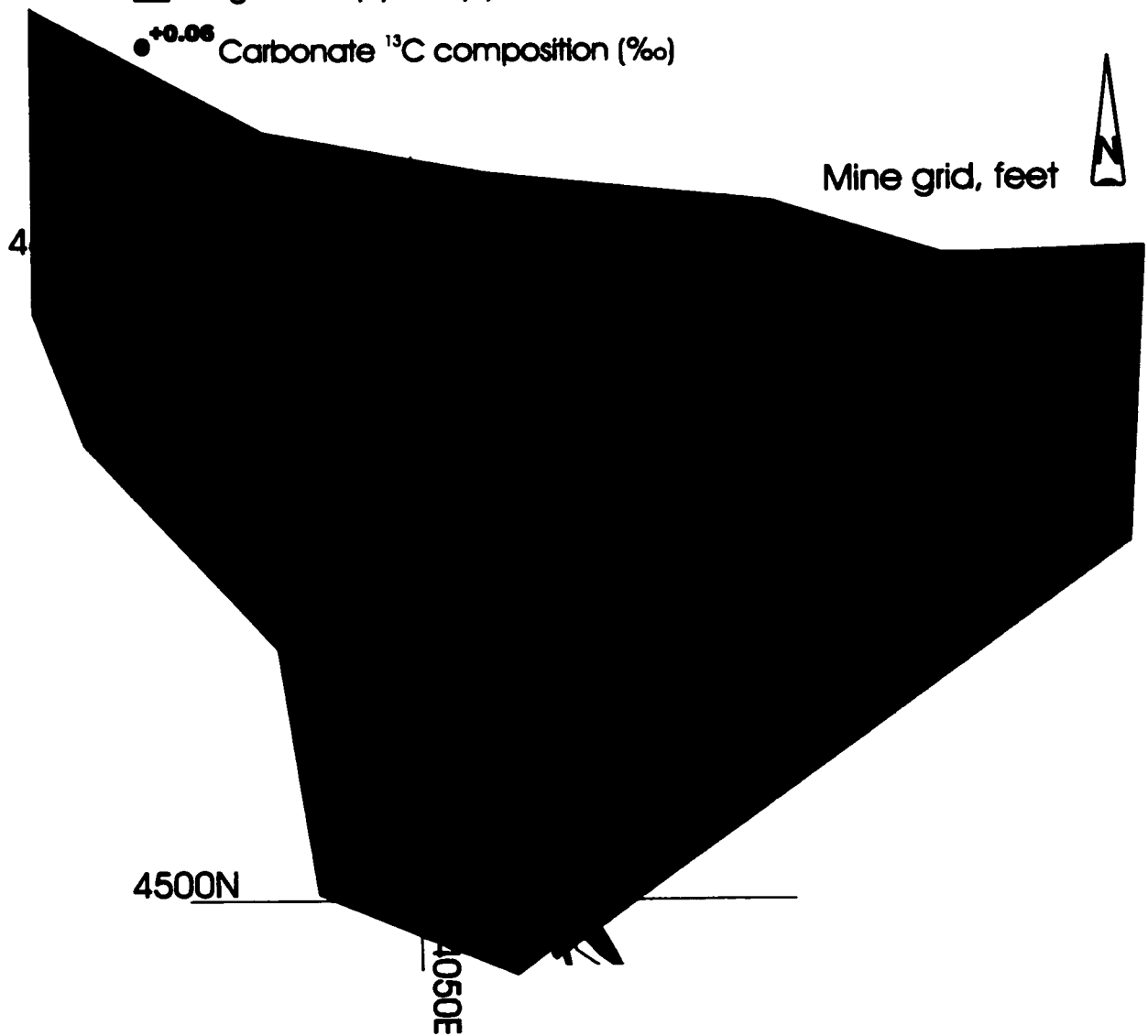
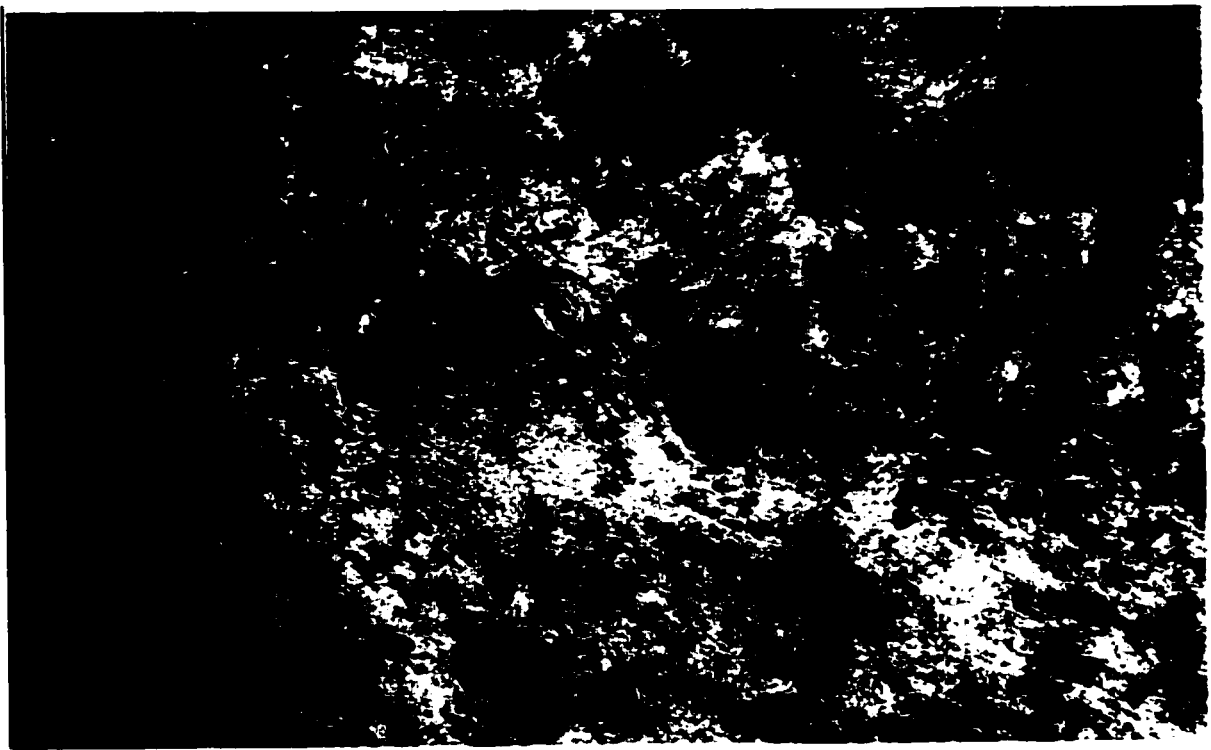
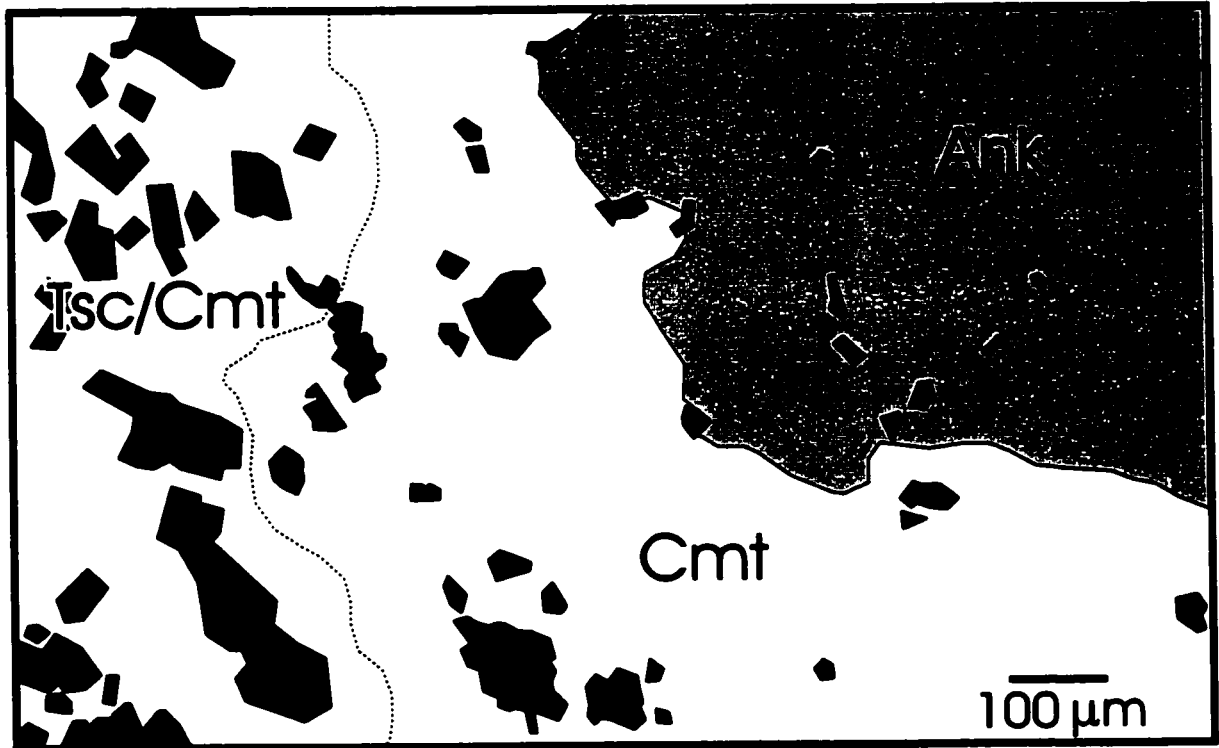


Figure 7.3

Ankerite (Ank) is rimmed by fine-grained cummingtonite (Cmt). Tschermakite occurs with cummingtonite, to the left of the dashed line. Opaque phases are pyrite and magnetite. Fine-grained quartz and plagioclase occur in the matrix.

The overprinting of ankerite by cummingtonite suggests metamorphic decarbonation producing cummingtonite. Sample 95a, from the wallrocks of the NL ore zone at the Campbell mine.



8.0 Strontium isotopic composition of scheelite

The radioactive decay of ^{87}Rb yields ^{87}Sr , and mantle and crustal rocks display large variations in Rb/Sr ratios. It is therefore possible to investigate the rocks the hydrothermal fluid may have interacted with. This can be accomplished by determining the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of a Rb-poor hydrothermal mineral and assuming that the mineral has remained a closed system to Sr and Rb since hydrothermal deposition.

Scheelite is a relatively common hydrothermal mineral in Archean vein Au deposits, Rb-poor, and relatively resistant to post-depositional isotopic exchange. At the Campbell mine, scheelite occurs in auriferous veins and black line faults in the eastern part of the mine. The Sr isotopic compositions for scheelite are presented in Table 8.0. All scheelite samples except one have low Rb/Sr ratios (≤ 0.008). Details of the analytical method and sample selection are provided in Appendix 5.

8.1 Sample selection and results

Four scheelite separates were obtained from auriferous veins at the Campbell mine, along with one from an auriferous vein in the H.G. Young mine (Fig. 2.1). Three of the samples were obtained from auriferous veins hosted by basaltic rocks, and one from an auriferous vein at a mafic – ultramafic rock contact. An additional scheelite separate was analyzed from a post-ore black line fault at the Campbell mine. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ values for scheelite from auriferous veins ranged between 0.7013 and 0.7023. The $^{87}\text{Sr}/^{86}\text{Sr}$ value of the scheelite from the black line fault is higher, at 0.7026. Although the timing of Au mineralization in the eastern part of the belt is in question, it is generally accepted that it occurred close to 2.7 Ga (Corfu and Andrews, 1987). Accordingly, the scheelite $^{87}\text{Sr}/^{86}\text{Sr}$ values at 2.7 Ga was calculated as the initial value. Because of low Rb/Sr ratios, the age correction is very minor and the initial $^{87}\text{Sr}/^{86}\text{Sr}$ values were not affected by the choice of age. Scheelite from auriferous veins are 0.7012 to 0.7017 at 2.7 Ga. The black line fault scheelite has an $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.7024 at 2.7 Ga.

Table 8.0. Sr isotopic composition of scheelite from the Campbell mine area, and predicted $^{87}\text{Sr}/^{86}\text{Sr}$ values at 2.7 Ga.

Sample	Sr ppm	Rb ppm	Rb/Sr	$^{87}\text{Sr}/^{86}\text{Sr}$ measured	Error	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ 2.7 Ga
average Basalt	62	14	0.2258065			0.63601	0.703729 [#]
average UM rocks	43	6	0.1395349			0.393016	0.702686 [#]
2451.6s	342	0.280	0.0008187	0.701624	21	0.023	0.701530
2451.6s	342	0.280	0.0008187	0.701619	16	0.023	0.701525
921112-01	629	0.486	0.0007727	0.702529	13	0.00223	0.702442
CS1				0.701289	11		0.701289
920629-10	101	4.840	0.0479208	0.706548	14	0.1381	0.701153
920629-20	374	3.000	0.0080214	0.702472	16	0.0232	0.701462
920630-2	403	0.090	0.0002233	0.701374	15	0.0006	0.701261
920630-3	478	0.229	0.0004791	0.701733	16	0.0014	0.701675

[#] calculated assuming an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.701 at 3.0 Ga.

8.2 $^{87}\text{Sr}/^{86}\text{Sr}$ values for the host rocks

The host rocks at the mine are primarily basaltic volcanic rocks with subordinate komatiitic rocks. The age of the host rocks is approximately 3.0 Ga, determined by U-Pb zircon geochronology from interbedded felsic volcanic rocks (Corfu and Andrews, 1987). Assuming the basaltic rocks had an initial $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.701 at 3.0 Ga, their $^{87}\text{Sr}/^{86}\text{Sr}$ value at 2.7 Ga is calculated to be 0.7037 using the Rb and Sr concentrations in relatively unaltered basalt (Table 8.0). Ultramafic rocks typically contain very low concentrations of Rb and Sr (Faure, 1986), but even relatively unaltered ultramafic rocks in the mine contain significantly high concentrations due to cryptic alteration. Ultramafic rocks typically have very low $^{87}\text{Sr}/^{86}\text{Sr}$ values. Therefore, the minimum Rb and Sr concentrations for ultramafic rocks at the mine are used here, and a $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.7027 is obtained at 2.7 Ga. The values were most likely higher because we used the minimum Rb/Sr ratio. The values listed in Table 8 reasonably reflect the minimum $^{87}\text{Sr}/^{86}\text{Sr}$ value of relatively unaltered 2.7 Ga basalt and ultramafic rocks at the mine.

The $^{87}\text{Sr}/^{86}\text{Sr}$ values of scheelite from auriferous veins (0.7012 to 0.7017) are slightly higher than 2.7 Ga mantle values (0.7011; Hattori et al., 1996), but are significantly lower than the $^{87}\text{Sr}/^{86}\text{Sr}$ value of ultramafic (>0.7027) and basaltic (>0.7037) host rocks. The data suggests the hydrothermal fluid experienced variable but minimal interaction with the mafic to ultramafic host rocks, and reflects either significant interaction with the mantle, or deep crustal to mantle intrusions. The scheelite along the black line fault shows $^{87}\text{Sr}/^{86}\text{Sr}$ similar to the minimum value of ultramafic rocks. This may reflect interaction of the hydrothermal fluid with the host rocks during the development of black line faults after Au deposition.

9.0 Fluid Inclusion Study

9.1 Sample selection

Samples for the fluid inclusion study were selected from quartz + arsenopyrite breccia mineralization from the G, P, L, and F2 ore zones (2061 and 1861, 2451-6, 1854, and 1925W stopes, respectively) associated with locally high Au grades. A sample was also obtained from a low gold grade (< 0.35 g/t) quartz veinlet with wallrock alteration containing arsenopyrite, from drill core beneath the 27th level. These samples include both foliation-oblique (G, P, and L ore zones) and basalt-hosted foliation-parallel auriferous veins (F2 ore zone).

9.2 Petrographic description of fluid inclusions

The criteria for identifying primary and secondary inclusions followed Roedder (1984) and apparently secondary inclusions are not examined in this study. Inclusions were examined from both carbonate and quartz grains. Thirty-six doubly polished 100 μm thick chips were examined during the course of this study.

All fluid inclusions are very small (generally $< 10 \mu\text{m}$), and those larger than 3 μm were chosen for heating - cooling experiments. In quartz, two types of fluid inclusions were identified; Type I inclusions containing high-density CO_2 with variable CH_4 , and Type II $\text{CO}_2 + \text{CH}_4$ bearing (< 70 vol.%) aqueous inclusions. Water contents of < 20 vol.% are difficult to detect in small, Type I inclusions, as it adheres to the inclusion walls, making an estimate of $\text{CO}_2/\text{H}_2\text{O}$ ratios difficult. Clathrates observed during petrographic study of Type II inclusions are probably composed of a mixture of CH_4 and CO_2 . The addition of CH_4 to the CO_2 clathrate will increase the clathrate melting temperature, whereas increasing NaCl reduces the melting temperature (Roedder, 1984). Salinity data is interpreted with caution, and salinities for Type II inclusions with low CH_4 contents are considered here to provide a better estimate of true salinity.

Type I and II inclusions are primary and pseudosecondary, by the criteria of Roedder (1984). Secondary inclusions are not discussed. Primary inclusions are equant, rounded, with rare negative crystal shapes, and occur individually, or in small three-dimensional clusters in quartz. Pseudosecondary inclusions form planar trails, and have equant, rounded to cylindrical shapes with rare cusped edges.

Some auriferous veins and veinlets show a variation in the occurrence of Type I inclusions. Pseudosecondary inclusion trails are abundant in the margins of veins, but similar Type I inclusions occur as primary inclusions in the center of the veins. Inclusion trails terminate within quartz grains at grain margins, locally cutting adjacent quartz grains and garnet grains at the vein/wallrock contact. This gradation in Type I inclusions suggests that vein quartz and wallrock alteration minerals were fractured during episodic vein formation. Quartz in the interior of veins was deposited during and after the formation of vein margin quartz, and the formation of some wallrock alteration minerals. Native gold occurs along some pseudosecondary trails.

9.3 Description of Type I inclusions

Type I inclusions occur in all high Au-grade quartz + arsenopyrite breccia samples from the G, L, F2, and P ore zones, and locally make up > 90% of the inclusion population. In quartz, the inclusions are monophasic liquid at room temperature, and locally two-phase liquid + vapor, the vapor forming ≤ 30 vol.%. Inclusions rapidly homogenized to the liquid phase (commonly 5 °C to 18 °C) during heating, and melting occurs close to that of pure CO₂ (-56.6 °C to -58.8 °C), indicating the fluid is predominantly high-density CO₂ (0.80 to 0.88 g/cm³). The depression of the freezing point below that of pure CO₂ suggests the presence of another phase, such as N₂, H₂S, SO₂, or CH₄. Methane is considered the most probable gas species because it is common in fluid inclusions from many gold deposits (Spooner et al., 1985; Sherlock et al., 1993). Melting and homogenization temperature data are presented in Table 9.0.

Carbonate grains in quartz + arsenopyrite breccias contain abundant Type I liquid and liquid + vapor inclusions with negative crystal shapes. Two-phase inclusions are more common in carbonate than co-existing quartz. Zonations in liquid + vapor and liquid inclusions are displayed across some large carbonate grains, with vapor-bearing inclusion-rich cores, and liquid-only, inclusion-poor rims. In two-phase inclusions, CO₂ vapor forms 30 to 60 vol.% of the inclusion. As expected, Th (temperature of homogenization) decreases from cores to rims, from 24 °C to 27 °C, to 20 °C to 22 °C.

Table 9.0 Melting and homogenization temperatures (°C) for primary, Type I CO₂-rich inclusions in quartz

Stope,Zone	Incl. ¹	Th ²	Tm ³	Stope,Zone	Incl. ¹	Th ²	Tm ³
2451, P	16	8.6	57.6	1861, G	3	14.2	-57.3
	17	15.0	-57.7		4	14.7	-57.5
	18	12.0	-57.9		5	14.4	-58.3
	19	12.8	-58.2		6	11.1	-58.4
	20	15.7	-58.8		7	12.6	-58.3
	22	13.1	-57.2		9	-1.3	-57.9
	23	11.8	-57.6		10	-1.5	-57.9
	24	13.6	-57.9		11	9.8	-57.9
1854, L	1	14.8	-57.5		12	11.4	-59.9
	2	20.3	-57.4		13	9.4	-57.8
	3	10.6	-57.4		16	21.0	-53.6
	5	17.5	-57.0		17	10.0	-58.1
	6	11.6	-57.0		18	9.6	-58.1
	7	10.6	-56.9		19	5.2	-57.8
	8	-2.5	-56.9		20	5.2	-57.8
	9	-0.8	-57.1		21	10.7	-57.9
	9	0.1	-57.0		22	15.9	-57.3
	10	10.6	-57.7		23	15.2	-57.0
	12	27.2	-56.8		24	3.2	-56.9
	12	28.6	-56.0		25	15.9	-57.3
DDH27-264 ⁴	1	0.8	-59.0		26	9.0	-57.4
	2	1.8	-57.1		27	9.5	-57.4
	3	2.2	-56.8		28	12.2	-57.0
	16	9.2	-56.6		42	9.5	-57.4
	17	9.8	-56.6		43	10.7	-57.0
	18	10.2	-56.6		44	18.2	-57.0
	19	10.1	-56.6		45	10.0	-58.1
	20	10.3	-56.6		46	9.6	-58.1
	21	10.5	-57.0		47	9.7	-57.8
	22	11.8	-57.0		48	9.6	-58.1
	23	12.0	-57.0		49	5.2	-57.8
	24	12.8	-57.0		50	5.2	-57.8
	25	13.5	-57.0		51	15.9	-57.3
	26	16.8	-53.7		53	3.2	-56.9
	27	8.8	-55.9	1925, F2	4	22.5	-58.1
	28	0.4	-55.9		5	25.0	-58.1
2061B, G	1	9.0	-58.5		7	18.9	-58.2
	3	18.2	-58.2		7	18.9	-57.9
	10	14.5	-57.8		8	23.7	-57.8
	12	20.3	-58.8		9	18.1	-58.2
					12	25.0	-57.9
					13	19.0	-57.3

1 Inclusion number from the sample.

2 Homogenization temperature.

3 Melting temperature.

4 Drill hole number.

The CH₄ contents of Type I inclusions are evaluated from T_m (melting temperature) and T_h data (Fig. 9.0), using the graphical method of Heyen et al. (1982). Inclusions in quartz from the low gold grade veinlet contain ≤ 3 mole% CH₄, whereas inclusions from auriferous veins contain 4 to 16 mole% CH₄. Inclusions in carbonate fragments from the footwall of the G zone contain ≈ 2 to 3 mole% CH₄, whereas matrix quartz (containing native gold) is methane-rich (2 to 17 mole%, Fig. 9.0). Type I inclusions in quartz show that auriferous mineralization is associated with high CH₄ contents.

9.4 Description of Type II inclusions

Two-phase CO₂ (40 to 70 vol.% at room temperature)-bearing aqueous inclusions are present only in samples from a low Au-grade veinlet, representing < 5 vol.% of the Type I and II fluid inclusion population. Following the method of Burruss (1981), fluid XCO₂ was roughly estimated at ≈ 0.40 to 0.50. Accurate estimates were very difficult given the size of the inclusions. CO₂ melting temperatures of -56.7 to -57.9 indicate minor CH₄ (≤ 6 mole%, generally < 3 mole%) in the CO₂ phase. Salinities are estimated from clathrate melting temperatures using the diagram of Collins (1979), assuming that the clathrate is CO₂. CO₂ clathrate melts at ≈ 6.0 °C to 10.0 °C, corresponding to a salinity of ≈ 7 to 1.4 wt.% NaCl equivalent, using the diagram of Collins (1979). These inclusions are randomly distributed in quartz, with Type I inclusions, suggesting entrapment at roughly the same time.

9.5 Origin of CO₂+CH₄ inclusions

The Type I inclusions may represent a separate fluid from Type II inclusions, or a product of modification of Type II fluids through loss of most or all the H₂O component. Modification of an original H₂O+CO₂+CH₄ fluid may arise from fluid immiscibility and selective removal of the H₂O phase owing to its greater mobility (Santosh et al., 1991), selective loss of H₂O from the inclusion into crystallographic sites in quartz, or leakage of H₂O following dynamic strain of the inclusion-hosting mineral. Type I inclusions occur in relatively strain-free quartz, suggesting that strain-induced loss of H₂O is unlikely. Further, Type I and II inclusions are both hosted by quartz, and both

populations occur in quartz veins from deeper, more highly strained rocks of the Campbell mine. Post mineralization ductile strain is also considered minor, as reported above, and supporting Zhang et al. (1997). Selective loss of H₂O to crystallographic sites in quartz is also considered unlikely as all inclusion populations examined in this study are hosted by quartz, and selective loss from high gold grade breccias would be considered extremely fortuitous.

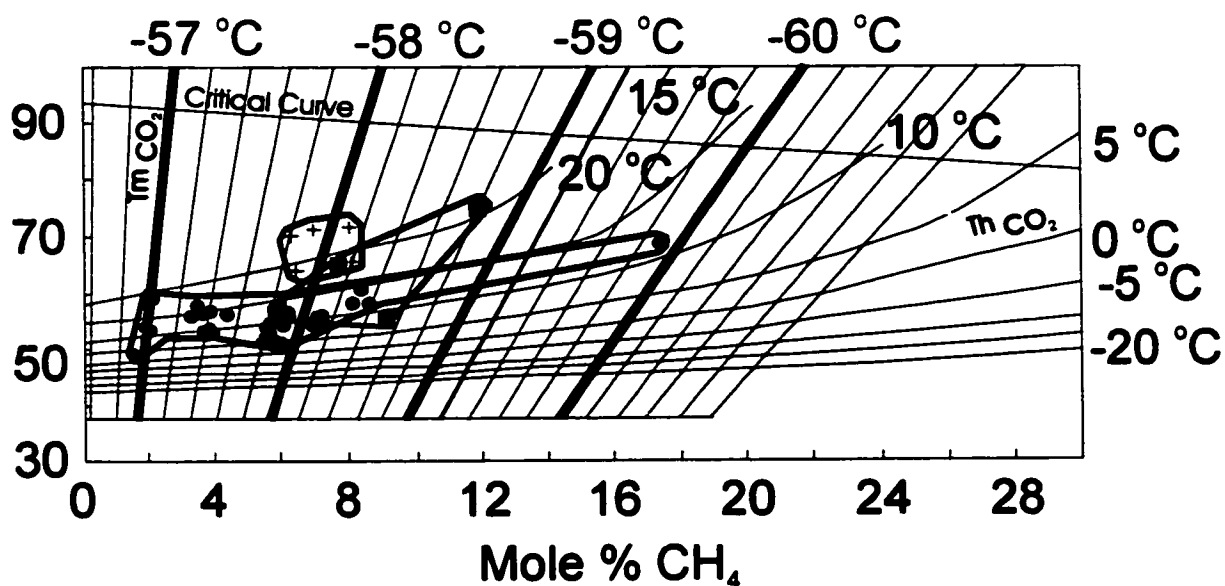
The predominance of primary and pseudosecondary Type I inclusions in high gold grade hydrothermal breccias suggest they are a product of fluid miscibility during gold deposition. The aqueous fluid phase may have been lost from the site of hydrothermal brecciation, as a result of its higher mobility (Santosh et al., 1991). Type II inclusions may represent the auriferous hydrothermal fluid, or the aqueous phase after phase separation. The author favors the former hypothesis.

9.6 Summary of fluid inclusion results

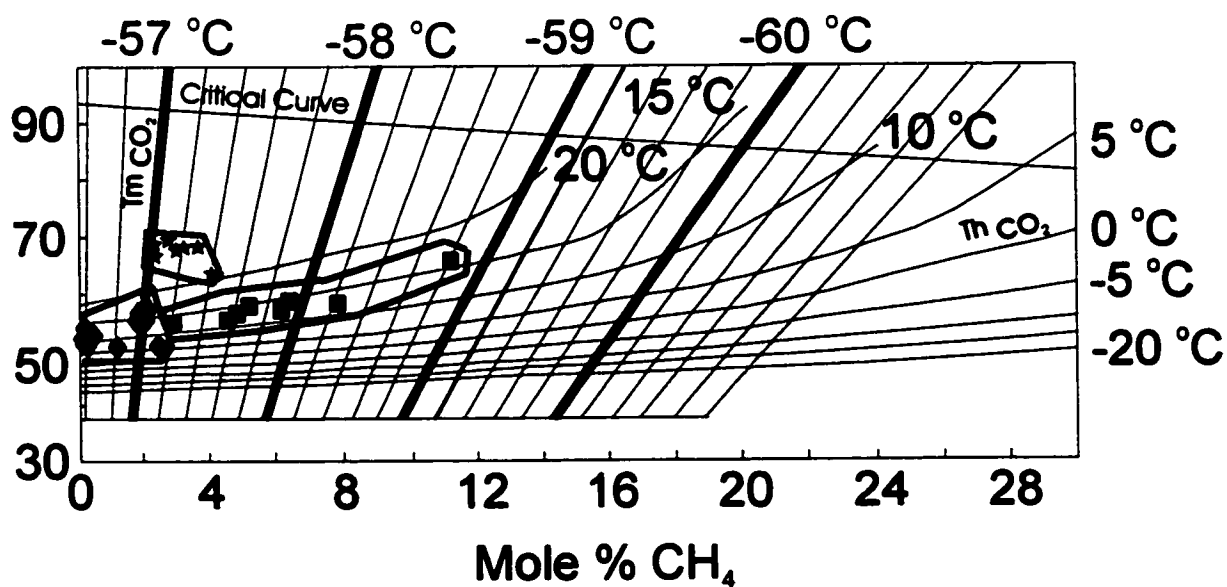
Type I inclusions are ubiquitous in high Au grade quartz + arsenopyrite hydrothermal breccias. Native gold grains occur as fine-disseminated blebs along Type I inclusion trails and clusters of Type I inclusions, suggesting Type I inclusions were trapped during Au deposition. It is unlikely that non-aqueous fluids of Type I inclusion composition were responsible for Au transport, although films of H₂O (<20 vol.%) on inclusion walls could remain undetected by visual means. Type I and II inclusions co-exist in non-ore veins. The composition of Type II inclusions is similar to low salinity, CO₂ + H₂O auriferous fluids described from other Archean Au deposits (Spooner et al., 1985; Colvine et al., 1988; Naden and Shepard, 1989; Bowers, 1991). Therefore, the author concludes that fluids in Type II inclusions were responsible for Au transport, but the fluids were not trapped as inclusions in ore zones.

Figure 9.0

CH₄ content of Type I high-density CO₂ inclusions. The mole% of CH₄ is given by the intersection of the homogenization temperature (Th_{CO₂}) and the melting temperature (Tm_{CO₂}) of a given inclusion. All CH₄ estimates are from inclusions hosted by quartz, except for inclusions within carbonate grains, indicated by stars. Diagram modified after Heyen et al. (1982).



- 1925 stope, F2 Zone
- + 2061 stope, G Zone
- 1861 stope, G zone NW trending footwall vein (quartz hosted, auriferous)



- 2451 P Zone
- ◆ Sub 27
- ★ 1861 stope, G zone NW trending footwall vein (carbonate hosted)

10.0 Wallrock alteration

Four episodes of fracture/fault and pervasive alteration occur in the mine area. The earliest episode involved pervasive bleaching and aluminous alteration in basaltic rocks throughout the Campbell and Goldcorp mine areas. Previous workers have interpreted this event as syn-volcanic or seafloor-related alteration (Rigg, 1980; MacGeehan and Hodgson, 1982; Pirie, 1982; Christie 1986), or early stage aluminous alteration related to later auriferous hydrothermal activity (Penczak and Mason, 1997; Penczak and Mason, 1999). This alteration event was followed by pervasive carbonate alteration, development of early stage banded to crustiform texture carbonate veins along foliation-parallel and oblique shear zones. Silicification, K-metasomatism, As-sulphides, and Au mineralization overprints the widespread carbonate alteration event and pervasive aluminous alteration. Hydrothermal activity continued after gold deposition, and involved volumetrically minor fracture controlled veinlet quartz, tourmaline, and carbonate. Below, the alteration events are described by their timing relative to auriferous veining, as pre-ore, ore-stage, or post-ore.

10.1 Pre-ore pervasive aluminous alteration

The earliest alteration event has resulted in pervasive irregular bleached areas and bleached pillow rims in basalt (Fig. 10.0). The altered basalt is light brown to cream colored, siliceous, and composed of fine-grained white mica (50 to 70 vol.%), anhedral carbonate (< 10 vol.%), and quartz (30 to 50 vol.%). In amphibolite facies domains, andalusite, garnet, and chloritoid commonly formed during the alteration.

This alteration facies is equivalent to the “early alteration” of Rigg (1980) and Christie (1986) and “feldspar destructive alteration” of Mathieson and Hodgson (1984). It is also equivalent to the “early” alteration of Penczak and Mason (1997) and Penczak and Mason (1999). The previous workers have noted that the light brown bleached basalt does not bear a spatial correlation to later auriferous mineralization (Rigg, 1980; Mathieson and Hodgson, 1984; Christie 1986). The “late buff alteration” of Christie (1986) is described as petrographically similar to “early buff alteration”, except that it is auriferous, contains disseminated pyrite, and is brecciated and sheared. This is the auriferous disseminated pyrite mineralization described in section 4.5.1. Sporadic

occurrences of non-auriferous disseminated pyrite mineralization hosted by light brown basalt with intense fine-grained white-mica alteration occurs throughout the mine, adjacent to some auriferous zones. Aluminous, pyritic alteration grades into light brown pervasively bleached basalt, distal to the ore zones (in drifts and cross cuts). It is suggested that auriferous disseminated pyrite mineralization is superimposed on the pre-ore aluminous alteration, during later shearing and brecciation. This is supported by SEM studies of samples from the 07 auriferous zone, which show that native gold occurs in the brecciated and foliated matrix of aluminous altered basalt, and cuts pyrite (Fig. 5.1).

Rhyolitic rocks are also altered to form abundant fine-grained white mica, and rare disseminated blebs of fuchsite and pyrite during the pre-ore aluminous alteration event. In amphibolite facies domains, rhyolitic rocks also contain andalusite.

10.2 Pre-ore carbonate alteration event

This is the most conspicuous alteration in the mine area, producing pervasive Fe-carbonate (Stage 1) as blebs and veinlets. The alteration is intense in mafic and ultramafic rocks, and minor carbonate veinlets occur in the sediments and felsic rocks. This event resulted in pervasive chloritization and destruction of the original mineralogy of basaltic rocks. Early banded to crustiform texture carbonate veins may have been developed during this alteration event, as regional pervasive carbonate alteration shows a broad spatial relation to the important faults controlling hydrothermal alteration, and hosting the early carbonate veins. These veins do not display clearly visible wallrock alteration halos, suggesting that either the hydrothermal fluid which deposited carbonate in the veins was in equilibrium with the wallrocks, or was responsible for the pervasive chlorite and carbonate alteration. Early stage carbonate veinlets cut patchy pre-ore pervasive aluminous alteration.

In both mafic and ultramafic rocks, carbonate occurs as pervasive disseminated grains, narrow veinlets along the foliation, and folded veinlets cutting the foliation. Veinlets cutting the foliation are typically symmetrically folded about the foliation, suggesting they were developed as tension veins at early stages of NE-SW shortening. Carbonate veins occur parallel to the cleavages in rhyolite. This cleavage is subparallel

to the foliation in mafic and ultramafic rocks. Together, these observations suggest that pervasive carbonate alteration occurred early during NE-SW shortening, and was coeval with the development of early crustiform to banded texture quartz + carbonate veins along foliation-parallel zones.

Pervasive carbonate alteration is most intense in the margins of ultramafic rock units. The margins of ultramafic units are green to light green, and contain abundant quartz, carbonate, and chlorite, with minor biotite and fuchsite. Chlorite + quartz + carbonate altered rocks grade inwards to chlorite + quartz + carbonate + talc altered ultramafic rock, with decreasing quartz and carbonate contents. The cores of the ultramafic units consist of talc + serpentine. The progressive inward change from carbonate dominated, through talc + chlorite, to serpentine dominated assemblages reflects decreasing fluid XCO_2 with infiltration, during the pervasive carbonate alteration event.

10.3 Ore stage alteration

Auriferous hydrothermal activity post-dated the emplacement of both foliation-parallel and oblique banded to crustiform texture quartz + carbonate veins. Gold occurs in breccias and quartz veinlets that overprint these vein textures. Visible alteration halos typically extend into the wallrocks 1 to 3 times the vein width. Ultramafic rocks show alteration zones typically several times wider (< 3 to 5 m) than spatially related basaltic rocks (generally < 0.5 m, commonly less than or equal to the vein width). Alteration mineralogy and style show variations with both rock type, and host rock metamorphic grade. The widest alteration envelopes are developed along foliation-oblique auriferous veins.

Throughout the deposit, the style of alteration is well correlated with vein mineralogy. Weakly brecciated auriferous carbonate veins are flanked by narrow envelopes of disseminated fine-grained biotite and pyrite alteration, with coarser-grained biotite and pyrite in fractures, and parallel to the foliation. Intensely brecciated and silicified veins and podiform high Au-grade breccia bodies are flanked by abundant fine-grained disseminated arsenopyrite, and pervasive and fracture controlled silicification, biotite, disseminated pyrite and pyrrhotite, $\pm$ white mica. Outboard of the quartz +

arsenopyrite alteration is a zone of disseminated biotite and pyrite alteration. This alteration zonation is typical for auriferous veins above the 27th level of the mine, although biotitic alteration is more intense adjacent to ore in amphibolite facies domains, and white mica in greenschist facies domains.

10.3.1 Ore stage alteration in greenschist facies domains

In the western parts of the mine, basaltic wallrocks adjacent to auriferous veins are creamy-white, strongly silicified, and contain fine-grained white mica (10 to 30 vol.%), carbonate, ≤ 5 vol.% arsenopyrite, 3 to 8 vol.% subhedral to euhedral pyrite, and minor biotite (< 10 vol.%) (Fig. 4.3). Margarite was identified by SEM-EDS in silicified wallrocks in the G and P ore zones. Veins containing abundant arsenopyrite are bordered by disseminated arsenopyrite in intensely silicified rocks, inboard of silicified, white mica and pyrite altered basalt.

Macroscopic silicification of basalt immediately adjacent to the veins is manifested as grey to grey – green, foliation-parallel wisps and irregular patches. Pyrite, pyrrhotite, and arsenopyrite occur as randomly oriented, finely disseminated grains, and foliation-parallel wisps. Biotite has a similar occurrence to the sulphides. Arsenopyrite + quartz altered wallrocks grade outwards to biotite + pyrite + carbonate assemblages. Biotite, pyrite and carbonate are disseminated in less silicified rocks, and coarser-grained biotite and pyrite occur in foliation-parallel bands and wisps. Distal to the biotite + pyrite zone, light grey-green basalt contains chlorite + white mica + plagioclase + pyrite + pyrrhotite + carbonate, and decreasing modal abundances of quartz. Relatively unaltered basalt distal to the auriferous veins in the western part of the mine contains chlorite + plagioclase + actinolite + quartz + carbonate (Fig. 10.1a), with rare epidote.

Auriferous NW, EW, and NNW veinlets in basalt display common biotite + pyrite assemblages, suggesting these veins are coeval. Boudin necks in brecciated, mineralized, banded carbonate veins contain quartz, arsenopyrite, and coarse-grained pyrrhotite, indicating shortening occurred during auriferous mineralization.

In ultramafic rocks, the wallrocks adjacent to carbonate + quartz + chlorite sheeted veinlets are chlorite + carbonate altered, similar to non-veined margins of the ultramafic units. Grey to green-grey sheeted quartz veinlets with minor carbonate have

biotite + fuchsite + arsenopyrite + pyrite altered wallrocks. Disseminated arsenopyrite (< 35 vol.%), wispy fuchsite, and intense wispy to pervasive silicified zones are developed adjacent to arsenopyrite-bearing sheeted veinlets. Patchy biotitic alteration occurs in ultramafic wallrocks to auriferous veins. Biotite occurs as narrow veinlets and fracture fillings parallel to the foliation, and within the boudin necks of deformed ultramafic rocks. Biotite grains are parallel to the foliation, and oriented parallel to the extension direction in boudin necks, supporting biotite growth during deformation (Fig. 10.2).

Wallrock alteration in basalt adjacent to high gold grade quartz + arsenopyrite breccias is similar to vein-related alteration, with intense patchy and pervasive grey-green silicified zones containing abundant fine-grained disseminated and fracture controlled arsenopyrite (< 30 vol.%) grading outward to a disseminated biotite + pyrite altered basalt. Grey quartz veinlets, and brecciated wallrock with a matrix consisting of grey quartz and white Fe-carbonate are developed adjacent to the podiform ore bodies. Wallrock fragments within podiform ore bodies are angular and occur in a matrix of grey to milky-grey quartz, and locally have a rim of carbonate. The wallrock fragments in quartz are intensely silicified, with 3 to 10 vol.% disseminated pyrite, and < 20 vol.% disseminated arsenopyrite. Biotite alteration is both pervasive and fracture controlled. Fragments enveloped by carbonate display weak visible chlorite alteration, and appear less intensely altered than quartz-hosted fragments.

Ultramafic rocks adjacent to high gold grade podiform quartz + arsenopyrite breccias, and within zones of arsenopyrite-bearing sheeted veinlets, are granular textured, creamy green in colour, and strongly deformed. The granular texture is imparted by coarse-grained carbonate blebs. Biotite occurs along foliation planes, and cuts anhedral grains of pervasive carbonate and chlorite alteration. Pervasive, irregular silicification occurs along biotite-rich zones, and cuts earlier carbonate and chlorite. Wispy fuchsite shows mutual overprinting relations with fracture controlled biotite, suggesting they are coeval. Minor late chlorite overprints the fracture controlled biotite. The mine term for this alteration facies is "altered rock". At the adjacent Goldcorp mine, this rock is termed "chickenfeed".

10.3.2 Ore stage alteration in amphibolite facies domains

In the eastern and deeper (> 850 m below surface) portions of the mine, dark green basaltic rocks distal to the auriferous veins contain chlorite + plagioclase + cummingtonite or anthophyllite + quartz, and chlorite + hornblende + plagioclase with minor epidote and carbonate. In both relatively unaltered basalt and garnetiferous basalt, cummingtonite and orthoamphibole commonly mantle carbonate (Fig. 7.3). In garnetiferous basalt, cummingtonite and anthophyllite occur as inclusions in garnet (Fig. 3.25, 3.29), mantling garnet, and in the basalt matrix. Carbonate contents in basalt increase towards the auriferous veins, but commonly decrease immediately adjacent to the ore veins. Staurolite, chloritoid, anorthitic plagioclase, and rare cordierite and andalusite occur in grey-brown to grey-green weakly silicified basalt immediately adjacent to some low Au-grade quartz veins. The occurrence of cordierite may be related to pre-ore aluminous alteration, overprinted by later hydrothermal alteration related to auriferous veining. Evidence for an early cummingtonite-bearing assemblage is locally preserved in staurolite-bearing rocks as cummingtonite inclusions within garnet. Intense silicification adjacent to some ore veins overprints garnetiferous basalt. In these zones, garnet is anhedral and resorbed.

Basaltic wallrocks adjacent to auriferous veins are steel-grey, grading laterally to brown to grey-brown coloured alteration. Steel-grey altered zones (1 to 5 times the vein width) are intensely silicified (< 65 vol.%), with biotite, carbonate, and abundant fine-grained arsenopyrite (≤ 35 vol.%), and wispy sphalerite. Biotite and carbonate occur principally filling fine fractures cutting arsenopyrite and siliceous zones in this style of alteration. The brown alteration is biotite-rich (≤ 15 vol.%), with subhedral disseminated pyrite (< 10 vol.%), minor disseminated arsenopyrite (< 4 vol.%), and white mica (≤ 5 to 7 vol.%). Auriferous veinlets in basalt from the L zone locally have narrow (1 to 2 cm) wallrock alteration envelopes containing gedrite (20 to 25 vol.%) and garnet (55 to 65 vol.%), with minor anorthitic plagioclase, chlorite, carbonate, and quartz (Fig. 10.3). Mineralogical changes in basalt adjacent to the auriferous veins are shown in Fig. 10.1b.

Auriferous quartz veinlets with <10 vol.% irregular white carbonate, in drill core (27-510, 670' to 680') from beneath the 27th level of the mine, display zonations from biotite, to amphibole altered wallrocks. Grey-brown wallrock alteration adjacent to some

veinlets contains abundant medium-grained biotite with interstitial carbonate and arsenopyrite, grading outwards to a zone of dark green medium to coarse-grained amphibole (30 to 40 vol.%) intergrown with carbonate. Amphibole (typically gedrite) contents decrease sharply into relatively unaltered wallrocks. The boundary between biotite and amphibole alteration is irregular, and clots of amphibole are overprinted by biotite. The relatively unaltered wallrocks are light green-brown to brown basalt, and contain chlorite + biotite + quartz + plagioclase, with pyrrhotite, magnetite (≤ 1 to 2 vol.%), almandine garnet (≤ 2 to 3 vol.%), and minor staurolite, chloritoid, and pyrite.

10.4 Late to post-ore fracture-controlled alteration

Volumetrically minor silica and carbonate alteration (Stage 3) occurred along fractures and faults after Au deposition. Narrow carbonate veinlets cut breccia hosted auriferous quartz + arsenopyrite mineralization. White quartz veinlets, with variable black tourmaline and pyrrhotite, cut auriferous mineralization, occur in boudin necks in brecciated, auriferous early quartz + carbonate veins, and in fractures in post-ore lamprophyre dykes. These quartz veinlets are north-trending, and shallow to moderately E and W dipping. Locally, they are cut by carbonate veinlets (Fig. 3.20), suggesting that late carbonate and quartz veining was synchronous.

10.5 Alteration mineralogy adjacent to auriferous veins

10.5.1 White mica

White mica is fine-grained (< 200 to $400\ \mu\text{m}$) muscovite to paragonite, and is abundant in altered basalts adjacent to the auriferous veins in the western and central parts of the mine. In the eastern and deeper parts of the mine, white mica is less common (≤ 5 to 10 vol.%) in altered basaltic wallrocks.

10.5.2 Biotite

Red-brown biotite occurs in basalt and ultramafic rocks adjacent to ore throughout the deposit, but is most abundant in eastern and deeper levels (≤ 20 vol.%). Wide (1 to 5 m) biotite-rich halos are developed in intensely foliated ultramafic wallrocks, whereas narrow halos are present in basalt. Biotite occurs along the foliation,

and in boudin necks in altered ultramafic rocks, with grains oriented parallel to the foliation. Rarely, biotite is overprinted by randomly oriented chlorite. Biotite displays increasing Al^{VI} and Si^{IV} with decreasing $(\text{Fe}/[\text{Fe}+\text{Mg}])$ and $(\text{Fe}^{+2}+\text{Mg})$ toward the auriferous veins.

10.5.3 Amphiboles

Amphiboles in the wallrock adjacent to auriferous veins and within veins in the eastern and deeper parts of the mine include calcian gedrite, gedrite, and magnesio-tschermakite (2 to 5 in Table 10.0). In staurolite-bearing altered basalt the amphiboles are typically absent, or restricted to inclusions within garnet. In basalt distal to ore, cummingtonite and anthophyllite overprint carbonate (Fig. 7.3), and cummingtonite occurs as inclusions in garnet, as well as mantles on garnet grains. Both cummingtonite and tschermakite are replaced by biotite and carbonate in the wallrocks near auriferous veins. Cummingtonite is locally present within auriferous veins as yellowish fine-grained rims on carbonate fragments from the eastern parts of the mine.

10.5.4 Garnet

Garnet is subhedral to anhedral, and tabular garnet porphyroblasts are developed along the foliation. Coarse-grained garnet is compositionally zoned, with Mn- (spessartine $\leq 22.5\%$) and Ca- (grossular $\leq 15.3\%$) rich cores containing cummingtonite, quartz, magnetite, and pyrrhotite inclusions, and Fe- and Mg- rich, inclusion poor rims. In the wallrocks in the central parts of the deposit, garnet contains inclusions of carbonate, arsenopyrite (Fig. 10.4), tetrahedrite, chalcopyrite, scheelite, tourmaline, and gersdorffite. Garnet is cut by fractures containing pyrrhotite, carbonate, and tetrahedrite (Fig. 10.5). Garnet is replaced by chlorite (Fig. 10.6) and quartz adjacent to some auriferous veins. This suggests that hydrothermal activity continued after garnet growth.

10.5.5 Plagioclase

Plagioclase (Table 6.2) in basalt is andesine in composition (An_{30-45}) and is locally twinned, whereas plagioclase in auriferous veins (Fig. 10.7) and in wallrocks

adjacent to veins is distinctly Ca-rich (An_{63-92}) without twinning and it contains numerous fluid inclusions. Plagioclase occurs with andalusite, and is cut by carbonate.

10.5.6 Andalusite and cordierite

Andalusite occurs in altered basaltic rocks with cordierite or staurolite adjacent to ore veins in the deeper and eastern parts of the mine. Cordierite occurs in ultramafic rocks, and rarely in basaltic wallrocks. Cordierite is deformed and cross-cut by biotite and chlorite, and encloses fine-grained inclusions of quartz and minor Fe-sulphide, in a matrix of coarse-grained biotite, quartz, and Fe-sulphide (Fig. 3.30). These observations suggest that cordierite preserves the original pre-metamorphic fine-grained texture of the rock, and pre-dates much of the Fe-sulphide deposition. The occurrence of chlorite and biotite cutting cordierite adjacent to auriferous veins in basalt, suggests hydrothermal alteration post-dating cordierite growth. Cordierite in basalt is part of the pre-ore, aluminous alteration assemblage.

10.6 Fluid XCO_2 determined from the alteration minerals

Basaltic rocks distal to the auriferous veins contain carbonate, cummingtonite, anthophyllite, chlorite, quartz, with, or without garnet. Cummingtonite, garnet, and locally anthophyllite overprint carbonate (Fig. 7.3). The XCO_2 of the fluid in equilibrium with this assemblage is estimated at ≈ 0.025 , using the TWEEQU software of Berman (1991).

In the eastern part of the mine, closer to the auriferous veins, garnet becomes abundant and contains inclusions of cummingtonite and anthophyllite. Cummingtonite is absent from the matrix. Adjacent to the auriferous veins, carbonate overprints the hornblende, gedrite and anthophyllite. Garnet and amphibole are absent from intensely altered (silicified, arsenopyrite-rich) rocks adjacent to the auriferous veins above the 24th level in the mine. The zonation of alteration mineral assemblages in basalt is illustrated in Figs. 10.1a and 10.1b. Drill core from beneath the 27th level shows wallrock alteration zoning from biotite + carbonate, to amphibole, outwards from quartz veins. Biotite overprints amphibole at the boundary between biotite and amphibole alteration zones. The spatial variations in the occurrence of biotite, garnet, and

Table 10.0. Representative composition of amphibole from auriferous hydrothermal alteration

	1	2	3	4
SiO ₂	41.93	46.90	49.48	46.53
Al ₂ O ₃	14.84	9.25	5.76	9.99
FeO*	26.07	13.85	14.73	15.33
MgO	4.75	12.72	14.42	12.21
TiO ₂	0.12	0.26	0.00	0.23
MnO	1.05	0.68	0.56	0.53
CaO	6.55	11.15	10.65	11.05
Na ₂ O	0.91	0.83	0.00	1.16
K ₂ O	0.16	0.00	0.00	0.10
Cl	0.06	0.00	0.11	0.00
Total	96.44	95.64	95.71	97.13
Σ (O basis)	96.43	95.64	95.69	97.13
Formula (Basis 23 O, Cl)				
Si ⁺⁴	6.511	6.987	7.346	6.879
Al ⁺³	1.489	1.013	0.654	1.121
Σ T site	8.000	8.000	8.000	8.000
Al ⁺³	1.228	0.611	0.353	0.620
Fe ⁺²	3.386	1.726	1.828	1.896
Mg ⁺²	1.099	2.824	3.190	2.691
Ti ⁺⁴	0.013	0.029	0.000	0.026
Mn ⁺²	0.138	0.086	0.070	0.067
Σ C site	5.864	5.276	5.441	5.300
Ca ⁺²	1.090	1.779	1.695	1.750
Na ⁺	0.275	0.239	0.000	0.332
K ⁺	0.032	0.000	0.000	0.018
Σ B site	1.397	2.018	1.695	2.100
Cl ⁻	0.017	0.000	0.029	0.000

Cr₂O₃ below detection limit, = 0.01 wt.%* All Fe calculated as Fe⁺²

1 Calcian gedrite with garnet and anorthite, in the wallrocks adjacent to an auriferous quartz vein, L zone, 1851 stope.

2,3,4 Magnesio-hornblende with anorthitic plagioclase in an auriferous vein. From drill core beneath the 27th level. Drill core 27-267.

amphibole are consistent with increasing X_{CO_2} towards the ore veins (Fig. 3.35), and diagrams in Will et al. (1990).

Silicified wallrocks to the P zone, containing margarite + anorthitic plagioclase + muscovite + garnet + carbonate + biotite + chlorite provide an X_{CO_2} estimate of ≈ 0.45 (Fig. 10.8), at temperatures of $\approx 450^\circ\text{C}$. The variations in the mineral assemblages in basaltic rocks distal to auriferous veins, and in altered basalt proximal to auriferous veins, suggest that fluid X_{CO_2} decreased into the wallrocks. Textural relations between biotite, amphibole and carbonate outwards from auriferous veins support this interpretation.

The estimate of X_{CO_2} from Type II inclusions in auriferous quartz veinlets is compatible with the X_{CO_2} from the alteration assemblage adjacent to the P zone (Fig. 10.8), suggesting that a fluid of this composition may be related to the auriferous alteration event.

10.7 Summary of time relations for metamorphism and hydrothermal alteration from mineral textures

The textural evidence suggests that auriferous hydrothermal alteration was broadly synchronous with the peak metamorphism. Pre- to syn-metamorphic hydrothermal activity is suggested by almandine garnet enclosing carbonate and ore-related mineral inclusions (Fig. 10.4). Syn to post-peak metamorphic hydrothermal activity is suggested by metamorphic minerals cut by veins of tourmaline and metallic minerals. Auriferous hydrothermal alteration continued after the peak metamorphism and resulted in retrograde alteration of garnet and cordierite.

Tourmaline is closely associated with native gold and aurostibite, and biotite is common in ore veins and wallrock alteration assemblages. Tourmaline and biotite show decreasing $\text{Fe}/[\text{Fe}+\text{Mg}]$ with increasing sulphide contents near auriferous veins (Fig. 3.27), suggesting co-deposition of these minerals. Results from garnet-biotite thermometry suggest the formation of biotite during the peak metamorphism. Textural relations between biotite, cummingtonite, garnet, anthophyllite, and carbonate indicate increasing fluid X_{CO_2} towards auriferous veins.

10.8 Changes in mineral assemblages during alteration, and the timing of metamorphism •

Following the concepts of Korzhinskii (1959, 1965) and Thompson (1959), alteration can be viewed as a composite of quasi-equilibrium systems, where each system is represented by individual alteration envelopes. During progressive alteration, reaction fronts will shift; early minerals become unstable and replaced by new minerals in response to the change of bulk chemical composition of rock and fluid. Early mineral assemblages may be preserved in less altered rocks or as inclusions within relatively resistant minerals.

In altered basaltic wallrocks containing garnet and staurolite, cummingtonite is restricted to inclusions in garnet cores. Relatively unaltered basalt preserves cummingtonite in the matrix. The relationship, observed in drill core samples from the rocks below the 27th level in the mine, suggests that cummingtonite was a relatively early phase, together with Ca- and Mn-rich garnet. During deformation and hydrothermal alteration, cummingtonite in the matrix became unstable, but cummingtonite enclosed by garnet was effectively isolated from the reacting matrix.

Change in P-T conditions: Cummingtonite appears in some basaltic rocks at the uppermost greenschist facies and is stable during the amphibolite facies at high and low pressures (Spear, 1993). Significant changes in metamorphic pressure are geologically unlikely and not supported from mineral assemblages.

Changes in bulk composition: Staurolite, biotite, and chloritoid are unusual minerals in basaltic rocks, and indicates that alteration with Al and K enrichment prior to, or during metamorphism. In contrast, cummingtonite is stable in relatively Fe-Mg rich, and Ca, Al and alkali poor rocks. The occurrence of cummingtonite as inclusions in garnet, and its absence in the rock matrix suggest that Al and K enrichment occurred during amphibolite facies metamorphism, as both garnet + cummingtonite, and garnet + staurolite + chloritoid reflect amphibolite grade metamorphic conditions. This suggests that a change in the bulk-rock composition during hydrothermal alteration led to the instability of cummingtonite, and must have occurred during amphibolite-facies metamorphism.

10.9 AFM representation of change in bulk-rock composition during hydrothermal alteration •

Here, the author uses AFM(Ca) diagrams to show the change in bulk composition of basalt during auriferous hydrothermal activity. As a result, samples affected by pre-ore aluminous alteration were excluded from this study. Phase relations in basaltic rocks in the system $K_2O-FeO-Fe_2O_3-MgO-Al_2O_3-SiO_2-CaO-Na_2O-TiO_2-H_2O$ can be displayed on AFM diagrams projected from anorthitic plagioclase with the presence of magnetite, ilmenite, quartz, and H_2O (Froese, 1981). Typical bulk rock compositions of basalts are provided in Table 3.0 and the composition of the samples used for the projection are provided in Appendix 3. Plagioclase is common to all assemblages, and an average plagioclase composition (An_{66}) is used for the projection. The Fe_2O_3 and TiO_2 components are minor, occurring primarily in magnetite, and ilmenite and rutile, respectively. The $FeO/[FeO + Fe_2O_3]$ for basaltic rocks in the mine is generally 0.88 to 0.95 (wt.% basis), (Christie, 1986), and is uniform between relatively fresh and intensely altered rocks. Whole rock Fe was recalculated assuming a $FeO/[FeO + Fe_2O_3]$ ratio of 0.92, following determinations made by Christie (1986). Na_2O is also a minor component, occurring predominantly in plagioclase and tschermakitic amphiboles, and in paragonite in the western portions of the mine. Bulk rock compositions are plotted on a carbonate and biotite-free basis and removing K_2O from the system. Carbonate contents were estimated from CO_2 analysis, or from modal abundance in thin sections. Where the data was unavailable, whole rock data was recalculated assuming 3 vol.% carbonate, from the typical abundance of carbonate in basalt from the mine. The contents of K_2O in the bulk rock compositions were assumed to be represented by biotite because the orthoclase component in plagioclase is negligible (Table 6.2) and because biotite predominates over muscovite in the samples for this plot collected below the 17th level. Average biotite and carbonate compositions are shown on figures 3.27 and 6.0, respectively. Constant H_2O activity is assumed, which reduces the system to four principle components ($FeO-MgO-Al_2O_3-CaO$) for the graphical display on an AFM diagram. As shown previously, XCO_2 varied significantly, and this assumption is evaluated below.

On AFM diagrams, bulk rock compositions of basalt generally plot at and below the F-M join. With increasing Al, samples plot closer to the A-F join (Fig. 10.9). Relatively unaltered basalts (basalts and chloritic basalt) plot at, and below the point for chlorite in the diagram. Altered basalts (silicified and some garnetiferous basalts) plot above the position for chlorite, towards the Al apex. The trend from chloritic, to garnetiferous and silicified basalt shows the progressive change in bulk composition on the AFM projection as a result of alteration. The distribution of rock compositions on the AFM diagram, and the mineralogy of the rocks, indicates either pre- or syn-peak metamorphic hydrothermal alteration. Garnet compositions form a J shaped array, due to higher grossular component in pyrope-poor garnet relative to pyrope-rich garnet.

Overlap between phase triangles on AFM diagrams indicates changes in P, T, fugacity of CO_2 or H_2O , or the influence of an extra component. Significant change in P across the mine is not likely, but variations in T and $a_{\text{H}_2\text{O}}$ are expected. Phase relations from samples 16 and 22 are presented in Fig. 10.10. Both samples were obtained from drill core from deeper than the 27th level, and have experienced similar grades of metamorphism. The early assemblage observed in garnet cores (garnet + cummingtonite + chlorite) in sample 22 is compatible with the composition of relatively unaltered basalt and chloritic basalt compositions. This assemblage occurs distal to the auriferous veins. The later assemblage garnet (rim) + staurolite + chlorite occurs in altered basaltic rocks adjacent to auriferous veins. Staurolite-bearing assemblages in samples 16 and 22 show minor overlap in Fe-Mg that may be due to the relatively sulfide-rich assemblage in sample 16, but does not indicate significant differences in the intensive variables between these samples.

Two assemblages, garnet (core) + cummingtonite + chlorite and garnet (rim) + staurolite + chlorite, do not overlap, indicating that there is no significant changes in intensive variables between the two assemblages. Investigation of XCO_2 during alteration shows that significant variations in XCO_2 occurred during alteration, which are not reflected in the phase relations. Although this is considered unusual by the author, it suggests that variations in XCO_2 may not have significantly affected the phase relations on the AFM projection. Instead, the change in assemblages suggests a change in bulk composition resulting in the consumption or disappearance of cummingtonite in the

staurolite-bearing assemblages. This proposal is supported from the spatial occurrence of cummingtonite, and the occurrence of cummingtonite as inclusions in garnet, in altered basalt. The bulk composition progressed through chlorite, towards staurolite, to crystallize early cummingtonite, and later staurolite, in a garnet + chlorite assemblage. The occurrence of cummingtonite in altered basalt, as inclusions in garnet cores with almandine and pyrope rich rims, suggests syn-peak metamorphic hydrothermal alteration, consistent with phase relations observed in sample 22, and variations in bulk rock compositions with increasing alteration intensity.

Figure 10.0

Pervasive aluminous alteration in pillow basalt. This early aluminous alteration has resulted in selective bleaching of pillow rims (visible in the center of the photo), and flow top breccias. 30th level, 3061 stope area.

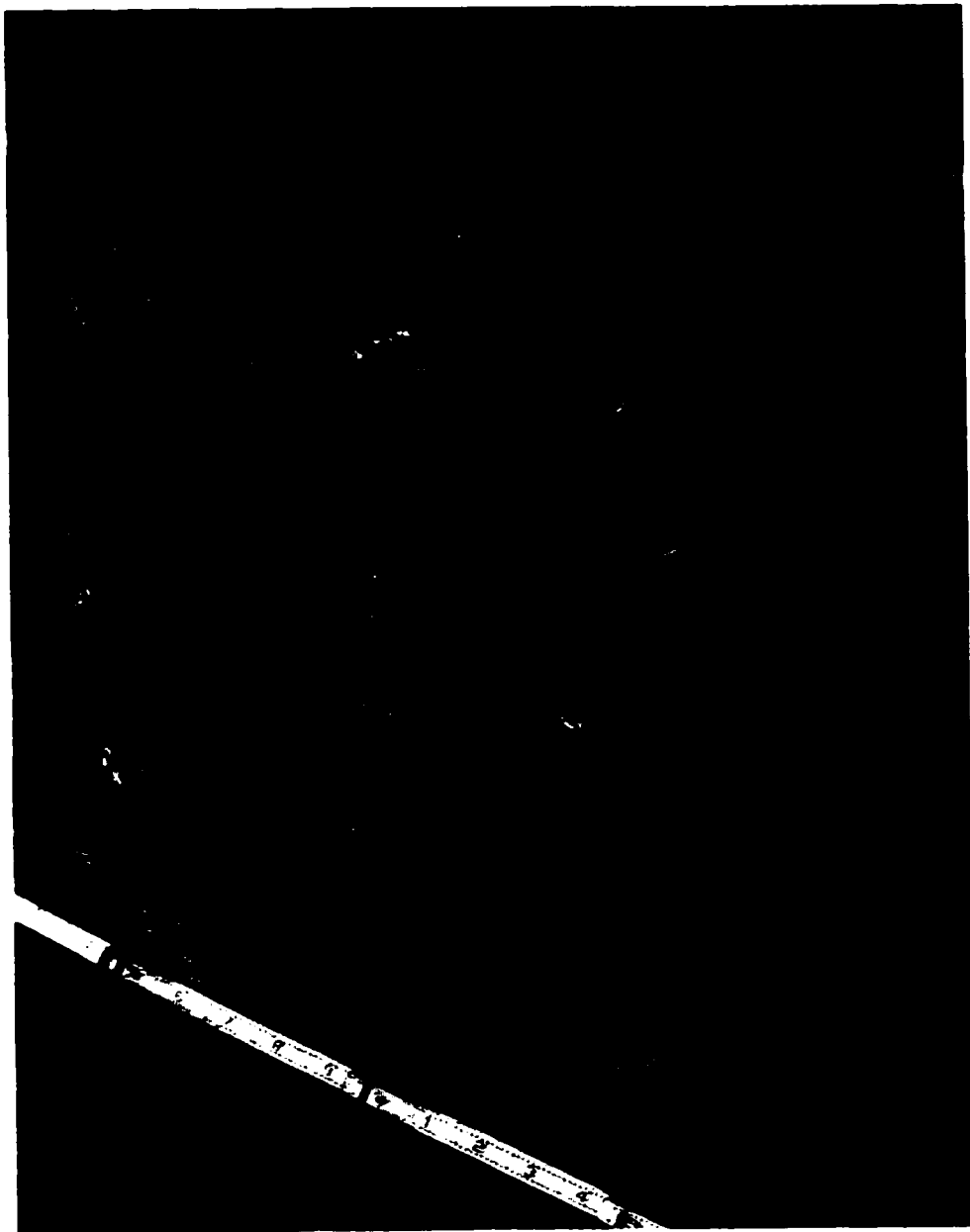


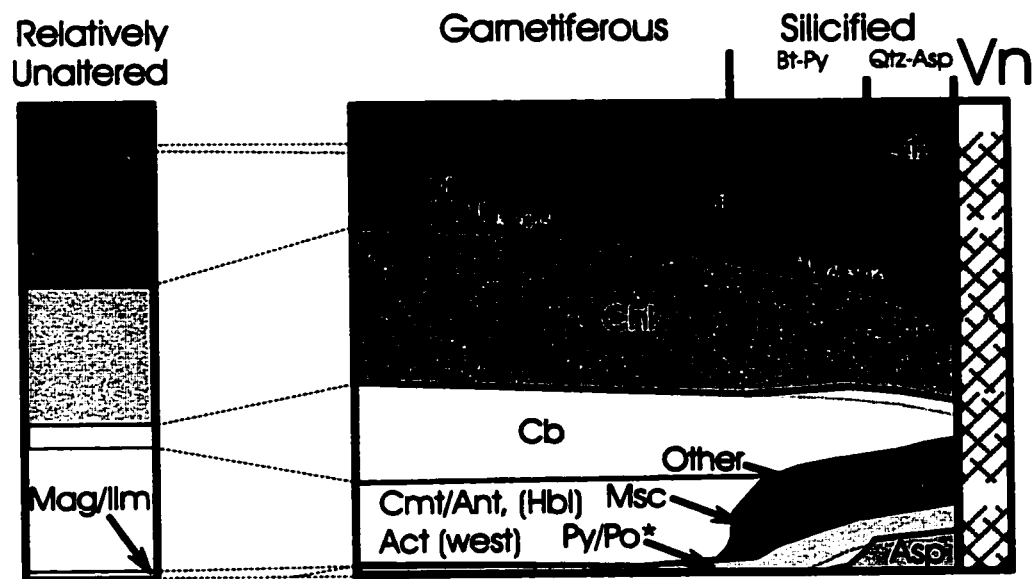
Figure 10.1

Schematic diagram showing wallrock alteration zoning of basalts in the amphibolite facies domain.

10.1a shows the alteration zonation in the wallrocks to auriferous veins and breccias at lower amphibolite facies. Away from auriferous veins, wallrock alteration progresses from a silicified zone with quartz (Qtz)+ arsenopyrite (Asp), outwards to a zone with biotite (Bt)+ pyrite (Py), to weakly altered, garnetiferous wallrocks. The garnetiferous wallrocks contain garnet (Grt), cummingtonite (Cmt), anthophyllite (Ant), or rare hornblende (Hbl). Actinolite (Act) occurs in the western part of the mine, in garnet-free basalt. These amphiboles, and occasionally garnet, are overprinted by silicification associated with the auriferous mineralization. Fine-grained white mica (Msc) occurs in both greenschist and lower amphibolite facies metamorphic domains, but is more abundant in the wallrocks of veins hosted by lower metamorphic grade rocks. Pyrrhotite (Po) occurs with Py, and is relatively more abundant closer to auriferous veins. Magnetite (Mag) and ilmenite (Ilm) occur distal to the veins in relatively unaltered rocks. Metasomatic plagioclase (Pl) in the wallrocks is An-rich adjacent to auriferous veins, and plagioclase in less altered basalt distal to the veins is andesine to labradorite in composition. This style of alteration is associated with high Au grade quartz + arsenopyrite mineralization, and auriferous veinlets from the western part of the deposit.

10.1b shows alteration zoning associated with narrow auriferous quartz veinlets in basaltic wallrocks below the 27th level of the mine, in the amphibolite facies domain. Narrow almandine garnet, biotite, and gedrite (Ged) alteration halos are developed adjacent to the veins. Biotite rich selvages occur at the margins of the veins, and overprint gedrite rich alteration outboard of this zone. Outboard of the silicified zone, less altered wallrocks contain a greater modal abundance of carbonate and chlorite. Garnet is sporadically developed in this assemblage, and amphiboles are typically cummingtonite or anthophyllite, with minor hornblende.

A



B

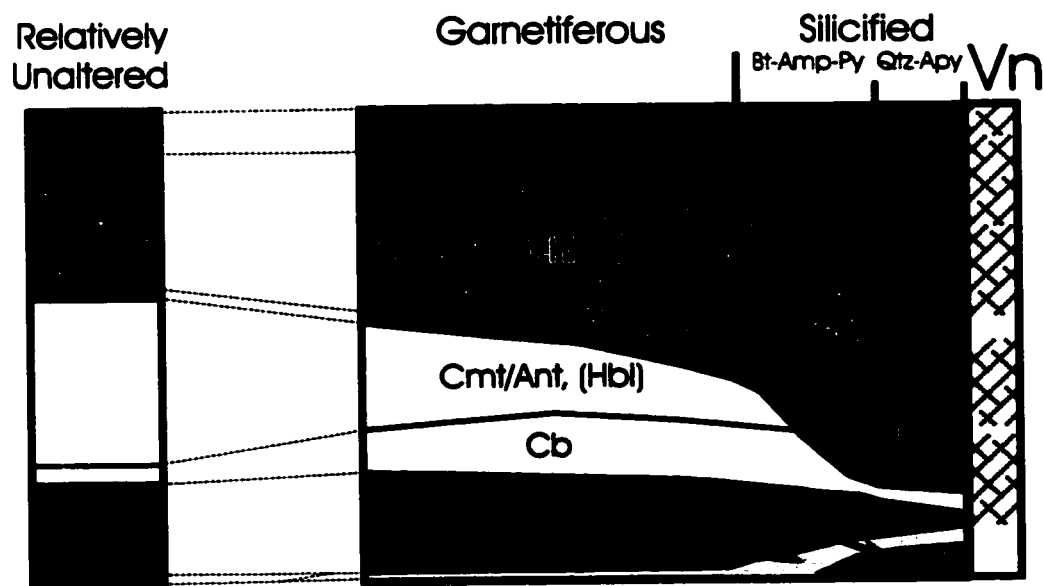


Figure 10.2

Photomicrograph showing biotite (brown) replaced by chlorite (green) at the margins of boudinaged, carbonate altered fine-grained ultramafic rock (light grey). Biotite and chlorite are aligned along the foliation, and are oriented parallel to the extension direction in the boudin neck. This suggests that biotite alteration occurred during boudinage, and was overprinted by later chlorite during the deformation. The width of the photo is approximately 2.5 mm.

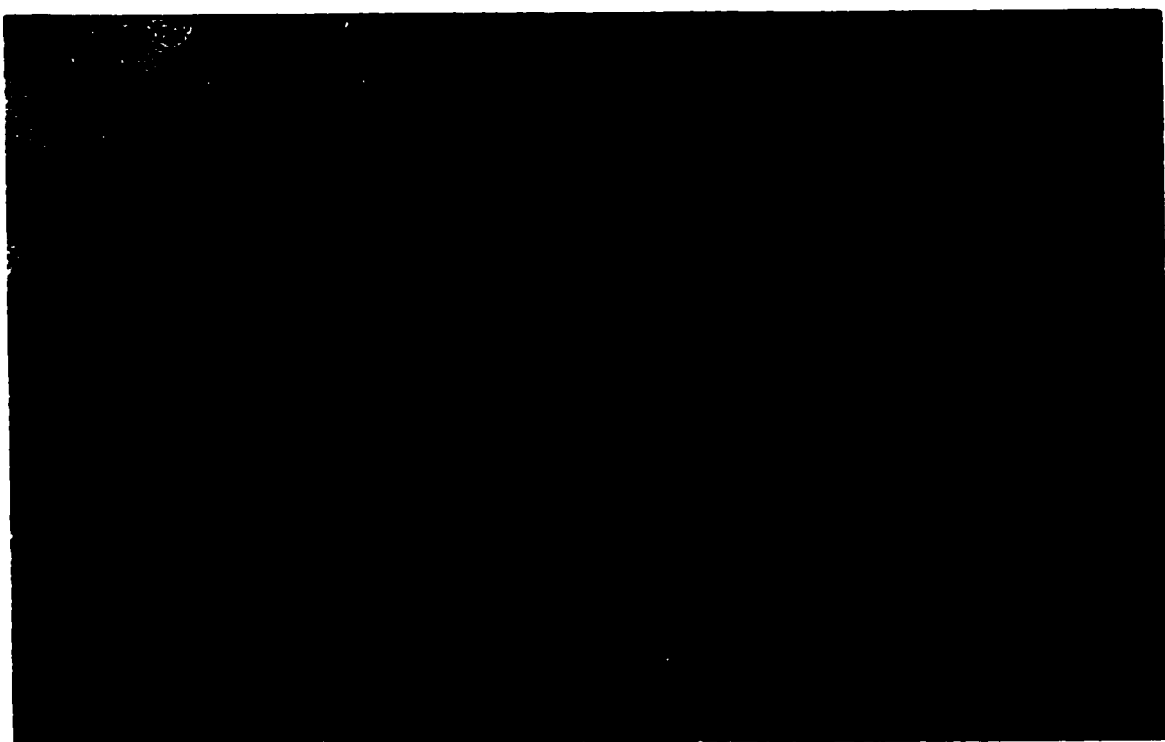


Figure 10.3

Photomicrograph of an auriferous quartz vein from the L zone, 1851 stope. Garnet (Grt) and gedrite (Ged) occur in the quartz vein (qv) and in the wallrocks immediately adjacent to the vein, and are intergrown with quartz in the vein. Fluid inclusions are CO₂-rich, forming secondary and pseudosecondary planar arrays perpendicular to the foliation in the wallrocks. Randomly oriented primary inclusions are more abundant toward the center of the vein. Native gold (Au) grains occur along some pseudosecondary arrays.

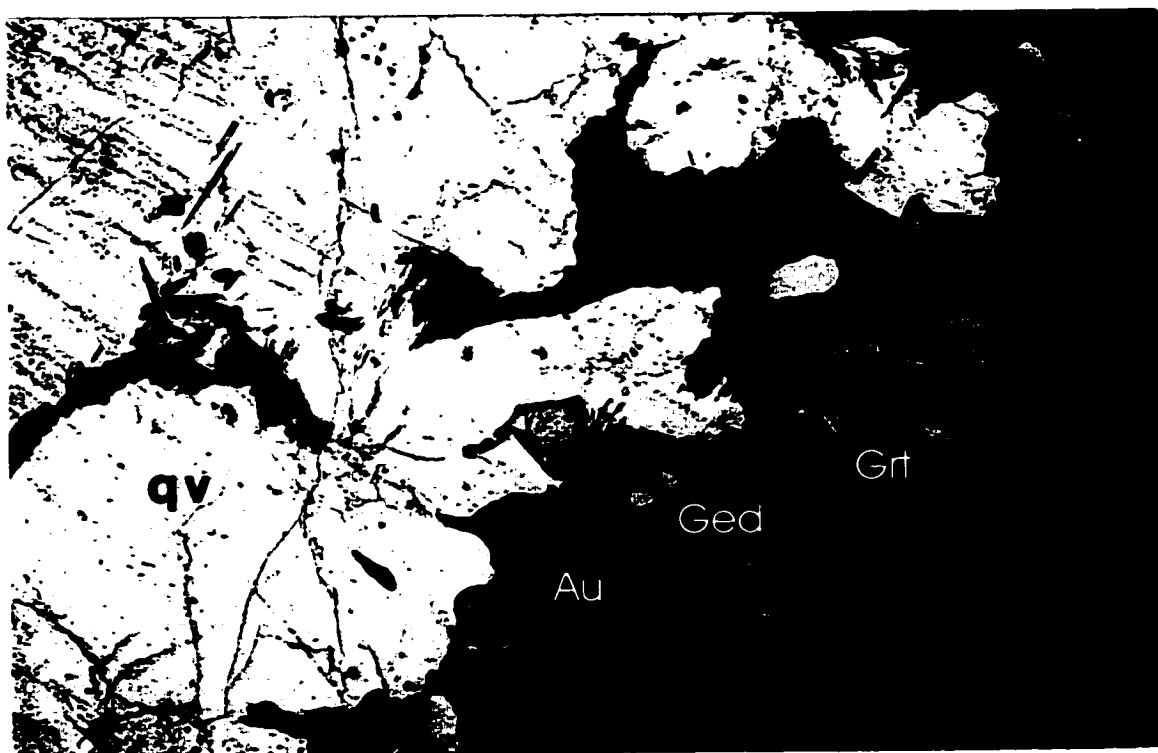


Figure 10.4

Reflected light photomicrograph of arsenopyrite (Apy) and pyrrhotite (Po) enclosed by garnet (Grt), indicating arsenopyrite deposition was pre- to syn-peak metamorphism. Sample from the P zone, 2551 stope.

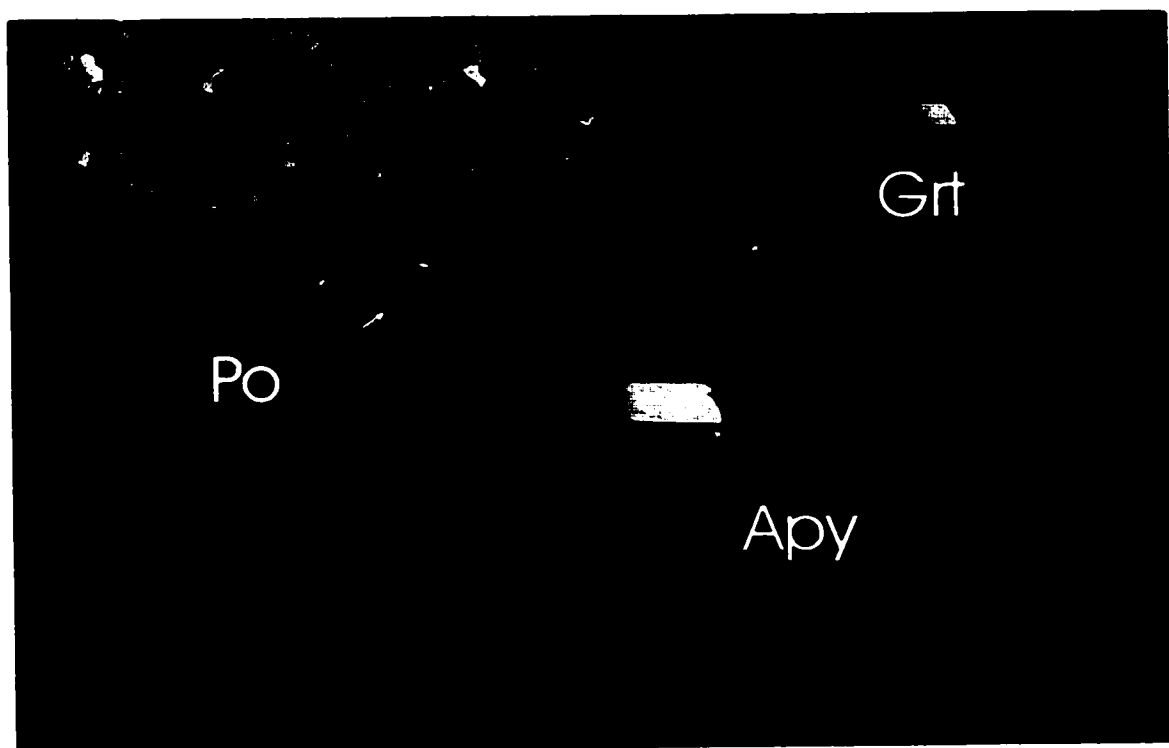


Figure 10.5

Reflected light photomicrograph of garnet (Grt, light grey) cut by a fracture containing tetrahedrite (tet). The fracture filling minerals are verified by SEM-EDS. The bright phases are arsenopyrite and pyrite. The field of view approximately 3 mm.



Figure 10.6

Photomicrograph showing a pseudomorph of garnet, composed of chlorite, quartz, and pyrrhotite (opaque phase) in the wallrocks to an auriferous quartz vein. The field of view is 4 millimeters.



Figure 10.7

Photomicrograph of an auriferous quartz veinlet from a drill core. Quartz grains are polygonal with weak undulose extinction, and form platy grains parallel to the vein margin (Qv, inset). Anorthitic plagioclase (low birefringence, Cb/An in inset) is anhedral, intergrown with quartz, and cut by carbonate (high birefringence, Cb/An in inset). The wallrocks (Wlrk, inset) are composed of fine-grained quartz (grey), biotite (brown), and pyrite (opaque phases). The field of view is approximately 3 mm. Abbreviations; Sample crl-16.

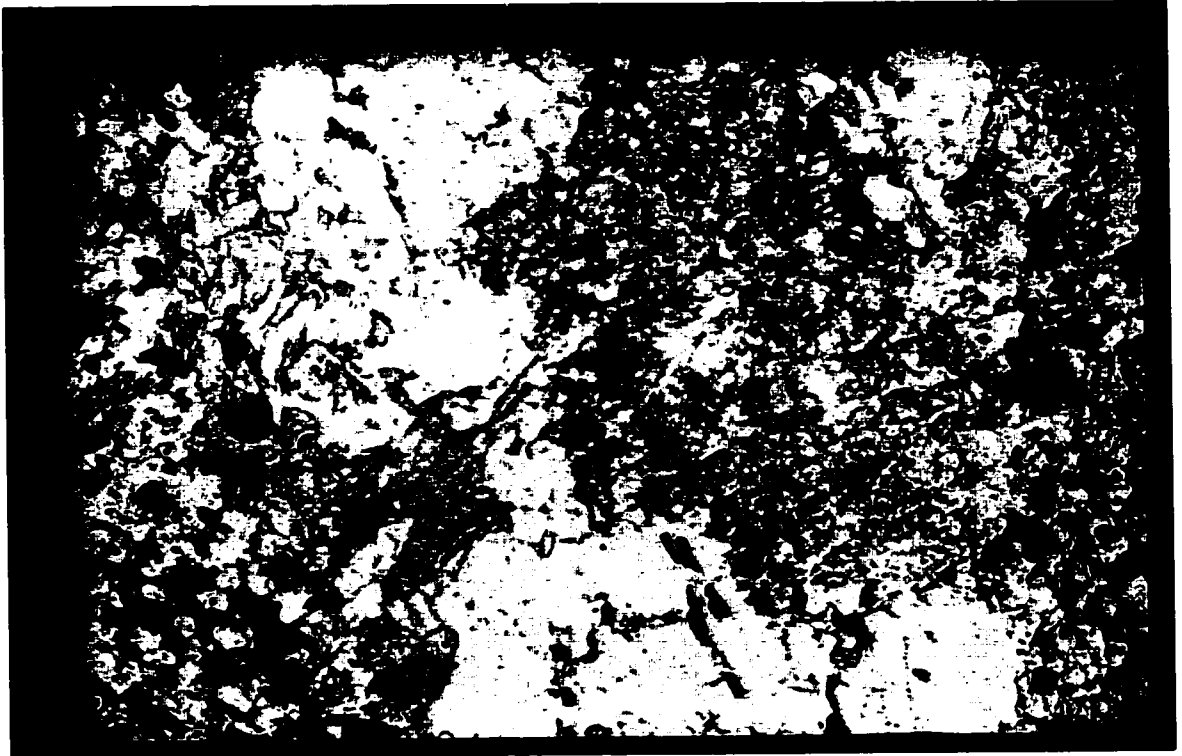


Figure 10.8

Temperature vs. X_{CO_2} diagram at 3 Kbars pressure for the wallrock alteration mineral assemblage adjacent to auriferous veins in the P ore zone. The assemblage of quartz+dolomite+chlorite+white mica+biotite+margarite suggest X_{CO_2} of ≈ 0.45 at ≈ 450 °C, which is comparable to the temperature estimated from mineral assemblages and garnet-biotite thermometry (Table 3.6, samples 7a and 7b). Diagram calculated using the TWQ 1.02 program of Berman (1991), incorporating the compositional data of biotite, plagioclase, garnet, and muscovite, and the mixing models of Berman (1990) for garnet, Furhman and Lindsey (1988) for plagioclase, Chatterjee and Froese (1975) for muscovite, and McMullin et al. (1991) for biotite. Labeled reactions are 1) $Chl + 11An + 2W + 10CO_2 = 5Do + 6Mrg + 13aQz$, 4) $13Ann + 7An + 3Chl + 8Mrg + 30CO_2 = 13Ms + 15Do + 13Alm + 20W$, 5) $3Ann + 5An + Chl + 10CO_2 = aQz + 3Ms + 5Do + 3Alm + 4W$. Abbreviations: Alm=almandine, Ann=annite, aQtz=alpha quartz, Chl=chlorite, $CO_2=CO_2$, Do=dolomite, Mrg=margarite, Ms=muscovite, $W=H_2O$

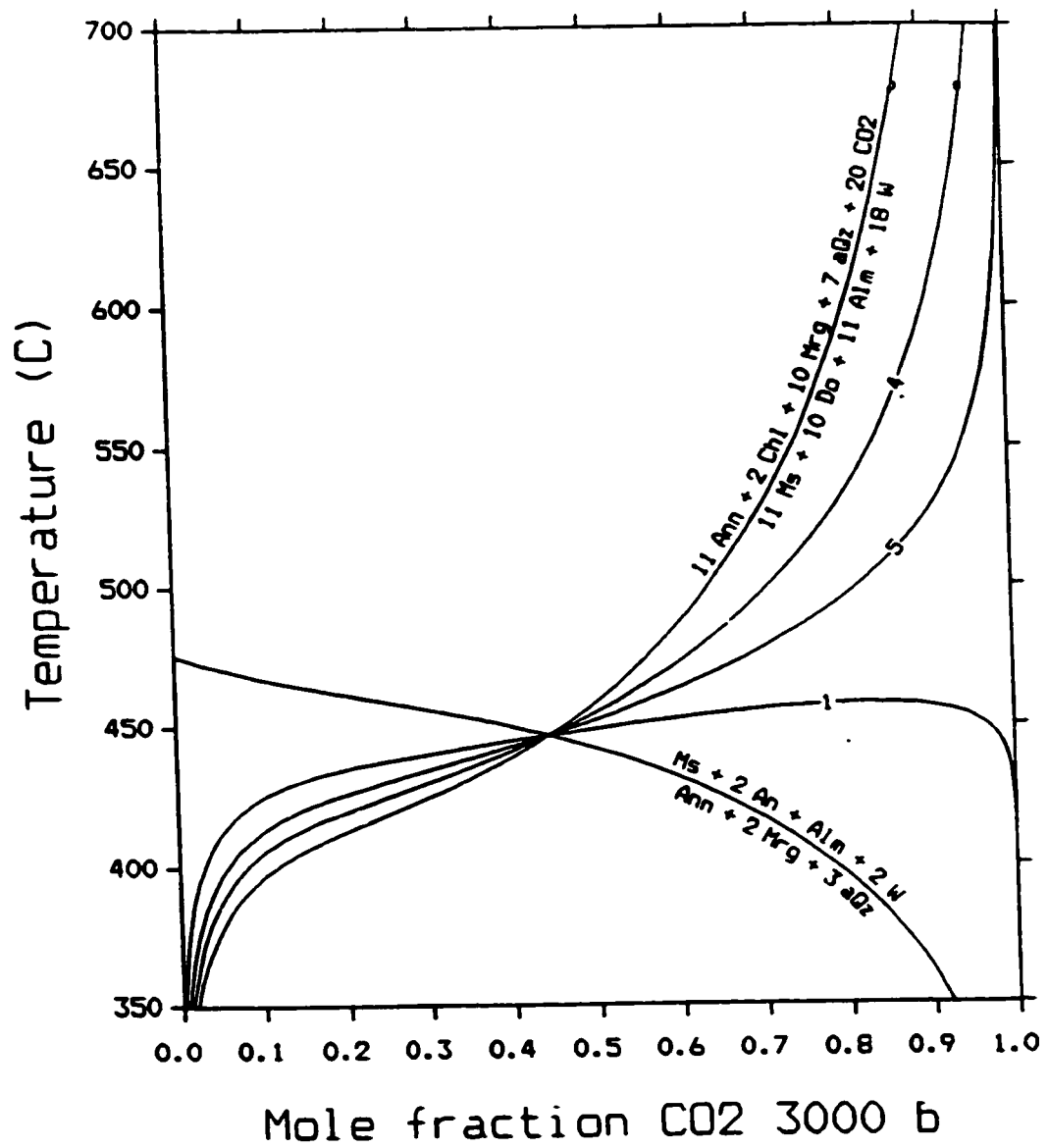


Figure 10.9.

AFM diagram for altered basaltic rocks (solid circles) projected from the phases shown on the right.

Ranges in mineral compositions at the mine are shown as shaded grey areas. Basalt described as; silicified, garnetiferous, relatively unaltered basalt and chloritic are shown as fields separated by light dashed lines, and described in the text. Average compositions for relatively unaltered basalt (⊙) distal to the auriferous veins, n=15; garnetiferous basalt (*), n=3; and silicified basalt (◆), n=6; from Placer Dome Exploration Ltd. Samples are listed in Table 3.0. Samples with abundant veining and sulfide mineralization, and those representing obvious pre-ore pervasive aluminous alteration, were excluded from consideration. The data include the bulk rock compositions of altered basalt determined by Placer Dome Canada Ltd. (1994), and Christie (1986).

Increasingly altered and silicified basalt compositions plot towards the A apex and the A-F join. Details of the construction of the AFM diagram are provided in the text. Abbreviations; Chl=chlorite, Cmt=cummingtonite, Grt=garnet, Ilm=ilmenite, Mag=magnetite, Qtz=quartz, St=staurolite, Tsc=tschermakite.

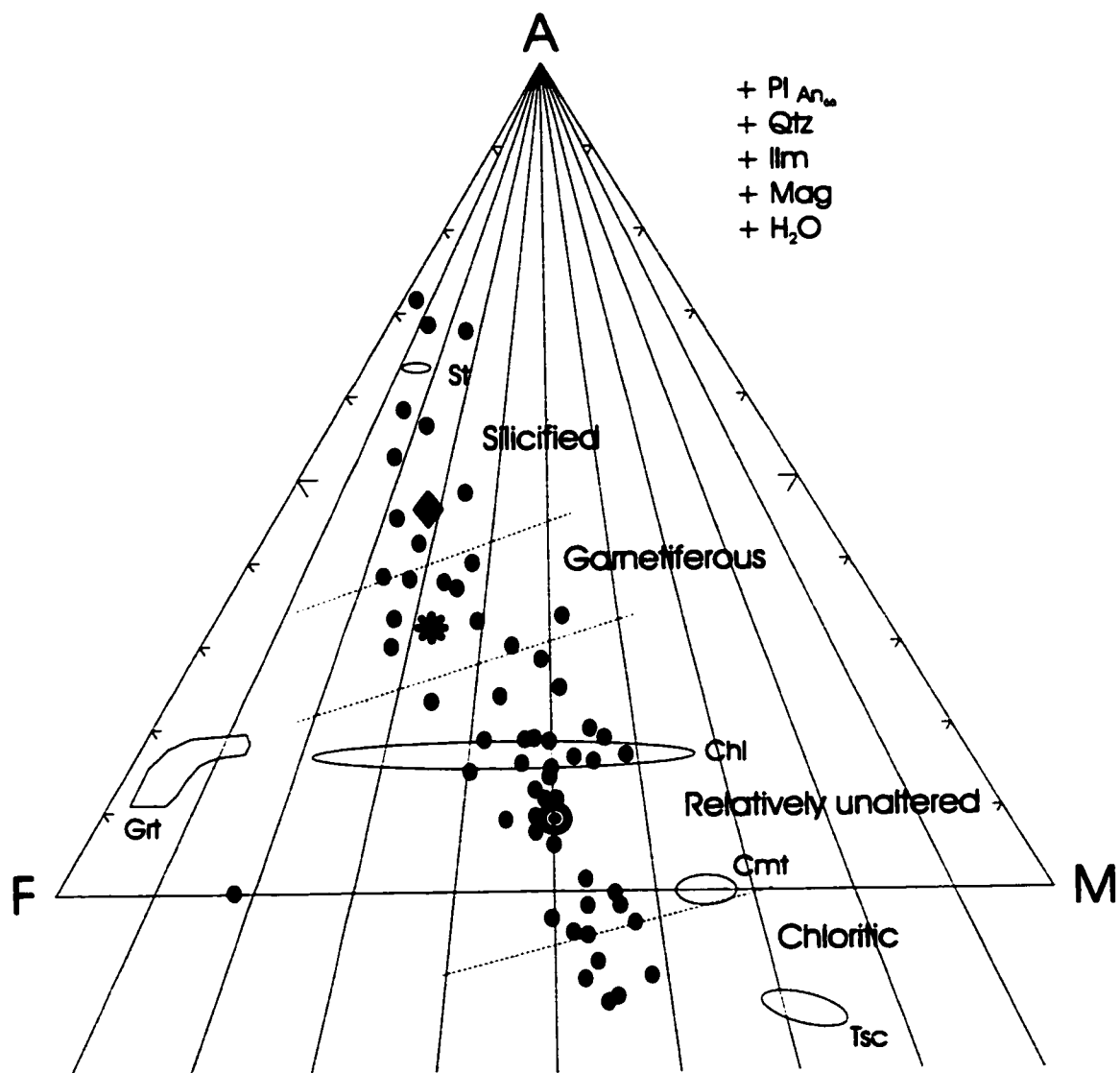
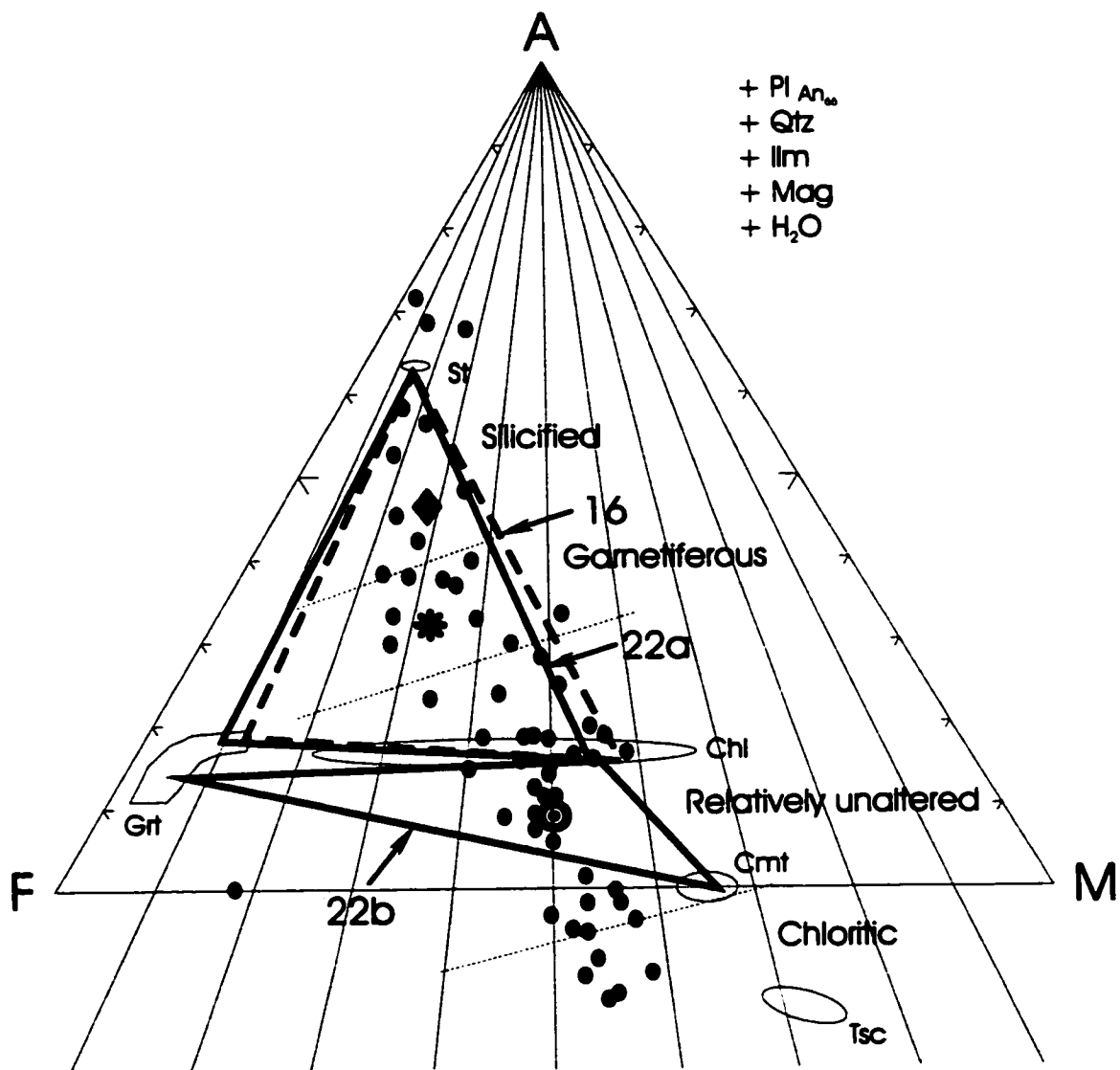


Figure 10.10

AFM diagram showing mineral assemblages for sample 22 (solid lines) and 16 (dashed lines). The triangles shown with solid lines represent garnet (core) + chlorite + cummingtonite (22b) and garnet (rim) + chlorite + staurolite (22a, Assemblage 8 in table 3.1) assemblages in sample 22. Tie lines between the assemblages do not cross. This may indicate that changes in XCO_2 during progressive alteration may not have significantly affected the phase relations, and the AFM is appropriate for the evaluation of changes in bulk composition during metamorphism. Heavy dashed line shows the assemblage garnet + chlorite + staurolite for sample 16.

The occurrence of early cummingtonite in garnet cores, and Fe-Mg rich garnet rims with staurolite, indicates cummingtonite became unstable during amphibolite facies metamorphism. This occurred as the bulk-rock composition progressed through chlorite, and suggests that hydrothermal alteration occurred during peak metamorphism.

Abbreviations; Chl=chlorite, Cmt=cummingtonite, Grt=garnet, Ilm=ilmenite, Mag=magnetite, Qtz=quartz, St=staurolite, Tsc=tschermakite.



11.0 Discussion

11.1 Regional deformation and metamorphism

The eastern Red Lake belt has undergone at least three deformation events after ≈ 2731 Ma (Table 2.0). In addition, Sanborn-Barrie et al. (2000) suggested an episode of tilting in the Lower sequence rocks resulting in an unconformity between the Upper and Lower sequences. This was not observed by the author in the study area. The oldest event reported here, D_1 , affected the entire eastern part of the belt and produced NE-trending F_1 folds and deformation zones, and a weak S_1 foliation, prior to the intrusion of the Dome and McKenzie stocks. D_1 , and possibly M_1 , may be temporally related to the emplacement of the Little Vermillion Lake batholith (2731 ± 3 Ma, Corfu and Andrews, 1987) to the NE of Cochenour, and F_1 folds are parallel to the batholith contact. Alternatively, D_1 may be related to regional NW shortening during the initial stages of the Kenorean orogeny, occurring at roughly the same time as emplacement of the late phase of the Little Vermillion Lake batholith.

M_2 is a minor (< 30 m wide) contact metamorphic event along the margins of the Dome and McKenzie stocks (Table 2.0). Metamorphism during this time period was more significant in the western part of belt, with extensive effects adjacent to the 2717 Ma Hammell Lake batholith at (Corfu and Andrews, 1987). Menard et al. (1999) suggested that metamorphism with M_2 timing was also related to the emplacement of the Trout Lake batholith. The age of a marginal phase of the Trout Lake batholith is 2699 ± 1 Ma (Noble, 1989; Corfu and Andrews, 1987). As no known intrusive ages for the Trout Lake batholith correspond to M_2 timing, this hypothesis is unlikely.

M_3 metamorphism is interpreted to be associated with the emplacement of marginal phases of the Trout Lake batholith at 2699 ± 1 Ma (Corfu and Andrews, 1987) (Table 2.0). Intrusion was accompanied by F_2 folding in the eastern part of the study area (Fig 2.3). In the supracrustal rocks adjacent to the batholith, F_2 fold axial traces and S_2 axial planar foliations are parallel to the margins of the batholith. Garnet and amphibole extension lineations on F_2 axial planar foliations form a cluster on stereonet, across the F_2 McDougal Lake Antiform (Fig 2.8). Field relations between the intrusion and foliated country rocks (Fig 2.9, 2.10), and the apparent increase in strain with metamorphic grade (Hugon and Schwerdtner, 1985), indicate that the intrusion was coeval with at least the latest D_2

deformation. These data indicate M_3 is contemporaneous with D_2 (Table 2.0, Fig. 2.5). F_1 and S_1 were folded by F_2 , and D_2 NW, NNW, and E-W shear zones were developed. NE trending D_1 shear zones were reactivated, decoupling NW trending folding from the western part of the belt.

The final deformation event, D_3 , post-dated the emplacement of the marginal phase of the Trout Lake batholith (Table 2.0). Sinistral strike-slip shear occurred along the NE trending Post Narrows and East Bay deformation zones during D_3 and cut the Dome and McKenzie stocks, as well as F_2 folds. D_3 is apparently unrelated to an intrusive or metamorphic event in the Red Lake belt.

11.2 Relationship of metamorphism and deformation at the Campbell\Goldcorp Red Lake deposit

Pelitic rocks are useful in defining P-T condition of metamorphism, but they are uncommon at the Campbell and Goldcorp mines. As a consequence, previous metamorphic studies have focused on mineral assemblages in aluminous altered mafic volcanic rocks (Mathieson and Hodgson, 1984; Christie, 1986; Penczak and Mason, 1997). The author has evaluated the P-T conditions for metamorphism from mineral assemblages of relatively unaltered basalts and interflow sedimentary rocks, and from aluminous altered rocks and ore-stage-altered rocks. The data are consistent with previous work showing that the eastern Red Lake belt has undergone low P and high T metamorphism, and the Campbell and Goldcorp mines lie approximately at the amphibolite-greenschist facies metamorphic transition.

The author has shown that metamorphism occurred during D_2 shortening, from the occurrence of synkinematic garnet (Fig 3.24) and andalusite (Fig 3.26) containing sigmoidal F_2 related foliation trails. Bands of garnet grains occur as aggregates along D_2 foliations. Garnets are boudinaged, with euhedral growth faces developed in boudin necks (Fig. 3.25). This supports garnet growth during D_2 shortening. Coarse-grained, zoned garnets are deformed, and locally fragmented after the development of almandine and pyrope-rich rims, suggesting deformation continued after the growth of garnet. The random orientations of cummingtonite inclusions in garnet (Fig. 3.2, 3.29), and fine-grained inclusions in cordierite, hosted by a coarse-grained foliated matrix (Fig. 3.30),

suggests metamorphism commenced prior to significant flattening in the rocks. The occurrence of radial aggregates of orthoamphibole within the foliation is consistent with their growth during deformation under predominantly flattening strain.

The evidence indicates that ductile to brittle-ductile deformation occurred synchronous with metamorphism. The steep temperature gradient in the area is consistent with the observation of a sharp increase in D_2 strain through the Campbell and Goldcorp mine properties, eastward.

11.3 Structural controls for auriferous veining

Mapping by the author at the Campbell mine shows that F_2 folds have moderately SW plunging axes (Fig 3.9). The penetrative foliations in the mine were formed during F_2 folding. Auriferous mineralization is hosted by foliation-parallel and foliation-oblique vein systems. Foliation-oblique auriferous veins are E-W and NNW trending, and occur along the F_2 fold limbs, at mafic-ultramafic rock contacts. They have been interpreted as splays and releasing bends controlled by lithologic contacts, related to NW trending shear zones hosting auriferous veins (Penczak and Mason, 1997; Penczak and Mason, 1999), and conjugate shear zones (Wilson et al., 1984; Zhang et al., 1997) developed under a pure shear regime (Zhang et al., 1997).

Penczak and Mason (1997) and Penczak and Mason (1999) suggested that NNW-trending auriferous veins and shear zones (L and G ore zones) are releasing bends related to a sinistral strike-slip fault and that the splay faults were controlled by the NNW trending contacts between basalt and ultramafic rocks. These foliation oblique shear zones cannot represent releasing bends along a sinistral fault. Instead, the geometry is compatible with a restraining bend (i.e., Woodcock and Fischer, 1986). Furthermore, data presented by the author, and by previous workers (MacGeehan and Hodgson, 1982; Wilson et al., 1984; Zhang et al., 1997) support that these auriferous veins are hosted by dextral simple shear zones, rather than sinistral oblique slip releasing bends. The author has shown that NW trending folds, which are spatially related to the foliation-oblique auriferous veins were developed during the D_2 , NE-SW shortening episode. However, Penczak and Mason (1997; 1999) interpreted folding to post-date auriferous veining. Therefore, releasing bends and fault splays hosting auriferous veins, controlled by F_2 fold limbs, are incompatible with their

interpreted time relations between folding and auriferous veining. The author's work support that the E-W trending (sinistral) and NNW trending (dextral) foliation-oblique veins represent a local conjugate shear array.

Folding is commonly accommodated by ductile flow of layers and beds, and by shear between layers. Folds developed by flexural slip display shear along rock layering, with opposite senses of motion on opposing limbs. These shear zones on opposite limbs will display a conjugate relationship. Dilation may occur in the hinge areas during flexural slip folding (Ramsay, 1967). Ramsay (1967) has shown that during progressive shortening of an asymmetric fold, the limb forming the larger angle to the axial plane will provide dilation.

The occurrence of quartz+carbonate veins differ between E-W and NNW shear zones, proximal to F_2 hinges. This suggests that veins in E-W and NNW do not represent pre- F_2 , folded carbonate vein system which would be expected to have a similar occurrence. However, it also suggests that the timing for vein in E-W and NNW shear zones may be different, or that they are synchronous but the mechanism for their formation may differ. This study has demonstrated that large auriferous veins at the Campbell mine occur along NNW to N-S trending F_2 fold limbs (Fig 3.17), and the angles between the axial plane (NW trending foliations) and the NNW limbs and shear zones are greater than those for E-W limbs and shear zones. The author suggests that foliation-oblique auriferous veins formed under shortening and flexural slip folding during the F_2 deformation, and that NNW shear zones experienced greater dilation than E-W limbs.

Shear zones hosting foliation-parallel auriferous veins in the eastern part of the Red Lake belt have been interpreted as strike-slip shear zones (MacGeehan and Hodgson, 1981; Wilson et al., 1984; Andrews et al., 1986; Penczak and Mason, 1997; Penczak and Mason, 1999), and pure shear zones by Zhang et al. (1997). The author's study shows that foliation-parallel shear zones are spatially related to highly attenuated F_2 fold limbs, and characterised by increasing strain towards the auriferous veins, and strong flattening with symmetrical folds. Symmetrical folds are not compatible with a simple shear model. The observations therefore support the interpretation proposed by Zhang et al. (1997), that foliation-parallel shear zones accommodated significant shortening. Boulter et al., (1987) have similarly noted that auriferous mineralization at the Golden Mile mine occurs partially in high strain

portions of shear zones, displaying predominantly pure shear deformation.

11.4 Relative timing of alteration and metamorphism

A variety of opinions have been proposed for the timing of gold mineralization at the Campbell/Goldcorp deposit, including syn-peak metamorphic (Andrews et al., 1986; Tarnocai et al., 1997; Zhang et al., 1997) and syn- to post-peak metamorphic (Wilson et al., 1984). MacGeehan and Hodgson (1982) suggested that the main phase of mineralization occurred synchronous with the regional metamorphism. In contrast, Penczak and Mason (1997) and Penczak and Mason (1999) interpreted the Au mineralization to pre-date the metamorphism, in part based on the assumption that pervasive aluminous alteration is associated with the Au mineralization. The assumption is clearly not warranted. Previous workers have noted that pervasive, pre-ore aluminous alteration does not bear a spatial relation to Au mineralization (Rigg, 1980; Mathieson and Hodgson, 1984; Christie, 1986). Therefore, the aluminous alteration event does not reflect the timing of Au mineralization.

Pre-ore aluminous alteration contains a low variance mineral assemblage in amphibolite facies domains. This prompted Damer (1997) and Penczak and Mason (1999) to suggest that the assemblages reflected metamorphism of an aluminous clay-rich precursor. However, pre-ore aluminous alteration does not show a spatial relation to auriferous mineralization, is pervasive, and is commonly controlled by features such as flow tops and pillow rims, suggesting a syn-volcanic timing (Mathieson and Hodgson, 1984; MacGeehan and Hodgson, 1982; Christie, 1986).

11.5 Temporal relation between gold mineralization, deformation, and metamorphism

High Au grade quartz + arsenopyrite breccias host the most significant Au mineralization in the Campbell mine, and occur in both foliation-parallel and oblique vein systems. Although auriferous veins show various trends (E-W, NW, and NNW), high Au grade breccias show consistent moderate to steep SW plunges (Fig 2.7, 3.11) in all vein systems. This orientation is parallel to the orientation of F_2 fold axes, and the intersection of E-W and NNW shear zones hosting auriferous veins (Fig 3.11). Along foliation-oblique veins, the breccias also show a strong spatial relation to F_2 fold hinges and the intersections

of auriferous E-W and NNW veins (Fig 3.15, 3.16). The plunge orientation is parallel to the intermediate stress direction, inferred from the intersection of conjugate shear zones, the orientation of F_2 fold axes, and moderate to steep-plunging boudin axes. This suggests auriferous brecciation may have occurred along foliation-parallel and oblique vein systems simultaneously. Auriferous mineralization overprints F_2 foliations, and F_2 foliations rotate progressively into the shear zones hosting auriferous, foliation-oblique veins. Therefore, auriferous mineralization occurred during the F_2 event.

Progressive shortening during F_2 resulted in attenuation and shearing of the western limb of the F_2 Balmer Synform (Fig 2.4). The largest and most productive gold deposits in the Red Lake belt (the Campbell and Goldcorp mines) are spatially related to the attenuated fold limb, and some auriferous veins occur along the attenuated limbs of parasitic folds (Fig 2.4, 3.17). At the Marcus property, gold mineralization occurs along F_2 foliation-parallel shear zones, suggesting a syn- to post- D_2 timing for Au mineralization. At the Cochenour mine, the auriferous shear zone hosting the West Carbonate zone cuts F_1 folds, and S_2 foliations undergo progressive rotation into the W dipping, NNW trending West Carbonate shear zone (Sanborn, 1987). This suggests that motion along this shear zone occurred during NE-SW flattening and F_2 folding. These observations suggest that auriferous mineralization occurred in other areas of the eastern Red Lake belt, coeval with gold deposition at the Campbell and Goldcorp mines.

The spatial variations in the occurrence of gahnite, and textural relations between gahnite and arsenopyrite, indicate that auriferous mineralization occurred during the peak of metamorphism (Tarnocai et al., 1997). This is further supported by the spatial occurrence of Au-bearing minerals. The occurrence of aurostibite is restricted to the western, greenschist facies hosted ore bodies. This is consistent with the stability of aurostibite within greenschist facies temperature conditions. Aurostibite decomposes to stibnite and native gold at temperatures greater than 460 °C (Hansen, 1958; Nekrasov, 1996). If aurostibite deposition occurred significantly after the peak metamorphism, it may be expected to occur in ore throughout the mine. Aurostibite is absent from ore hosted by wallrocks with metamorphic temperature greater 475 °C to 500 °C. In contrast, wallrocks hosting aurostibite-bearing ore have been metamorphosed to ≤ 460 °C to 480

°C, compatible with the thermal stability of aurostibite. This indicates aurostibite deposition occurred either prior to, or during peak metamorphism.

It could be argued that the spatial variation of Au mineralogy suggests metamorphism of pre-existing gold deposits. This is unlikely because native gold from the amphibolite facies domain is relatively Ag rich compared to greenschist facies domains. If native gold is partly a product of thermal decomposition of aurostibite, native gold should not contain Ag. This suggests that aurostibite was deposited during peak metamorphism.

It was shown that randomly oriented cummingtonite grains occur in the matrix of relatively fresh basalt. In foliated, hydrothermally altered basalt, cummingtonite occurs only as inclusions in garnet. The occurrence of cummingtonite is interpreted here to suggest that cummingtonite within garnet was isolated from the reacting matrix during alteration. The occurrence of cummingtonite and almandine garnet in basalt reflects amphibolite facies metamorphic conditions. These observations suggest that the bulk composition of basalt changed under amphibolite grade metamorphic conditions, and loss of cummingtonite from basalt is compatible with increasingly aluminous bulk compositions towards auriferous veins. This moderate, syn-ore aluminous alteration is spatially related to auriferous veins and distinct from the pervasive pre-ore aluminous alteration and bleaching event.

The proposed interpretation that auriferous hydrothermal activity occurred during peak metamorphism is also supported by delicate compositional zoning, including Au zoning, in arsenopyrite and pyrite, and compositional zoning in carbonate minerals (Tarnocai et al., 1997). These textures would have been obliterated if the ores had undergone amphibolite grade metamorphism (Tarnocai et al., 1997). Hydrothermal activity continued after the peak metamorphism based on textural evidence including the chlorite replacement of garnet (Fig 10.6) and biotite (Fig. 10.2), and carbonate cutting hornblende and plagioclase, arsenopyrite veinlets in fractures of garnet, and chlorite replacing amphibole in post-ore lamprophyre dykes.

Fluid inclusion studies indicate that the density of CO₂ is between 0.8 and 0.88 g/cc. The trapping pressure of inclusions using the temperature estimate from wallrock assemblages agrees well with pressure estimates using GASP and metamorphic mineral

assemblages (Fig. 11.0). This indicates either Type I inclusions are related to auriferous fluids during syn-peak metamorphism, or are metamorphic fluids which post-date Au mineralization. Metamorphic fluids at upper greenschist / lower amphibolite facies are typically CO₂ bearing, but H₂O-rich (i.e., Hollister and Burruss, 1976). Type I inclusions are non-aqueous. Therefore, Type I inclusions are unlikely to represent a metamorphic fluid which is generally regarded to have a significant aqueous component. The interpretation that Type I inclusions are a product of phase separation during gold deposition, under amphibolite facies metamorphic conditions, is favored here.

11.6 Mechanism for gold deposition

High grade Au mineralization at the Campbell mine occurs in hydrothermal breccia bodies that replace quartz + carbonate veins. The close spatial association of breccias with high grade Au suggests that pressure fluctuations related to brecciation may have played an important role in Au deposition. Rapid fluid pressure changes may arise from fault movement accompanying seismic activity (Sibson et al., 1988). Pressure decrease would have caused fluid immiscibility, resulting in a sharp drop of fluid temperature, increase in fO₂ and pH. All of these are conducive for Au deposition (Naden and Shepard, 1989; Bowers, 1991). Fluid immiscibility has been proposed as the principal mechanism for Au deposition in Archean Au vein systems (Guha et al., 1991; Smith and Kesler, 1985; Robert and Kelly, 1987) and low-sulphidation epithermal vein systems (Hedenquist et al., 1996).

Type I CO₂ + CH₄ inclusions are common in auriferous quartz + arsenopyrite hydrothermal breccias. Type II aqueous inclusions are observed with Type I inclusions in weakly auriferous veinlets. The co-occurrence of these inclusion populations suggests that Type I and II inclusions coexisted. As auriferous fluids are typically interpreted to have a significant aqueous component (i.e. Colvine et al., 1988), the non aqueous fluids of Type I inclusions are considered unlikely to represent the auriferous fluid. The compositions of Type II inclusions are similar to auriferous fluids from other Archean Au deposits (Colvine et al. 1988) and CO₂ –bearing aqueous fluids are generally considered responsible for Au transport. Therefore, Type II inclusions are considered to represent the auriferous fluid. Type I inclusions likely represent the CO₂-rich phase after the

immiscibility of auriferous fluids. The proposed interpretation requires co-existing H₂O-rich fluids, but aqueous inclusions were not observed together with CO₂-rich Type I inclusions in high Au grade breccias. The absence of Type II inclusions suggests the preferential trapping of H₂O following phase separation, owing to the higher mobility of H₂O over CO₂ (Santosh et al., 1991).

The $\delta^{13}\text{C}$ values for all stages of carbonate from the eastern Red Lake belt are significantly higher than those from other Archean Au deposits. Stage 2 (syn-gold) carbonate is enriched in ^{13}C (Fig 7.2), relative to Stage 1 and 3 carbonate. Stage 2 carbonate deposited from fluids during and after fluid immiscibility would have higher $\delta^{13}\text{C}$ values because of preferential loss of CH₄ from aqueous fluids. The data suggest that hydrothermal fluids may have lost significant CH₄ during Au deposition, and deposition of Stage 2 carbonate.

11.7 Origin of quartz + carbonate veining

Quartz + carbonate veining preceded Au deposition at the Campbell and Goldcorp mines, and other deposits in the eastern Red Lake belt. The author has shown that quartz + carbonate veining occurred during early D₂ in the eastern part of the greenstone belt. It was synchronous with the emplacement of a marginal phase of the Trout Lake batholith.

Banded to crustiform quartz + carbonate veins and wallrock breccias are common at the Campbell mine, but are not common in other Archean Au deposits. Penczak and Mason (1997) and Penczak and Mason (1999) suggested that the carbonate vein textures are compatible with near surface hydrothermal activity, possibly related to a low sulphidation (LS) epithermal system. Other evidence used to support epithermal mineralization by Penczak and Mason (1997) and Penczak and Mason (1999) includes the occurrence of "stockwork" and sheeted veins, implosive breccias and open space filling vein textures, and enrichment of Ag+As+Sb+Hg+Zn with Au.

Implosive brecciation and open-space filling textures can form at depths equivalent to that inferred for peak metamorphism at the Campbell mine (Tarnocai et al., 1998). Sibson (1986) notes that fault rupture and implosion breccias may form throughout the seismogenic regime (to a depth of 10 to 15 km). Thus, such breccias are by no means unique to epithermal environments. For example, Au-bearing veins in the

Bendigo-Ballarat area in Australia display crustiform and vuggy textures, and are considered to have formed by a fault-valve mechanism at 6 to 10 km depth (Cox, 1995). In addition, dilational breccias and stockwork veins with "sheeted veinlet" zones are common in many other lode Au deposits, such as the Sigma mine in the Val d'Or camp, Québec (Robert and Brown, 1986), the San Antonio mine in Manitoba (Poulsen, 1989; Robert et al., 1994), Bendigo-Ballarat area (Cox et al., 1991), Mount Charlotte, Kalgoorlie (Ridley and Mengler, 2000) and the Victory mine in Western Australia (Clark et al., 1986). Therefore, the vein textures at the Campbell-Red Lake deposit may simply indicate high fluid pressures relative to confining pressures, allowing open spaces to be maintained, and also rapid fluctuation in fluid pressure. The occurrence of carbonate vein-hosted wallrock breccias and crustiform banding are not unequivocal criteria for near surface veining. Tarnocai et al. (1998) indicate other inconsistencies between the deformation fabrics, and the interpreted timing and origin of veining, as suggested by Penczak and Mason (1997).

Penczak and Mason (1997) used the enrichment of Ag, As, Sb, Hg and Zn as evidence for epithermal mineralization, but the elemental assemblage of Au+Ag+As+Sb+Hg+Zn, and K metasomatism are not unique to low sulphidation epithermal systems. These metals share similar geochemical properties, leading to their cohesive behavior in nature. This explains high concentrations of these elements in many gold deposits (e.g., Boyle, 1979). Mercury is not commonly analyzed, but the concentration is always high when it is reported from Archean gold deposits. Examples include the Con and Giant mines, Yellowknife, Dome and Kerr Addison mines, Ontario (Kerrich, 1983; Fyfe and Kerrich, 1984), Ross mine, Ontario (Boyle, 1979), Cameron Lake deposit, Ontario (Melling et al., 1986), Hemlo deposit, Ontario (Kuhns et al., 1986) and Gwillim mine, Québec (Bouchard et al., 1984). The occurrence of Hg minerals in close association with gold are reported from mineralogically well studied Archean camps, including Kalgoorlie in Western Australia (Boyle, 1979), Kirkland Lake in Ontario (Hawley, 1950; Hicks, 1990; Hattori, 1993) and Hemlo (Harris, 1989). The enrichment of As, Sb, and Ag is also widely known in many Archean gold deposits (e.g., Fyfe and Kerrich, 1983; Pattison et al., 1986; Wyman et al., 1986). Some gold veins in the Bendigo-Ballarat area, Australia, are so high in Sb and As that they are called Sb-Au

and As-Au veins (e.g., Changkakoti et al., 1990; Phillips and Hughes, 1996). As mentioned above, it is well established that these Australian veins formed at a mid-crustal level (Cox, 1995). Thirty-one mineralized samples from the Campbell mine (Appendix 7) report an average of 15.28 ppm Au, 1.82 ppm Ag, 771 ppm Zn, and 3299 ppb As. Although Ag, As, and Zn are enriched relative to some Archean lode-type deposits (e.g. Kerrich and Hodder, 1982), the Au:Ag ratio is 8.4:1, typical of many deposits of this class (Groves and Foster, 1991), and much higher than low sulphidation epithermal systems, which often have ratios of 1:1 or lower. Higher Zn, and much higher As values are reported from other mafic-rock hosted deposits of this class (Dome mine, Kerrich and Hodder, 1982; Con and Giant mines, Kerrich and Hodder, 1982; van Hees et al., 1999). The enrichment of these elements at the Campbell and Red Lake mines, therefore, does not support or refute the depth nor the timing of mineralization. Potassium metasomatism is an exceptionally common feature of Archean auriferous veins systems (i.e., Colvine et al., 1988), and implies neither a genetic origin or depth for auriferous mineralization.

Quartz + carbonate veins at the Campbell mine lack electrum and adularia, which are common in LS veins (Hedenquist et al., 1996). Bladed tabular carbonate, opaline quartz, and sulphates are not associated with auriferous mineralization at Campbell, but are common features of LS type deposits (i.e. Hedenquist et al., 1996). Both Type I and Type II fluid inclusions are temporally related to gold and contain high density CO₂, incompatible with a LS-type origin (Bodnar et al., 1985).

11.8 Alteration assemblages at the Campbell mine and other gold deposits in amphibolite facies terranes

Several stages of hydrothermal alteration are recognized at the Campbell mine. The origin of pervasive, pre-ore aluminous alteration has been discussed, and bears neither spatial nor temporal relation to the gold mineralizing event. Alteration related to auriferous veins forms relatively narrow envelopes adjacent to the veins. Alteration assemblages vary with metamorphic grade of the host rocks. The assemblages in greenschist facies domains contains white mica, whereas those in amphibolite facies domains contain garnet + amphibole and biotite + amphibole. Biotite + amphibole

wallrock alteration shows zonations with progressive replacement of amphibole by biotite, forming biotite-rich domains closer to the veins. Although amphibole alteration occurs only in the western and deeper parts of the Campbell mine, dark green amphibole-rich alteration selvages are common in the Goldcorp mine (Rodgers, 1992; D. Leroux, pers. comm., 1993).

The alteration assemblage at the Campbell mine is comparable with other gold deposits hosted by amphibolite facies metamorphic rocks. Examples include Missouri and Sand King in Western Australia (Mernagh and Witt, 1992), and deposits in the Norsman District (Mueller, 1992; Knight et al., 2000). Witt (1991) described a variation of alteration minerals in Au deposits hosted by mafic volcanic rocks, from muscovite in the greenschist facies to biotite in amphibolite facies domains in the Eastern Goldfields province. The deposits are considered by Witt (1991) to have formed during metamorphism. Deposits in the Norsman district and at Missouri and Sand King, are interpreted to have formed during peak lower to middle amphibolite facies metamorphism (Mueller, 1992; Mernagh and Witt, 1992; Knight et al., 2000). They show alteration of amphibole + biotite + carbonate + plagioclase in mafic rocks adjacent to auriferous veins. Metasomatic amphibole and biotite, and the zonation from biotite to amphibole wallrock alteration away from the auriferous veins, are reported from deposits hosted by lower amphibolite facies rocks (Mernagh and Witt, 1992; McCuaig et al., 1993; Mikucki and Ridley, 1993; Neumayr et al., 1995). Garnet occurs in the wallrocks and plagioclase occurs in auriferous veins (Bloem et al., 1994; Neumayr et al., 1995). The wallrock alteration at the Campbell mine is very similar to other syn-metamorphic Archean gold systems, formed at 450 °C to 550 °C.

11.9 Absolute timing for auriferous mineralization

The author proposes that the Au mineralization took place at around 2700 Ma based on the synchronous nature of Au mineralization with F₂ folding and the emplacement of the marginal phase of the Trout Lake batholith (2699±1 Ma, Corfu and Andrews, 1987). The age is significantly younger than the age of Au mineralization (2720 to 2714 Ma) proposed by Corfu and Andrews (1987). They interpreted that the Au mineralization is younger than the intrusion of the McKenzie stock (2720 Ma), but older

than the age of quartz+feldspar porphyry dykes of 2714 ± 4 Ma at the Goldcorp mine (Andrews et al., 1986; Corfu and Andrews, 1987) because some dykes cut auriferous mineralization.

Corfu and Andrews (1987) recognized that this is not consistent with the observations that metamorphism related to the marginal granite of the Trout Lake batholith was accompanied by deformation and auriferous mineralization. As a compromise, they speculated on the presence of an earlier intrusive phase (≈ 2720 to 2710 Ma) and postulated that the 2699 ± 1 Ma age reflects a later intrusive phase of the Trout Lake batholith (Corfu and Andrews, 1987). The marginal phase of the Trout Lake batholith is a K-feldspar megacrystic granodiorite and these rocks intruded during D_2 shortening (Fig 2.9), suggesting the 2699 ± 1 Ma age accurately reflects the timing for M_3 and F_2 .

The author suggests that the dyke used for age determination in the Goldcorp mine is older than the Au mineralization. Corfu And Andrews (1987) recognized strong deformation and intense alteration in the dyke sample that was provided by the staff of the A.W. White mine (currently the Goldcorp Red Lake mine) (Andrews, pers. Comm., 1994).

A NW trending quartz + feldspar porphyry dyke exposed at surface at the Campbell mine (Fig 3.4) is presumably related to similar NW trending dykes at the adjacent Goldcorp mine. The dyke contains a well-developed foliation that is parallel to the foliation in the adjacent mafic rocks. The dyke is veined and altered (sericitized, carbonatized), and offset across the auriferous NNW shear zone which hosts the G ore body. Apparent strain, inferred from the intensity of foliation development, increases in the dyke as it thins towards this NNW shear zone. This indicates that the dyke had undergone hydrothermal alteration and deformation by the shear zones hosting auriferous veins (Fig 3.4), and is clearly not post-tectonic, as reported by Sanborn-Barrie et al. (2000). Dextral motion along NNW trending shear zones hosting auriferous veins has a syn- to late F_2 timing. The evidence suggests that the dyke predated auriferous mineralization during F_2 . Therefore, a proposed age for auriferous mineralization between ≈ 2720 and 2714 Ma is inconsistent with the geologic relationships, and the dyke must pre-date, or be coeval with, auriferous mineralization related to foliation oblique

shear zones and veining. Therefore the author suggests that the 2714 Ma age of quartz+feldspar porphyry provides a maximum age for Au mineralization in the area.

Corfu and Andrews (1987) reported a reset U-Pb zircon age of 2701 ± 1.5 Ma in the >2745 Ma Wilmar granodiorite at the Wilmar mine (Fig. 2.1). This intrusion has been intensely altered and the reset age likely represents the timing of auriferous hydrothermal alteration. The age is the same as that of the marginal phase of the Trout Lake batholith. This suggests that an auriferous hydrothermal event took place in the eastern part of the belt at ≈ 2700 Ma.

11.10 Thermal regime and source of heat for auriferous hydrothermal activity

Garnetiferous basalt is extended laterally along the foliation-parallel auriferous veins, and abundant garnet occurs in rocks adjacent to some foliation-oblique auriferous veins. Garnet formation is favored in rocks with high Ca, Mn, or Al, but Mn, Ca and Al contents are similar between garnetiferous and garnet absent rocks (Table 3.0). Therefore, the occurrence of garnet at the mine is not primarily a function of bulk rock composition variations. For the mine area, the author has argued that the steep thermal gradient in the mine area would minimize the local effects of variations in bulk composition in controlling the occurrence of garnet. In addition, the garnet-in line crosses lithologic units (Figs. 3.36, 3.37). The author concluded that the garnet-in isograd indicates amphibolite facies.

The garnet isograd dips steeply westward, and is extended along the foliation-parallel auriferous veins (Fig. 3.36, 3.37). The attenuation of the isograd may be due to; 1) folding of a pre-existing isograd surface, 2) intrusive rocks emplaced along the deformation zone, or, 3) preferential heat transfer along the deformation zone.

Folding of the isograd surface: Although deformation continued after garnet growth, significant folding after garnet formation is rejected because the deformation is shown to have been synchronous with metamorphism, and garnet extension lineations are not folded across syn-metamorphic F_2 fold axial traces. Secondly, the isograd crosscuts F_2 fold axes (Fig. 3.32).

Intrusive rocks along the deformation zone: Minor felsic stocks, sills, and dykes are present along the Cochenour-Gullrock Lake deformation zone, and may have supplied

heat to the surrounding rocks. The abundance of intrusive rocks, including lamprophyres and diorite, appear to be similar in both greenschist and amphibolite facies domains. It is therefore unlikely that minor intrusions are responsible for the attenuated isograd. The presence of intrusive bodies below the deepest levels of drilling at the mine cannot be discounted, but a steep west-dipping to sub-vertical garnet isograd at the mine is not easily explained by heating from below. Instead, a heat source to the east and northeast explains the distribution of metamorphic facies.

Heat transport along the deformation zone: The distribution of metamorphic facies in the study area (Fig. 3.32) suggests that emplacement of the a marginal phase of the Trout Lake batholith was related to metamorphism. The conductivity of deformed and relatively undeformed rocks is presumed to be roughly comparable, and heat transferred by conduction would not account for the attenuated isograd. The extension of the garnet isograd along the limb faults (Fig. 3.32), and auriferous veins at the mine (Figs 3.33 and 3.34) therefore requires significant heat transport by fluids.

Narrow zones of local high strain (Zhang et al., 1997) with significant strike and dip extents host the foliation-parallel auriferous veins. These high strain zones are spatially related to attenuated F_2 fold limbs. These shear zones host the extensive foliation-parallel vein systems at the Campbell and Goldcorp mines. Hydrothermal flow was laterally directed along the shear zones, inferred from the shape of the garnet isograd. Furthermore, the hydrothermal flow is temporally related to F_2 folding and emplacement of a marginal phase of the Trout Lake batholith. This suggests that a phase of the Trout Lake batholith was the source of heat for hydrothermal flow.

The initial $^{87}\text{Sr}/^{86}\text{Sr}$ values of scheelite from auriferous veins at the Campbell mine are significantly lower than those predicted for the host volcanic rocks at ~ 2.7 Ga, suggesting that the hydrothermal fluids were not derived from, or did not interact significantly with, the old host rocks. Similarly, $^{87}\text{Sr}/^{86}\text{Sr}$ of scheelite associated with auriferous mineralization, and carbonate $\delta^{13}\text{C}$ values do not support significant fluid interaction with the sedimentary rocks, or derivation from anatectic granites in the eastern part of the belt. The scheelite $^{87}\text{Sr}/^{86}\text{Sr}$ data supports hydrothermal fluid derivation from, or interaction with, contemporaneous non-anatectic igneous rocks. This supports the

proposal that heat driving hydrothermal flow was related to emplacement of igneous intrusions. •

11.11 The Campbell/Goldcorp deposit, and application to genetic models for Archean lode-type deposits.

Genetic models for Archean lode-type Au systems either suggest derivation of the constituents during regional metamorphism or from contemporaneous magmas. The metamorphic model is further divided into two models involving the derivation of fluids and metals during amphibolite facies prograde metamorphism (e.g. Kerrich and Fryer, 1979), and during granulitization of lower crustal rocks (e.g. Colvine et al., 1988). The magmatic model suggests that auriferous hydrothermal fluids originated from spatially associated felsic plutons (Burrows and Spooner, 1985, 1988), or from lamprophyres (Rock and Groves, 1988).

Metamorphic models were proposed because significant amounts of fluids are liberated during prograde metamorphism at the greenschist/amphibolite facies transition and because many well-studied Au deposits occur in greenschist facies volcanic rocks. However, this model fails to explain the occurrence of numerous deposits in high grade metamorphic rocks.

Jolly (1978) recognized that late metamorphism of greenstone belts is of contact type, associated with the emplacement of large batholiths, and overprints earlier regional burial metamorphic assemblages. It is reasonable to assume that the magmas for the intrusions may have either driven large hydrothermal convection cells or influenced the flow of deep hydrothermal fluids.

Many Archean greenstone belts show a spatial and temporal association between Au deposits and late to post-tectonic granitoids. Examples include Au deposits in the Harare-Shamva greenstone belt in Zimbabwe (Vinyu et al., 1996), Red Lake belt (Andrews et al., 1986) Coolgardie Au field in the Yilgarn block, Western Australia (Knight et al., 1993; Knight et al., 2000), Pilbara craton, Western Australia (Neumayr et al., 1995), and Murchison Province, Western Australia (Wang et al., 1993). These syn-peak metamorphic Au deposits are often located within the thermal aureoles of batholiths at the margins of the greenstone belt (Witt, 1991; Nuemayr et al., 1998) and the

intrusions have undergone hydrothermal alteration (Mueller et al., 1991; Wang et al., 1993). It is, however, unclear if the magmas for the intrusions supplied the auriferous fluids or the ore constituents. Some intrusions may have been associated with granulitization of the lower crust at the final stages of cratonization. However, Archean lode-type deposits in other camps do not share these spatial and temporal relations with intrusions, particularly those occurring in greenschist facies metamorphic terranes.

Direct dating of hydrothermal minerals by U-Pb (rutile, zircon) and Sm-Nd (scheelite) appears to support the interpretation that, in many cases, gold mineralization post-dates local intrusions. In the Val d'Or camp (Jemielita et al, 1990, Camflo mine; Wong et al., 1991, Sigma mine; Anglin, 1992, Sigma, Pascalis, and Siscoe mines) of the Abitibi Subprovince, Au mineralization is generally regarded to have occurred at approximately 2599 to 2611 Ma, 50 to 80 Ma after significant metamorphism and intrusion (Jemielita et al, 1990; Wong et al., 1991; Hanes et al., 1992). Kerrich and King (1993) dispute these findings, providing data which suggests that the young age reflects secondary remobilization, with auriferous veins developed during late magmatism and metamorphism in the area. However, auriferous mineralization significantly post-dating metamorphism and magmatism is also reported for the Yilgarn (Kent et al., 1996, Westonia, Nevoria, and Scotia deposits) and the Zimbabwe craton (Darbyshire et al., 1996). It therefore seems reasonable that many lode-type deposits were developed significantly after, and therefore unconnected with, magmatism and metamorphism in the upper crust in their local areas. For these deposits, mineralization cannot be a product of metamorphism and magmatism expressed in the belt in these areas, and auriferous hydrothermal activity must have an origin external to the belt.

Vervoort et al. (1993) have documented late hydrothermal activity in the Noranda (2638 ± 14 Ma) and Matagami ($2584 \pm 19/-20$ Ma, and 2647 ± 14 Ma) areas from a whole rock Pb isotopic investigation. These dates are significantly younger than volcanism, metamorphism, and magmatism, and broadly compatible with the ages for mineralization at Sigma, Pascalis, Siscoe, and Camflo. These data suggest widespread "late" hydrothermal activity affecting the entire Abitibi belt. Previous workers (Corfu, 1987; Krogh, 1993) suggested that these "young" dates represent auriferous hydrothermal activity reflecting magmatism and metamorphism in the lower crust. The late

hydrothermal event, recorded in the Kapuskasing Zone, occurred between 2670 to 2630 Ma, and possibly extended to 2585 Ma, as a product of mantle underplating and downward accretion (Corfu, 1987; Krogh, 1993). The occurrence of late Au deposits in the hanging wall of regional fault structures (i.e., Destor-Porcupine fault, Cadillac-Larder Lake Break) interpreted as deep structures marking terrane boundaries, suggests hydrothermal fluids ascended to the upper crust along the major structures. These faults have long and complex histories, and it is reasonable to assume that they have also had similarly complex hydrothermal histories, acting as regional conduits for hydrothermal flow during and after Archean accretion. Many of these “young” Archean Au deposits are clearly related to a compressional event (i.e. Sigma mine, Robert and Brown, 1986). In the Abitibi, this may be related to renewed tectonism external to the belt, possibly linked to magmatism and metamorphism in the adjacent Pontiac Subprovince.

Anglin (1992) reported two episodes of Au mineralization in the Timmins camp based on Sm-Nd dating of scheelite. At the Dome and Preston East Dome mines, Au mineralization is dated at 2688 ± 172 Ma, and 2425 ± 68 Ma at McIntyre-Hollinger-Coniarium. Therefore, mineralization in the Dome mine area *may* be correlative with magmatism and metamorphism in the Timmins area. The young date for the Hollinger system is contentious, as Au deposition is regarded to have occurred at ≈ 2673 Ma, immediately post-dating albitite dyke emplacement. The young date may be temporally related to the Matachewan dyke event at 2454 ± 2 Ma (Krogh et al., 1984; Hearn 1988). Two episodes of Au mineralization are also recognized in the Val d’Or area. The Kiena mine (Morasse et al., 1995), and Norlartic mine (Couture et al., 1994) may have developed prior to approximately 2690 Ma, roughly synchronous with late magmatism and metamorphism in the region. The data suggest multiple stages of Au mineralization, each of which probably represents a greenstone belt-scale orogenic event and reactivation of deep crustal structures.

These data suggest that Archean lode-type Au mineralization may have occurred during either syn-metamorphic/magmatic, or post metamorphic tectonic events. The author suggests the Campbell and adjacent Goldcorp mines belong to a class of syn- to late-magmatic/metamorphic Au deposits that include others described in the Pilbara craton (Mt. York area; Main Hill, Breccia Hill deposits, Neumayr et al., 1993) and

Yilgarn craton (Southern Cross belt; Corinthia, Frasers, Barnicoat et al., 1991; Bloem et al., 1994; Nevoria, Barnicoat et al, 1991; Coolgardie area; Knight et al., 1993; Knight et al, 2000; Norsman – Wiluna belt; Regent and North Royal mines, Mueller, 1991). This is based on the similarity between alteration assemblages, and timing of Au deposition with respect to magmatism and metamorphism. Witt (1991) has suggested that heat associated with the emplacement of large batholith complexes drove auriferous hydrothermal flow.

The author has shown that auriferous hydrothermal activity occurred during peak metamorphism related to the intrusion of a late granitic phase of the Trout Lake batholith, at the eastern margin of the belt. Deformation was concurrent with contact metamorphism. The high temperature gradient in the mine area suggests hydrothermal fluid flow along the auriferous shear zones during metamorphism. Andrews et al. (1986) suggested that hydrothermal flow during magmatism resulted in rapid cooling of the Trout Lake batholith, from the close magmatic and metamorphic ages. Observations by Corfu and Stone (1998a) appear to support rapid cooling of the Red Lake belt after peak metamorphism. Further, Sr isotopic compositions of scheelite are compatible with the auriferous fluids having originated from a contemporaneous intrusion, and rule out evolved meteoric water, or a fluid that interacted with, or was derived from, the supracrustal sequence or anatectic melts. Because metals typical of porphyry systems have not been documented in the area, Au was unlikely to have been derived from the parental magmas for the intrusion. Rather, the author favors that emplacement of voluminous intrusives at the margin of the Trout Lake batholith drove hydrothermal activity in the eastern part of the belt.

11.12 Comparison of pressure and temperature conditions for the mineralization of Archean gold deposits

Pressure and temperature conditions of mineralization in Archean Au deposits are compared to those of the Campbell and Goldcorp Red Lake deposits (Fig. 11.1; Appendix 8). The diagram includes deposits interpreted as syn-magmatic/metamorphic (i.e., Marvel Loch), and those post-dating these events (i.e., Sigma mine). The P-T conditions for auriferous hydrothermal activity for Archean Au deposits fall along a low P/ high T thermal path. The path is very different from the Archean cratonic geotherm,

which is generally considered to have been similar to the modern cratonic geotherm (Ballard and Pollack, 1988; Percival, 1994; de Wit, 1998). The data sets rule out regional prograde burial metamorphism of the greenstone sequence as a source of the auriferous hydrothermal fluid. The path may suggest either a link between magmas and auriferous hydrothermal activity, or rapid heat transfer by hydrothermal fluids from deep levels. This also implies rapid channelized fluid flows to the upper crust are important for Au mineralization.

Figure 11.0

CO₂ isochore diagram, modified after Roedder (1984). The shaded area shows the pressure - temperature estimate for the auriferous fluids at the Campbell mine, based on the densities of primary Type I CO₂ inclusions from a sample below the 27th level of the mine. Inclusions have densities of 0.80 to 0.88 g/cm³, and contain < 3 mole% CH₄. Vertical and horizontal bars indicate the pressure and temperature estimates based on metamorphic mineral assemblages in the wallrocks for the area from which the sample was obtained. Excellent agreement is obtained between conditions of metamorphism, and pressure and temperature of trapping for the auriferous fluid.

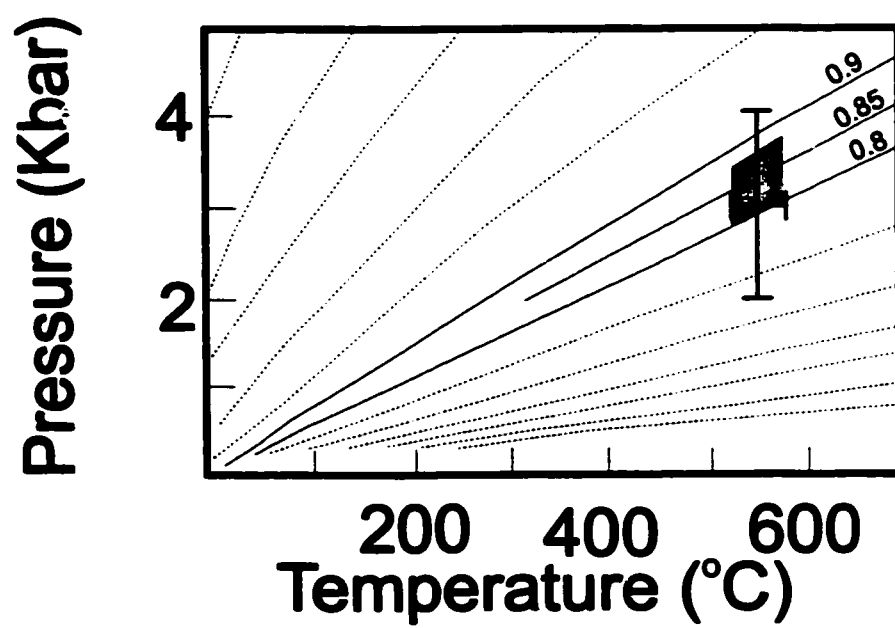
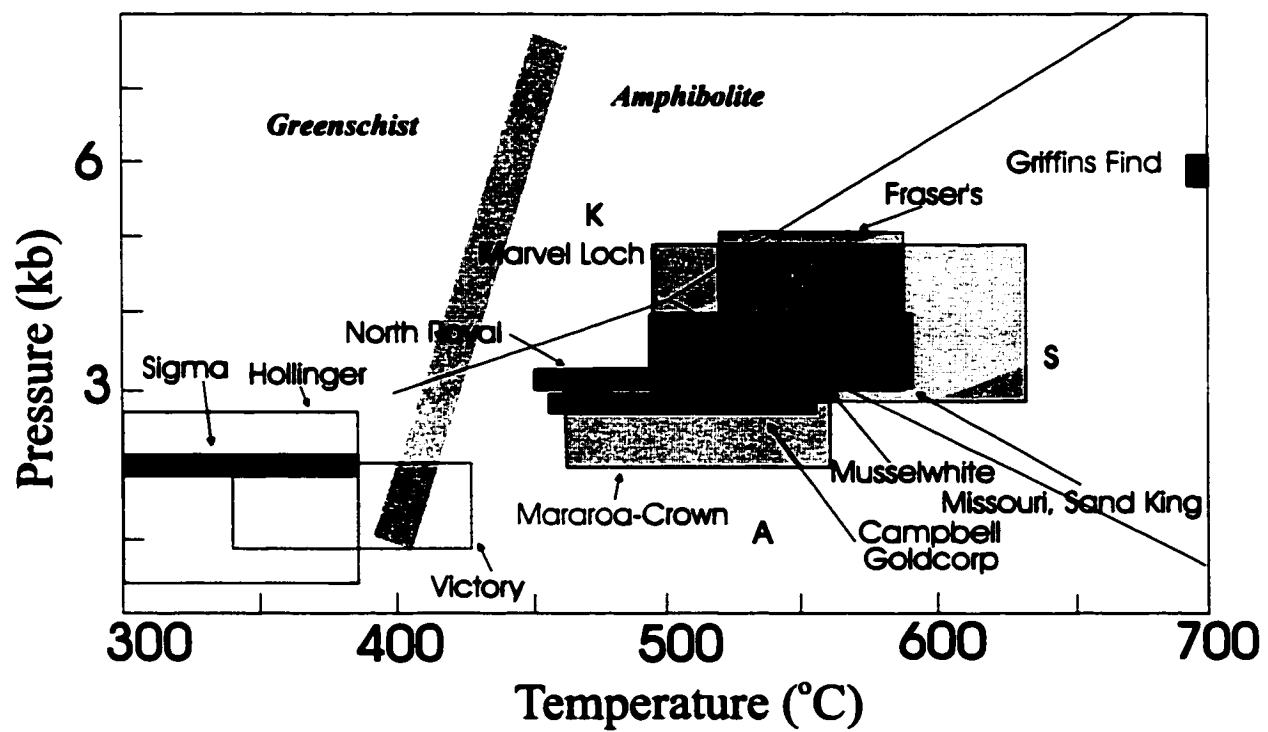


Figure 11.1

Pressure-temperature conditions for gold deposition from various Archean lode-type deposits. The green bar represents the approximate transition from greenschist to amphibolite facies. The data show a low pressure / high temperature path. This suggests either significant transport of heat to the upper crust by the hydrothermal fluid, or the influence of intrusions supplying heat during hydrothermal flow.

Data sources: Sigma mine (Robert and Kelly, 1987), Hollinger mine (Smith et al., 1984), Musselwhite mine (Hall and Rigg, 1986), North Royal mine (Mueller, 1992), Marvel Loch, Frasers and Nevorla deposits (Groves et al., 1990), Victoria mine (Clark et al., 1986).



12.0 Conclusions

- The Campbell mine and eastern Red Lake belt has experienced at least three deformation events. D_1 , and related NE trending F_1 folds, occurred prior to 2720 Ma, and may have formed at ca. 2731 Ma during the emplacement of a late phase of the Little Vermillion Lake batholith or the initial phase of Kenoran NW-SE shortening. NE trending shear zones, such as the East Bay, Post Narrows, and Flat Lake-Howey Bay deformation zones, may have formed at this time. D_2 is interpreted to have occurred at ca. 2700 Ma, synchronous with the emplacement of a late phase of the Trout Lake batholith, and it produced NW trending F_2 folds and foliations. D_2 shear zones were developed along F_2 fold limbs and hinges in the Campbell and Goldcorp mine areas, and some NE trending shear zones along the East Bay and Post Narrows deformation zones were active. D_3 deformation resulted in sinistral shear along the East Bay deformation zone after 2700 Ma, cutting F_2 folds and producing NE to NNE trending folds (F_3) within the shear zone.
- Greenschist facies M_1 metamorphism may have occurred during D_1 , from the occurrence of chlorite and rare actinolite along S_1 . It may have been related to the emplacement of a phase of the Little Vermillion Lake batholith at ≈ 2731 Ma. Minor (generally < 10 m) amphibolite grade M_2 metamorphic aureoles associated with the Dome and McKenzie stocks (≈ 2720 Ma) overprint greenschist facies rocks. The final metamorphic event (M_3) in the eastern part of the belt involved the development of wide haloes (> 2 to 4 km) of contact metamorphism around the marginal phase of the Trout Lake batholith, at ≈ 2700 Ma. M_3 occurred during D_2 - F_2 , from the increase in strain with metamorphic grade (inferred from foliation intensity), and the lack of rotation of M_3 garnet and amphibole extension lineations across F_2 folds.
- Rocks at the Campbell and Goldcorp mines experienced low P/high T metamorphism at the M_3 greenschist / amphibolite facies transition. Garnet-biotite thermometry and metamorphic mineral assemblages indicate a sharp, eastward increasing temperature gradient. Rocks in the western part of the Campbell mine were metamorphosed ≤ 460 °C, and rocks at the Goldcorp mine were metamorphosed to ≈ 565 °C. Metamorphic pressure was ≤ 3 Kbars, from mineral assemblages and GASP barometry.

- Spatial variations and textural relations between ore and gangue minerals and the metamorphic grade of the host rocks, support Au deposition during the peak metamorphism. Gahnite is rimmed by, and contains inclusions of, auriferous arsenopyrite. Garnet contains arsenopyrite inclusions, and is cut by tetrahedrite. Aurostibite is restricted to greenschist facies domains, and spatial distribution of aurostibite and Ag content of native gold suggest deposition during metamorphism of the wallrocks.
- The Campbell and adjacent Goldcorp mines occur on the western, highly attenuated limb of the Balmer synform that was formed during D₂.
- Auriferous veins occur in local high strain zones of D₂. Auriferous veins parallel to the foliation are spatially associated with attenuated F₂ NW fold limbs, and are characterized by strong flattening strain. Foliation-oblique, NNW and E-W auriferous veins occur in shear zones at the contacts between mafic and ultramafic rock units, developed during flexural slip folding..
- High Au grade hydrothermal breccias plunge moderately SW, parallel to F₂ fold axes, and boudin axes in carbonate veins. Breccia bodies are also spatially associated with these fold hinges, and the intersection of foliation oblique shear zones.
- Deformation continued after auriferous mineralization. Black line faults, which are common in the mine, displaced auriferous veins. NW trending brittle-ductile oblique-slip (left-lateral reverse) are related to decollement along the western limb of the Balmer synform and they displaced the auriferous zones during late D₂, and transpose the F₂ limbs to develop hook-shaped folds. These faults may have offsets of > 100 m.
- Colliform to banded quartz + carbonate veins, formed during D₂-M₃, are overprinted by later auriferous mineralization that took place near the peak of M₃.
- Several stages of veining and alteration are identified.
 1. Synvolcanic, pervasive alteration resulted in patchy bleached zones and Al-enrichment in basalt. This alteration is not related spatially or temporally to Au mineralization.

2. Pre-ore carbonate alteration resulted in the formation of ankerite and dolomite as veins and in the matrix of basalt and ultramafic rocks. This alteration occurred during an early stage of D₂.
 3. Gold was deposited with quartz and arsenopyrite as veins, and as breccia matrix within earlier formed carbonate veins. High Au grade hydrothermal breccias plunge moderately SW within the shear zones hosting carbonate veins and veinlets.
 4. Post Au alteration involved minor retrograde chlorite and late quartz and carbonate veinlets.
- Wallrock alteration adjacent to auriferous veins contains chlorite+white mica+carbonate in greenschist facies domains, and biotite+amphibole+garnet+anorthitic plagioclase in amphibolite facies domains. They formed at approximately 3 Kbars pressure, and 460 °C to 565 °C. Pyrite, pyrrhotite, and arsenopyrite occur throughout the deposit, and carbonate and white mica decrease in abundance with increasing metamorphic grade. The alteration mineralogy is compatible with the metamorphic grade of the host rocks. AFMC diagrams and textural relations show that alteration occurred during the peak of metamorphism.
 - Syn-metamorphic hydrothermal fluid flow was focused along the foliation parallel and oblique shear zones. Auriferous hydrothermal fluids flowed westward in the area, suggested from the form of the garnet-in isograd.
 - All vein carbonates from the eastern part of the Red Lake belt are anomalously enriched in ¹³C, compared to carbonates from other Archean gold camps. Carbonate associated with Au is mildly enriched in δ¹³C (≤ 2.69‰) relative to pre- and post gold carbonate. The author suggests that phase separation resulted in discharge of CH₄ and the enrichment of ¹³C in gold related carbonate.
 - Primary and pseudosecondary fluid inclusions in high gold grade, quartz + arsenopyrite hydrothermal breccias are predominantly Type I, CO₂+CH₄ (4 to 16 mole% CH₄) bearing. Inclusions in weakly auriferous quartz veins are Type I and co-existing Type II H₂O + CO₂ populations. Type I inclusions from weakly auriferous quartz have CO₂ densities between 0.8 and 0.88 g/cm³, and are relatively methane-poor (≤3 mole%). These observations are interpreted to indicate phase separation of

the auriferous fluid (represented by the Type II population) in the hydrothermal breccias, with methane enrichment in the Type I inclusions. Selective transport of the aqueous phase resulted in the ubiquitous occurrence of Type I inclusions in high gold grade ore.

- Scheelite associated with Au yields initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios between 0.7012 and 0.7017 at 2.7 Ga. These data suggest that the fluid, at 2.7 Ga, did not interact significantly with the old (≈ 3.0 Ga) host rocks. The data do not support fluid derivation from prograde metamorphism of the supracrustal rocks, or an evolved meteoritic water source. Instead, the data suggest the fluid interacted with, or was derived from, contemporaneous, non-anatectic igneous rocks. The Sr isotopic composition of scheelite from post-gold black line faults indicates interaction of the fluid with the wallrocks or formation long after the Au mineralization.
- The alteration mineralogy at the Campbell mine is similar to that from other syn-metamorphic, Archean lode-type Au deposits in mafic/ultramafic rocks.

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Appendix 1

Bulk chemical composition of ultramafic rocks

SAMPLE	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ (T)*	Cr ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	CO ₂	S	LOI**	Total
E57013	46.30	0.345	6.53	11.30	0.22	0.24	7.87	11.00	0.12	0.12	0.05	15.10	0.04	16.30	100.40
E57015	45.70	0.517	8.17	11.30	0.45	0.24	8.30	9.54	0.18	0.31	0.08	12.30	0.05	14.70	99.50
E57016	38.90	0.352	6.64	13.50	0.37	0.28	10.30	11.60	0.05	0.03	0.04	17.10	0.11	18.20	100.30
E57017	39.60	0.300	5.55	11.70	0.37	0.32	10.80	12.30	0.03	0.05	0.04	17.90	0.13	19.40	100.50
E57018	48.30	0.318	6.21	8.48	0.52	0.26	8.32	11.10	0.28	0.58	0.07	16.00	0.26	15.30	99.70
E57019	53.90	0.357	6.28	7.63	0.50	0.24	7.72	9.26	0.35	0.42	0.05	12.20	0.09	13.60	100.30
E57020	45.20	0.342	5.99	10.50	0.39	0.27	10.20	10.40	0.05	0.68	0.04	15.20	0.10	16.40	100.50
E57021	51.20	0.391	6.87	9.65	0.37	0.21	10.50	7.25	0.01	0.37	0.03	10.40	0.05	13.30	100.10
E57022	42.80	0.399	6.82	13.10	0.30	0.24	13.10	7.82	0.02	0.98	0.03	11.60	0.24	13.80	99.40
E57023	40.70	0.300	5.18	14.90	0.31	0.28	13.90	9.32	0.04	0.95	0.03	13.50	0.49	14.30	100.20
E57024	53.70	0.541	9.03	12.80	0.42	0.15	4.97	6.70	1.31	1.75	0.02	7.28	3.77	3.35	94.70
E57028	32.00	0.393	6.73	13.40	0.23	0.21	12.00	13.00	0.15	0.52	0.03	19.30	2.36	17.00	95.70
E57029	44.80	0.507	8.55	9.94	0.17	0.20	9.48	9.72	0.58	0.82	0.04	13.90	0.31	15.10	99.90
E57030	45.00	0.367	6.18	10.20	0.38	0.18	14.00	7.83	0.04	0.30	0.03	11.90	0.16	14.30	98.80
E57031	41.80	0.339	5.78	12.70	0.46	0.28	14.40	9.02	0.03	0.58	0.04	13.00	0.25	14.90	100.30
E57033	34.80	0.403	6.70	11.80	0.36	0.21	15.90	9.95	0.02	0.97	0.04	15.70	0.09	17.90	99.10
E57037	38.00	0.460	7.89	11.40	0.27	0.26	10.60	12.20	0.24	0.47	0.05	17.60	0.05	18.60	100.40
E57038	32.50	0.323	5.83	10.50	0.40	0.22	10.40	16.20	0.23	0.21	0.04	22.80	0.14	23.60	100.50
E57039	43.50	0.420	7.37	10.70	0.33	0.23	8.47	11.60	0.19	0.28	0.04	15.30	0.07	16.80	99.90
E57040	46.00	0.458	7.71	11.00	0.38	0.22	8.28	10.50	0.27	0.41	0.05	13.80	0.17	14.70	100.00
E57041	48.80	0.453	7.42	10.60	0.36	0.28	7.25	9.97	0.23	0.64	0.04	12.80	0.03	14.40	100.40
E57042	50.90	0.386	6.86	11.60	0.31	0.26	6.44	8.04	0.10	0.21	0.04	11.70	1.38	10.70	95.80
E57043	41.00	0.426	7.50	12.60	0.34	0.16	12.70	8.99	0.02	0.01	0.04	13.50	0.08	16.10	99.90
E57044	32.50	0.370	6.14	8.95	0.26	0.20	10.90	15.50	0.20	1.25	0.04	23.60	0.03	24.00	100.30
E57045	35.70	0.456	7.57	12.70	0.34	0.20	12.50	11.30	0.02	0.02	0.05	17.20	0.09	19.40	100.30
E57054	39.20	0.425	7.08	11.00	0.32	0.18	13.50	10.60	0.01	0.16	0.04	16.00	0.07	17.80	100.30
E57056	43.40	0.383	7.06	10.00	0.27	0.19	8.28	12.80	0.25	0.60	0.05	17.40	0.04	17.30	100.60
E57057	39.90	0.382	6.63	9.99	0.37	0.21	8.74	14.00	0.28	0.58	0.06	18.90	0.15	19.10	100.20

* Total Fe expressed as Fe₂O₃

** Loss on Ignition.

Appendix 2

Bulk Chemical Composition of "Chloritic" Ultramafic Rocks

SAMPLE	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ (T)*	Cr ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	CO ₂	S	LOI**	Total
E57034	39.00	0.467	7.82	14.00	0.37	0.19	17.10	6.58	0.03	1.35	0.04	9.61	0.32	12.20	99.20
E57035	40.60	0.480	7.85	12.70	0.34	0.24	13.10	9.12	0.02	0.62	0.05	13.40	0.30	15.00	100.10
E57036	35.10	0.457	7.96	13.10	0.29	0.25	12.50	11.00	0.02	0.09	0.04	16.40	0.15	18.90	99.70
E57046	40.50	0.394	6.76	11.70	0.32	0.15	13.00	9.59	0.03	0.02	0.04	14.40	0.19	16.80	99.30
E57047	38.80	0.420	7.40	11.50	0.31	0.17	14.50	9.78	0.02	0.24	0.04	14.70	0.04	17.20	100.40
E57048	44.10	0.476	7.07	9.94	0.36	0.17	13.80	8.50	0.07	0.27	0.04	12.80	0.04	15.20	100.00
E57049	38.30	0.371	6.48	10.80	0.28	0.18	15.10	9.78	0.16	0.36	0.04	14.80	0.03	17.20	99.10
E57050	40.60	0.451	7.24	13.60	0.39	0.12	16.40	6.37	0.19	0.39	0.05	9.76	0.10	13.00	98.80
E57051	42.80	0.471	7.47	12.30	0.40	0.13	14.90	6.83	0.17	0.51	0.05	10.20	0.08	13.20	99.20
E57052	39.60	0.539	8.53	12.70	0.23	0.13	13.80	7.47	0.05	0.74	0.05	11.50	0.03	14.60	98.40
E57053	39.40	0.462	7.43	12.30	0.30	0.15	14.80	8.99	0.04	0.20	0.05	13.30	0.03	16.10	100.20
E57058	35.40	0.385	6.29	11.30	0.32	0.22	12.70	12.20	0.03	0.03	0.04	18.80	0.11	20.20	99.10
E57060	38.80	0.413	6.24	10.80	0.38	0.18	12.30	11.90	0.01	0.28	0.03	17.70	0.42	18.30	99.60
E57061	41.20	0.438	6.97	11.60	0.34	0.21	11.30	10.20	0.06	0.38	0.04	15.50	0.36	16.10	98.80

* Total Fe expressed as Fe₂O₃

** Loss on Ignition.

Appendix 3

Bulk chemical compositions of samples used for the AFM(Ca) diagrams, and analytical methods

“E” Series Samples

Samples with names starting with “E” were collected by Placer Dome Canada Ltd. Exploration in 1993, as part of a Campbell mine exploration program. Drill holes were advanced through the A, F zones from the 13th level, and G zone from the 19th level of the mine, to test geochemical zoning adjacent to the ore zones. Selected and composite samples were obtained from split, AQ and BQ core from these holes.

Selected samples were collected to examine specifically, veins or wallrock alteration. Veining was excluded from wallrock samples. These samples were 15 to 20 cm long. The chemical composition was obtained by XRAL Laboratories in Don Mills, Ontario. Major oxide elements, Ba, Sr, Rb, Nb, Y, Zr were determined by XRF. Au, Ag, As, Bi, Cd, Cu, Pb, V, Ni, Mo were analyzed by ICP. Samples containing >10,000 ppb Au by ICP were re-analyzed by fire assay method and the results were reported in grams per ton. Sb was determined by instrumental neutron activation analysis, CO₂ and S by Leco furnace.

Composite samples ranged from 1.5 to 0.3 m in length, and sampled grossly similar styles of wallrock alteration and veining, with end-of-interval based on changes in these characteristics. Geochemical determination was made by the Placer Dome Research Centre in Vancouver, British Columbia. Ag, As, Co, Ca, Cu, Mo, Pb, Zn, Sb, Cd, Ni, Mn, Bi, Cr, V, Ba, W, Be, La, Sr, Ti, Al, Fe, Mg, K, Na, and P were analyzed by ICP following digestion of a 0.5 g sample Au was determined by graphite-furnace AAS, following digestion of a 10.0 g sample by the same method.

Samples from Christie (1986)

Christie (1986) reported the composition of selected rock types and alteration facies at the Campbell mine. Details of the sample selection and analysis method are reported therein.

Samples from Christie, 1986

sample	Al ₂ O ₃	Fe ₂ O ₃ (T)*	MgO	CaO
c4-2	14.23	10.60	5.21	7.83
c18-19	14.89	12.25	6.30	7.49
c18-23	15.70	10.26	5.13	2.79
c18-62	13.06	11.85	5.83	8.24
c18-63	13.53	11.92	5.10	6.25
c18-75	15.18	12.83	5.23	4.53
c18-77	12.85	13.83	8.10	9.28
c18-82	13.88	15.67	5.53	5.35
c18-83	13.50	19.48	4.66	3.28
80c4	14.07	8.32	4.98	7.11
c19-35	14.86	11.70	5.60	7.00
c19-61	11.47	12.13	6.88	7.84
c21-2	12.63	13.87	3.46	6.41

crl-116	12.93	1.84	2.06	3.95
crl-123	13.04	2.39	1.19	1.06
crl-137	13.94	3.31	3.33	6.74
crl-85	3.75	12.97	27.49	8.65
crl-86	14.07	13.53	6.81	10.30
crl-87	14.21	16.75	5.43	3.26
crl-88	13.93	2.52	2.51	3.88
crl-104	14.16	11.28	4.51	8.31
crl-81-2	13.20	0.96	0.16	0.72
crl-83	8.18	10.71	13.79	6.70

Samples analyzed by C. Tarnocai

crl02	12.94	8.49	2.73	2.42
crl03	6.23	10.45	13.24	11.02
crl05	1.76	4.90	4.88	9.76
crl06	12.32	1.40	0.81	1.48
crl09	14.55	10.66	5.59	8.10
crl12	13.49	0.28	0.26	0.13
crl13	14.36	11.64	5.24	10.10
crl40	3.97	7.35	8.18	9.50
crl49a	16.57	8.07	4.06	7.43
crl19	17.65	11.92	5.14	5.69
crl04	4.77	5.33	2.61	4.70
crl65	7.47	0.48	5.49	10.19

Samples from Placer Dome Canada Ltd.

sample	Al ₂ O ₃	Fe ₂ O ₃ (T)*	MgO	CaO	MnO	
E57062	21.30	10.10	1.79	0.71	0.24	
E57063	14.30	10.90	2.53	0.93	0.23	
E57064	14.50	10.40	2.99	2.96	0.22	
E57065	15.10	6.04	1.34	0.73	0.13	
E57066	16.30	6.36	1.19	0.25	0.14	
E57067	14.30	11.10	2.11	0.36	0.27	
E57068	15.20	17.40	3.24	0.80	0.40	
E57069	13.70	18.70	3.52	0.85	0.39	
E57070	12.70	7.05	1.33	0.52	0.15	
E57071	12.60	15.30	2.93	0.43	0.34	
E57072	12.30	12.40	2.54	0.57	0.24	
E57074	12.90	8.32	1.64	0.31	0.19	
E57073	12.80	6.07	1.22	0.24	0.12	
E57075	11.30	4.89	0.97	0.17	0.11	
E57076	11.90	11.60	2.47	0.46	0.24	
E57077	5.42	17.80	1.71	2.65	0.11	
E57078	12.90	12.00	2.40	0.52	0.22	
E57079	15.10	10.20	2.03	0.30	0.24	
E57080	13.50	9.29	2.53	0.50	0.16	
E57081	11.40	12.10	3.34	0.62	0.20	
E57082	8.59	5.73	1.77	1.36	0.06	
E57083	8.64	7.12	2.20	1.15	0.08	
E57084	8.82	11.80	3.79	1.32	0.18	
E57085	10.10	7.95	2.36	0.97	0.11	
E57086	9.58	5.91	1.90	1.30	0.06	
E57087	0.70	10.20	9.55	16.90	0.23	
E57088	7.77	6.03	1.31	1.44	0.04	
E57089	9.29	5.07	1.38	0.25	0.07	
E57090	7.93	7.91	1.98	1.47	0.08	
E57091	0.91	7.13	5.88	9.32	0.20	
E57092	0.67	7.30	4.33	7.02	0.17	
E57096	10.30	10.30	1.52	0.54	0.09	
E57097	12.20	14.30	4.12	1.18	0.27	
E57098	7.90	16.40	4.16	2.87	0.27	
E57099	11.50	16.10	4.14	0.98	0.34	
E57100	11.80	11.60	4.06	3.40	0.21	
E57101	6.03	8.10	2.63	3.02	0.16	
E57102	13.10	13.90	4.78	5.49	0.36	
E57103	13.80	8.03	4.35	7.20	0.19	
E57104	14.20	11.00	5.25	8.42	0.25	
E57105	15.00	13.40	5.88	8.30	0.26	
E57106	12.40	10.90	6.52	9.31	0.18	
E57107	12.00	12.20	7.55	8.84	0.19	
E57108	14.80	14.70	6.09	6.69	0.24	
E57109	10.50	12.50	8.87	18.00	0.33	
E57110	12.90	14.40	6.62	8.35	0.26	
E57111	12.60	10.50	4.34	6.01	0.23	
E57112	14.40	13.80	6.14	8.98	0.27	

sample	Al ₂ O ₃	Fe ₂ O ₃ (T)*	MgO	CaO	MnO	
E57113	13.80	15.00	6.44	7.91	0.29	
E57114	12.60	13.50	6.58	10.10	0.21	
E57115	11.60	12.50	8.34	9.85	0.18	
E57116	13.40	12.80	6.18	8.32	0.22	
E57117	13.50	15.50	5.32	5.52	0.24	
E57118	12.50	8.91	6.03	11.10	0.23	
E57120	11.50	10.70	6.64	13.80	0.26	
E57121	12.50	12.20	6.52	10.80	0.25	
E57122	11.60	11.50	6.67	10.40	0.26	
E57123	11.80	11.40	6.72	10.80	0.27	
E57124	13.50	11.90	6.55	10.90	0.24	
E57125	12.20	12.10	6.86	10.40	0.19	
E57126	15.80	9.01	5.02	10.60	0.23	
E57127	12.30	11.40	7.71	10.10	0.19	
E57128	12.80	12.00	6.44	8.65	0.20	
E57129	14.70	9.58	4.98	8.21	0.19	
E57130	15.60	12.00	5.81	8.53	0.25	
E57132	18.80	14.10	3.88	3.14	0.27	
E57133	19.40	11.10	3.16	1.75	0.20	
E57134	14.50	10.30	4.87	7.54	0.20	
E57135	13.80	8.80	5.32	11.20	0.21	
E57137	13.60	9.99	5.58	12.70	0.20	
E57138	14.20	8.42	4.29	7.59	0.18	
E57139	12.30	13.20	5.32	12.60	0.19	
E57140	13.60	8.79	5.27	10.70	0.21	

* Total Fe expressed as Fe₂O₃

Appendix 4

Scanning Electron Microscope Method

Mineral compositions were determined using a JEOL JSM-6400 SEM and a Link EXL energy dispersive spectrometer at Carleton University, Ottawa. Analytical conditions were counting time of 120s, 20 kV accelerating voltage, 0.8 nA beam current. Standards were natural quartz (Si), wollastonite (Ca), hematite (Fe), microcline (K), and synthetic Al_2O_3 (Al), MgO (Mg), Cr_2O_3 (Cr), and MnTiO_3 (Mn and Ti). Analytical uncertainties for major and minor (< 1 wt.% as oxides) elements are $\pm 2\%$ and $\pm 10\%$ of the quoted values, respectively.

Appendix 5

Analytical procedure for carbon and oxygen isotopic composition of carbonates.

For the determination of carbon and oxygen isotopic compositions of carbonates, CO₂ was released from the carbonate by reacting with 100% phosphoric acid, following the method of McCrea (1950). Isotopic compositions of CO₂ was determined using a VG-Sira 12 triple-collector mass spectrometer. The duration of the reaction was 1 day at 50 °C. For the calculation of O isotopic compositions of ankerite, isotope fractionation factor, α , of 1.010603 was at 50 °C (Rosenbaum and Sheppard, 1986). The analytical precision (2σ) is 0.1 ‰ for $\delta^{13}\text{C}$, and 0.5‰ for $\delta^{18}\text{O}$.

Appendix 6

Analytical procedure for determining the strontium isotopic compositions of scheelite

For isotopic analysis, scheelite grains were mechanically separated from a mixture of quartz, carbonate and scheelite using an ultraviolet lamp. The purity of scheelite separates was over 90% judging from their fluorescent light. Impurities are mostly quartz, which does not contain Sr, and does not affect the isotopic compositions of scheelite samples.

Approximately 10 mg of powdered scheelite samples were placed into Savillex screw-top Teflon vials and added a mixed spike of ^{84}Sr - ^{87}Rb . The mixtures were dissolved using HF-HNO₃ (4:1) overnight at 2000 °C. One sample was digested with HF-HNO₃ as well as HF-HClO₄ (10:1). The results were comparable within the analytical errors. After the digestion, samples were dried at 150 °C and raised to 200 °C and reacted with 6N HCl, followed by evaporation to dryness. Reaction with HCl and evaporation to dryness was repeated twice and the sample was dissolved in 2.5N HCl for cation separation using Bio Rad AG W-X8 (200-400 mesh). Rubidium and Sr were eluted with 2.5N HCl. Total chemical blanks are less than 0.05 ng for Sr, negligible in comparison to the amount of Sr present in the sample aliquots. All acids used were double distilled in Savillex Teflon distillation vessels at sub-boiling temperatures. Double distilled HF and HClO₄ were purchased from Seastar Chemicals.

Isotopic measurements were made on a multi-collector thermal ionization mass spectrometer (Finnigan MAT 261), and the ratios for Sr isotopes were normalized to $^{86}\text{Sr}/^{88}\text{Sr}$ of 0.1940. During the course of isotopic analysis, the measurements of NBS987 gave $^{87}\text{Sr}/^{86}\text{Sr}$ of $0.710275-0.71027 \pm 0.0000119$.

Appendix 7

Composition of vein and silica replacement ore at the Campbell Mine[▲]

Sample#	Description*	Drill Hole#	From**	To**	Ag	As	Au (g/t) **	Au (ppb) **	Pb	Rb	Sb	Sr	Zn	
E57131	Q/C STR	13-371	5.4	6.3	0.20	57			179	2	22	25	85	59
E57137	Q/C VLTS	13-371	25.3	27.9	0.50	103			1370	BD	38	21	96	76
E57142	Q/C VEIN	13-371	34.9	37.9	0.40	97			406	BD	33	27	126	56
E57145	Q/C VLTS MIN	13-371	40.2	41.5	1.80	1290			2430	3	54	48	76	68
E57147	Q/C VEIN	13-371	42.9	43.9	2.20	3490	28.5	> 10000	3	37	97	57	771	
E57148	Q/C VLTS	13-371	43.9	44.9	0.40	789			280	BD	39	25	28	146
E57149	Q/C VEIN	13-371	44.9	46.0	4.40	2840	714	> 10000	BD	59	31	84	107	
E57160	Q/C VLTS MIN	13-371	79.9	81.2	0.50	67			147	BD	32	24	85	115
E57161	Q/C STR	13-371	81.2	83.0	0.60	80			2	3	21	22	88	76
E57168	BX Q/C	13-371	105.1	108.9	0.60	144			376	BD	36	38	68	525
E57183	Q/C VEIN MIN	13-437	20.0	20.5	0.50	226			352	2	19	16	93	69
E57184	Q/C VEIN MIN	13-437	20.6	20.4	0.20	99			6	BD	30	22	82	43
E57187	BX Q/C MIN	13-437	30.7	31.4	1.20	117			1100	2	32	48	106	54
E57194	Q/C VLTS	13-437	48.2	51.0	1.30	162			21	4	10	19	25	309
E57025	Q/C STR	13-484	155.4	156.2	2.80	937			1380	3	48	21	275	15
E57026	Q/C STR	13-484	158.8	159.5	1.70	1160			828	4	45	21	181	11
E57027	Q/C STR	13-484	162.2	162.8	2.30	1140			2090	3	45	30	294	15
E57032	Q/C VEIN	13-484	170.1	170.7	0.50	675			46	BD	31	21	149	15
E57082	MIN REP	19-707	76.0	78.0	2.30	3300	44.82	> 10000	BD	57	45	53	34	
E57083	MIN REP	19-707	78.9	80.0	1.30	9420			6530	2	46	66	71	40
E57084	MIN REP	19-707	81.5	82.7	1.00	8310			6760	BD	40	58	27	43
E57085	MIN STR	19-707	85.0	86.4	1.60	8380			1710	3	51	66	64	30
E57086	MIN STR	19-707	88.0	89.5	1.10	6970	57.59	>10000	BD	57	43	62	24	
E57087	MIN REP	19-707	89.9	90.6	3.00	971			726	BD	12	6	31	32
E57090	BAS SIL MIN	19-707	97.8	99.0	0.80	11800			3590	4	54	46	37	28
E57091	MIN REP	19-707	100.8	101.8	2.00	9850	54.41	>10000	3	4	40	29	94	
E57092	MIN REP	19-707	103.6	104.5	5.40	152	129.00	>10000	3	5	8	24	62	
E57093	Q/C VEIN	19-707	105.0	105.7	0.70	71			1840	BD	6	19	43	23
E57094	MIN REP	19-707	107.4	109.0	13.00	16800	133.60	>10000	17	1	150	20	398	
E57095	Q/C VEIN	19-707	109.8	110.7	0.60	192			533	BD	11	24	46	67
E57101	BAS MIN VN	19-707	129.3	130.3	1.50	12600			4940	13	29	120	44	38

= number.

* BAS, basalt; BX, brecciated; MIN, mineralized with sulphides; Q/C, quartz+carbonate vein; REP, silica replacement body; SIL, silicified; STR, stringer; VLTZ, veinlets; VN, vein.

** feet.

BD = below detection limit

▲ Analysis in ppm, unless otherwise noted.

▲▲ Where Au exceeded 10000 ppb, samples re-analyzed by fire assay with AAS completion, and results reported as grams per ton (g/t).

Appendix 8

Pressure and Temperature of Au mineralization and metamorphism from Archean Gold Deposits

Location / Deposit	Production+Reserve	Au mineralization P (Kbar) / T (°C)		Metamorphism P (Kbar) / T (°C)	Timing (rel. to meta)	Host Rock	
Superior Province of Canada							
Sigma ¹	> 3.2 M oz.	2	/ 285-395	Greenschist	Post	andesite, porphyry	
McIntyre-Hollinger ²		1.3-2.9	/ 380-480	Greenschist	Post	mafic to felsic vol.	
Musselwhite ⁵	1.3 Mt@5.8g/t ⁶	3	/ 530-550		Syn	silicate SFBIF	
Campbell/Goldcorp ³	(1.5 Mt@5.8 ⁴ , 15.2 Mt @18.5g/t ¹¹) ^A (1.9Mt@34.8g/t ¹² , 8.3 Mt@11.8g/t ¹¹) ^B	1.5-3	/460-565	Grn./Amph.	Syn	Mafic to UM volc.	
Australia							
Norsman Dist. ^{4,7}	>140t, 10.3Mt@13.2g/t, res 1.8Mt@7.0g/t						
North Royal		3	/450±20	3	/500-550	Syn	Mafic-UM Int. Mafic vol.
Southern Cross Province⁸							
Marvel Loch ^{18,19}	0.62Mt@7.4g/t* 7.48Mt@2.3g/t*	3-5	/500-640	3	/ 550-630	Post	Komatiite, gabbro
Fraser's ⁷	.24Mt@3.7g/t .65Mt@5.0g/t (pp)	3-5	/ 520-590	3	/ 520-670	Syn	Mafic-UM
Griffins Find ⁷	.71Mt@3.8g/t	6	/ 700	5-7	/ 700-750	Syn	sed, Mafic vol.
Eastern Goldfields Province¹⁰							
Missouri	921 kg Au	3-4	500-600		Syn	Fe-tholeiite	
Sand King	8500 kg Au	3-4	500-600		Syn	Fe-tholeiite	

1. Robert and Kelly, 1987; 2. Smith et al, 1984; 3. This study; 4. Mueller, 1992; 5. Hall and Rigg, 1986; 6. Reserves, Placer Dome 1999 Annual Report; 7. Groves et al., 1990; 8. Compiled by McCuaig et al. (1993); 9. Mueller, 1997, 10. Mernagh and Witt, 1992. 11. Blackburn et al., 1998. 12. Reserves, Goldcorp website, www.goldcorp.com. Aug 2000.
A. Campbell mine reserves and production, B. Goldcorp Red Lake mine reserves and production. Abbreviations:
Amph.=amphibolite, Grn.=greenschist, Int.=intermediate, pp=proven and probable reserves, res.=reserves,
SFBIF=sulphide facies iron formation, UM=ultramafic, vol.=volcanic rocks. * = as of 1994

Appendix 9

**Invisible gold in sulfides from the Campbell mine, Red Lake greenstone belt,
Ontario: Evidence for syn-peak metamorphic mineralization.**

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"INVISIBLE" GOLD IN SULFIDES FROM THE CAMPBELL MINE, RED LAKE GREENSTONE BELT, ONTARIO: EVIDENCE FOR MINERALIZATION DURING THE PEAK OF METAMORPHISM

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ABSTRACT

The Au content of pyrite and arsenopyrite from the Campbell mine, Red Lake mining camp, Ontario, has been established by secondary-ion mass spectrometry (SIMS). All Au contents significantly exceed the minimum detection-limits of the technique (36 ppb in arsenopyrite, 193 ppb in pyrite). Arsenopyrite is characterized by high contents of "invisible" Au (up to 0.56 wt%). Coarse-grained arsenopyrite displays Au zoning, and fine-grained arsenopyrite generally displays a uniform Au content. Gold is present in low concentrations in pyrite (up to 168 ppm), and elevated Au is related to As-rich zones. The host rocks at the mine have experienced low-pressure amphibolite-grade metamorphism. We consider it unlikely that compositional zoning could survive metamorphic recrystallization. Textural relations between garnet and auriferous arsenopyrite, and the presence of auriferous arsenopyrite inclusions in garnet, indicate that the introduction of Au was synchronous with peak conditions of metamorphism.

Keywords: arsenopyrite, pyrite, metamorphism, "invisible" gold, secondary-ion mass spectrometry (SIMS), crystal chemistry, Archean gold mineralization, Campbell mine, Red Lake, Ontario.

SOMMAIRE

La teneur en or de la pyrite et de l'arsénopyrite de la mine Campbell, du camp minier de Red Lake, en Ontario, a été déterminée par spectrométrie de masse des ions secondaires (SIMS). Toutes les teneurs en Au dépassent largement le seuil de détection de la méthode (36 ppb dans l'arsénopyrite, 193 ppb dans la pyrite). L'arsénopyrite possède des teneurs élevées en or "invisible" (jusqu'à 0.56% en poids). L'arsénopyrite à grains grossiers fait preuve de zonation dans la distribution de l'or, tandis que l'arsénopyrite à grains fins possède en général une teneur uniforme. De plus faibles concentrations d'or sont présentes dans la pyrite (jusqu'à 168 ppm), et les teneurs les plus élevées coïncident avec les zones riches en As. Les roches-hôtes du gisement ont subi un métamorphisme dans le faciès amphibolite à faible pression. Nous n'entrevoions pas la possibilité que la zonation dans ces minéraux puisse survivre la recristallisation métamorphique. Les relations texturales impliquant la garnite et l'arsénopyrite aurifère, aussi bien que la présence d'inclusions d'arsénopyrite aurifère dans le grenat, indiquent que l'introduction de l'or a eu lieu lors du paroxysme métamorphique.

(Traduit par la Rédaction)

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Mots-clés: arsénopyrite, pyrite, métamorphisme, or "invisible", spectrométrie de masse des ions secondaires, chimie cristalline, minéralisation aurifère archéenne, mine Campbell, Red Lake, Ontario.

INTRODUCTION

The Campbell mine, at Red Lake, northwestern Ontario, exploits an Archean lode-Au deposit; past production and present reserves amount to approximately 344 tonnes of Au. The deposit is hosted within mafic-ultramafic volcanic rocks; it is situated at the transition between the upper greenschist and lower amphibolite facies (Andrews *et al.* 1986, Tamocai & Hattori 1994). The host rocks have been metamorphosed at temperatures ranging between 450 and 500°C, and approximately 2–3 kilobars pressure (Tamocai & Hattori 1996). Wallrocks in the upper and western portions of the deposit contain upper-greenschist-facies mineral assemblages, whereas those in the deeper and eastern portions of the deposit contain lower-amphibolite-facies assemblages. The upper-greenschist-facies alteration assemblages include chlorite and white mica (muscovite and paragonite); the lower-amphibolite-facies alteration assemblages include magnesian tschermakite, gedrite, almandine, and cordierite. The peak metamorphic assemblage does not exhibit significant retrogression, indicating that the post-peak hydrothermal activity was minor. Ore veins in amphibolite-facies domains may contain hornblende and gahnite (ZnAl_2O_4). Arsenopyrite occurs in veins and wallrocks, and displays a strong spatial association with ore at the Campbell mine.

Very high Au grades occur in quartz–arsenopyrite ore samples that contain few visible Au grains. Spectacular leaf gold occurs in fractures that cut quartz–arsenopyrite ore. This relationship suggests that ores contain significant refractory Au, and that the Au may be remobilized locally into late fractures.

Gold mineralization in greenschist-facies terranes is considered to postdate peak metamorphism (Robert &

Brown 1981, Wong *et al.* 1991), but the timing of Au mineralization in amphibolite-facies terranes is under debate. Pre-metamorphic Au mineralization has been reported at the Big Bell deposit, Yilgarn Block, Australia (Phillips & de Nooy 1988). Syn-peak metamorphic Au deposits have been documented in the Southern Cross Province of the Yilgarn Block, and at Main Hill and Breccia Hill in the Pilbara Block (Bloem *et al.* 1994, Neumayr *et al.* 1993), and at Musselwhite in the Superior Province (Hall & Riggs 1986). Syn- to post-peak metamorphic Au mineralization has been suggested for the Con mine in the Slave Province (McDonald *et al.* 1993), and post-peak metamorphic introduction of Au is indicated at the Detour mine (Marmont 1986), in the Superior Province.

Results from scanning electron microscopy (SEM) and secondary-ion mass spectrometry (SIMS) analysis of pyrite and arsenopyrite from the Campbell mine are presented in this paper. The occurrence of zoned auriferous arsenopyrite, and the textural relations between auriferous arsenopyrite and both garnet and gahnite, are used to constrain the timing of introduction of Au relative to metamorphism.

SAMPLE SELECTION

Ores at the Campbell mine are hosted by foliation-parallel, northwest-trending quartz–carbonate veins, and by foliation-oblique conjugate shears (Zhang *et al.*, in press), quartz–carbonate veins and pipe-like quartz–arsenopyrite replacement bodies. Both styles of mineralization have caused an envelope of wallrock alteration that contains disseminated arsenopyrite. Typical samples of ore were selected for SIMS and SEM analyses, from both foliation-parallel (1925 stope, F2 zone; and 2451–5 stope,

TABLE 1. SAMPLE LOCATIONS AND DESCRIPTIONS, ORE ZONES IN THE CAMPBELL MINE, RED LAKE BELT, ONTARIO

Ore Zone	Stope	Ore type	Mineralogy
F2	1925	Foliation-parallel quartz–ankerite vein.	Predominantly quartz, ankerite, pyrrhotite, minor pyrite, schistose, and tourmaline. 2–3 vol. % euhedral arsenopyrite.
P	2451-6 (EW)	Foliation-oblique, E/W-trending quartz veins.	Quartz, ankerite, pyrrhotite, pyrite, 3 vol. % fine-grained euhedral arsenopyrite.
P	2451-6 (NS)	Foliation oblique N/S trending quartz–arsenopyrite replacement zone.	Quartz with 10 vol. % euhedral arsenopyrite. Minor pyrrhotite and trace magnetite.
P	2451-5	Foliation-parallel quartz–ankerite vein.	Predominantly quartz, ankerite, 5 vol. % euhedral arsenopyrite, 3–5 vol. % patchy pyrrhotite aggregates, minor pyrite, trace sphalerite and staurolite.
L	1854	Foliation-oblique quartz–pyrite–arsenopyrite replacement zone.	Quartz, ankerite, 8 vol. % patchy pyrite, 3 vol. % fine-grained arsenopyrite, minor chromian muscovite, trace pyrrhotite and tourmaline.

P zone) and foliation-oblique (2451-6 stope N/S and EW structures, P zone; and 1854 stope, L zone) ore structures. The host rocks for these samples have been subjected to upper-greenschist- (P and F2 zone) to lower-amphibolite-facies (L zone) metamorphism. Sample locations and descriptions are provided in Table 1. Quantitative SIMS analysis was conducted on euhedral to subhedral grains of arsenopyrite, 40 to 150 μm in size, and on subhedral grains of pyrite 30 to 800 μm in size. Arsenopyrite used for the SEM study was also obtained from a quartz-arsenopyrite replacement zone in the 1756-2E stope, NL zone, which contains 5-10 vol.% gahnite, biotite, and ankerite, and from wallrocks in the 2451-5 P zone.

ANALYTICAL PROCEDURES

Rock samples were crushed and hand-sorted to produce pyrite and arsenopyrite separates. Mineral separates were mounted in a mixture of epoxy and 10 wt% graphite, and polished. Sulfide grains were checked for inclusions by

TABLE 2. SUMMARY OF THE SIMS EXPERIMENTAL PARAMETERS

Primary-Ion Beam and Polarity	Ca ⁺
Secondary-Ion Polarity	Negative
Matrix Masses Measured	¹⁹⁷ Fe, ³² S
Primary-Beam Current	300-350 nA
Primary-Beam Accelerating Voltage	10 kV
Impact Energy	14.5 kV
Field-Aperture Diameter	750 μm
Contrast-Diaphragm Diameter	400 μm
Raster	250 μm
Image Field	150 μm
Diameter of Analysis Area	62.5 μm
Acquisition rate	1 s cycle ⁻¹
Implanted standards	arsenopyrite, pyrite
Implantation fluence, species	2.5×10^{15} ions ¹⁹⁷ Au/cm ²
Mass resolution	~ 2,000 m/dm
Energy offset	none
Minimum detection limits: arsenopyrite	avg. 36 (28-50) ppb Au
pyrite	avg. 193 (119-266) ppb Au
Graphite-loaded polished section	carbon-coated
Depth of analyzed profiles	0.3 - 2.0 μm

reflected light microscopy, and with a JEOL JSM-6400 scanning electron microscope at up to 43,000 $\times$ magnification. Gold inclusions 100-200 nm across would be recognizable at this magnification by SEM.

The scanning electron microscope, with an attached Link EXL energy-dispersion spectrometer (EDS), was used to quantitatively determine major-element compositions of sulfides, and to qualitatively investigate the location of Au in zoned crystals of arsenopyrite. Major-element determinations were carried out for 150 s at an accelerating voltage of 20 kV, and a beam current of 0.8 nA, using pyrite for Fe and S, NiAs for As, and

CoAs₂ as standards. Accuracy is $\pm 1\%$ of the reported values for these elements. Because the minimum excitation-potential for elements with high mass-numbers is relatively large, and the K α line for Au (2.12 keV) is obscured by the K α line for sulfur (2.308 keV), the accelerating voltage was increased to 25 keV to qualitatively examine the AuL α peak at 9.711 keV in zoned arsenopyrite.

A Cameca IMS-4f double-focusing, magnetic-sector secondary-ion mass spectrometer was used to determine the Au contents of the sulfide samples. The operating conditions of the spectrometer are listed in Table 2. The collector was set at high mass-resolution ($m/\Delta m = 2000$) to alleviate molecular mass-interferences associated with ¹⁹⁷Au. Local calibrations of the magnet were made as described by Cabri & McMahon (1995), followed by depth profiles on implanted standards (e.g., Chrysosoulis *et al.* 1989). The ¹⁹⁷Au mass and one or two other masses (e.g., ⁵⁶Fe, ⁵⁷Fe + ³²S) were recorded, in order to monitor the system stability, the integrity of the sample, and to provide a reference mass for RSF (relative-sensitivity factor) calculations. Integrated counts for matrix mass ⁵⁷Fe + ³²S subsequently were used for the RSF calculations at the conclusion of the depth profile. The RSF is a multiplying factor used to convert the experimentally measured ion-count rate to atom density, and is unique to the sample matrix (pyrite and arsenopyrite in this study), the operating conditions, and the mass of calibrating matrix. Following depth-profiling of the implanted standard, the crater depths were measured across their lengths and widths using a Tencor Alpha-Step 200 profilometer. The RSF values were then determined using SIMS Instrument Control System software, version 4.0, obtained from Charles Evans and Associates (Cabri & McMahon 1995). Minimum detection-limits determined during calibration are provided in Table 2. Depth profiles were then acquired on each mineral selected for analysis, using the same analytical procedures as for the implanted standards. Counts for ¹⁹⁷Au and the reference matrix-mass from the mineral samples were obtained by depth-profiling, and the Au concentration calculated using the procedure described by Larocque *et al.* (1995a).

Direct-ion images of selected elemental- and molecular-ion species were acquired by operating the instrument in the ion-microscope mode (which is similar to optical microscopy in principle), while retaining high mass-resolution using a primary-ion beam of large diameter to illuminate the sample surface. Secondary ions are emitted from all points illuminated by the beam in a planar electrostatic field, and the lateral distribution of secondary ions is retained through the double-focusing mass spectrometer system. Following energy and mass filtering, the resulting image is recorded on a resistive anode encoder. The encoder is a pulse-counting, position-computing device capable of localizing ion signals in the image plane of the detector, as well as recording the ion intensity in the image. For the present study, acquisition times ranged from 30 to 60 s, depending

TABLE 3 Au CONTENT OF ARSENOPYRITE, AS DETERMINED BY SIMS

Grain No.	Ore Zone	Location ^a	Au (ppm) ^b
1	P2	1925	122
2	P2	1925	2942
3	P2	1925	1262
4	P2	1925	183
5	P2	1925	1357
6	P2	1925	1992
7	P2	1925	2716
8	P2	1925	5610
9	P2	1925	1113
10	P2	1925	1326
11	P2	1925	368
12	P	2451-4(EW)	1687
13	P	2451-4(EW)	4322
14	P	2451-4(EW)	1844
15	P	2451-4(EW)	639
16	P	2451-4(EW)	268
17	P	2451-4(EW)	118
18	P	2451-4(EW)	2096
19	P	2451-4(EW)	1925

^a Single location.^b ppm by weight.

on the concentration and sensitivity of the selected species. Minimum detection-limits for Au in arsenopyrite were 36 ppb and 193 ppb for pyrite. Analysis of the standard gives an accuracy of 10–15% of the reported value.

RESULTS

Pyrite and arsenopyrite from ore in the Campbell mine contain elevated concentrations of Au. Arsenopyrite has up to 0.56 wt% Au (Table 3), and pyrite has up to 168 ppm Au (Table 4). All grains contain Au at concentrations well above the minimum detection-limit established by SIMS. Grain 8, having 0.56 wt% Au, also shows a $AuLa_1$ peak by SEM-EDS. Grain 7, containing 2.715 ppm Au, lacks a clear $AuLa_1$ peak, indicating that greater than 2,700 ppm Au is required for detection of Au by SEM-EDS.

When viewed with back-scattered-electron imaging (BSE), coarse-grained crystals (>50 μ m) of arsenopyrite display compositional zoning (Fig. 1). Zoning generally



FIG. 1. Back-scattered electron image of arsenopyrite. Light grey margin is Au- and As-rich, and S-poor compared to the darker core. Gold inclusions were not observed by SEM at 43,000 $\times$ magnification. Note the sharp boundary between the core and rim. Image taken at an accelerating voltage of 25 kV, and a beam current of 3.06 nA.

is more complex in arsenopyrite from the ore veins, as compared to wallrock arsenopyrite. The boundary between core and rim is sharp and irregular. Rims are bright compared to cores, indicating that the average atomic mass of the rims is greater than that of the cores. The rim from grain 8 (Fig. 2) shows a $AuLa_1$ peak, which is absent in the core, and shows that the rims are relatively rich in Au. SIMS ^{197}Au imaging indicates that the rim of zoned grains of arsenopyrite is enriched in Au (Fig. 2), which is compatible with BSE images and the presence of the $AuLa_1$ peak. SIMS imaging also shows that small grains of arsenopyrite are uniformly enriched in Au (Fig. 2).

On the basis of similarities in electron reflectance between Au-rich rims of coarse-grained arsenopyrite and adjacent small grains of unzoned Au-rich arsenopyrite, we suggest that the deposition of fine-grained Au-rich arsenopyrite was synchronous with the growth of an Au-rich rim on the coarse grains. The composition of arsenopyrite (Table 5) indicates that the Au-rich rims are S-poor and As-rich compared to the Au-poor cores. The sum of As and S is relatively constant at ~66 atom % in all arsenopyrite grains. The arsenopyrite formula is calculated on the basis of one atom, and site allocations were based on the data of Wood & Sorens (1979). Antimony is included in the XY site, on the basis of the similarity of its geochemical properties to those of As. Sums of M cations and XY site ions are nearly ideal at ~33.3 and ~66.6 atom %, respectively.

Pyrite displays complex Au zoning, which is moderately correlated to elevated As (Fig. 3). Gold contents in pyrite are much lower than those in arsenopyrite (Table 4).

Gahnite occurs in quartz-arsenopyrite ores in the eastern and deeper portions of the Campbell mine. Gahnite is euhedral to subhedral, occurring as individual crystals

TABLE 4 Au CONTENT OF PYRITE, AS DETERMINED BY SIMS

Grain No.	Ore Zone	Location ^a	Au (ppm) ^b
1	P	2451-5	8.5
2	P	2451-5	1.5
3	P	2451-5	0.6
4	P2	1925	0.1
5	P2	1925	0.3
6	P2	1925	0.2
7	P2	1925	0.1
8	L	1834	9.8
9	L	1834	2.6
10	L	1834	1.36
11	L	1834	168
12	P	2451-4(EW)	0.2
13	P	2451-4(EW)	0.5

^a Single location.^b ppm by weight.

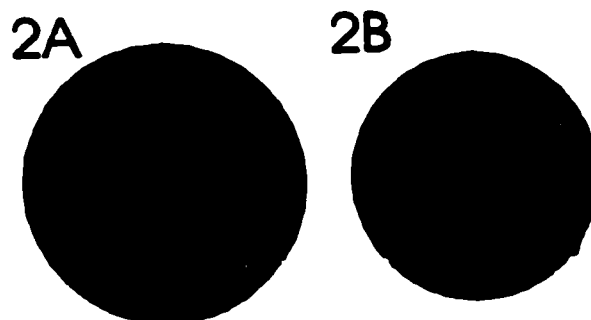


FIG. 2. SIMS direct-ion images of Au and S in arsenopyrite grains. The image of ^{34}S (Fig. 2A) outlines the arsenopyrite grains; the image in Figure 2B shows the distribution of ^{197}Au in arsenopyrite. Colored areas are relatively S and Au-rich, respectively. The large crystal of arsenopyrite in the center of the image (grain 8, Table 2) displays a Au-rich rim. The small crystal in the upper right is uniformly Au-rich. Sample from the P zone, 2451 stope. Acquisition times were 30 s for ^{34}S , 60 s for ^{197}Au . Fields of view are all 62.5 μm .

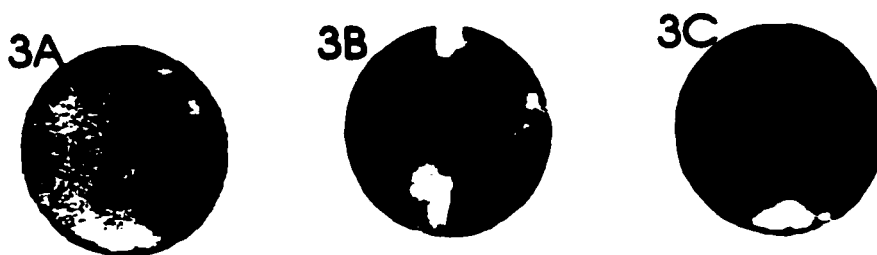


FIG. 3. SIMS direct-ion images of Au, S, and As in pyrite grains. Image 3A shows $^{57}\text{Fe}_2 + ^{32}\text{S}$, and outlines a pyrite grain that fills the field of view. Images 3B and 3C show the distribution of ^{197}Au and ^{75}As within pyrite, respectively. Yellow and red areas in the ^{197}Au and ^{75}As images are relatively Au- and As-rich, respectively. Note that the patchy zonations of Au enrichment are moderately well correlated with ^{75}As -rich areas. Sample from the L zone, 1854 stope. Acquisition times were 30 s for $^{57}\text{Fe}_2 + ^{32}\text{S}$, 45 s for ^{197}Au , 45 s for ^{75}As . Fields of view are all 62.5 μm .

TABLE 5. COMPOSITION OF ARSENOPYRITE, AS DETERMINED BY SEM-EDS

	65, rim	65, core	67, rim	67, core
Fe	36.22	36.71	35.22	36.32
Zn	0.00	0.00	0.22	0.00
Cu	0.00	0.17	0.00	0.00
Sb	0.00	0.03	0.00	1.10
As	41.10	41.37	46.37	41.67
S	22.47	23.61	20.30	22.90
Total	100.69	102.71	102.00	101.90
Formula (base 1 atom per formula unit)				
Fe	0.340	0.336	0.336	0.337
Zn	0.000	0.000	0.002	0.000
Cu	0.000	0.001	0.000	0.000
ΣM cat	0.340	0.337	0.336	0.337
Sb	0.004	0.004	0.000	0.005
As	0.288	0.282	0.327	0.288
S	0.368	0.177	0.337	0.370
ΣXY cat	0.660	0.663	0.664	0.663
Total	1.000	1.000	1.000	1.000
Au+S	0.636	0.639	0.664	0.638
Au+Fe	0.628	0.618	0.661	0.633
Au/Fe	0.047	0.039	0.970	0.855
Au/S	0.703	0.748	0.970	0.776

Counts and peak were below detection limits.
Gold not reported because of insufficient counts at 20 kV accelerating voltage. Refer to analytical procedures.

or as banded aggregates. Rosettes of euhedral arsenopyrite mantle garnite, and fine-grained euhedral to subhedral arsenopyrite forms inclusions in bands of garnite (Fig. 4). Arsenopyrite from garnite-bearing ore displays complex zoning, with a dark core and a bright, banded rim. A dark growth-band is present locally along the margin of zoned grains (Fig. 5).

Almandine occurs in altered wallrocks from the eastern portions of the deposit. Garnet growth was clearly syntectonic, as evidenced by helicoidal trails of inclusions (Fig. 6) of quartz, pyrrhotite, magnetite, and arsenopyrite. Arsenopyrite inclusions within almandine display a bright rim and a dark core in BSE images (Fig. 7). Their zoning is identical to the zoning in arsenopyrite grains outside of garnet. Arsenopyrite from the wallrocks and veins have a core that occasionally displays fractures, which are filled with arsenopyrite that is similar in brightness to that in the rim in BSE images (Fig. 7).

DISCUSSION

Invisible gold in sulfides

Invisible Au is defined as Au that occurs as a structurally bound component in a sulfide mineral, or as "colloidal" Au inclusions less than 0.1 μm in size (Boyle 1979). The presence of Au as a structural

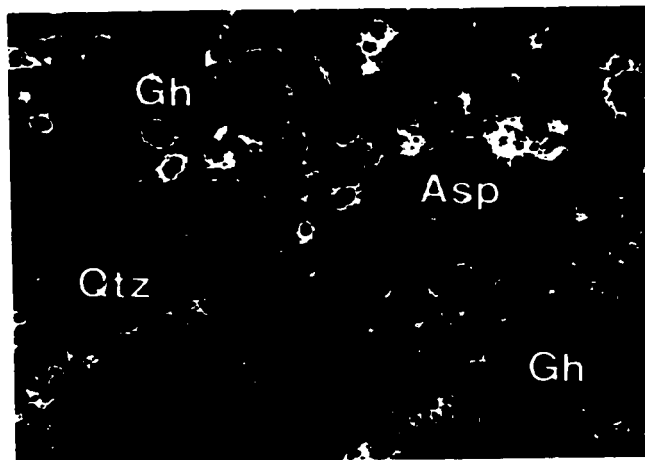


FIG. 4. Back-scattered electron image of arsenopyrite and garnite. Arsenopyrite (Asp) occurs as rosettes that envelop garnite (Gh), and also form inclusions within garnite crystals. Euhedral prisms of arsenopyrite also occur within bands of garnite in a quartz (Qtz) matrix. Quartz-arsenopyrite replacement ore from the 17th level, L zone.

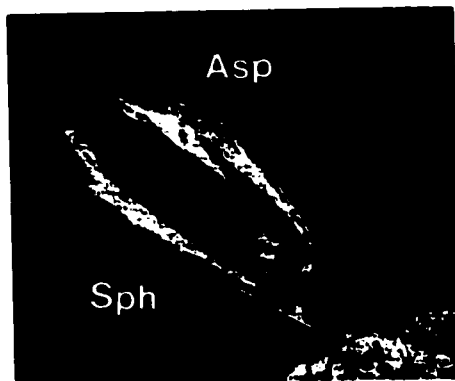


FIG. 5. Back-scattered electron image of arsenopyrite from gahnite-rich ore. Euhedral arsenopyrite (Asp) with associated subhedral to euhedral sphalerite (Sph). Arsenopyrite displays complex core-to-rim zoning, from dark, to light, with dark margins. Boundaries between zones are sharp and subparallel to grain margins. Arsenopyrite at the lower right of the image is an aggregate of fine grains, which display compositional zoning. Sphalerite also shows compositional zoning, with an Fe-poor core and Fe-rich rim. Image taken at a beam current of 3.05 nA, and an accelerating voltage of 25 kV.

component in sulfides has been documented by Wagner *et al.* (1986) and Cathelineau *et al.* (1989) using ^{197}Au Mössbauer spectroscopy, and by Zhang *et al.* (1987) using electron-spin resonance spectroscopy. Structurally bound Au can be present in arsenopyrite in amounts up to 1.5 to 1.7 wt%, although concentrations generally do not exceed 1,000 to 2,000 ppm (Johan *et al.* 1989, Cabri 1992). In many Au deposits in which arsenopyrite occurs with other Fe sulfides, arsenopyrite is the principal host for invisible Au. Pyrite can contain several hundred ppm Au (Cabri 1992), whereas pyrrhotite, chalcopyrite, and galena are relatively minor hosts for Au.

Pyrite and arsenopyrite samples from the Campbell mine were viewed with SEM at magnifications up to 43,000 times, and native Au inclusions were not observed. Gold values in pyrite grains 10 and 11 (Table 4) are high compared to other samples in this suite, but within the range of invisible Au contents reported elsewhere (Cabri 1992). Gold in pyrite and arsenopyrite from the Campbell mine is therefore present as either "colloidal" Au inclusions of less than 100 nm size, or as a structural constituent in the sulfide.

Residence of gold

The valency and residence of invisible Au in sulfides remain in question. The residence of invisible Au may involve a substitution mechanism (Johan *et al.* 1989, Wu & Delbove 1989), or presence as extremely fine "colloidal" inclusions (Cook & Chrysosoulis 1990). Pyrite and arsenopyrite contain metals in octahedral coordination with six paired anions (dianions) of Y_2 and XY , respectively (Wood & Strens 1979). In pyrite, the dianion is $(\text{S}_2)^{2-}$, and XY in arsenopyrite is $(\text{AsS})^{2-}$. This requires that Fe, which occupies the M site, be in the $2+$ state in pyrite, and in the $3+$ state in arsenopyrite, relations that are in accord with results from studies of magnetic susceptibility (Wood & Strens 1979) and crystal structure (Buerger 1936). Boyle (1979) first suggested that Au may be incorporated into arsenopyrite by substituting for As, on the basis of similarity of their ionic radii; a similar mechanism of substitution was suggested by Johan *et al.* (1989). However, imaging and analysis of sulfides for Au have shown correlations between As-rich growth bands in pyrite and arsenopyrite, and elevated concentrations of invisible Au (Cathelineau *et al.* 1989, Cabri *et al.* 1989, Archart *et al.* 1993, Fleet *et al.* 1993). Furthermore, Cathelineau *et al.* (1989) and Marcoux *et al.* (1989) showed that elevated Au in arsenopyrite is associated with low levels of Sb, S, and Fe. This suggests that Au may replace Fe in arsenopyrite, and is associated with increased As in both pyrite and arsenopyrite.

Exchange of Au for As in arsenopyrite was suggested by Johan *et al.* (1989). They showed that excess As calculated in the Fe site ($[M]$) was inversely correlated with $(\text{Au} + \text{Sb})$, and proposed that the substitution $2\text{As}[M] = (\text{Au}, \text{Sb}) + \text{Fe}$ applied to the incorporation of Au in arsenopyrite. However, as noted above, most data in the literature show an increase in As with increasing Au in auriferous arsenopyrite, and a decrease in Fe with increasing Au. These relations suggest that Au replaces Fe^{3+} in the M site in arsenopyrite. To maintain charge balance, it is likely that Au is present in the $3+$ state. Wu & Delbove (1989) recognized a near 1:1 correlation between $\text{Fe}_{\text{atomic}}$ and $\text{Au}_{\text{atomic}}$ in arsenopyrite, with a -0.79 correlation coefficient. This was interpreted to indicate the substitution mechanism $\text{Fe} = \text{Au}$. Recent work by Fleet & Mumin (1997) does not show this 1:1 correlation, but shows significant vacancies in the Fe site. They proposed that the arsenopyrite formula be expressed as $(\text{Fe}, \text{As}, \square)(\text{As}, \text{Sb})_{1.25}\text{S}_{1.25}$.

Studies of auriferous pyrite have shown a relatively constant atomic proportion of As + S at $\sim 66\%$, and a sympathetic increase in Au and As with a decrease in S (Fleet *et al.* 1989, Archart *et al.* 1993). The results suggest that As in pyrite forms dianions with S as $(\text{AsS})^{2-}$, as implied by Archart *et al.* (1993), and that Au^{3+} replaces Fe^{2+} . Incorporation of As into pyrite may cause a structural distortion, favoring an introduction of Au (Griffin *et al.* 1991). In addition, As may form $(\text{AsS})^{2-}$, which would favor Au^{3+} in the M site, to



FIG. 6. Reflected-light photomicrograph of arsenopyrite within garnet. Arsenopyrite (Asp) and magnetite (Mag) included within almandine garnet (Grt). Note the helicoidal trails of inclusions (dark grey) within garnet. Sample from drill core 27-264, beneath 27th Level.

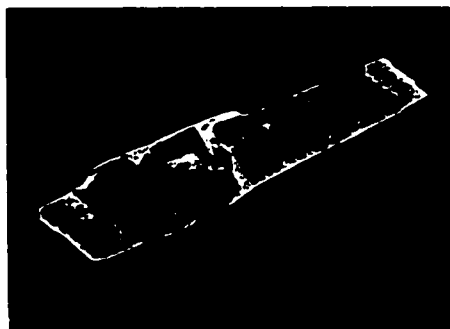


FIG. 7. Back-scattered electron image of arsenopyrite grain in garnet. Arsenopyrite with bright rim and dark core, and showing sharp boundary between rim and core. Arsenopyrite occurs as an inclusion within syntectonic almandine, and is aligned along foliation trails. Sample collected from wallrocks of the P zone.

maintain charge balance (Wood & Strens 1979, Cook & Chrysoulis 1990).

The $(As + S)_{\text{atomic}}$ in arsenopyrite from the Campbell mine is relatively constant at ~66 atom % (Table 5), and the $(As/S)_{\text{atomic}}$ ratio shows an increase in Au-rich rims. Incorporation of Au in arsenopyrite is accompanied by increasing As and decreasing S. The constant sum of As and S further implies that As replaces S, and that As may be incorporated to form the dianion $(As_2)^{4-}$. Both $(As/Fe)_{\text{atomic}}$ and $(As + Fe)_{\text{atomic}}$ increase in Au-rich rims, indicating that relative depletion in Fe occurs with Au enrichment. The data support the interpretation that Au substitutes for Fe, rather than As.

Variations in As and S, together with Au content, would be highly unlikely if Au occurs simply as very fine inclusions. Therefore, Au is likely bound as a structural component in arsenopyrite. Although Au-rich arsenopyrite rims are As-rich relative to cores, rims do not contain excess As ($As/S > 1$). Based on similar zoning between auriferous arsenopyrite and arsenopyrite enclosed in garnet and gahnite, and the fact that all arsenopyrite examined by SIMS in this study contains elevated concentrations of Au, we conclude that arsenopyrite inclusions also are auriferous.

TIMING OF HYDROTHERMAL ACTIVITY, AURIFEROUS MINERALIZATION AND METAMORPHISM

Gahnite occurs in quartz-arsenopyrite ore from the eastern portions of the deposit. Gahnite is most commonly associated with metamorphosed massive sulfide deposits and their wallrocks, and may form through the desulfidation of sphalerite during prograde amphibolite-facies metamorphism (Spry & Scott 1986). Gahnite-rich spinel also occurs in granitic pegmatite, and in aluminous metasedimentary rocks in the amphibolite facies (Dietvorst 1980). The formation of gahnite-rich spinel is favored in quartz-rich assemblages, or at relatively low temperatures within the stability field of spinel (Montel *et al.* 1986).

Gahnite grains from the Campbell mine contains inclusions of sphalerite, and also is overgrown by sphalerite. Contacts between sphalerite-pyrite and gahnite are smooth to weakly dentate, suggesting that sulfide formation was in equilibrium with gahnite. In addition, textural relations between gahnite and arsenopyrite (Fig. 4) indicate that arsenopyrite deposition spanned the period of gahnite formation and was synchronous with peak metamorphism.

Zoned arsenopyrite occurs as inclusions within almandine garnet (Fig. 6), indicating that arsenopyrite growth was pre- to syn-peak metamorphism.

This study indicates that auriferous arsenopyrite is ubiquitous in the Campbell mine. Gold zoning is present in pyrite and arsenopyrite samples from both upper-greenschist-facies and lower-amphibolite-facies domains. Zoning is more complex in arsenopyrite from the veins compared to arsenopyrite from the wallrocks. Complex zoning reflects variations in hydrothermal conditions, in fluid composition, or in both. The occurrence of sharp, anhedral boundaries between Au-rich rims and Au-poor cores indicates arsenopyrite resorption and growth. Irregular growth-zoning in synthetic arsenopyrite has been recognized by Wu & Delbove (1989), and has been interpreted to result from fluctuations in $a(\text{As})/a(\text{S})$. In BSE images, arsenopyrite grains from the wallrock occasionally display a fractured core, filled by arsenopyrite similar to that in the rim (Fig. 7). Veins and wallrocks were intensely deformed during vein emplacement, and episodic fracture and growth of arsenopyrite are anticipated.

Arsenic is transported as H_3AsO_3 in solutions that contain H_2S as the predominant sulfur species (Heinrich & Eadington 1986). The pH boundary between the fields in which H_2S and HS^- are predominant increases with increasing temperature, with HS^- favored under increasingly alkaline conditions at higher temperatures. An extrapolation of the dissociation constant of H_2S (Ellis & Giggenbach 1971) yields $\log K = -8.89$ at 450°C , indicating that HS^- is only significant under very alkaline conditions. Alteration assemblages in greenschist-facies domains at the Campbell mine contain chlorite, carbonate, and muscovite, and lack potassium feldspar, suggesting that fluid pH was near neutral to slightly alkaline. The predominant sulfur-bearing species in the auriferous

hydrothermal fluid most likely was H_2S ; As was probably transported as H_3AsO_3 . Deposition of arsenopyrite is facilitated by the breakdown of H_3AsO_3 during decreasing $f(\text{O}_2)$, or decreasing temperature (Heinrich & Eadington 1986). Arsenopyrite deposition may also result from increasing $a(\text{As})$, or through an increase in $a(\text{S}_2)$. Changes in $a(\text{S}_2)$, $f(\text{O}_2)$ and temperature likely occurred during circulation of an auriferous hydrothermal fluid, as a result of fluctuations in pressure. Phase separation due to a pressure decrease would result in short-term cooling of the fluid (Drummond & Ohmoto 1985), and increases in $f(\text{O}_2)$ and $a(\text{S}_2)$ (Bowers 1991). Griffin *et al.* (1991) have interpreted As zoning in pyrite as a consequence of brecciation of the host rocks and phase separation of the hydrothermal fluid. Periodic overpressuring and phase separation are supported by the common occurrence of brecciation in ore veins at the Campbell mine and the predominance of $\text{CO}_2 + \text{CH}_4$ inclusions in ore veins (Tamocai & Hattori 1996). This finding would account for the compositional zoning observed in the sulfides.

The application of arsenopyrite thermometry to metamorphosed sulfide deposits provides temperature estimates that generally agree with those of peak conditions of metamorphism (Sundblad *et al.* 1984, Lynch & Mengel 1995). Furthermore, arsenopyrite from metamorphosed deposits displays uniform compositions within individual grains (Sharp *et al.* 1985, Lynch & Mengel 1995). The evidence suggests that arsenopyrite easily re-equilibrates during peak metamorphism. In addition, metamorphic recrystallization of "invisible"-Au-bearing sulfides has been shown to expel Au from the pyrite (Larocque *et al.* 1995b) and arsenopyrite (Mumin *et al.* 1994) structures. Therefore, preservation of zoning in auriferous sulfides from the Campbell mine is interpreted to reflect deposition during, or after, the peak of metamorphism, as compositional zonation is unlikely to survive metamorphic recrystallization. Together with the textural relations of arsenopyrite with garnet and gahnite, the zoning in arsenopyrite suggests that auriferous hydrothermal activity was broadly synchronous with peak upper-greenschist-facies to lower-amphibolite-facies metamorphism.

CONCLUSIONS

- 1) Arsenopyrite and pyrite at the Campbell mine contain significant concentrations of "invisible" Au.
- 2) Arsenopyrite and pyrite display compositional zoning in Au, As, and S contents. Zoning indicates a peak to post-peak metamorphism introduction of Au, as zoning is not anticipated to survive upper-greenschist-facies to lower-amphibolite-facies metamorphic recrystallization.
- 3) The data suggest that pyrite and arsenopyrite incorporate "invisible" Au under upper-greenschist-facies to lower-amphibolite-facies metamorphic-hydrothermal conditions.

4) Occurrence of arsenopyrite in almandine garnet and auriferous arsenopyrite as inclusions within garnet and as rosettes mantling garnet indicate pre- to syn-peak metamorphism introduction of Au.

5) These data support a broadly syn-peak metamorphism introduction of Au at the Campbell mine.

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