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**Geology of the orogenic Cheminis gold deposit along the Larder Lake-Cadillac
deformation zone, Ontario**

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Draft

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Abstract

The Larder Lake - Cadillac deformation zone (LLCDZ) is one of two major, auriferous, deformation zones in the southern Abitibi subprovince of the Archean Superior Province. It hosts the Cheminis and the giant Kerr Addison – Chesterville deposits within a strongly deformed band of Fe-rich tholeiitic basalt and komatiite of the Larder Lake Group (ca. 2705 Ma). The latter is bounded on both sides by younger, less deformed, Timiskaming turbidites (2674 Ma - 2670 Ma). The earliest deformation features are F_1 folds affecting the Timiskaming rocks, which formed either during D_1 extensional faulting or during early D_2 north-south shortening related to the opening and closure, respectively, of the Timiskaming basin. Continued shortening during D_2 imbricated the older volcanic rocks and turbidites and produced regional F_2 folds with an axial planar S_2 cleavage. D_2 deformation was partitioned into the weaker band of volcanic rocks, producing the strong S_2 foliation, L_2 stretching lineation, and south-side-up shear sense indicators, which characterize the LLCDZ.

Gold is present in quartz-carbonate veins in deformed fuchsitic komatiites (carbonate ore) and turbiditic sandstone (sandstone-hosted ore), and in association with disseminated pyrite in altered Fe-rich tholeiitic basalts (flow ore). All host rocks underwent strong mass gains in CO_2 , S, K_2O , Ba, As, and W, during sericitization, carbonatization, and sulphidation of the host rocks, suggesting that they interacted with the same hydrothermal fluids. Textural relationships between alteration minerals and S_2 cleavage indicate that mineralization is syn-cleavage. Thus, gold was deposited as hydrothermal fluids migrated upward along the LLCDZ during contractional, D_2 south-side-up shearing. The gold zones were subsequently modified during D_3 reactivation of the LLCDZ as a dextral transcurrent fault zone.

Keywords: Orogenic gold deposit, Larder Lake – Cadillac deformation zone, Abitibi, flow ore, Cheminis mine, Kerr Addison mine, structural controls on gold, Superior Province

Introduction

Two major breaks or deformation zones are associated with gold deposits in the southern Abitibi subprovince: the Porcupine-Destor deformation zone to the north, which extends from Timmins in Ontario to Destor in Quebec, and the Larder Lake-Cadillac deformation zone (LLCDZ) to the south, which can be traced from Matachewan in Ontario to Val D'Or in Quebec (Fig. 1). The two breaks are further characterized by their spatial association with fluvial conglomerate and sandstone of the Timiskaming Group, intense carbonate alteration, and strongly foliated and lineated ultramafic rocks and Fe-rich tholeiitic basalts (Thompson 1941; Cameron 1993; Smith et al. 1993; Poulsen et al. 2000; Ispolatov et al. 2005, 2008; Dubé and Gosselin 2007). These features all played important roles in the formation of gold deposits along the breaks and their significance will be further explored here for the Cheminis and other deposits located along the LLCDZ in Ontario.

The LLCDZ is a 250 km long, curvilinear structure with an overall east-west trend. It consists of east-trending segments separated by shorter northwest- and northeast-trending segments. It is generally south-dipping west of the Cheminis mine (Fig. 1) and north-dipping from the Cheminis mine eastward to Val d'Or. From Kirkland Lake to Larder Lake (Fig. 2), it generally follows the contact between Timiskaming conglomerates and alkaline volcanic rocks (ca. 2680 Ma – ca. 2670 Ma; Ayer et al. 2005), and older volcanic rocks of the Larder Lake Group (ca. 2705 Ma; Corfu et al. 1989). East of Larder Lake, it leaves this contact to follow a narrow, 100 m – 350 m wide, band of ultramafic rocks and Fe-rich tholeiitic basalts of the Larder Lake Group, which is surrounded on both sides by deeper marine turbiditic sandstones of the Hearst assemblage (Hyde 1980). The LLCDZ can be traced up to 2.8 km west of the Ontario-Quebec provincial border before it disappears under cover rocks of the Paleoproterozoic

Huronian Supergroup only to reappear 33 km farther to the east, south of Rouyn-Noranda in Quebec.

In contrast to the Quebec segment of the LLCNZ, where deposits occur along both the main structure and higher order structures (Robert 1989), almost all deposits in Ontario occur along the main structure (Ispolatov et al. 2005, 2008). From Larder Lake to the Ontario-Quebec border, these include the giant Kerr Addison – Chesterville deposit and the smaller satellite Armistice, Barber Lake, Cheminis, Fernland and Omega deposits (Fig. 2). Excellent new exposures of the ore zones at the Cheminis mine, and a recent shallow drill program across the zones, provided excellent new sections that were used to characterize the structural controls on mineralization and the styles of alteration and mineralization along this prolific segment of the LLCNZ. Interpretation of this new data addresses several key questions on the structural evolution and timing of gold mineralization along the LLCNZ. Is the LLCNZ a transcurrent tranpression zone (Wilkinson et al. 1999), a reactivated syn-depositional extensional fault (Dimroth et al. 1982; Cameron 1993; Bleeker 2012), or a reverse, south- or north-side-up shear zone (Hodgson and Hamilton 1989; Hodgson et al. 1991; Smith et al. 1993; Ispolatov et al. 2008; Bleeker 2012; Zhang et al. 2014)? Was gold emplaced during early extension and formation of the Timiskaming basin (Cameron 1993; Bleeker 2012), during closure and shortening of the basin (Hodgson and Hamilton 1989; Cameron 1993; Ispolatov et al. 2008; Bleeker 2012), or during later dextral transcurrent movement (Neumayr et al. 2000, 2007; Bedeaux et al. 2014)? Was deep-seated alkaline magmatism the source of the gold-bearing hydrothermal fluids (Cameron 1993; Smith et al. 1993; Bleeker 2012)? These key questions on the formation of gold deposits in the southern Abitibi subprovince are discussed in the paper following a description of the host rocks, structures and ore zones at the Cheminis mine.

Analytical techniques

Geochemistry

Thirty-two samples of Fe-rich tholeiitic basalt (19), komatiite (7), syenite (3), and sandstone (3) were collected from drill core and outcrop on the Cheminis mine site. Representative analyses are listed in Table 1. All analyses were done at Activation Laboratories in Ancaster, Ontario, Canada. Lithium metaborate and tetraborate fused samples were analysed for major elements and trace elements (V, Sr, Ba) using the inductively coupled plasma emission spectrometry (ICP) technique, and for trace elements (Co, Ga, Ge, Rb, Y, Zr, Nb, Sb, Cs, Bi, Rare Earth Elements (REE), Hf, Ta, W, Tl, Th, U) using the inductively coupled plasma emission mass spectrometry (ICP-MS) technique. Precision is typically better than 5% for major elements (except K₂O) and REE and 10% for other trace elements. Base metals (Cd, Ni, Cu, Zn), Ag, and S concentrations were measured using a four acid digestion ICP technique with precision and accuracy better than 10% except for values close to detection limits. Au and trace elements (As, Cr, Sc, Sb) concentrations were determined by instrumental neutron activation analysis (INAA) with precision and accuracy better than 6%. CO₂ concentrations were measured by coulometry using the absorption of infrared energy. Their precision and accuracy are better than 2%. A subset of fourteen mineralized samples was prepared as aqua regia digestions and analyzed for Au, Ag, As, W, Zn, Cu, Pb, Bi, Te, Mo, Sb, Sn, Se using the ICP-MS technique.

Mineral chemistry

Mineral chemistries were determined using an Oxford Sight energy dispersive detector mounted on a JEOL 6400 scanning electron microscope at the Central Analytical Facility of Laurentian University. Data acquisition was done at an accelerating voltage of 15 kV, 1.005 nA beam

current, acquisition count times of 60 s, and a working distance of 15 mm. Well characterized jadeite, diopside, orthoclase, corundum, quartz, and pyrophanite were used as standards.

Rock description

Field and petrographic description

In the Cheminis area, the mafic and ultramafic rocks of the Larder Lake Group occur along the LLCDZ. They are all weakly to strongly hydrothermally altered. Least-altered mafic volcanic rocks are generally fine-grained and dark green to dark grey on outcrop surfaces. They consist of chlorite, albite, ankerite, and quartz $\pm$ rutile and pyrite. They occur as massive flows or intrusions, variolitic pillowed flows, and volcanoclastic rocks. The massive flows or intrusions are typically strongly lineated and weakly foliated. The variolitic pillowed flows consist of aphyric pillows that are strongly flattened parallel to foliation. The variolites occur along pillow margins as lighter grey to green, ellipsoidal lumps that are also flattened parallel to foliation. The volcanoclastic rocks comprise aphyric mafic clasts that are up to 9 cm in size. They are found south of the LLCDZ, where they are strongly lineated but only weakly foliated.

The ultramafic rocks invariably underwent strong hydrothermal alteration. Three types are present: black talc-chlorite, grey chlorite-carbonate, and green fuchsite-carbonate ultramafic rocks. The latter two are massive to strongly foliated with dark rusty brown, pitted, outcrop surfaces dissected by multiple cooling cracks. Chlorite-carbonate ultramafic rocks are grey to light yellow or pale brown on fresh surfaces. They consist of chlorite, albite, ferroan dolomite, magnesite, and quartz. Fuchsite-carbonate ultramafic rocks are apple green on fresh surfaces, and they have similar mineralogies as the chlorite-carbonate ultramafic rocks, except for the presence of fuchsite instead of chlorite $\pm$ apatite. Talc-chlorite ultramafic rocks are strongly foliated and

black on fresh and outcrop surfaces. They are composed of talc, chlorite, albite, ferroan dolomite, quartz and rutile. They are friable and recessive due to weathering, and are only exposed along trenches.

Timiskaming sedimentary rocks are represented by conglomerate and sandstone. The conglomerate comprises clasts of red and black chert, quartz vein, mafic volcanic rocks, reddish feldspar-phyric syenite, hornblende-phyric granitoid, and mudstone. The clasts are surrounded by a chloritic, green, sandy matrix. Conglomeratic beds are interlayered with massive sandstone beds. Turbidite sandstone of the Hearst assemblage consists of interlayered mudstone and well graded, locally trough cross-bedded, sandstone. It is locally interbedded with 10-20 cm thick, horizons of iron formation, consisting of black, laminated, magnetite-bearing layers and red jasper layers.

Timiskaming igneous rocks include trachyte flows, trachytic volcanoclastic rocks, and syenite dikes. In the Cheminis area, only the latter is present along the LLCDZ and is described herein. Syenite dikes are light grey on fresh surfaces and pinkish grey on outcrop surfaces. They are aphyric to plagioclase-phyric with a very-fine-grained to aphanitic matrix. Although the dikes occur along the LLCDZ, they are unfoliated to weakly foliated in contrast to their surrounding, more strongly foliated, mafic and ultramafic host rocks.

Geochemistry

The composition of all rock types is represented on the cation classification diagram of Jensen and Pyke (1982). The mafic rocks and ultramafic rocks plot in distinct fields (Fig. 3a). All ultramafic rocks plot in the komatiite and komatiitic basalt fields, whereas the mafic volcanic rocks plot almost entirely within the high-Fe tholeiitic basalt field. Timiskaming syenite dike

and turbiditic sandstone cluster together within the calc-alkaline basalt and andesite fields. A TiO_2 versus Zr binary diagram further illustrates the differences in their compositions (Fig. 3b). With one exception, ultramafic rocks have low TiO_2 (average of 0.40 weight %) and Zr (average of 22 ppm) concentrations. The mafic volcanic rocks have higher TiO_2 (0.8 – 2.2 weight %) and Zr (66 – 211 ppm) concentrations. The syenite dikes have low TiO_2 (0.5 – 0.8 weight %) and high Zr (150 – 190 ppm) concentrations. The turbiditic sandstones contain 0.7 – 1.0 weight % TiO_2 and 90 – 140 ppm Zr. They are intermediate in composition between the syenite dikes and the mafic and ultramafic rocks.

The trace element composition of the syenite dikes and turbiditic sandstones are further investigated on a rare-earth element (REE) diagram normalized to chondrite (Fig. 4a) and on an extended trace element diagram normalized to primitive mantle (Fig. 4b). On the REE diagram, syenite and turbiditic sandstone are both strongly enriched in light REE with $(\text{La/Yb})_N = 9 - 20$ and $5 - 6$, respectively. Their total light REE concentrations range from 20 – 200 times chondrite in syenite, and from 20 – 70 times chondrite in turbiditic sandstone. On the extended trace element diagram, syenite shows strong negative Nb, Ta, P, Ti, Sc anomalies, and turbiditic sandstone has the same but slightly less pronounced anomalies.

Structural Geology

The earliest deformation features in the Cheminis mine area are post-Timiskaming, map-scale, isoclinal, F_1 folds. The folds are located immediately east of a syenite stock in the western part of the map area (Fig. 5). These folds, which lack an axial plane cleavage, are defined by reversals in younging direction in turbiditic sandstone and conglomerate. They are overprinted by map-scale, F_2 folds with a strong axial planar S_2 cleavage, which is expressed as a sericitic

and chloritic slaty cleavage in mudstone and as a disjunctive cleavage defined by opaque minerals in sandstone. S_2 cleavage strikes east to east-northeast and dips steeply (75° - 85°) to the south or north (Fig. 5). F_2 fold axes and S_2 -bedding intersection lineations plunge 55° to 75° to the east or west (Fig. 5). Within 200 m of the LLCDZ, the folds are replaced by a south-younging, homoclinal panel of turbiditic sandstone with S_2 oriented consistently anticlockwise to bedding. North of the LLCDZ, beds remain south-younging but S_2 is oriented clockwise to bedding and therefore has an opposite angular relationship to bedding to that described in the homoclinal panel immediately south of the LLCDZ.

The LLCDZ is characterized by a strong S_2 cleavage and a down-dip, steeply-plunging (70° - 80°), L_2 lineation. Strain and alteration are heterogeneously distributed within the deformation zone. In weakly altered mafic volcanic rocks, S_2 is a wiggly disjunctive cleavage defined by opaque minerals and chlorite. Within more altered rocks, S_2 is defined by sericite and chlorite, by carbonate-rich microlithons with elongate carbonate grains, and by pyrite cubes with strain shadows infilled with carbonate, quartz, and sericite. In altered and foliated ultramafic rocks, S_2 is defined by fuchsite, talc, sericite, and chlorite, alternating with microlithons of crenulated talc and/or elongate carbonate grains. In strongly deformed conglomerate and mafic volcanoclastic rocks, S_2 is further defined by the flattening of the clasts. L_2 is usually expressed as a mineral lineation defined by alternating dark streaks of chlorite, fuchsite, and sericite and lighter streaks of carbonate. In weakly foliated to massive ultramafic rocks, L_2 is further expressed by stretched pseudo fragments bounded by deformed polygonal cooling joints or cracks.

Few D_2 shear sense indicators are present along the LLCDZ. All are observed on surfaces perpendicular to foliation and parallel to lineation and suggest south-side-up movement

up the plunge of the lineation (L_2). Shear sense indicators include shear bands, quartz fibers in asymmetrical strain shadows around pebbles in sheared conglomerate, oblique S-fabrics in S_2 -parallel chlorite bands (Fig. 6a), and S-shape fabrics defined by elongate recrystallized quartz grains in sheared quartz veins (Fig. 6b).

S_2 is overprinted by F_3 folds with an axial planar S_3 cleavage. The folds have Z-shaped asymmetry and plunge 30° - 60° to the east-northeast and west-southwest (Fig. 5). S_3 strikes 040° - 080° , dips steeply (70° - 90°) to the south or north (Fig. 5), and is typically oriented anticlockwise to bedding and S_2 . It is expressed as a chloritic and sericitic slaty cleavage in mudstone, as a disjunctive cleavage in interlayered sandstone, and as a crenulation cleavage overprinting S_2 in strongly foliated ultramafic rocks along the LLCDZ. Late dextral shear bands, which overprint S_2 along the LLCDZ, likely formed during the same deformation event as F_3 and S_3 .

Mineralization

In 1937 and 1938, Consolidated Mining and Smelting Company of Canada discovered the Cheminis deposit through diamond drilling (Thomson 1941). A shaft was sunk late in 1938 and mining was completed to the 525-foot level (Thomson 1941). The mine closed in 1940, reopened in 1947, and closed again after deepening the shaft to 1,085 feet and developing the 1035-foot mine level. From 1991 – 1996, the mine was briefly in operation and produced 17,530 ounces of gold from 179,013 tons of ore. In 2004 – 2006, extensive drilling programs were completed, including 2541 m from 35 shallow drill holes. The samples described herein come from these drill holes.

Ore zones occur along the LLCDZ where they are spatially associated with a south panel of ultramafic rocks with slivers of Fe-rich tholeiitic basalt, and a north panel of ultramafic rocks (Fig. 7). The two panels are separated by strongly lineated but otherwise texturally massive Fe-rich tholeiitic basalt and are bounded to the south and north by turbiditic sandstone (Fig. 7). The south panel hosts the A, B, and C flow ore zones (Figs. 7, 8). The latter are within slivers of Fe-rich tholeiitic basalt surrounded by green fuchsite-carbonate ultramafic rock, in turn flanked by an inner shell of grey chlorite-carbonate ultramafic rock and an outer shell of black talc-chlorite ultramafic rock (Fig. 7). Immediately south of the panel, minor mineralized zones, the South and North Sediment gold zones, are hosted by turbiditic sandstone along and within 20 m of their contact with the ultramafic rocks, respectively. The two zones, which lack surface expression, are shown on mine level plans (Fig. 8). The north panel hosts the No. 1 and No. 2 North Carbonate gold zones, which also lack surface expression and are shown in Figure 8.

Flow ore

The A and C flow ore zones are exposed at surface as 5-8 m thick, S_2 -parallel, slivers, which can be traced along strike for 75 m and 40 m, respectively. The zones are moderately to strongly S_2 foliated and L_2 lineated. S_2 is defined by sericite foliation planes interleaved with ankerite microdomains, and by Fe-Ti oxide and non-reflective opaque material along disjunctive foliation planes. The ore zones typically contain up to 15% pyrite surrounded by a fine-grained, dark to light grey, matrix. Few quartz-carbonate veins cut across S_2 .

Unmineralized and mineralized high-Fe tholeiitic basalts have similar mineralogy. Mineralized high-Fe tholeiitic basalt consists of ankerite, sericite, chlorite, albite, pyrite, quartz, Fe-Ti oxide and rutile. Mineralized basalt has undergone more extensive ankerite alteration (up

to 65% of the rock) than the unmineralized high-Fe tholeiitic basalt (“least-altered precursor”), and sericite has largely replaced chlorite, which is a major fabric-forming mineral in the least-altered precursor. Ankerite, sericite and chlorite differ in their mineral chemistry and are more Mg-rich (higher Mg numbers) in mineralized basalt relative to their least-altered precursor (Table 2). Following Grant (1986), gains and losses in major oxides and trace elements during mineralization are represented on a % mass change histogram that compares the composition of the mineralized basalt to that of its least-altered precursor (Fig. 9). Mineralized basalt underwent strong mass gains in K₂O, Rb, Ba, CaO, MgO, CO₂, S, As, and significant losses in SiO₂ and Na₂O (Fig. 9a). Binary molar diagrams of CaO/Al₂O₃, MgO /Al₂O₃, CO₂/Al₂O₃ versus K₂O/Al₂O₃ show strong positive correlations between K₂O, MgO, CaO, and CO₂ (Fig. 10). This reflects the strong ankerite and sericite alteration of the host rock and the increase in Mg# of ankerite, chlorite, and sericite in the mineralized basalt relative to the unmineralized basalt.

Pyrite is the dominant sulphide mineral in the ore zones. It occurs as small grains (< 1 mm in size) localized along S₂ and as large, euhedral, pyrite cubes (up to 5mm in size) that truncate S₂ (Fig. 11a) and are surrounded by S₂. Sericite and ankerite occur in strain shadows around pyrite cubes (Fig. 11b) and as inclusions in pyrite (Fig. 11c). The two textural types of pyrite lack Ni and Co but contain similar As concentrations (up to 0.43 atomic %) in both core and rim. Chalcopyrite, arsenopyrite and sphalerite are present as inclusions in pyrite and as disseminations in the matrix of the rock.

Gold is associated with both types of pyrite. Gold grains with Au/Ag ratios of 5 to 12 occur as inclusions in pyrite (Fig. 11d) and along pyrite grain boundaries (Fig. 11e). A later generation of gold grains with lower Au/Ag ratio of 4 to 5 occur with covellite in late fractures cutting across pyrite grains (Fig. 11f).

Ultramafic-hosted mineralization

Sulphide minerals constitute less than 2-3% of the ultramafic rocks. In addition to pyrite, barite and other sulphide minerals, such as cobaltite, chalcopyrite, siegenite, gersdorffite, ullmannite and millerite, occur as disseminated grains in the matrix of the rock. The green fuchsite-carbonate ultramafic rock typically contains more sulphide minerals than the talc-chlorite and chlorite-carbonate ultramafic rocks. This is manifested by strong mass gains in S and metals (As, W, Sb), as well as K_2O , Na_2O and Ba, in the fuchsite-carbonate ultramafic rock relative to that of the other two types of ultramafic rocks (Fig. 9b).

In contrast to the flow ore, multiple vein sets are present in the ultramafic rocks. The earliest vein set consists of grey quartz-carbonate veins that are strongly folded, boudinaged and transposed parallel to S_2 . The veins are a few cms to less than 1 cm thick and they commonly have fuchsitic selvages. They are zoned with blocky carbonate grains along the vein walls and anhedral quartz grains in the center of the veins. Drill core intervals with an abundance of those veins are typically enriched in gold. A second vein set consists of breccia veins with strongly carbonatized fragments of the host ultramafic rocks. One of these veins is 1 m thick and is exposed over 5 m along strike of S_2 in the southern ultramafic panel. A down-dip lineation is present along the vein wall and an oblique shape fabric, visible optically only, is defined by recrystallized, elongate, small quartz grains, suggesting reverse, south-side-up movement parallel to the vein wall. A third vein set consists of shallowly-dipping, fibrous to massive, quartz-carbonate veins cutting across S_2 . The veins are folded by Z-shaped F_3 folds suggesting that they were emplaced after D_2 but before D_3 . A swarm of those veins cut across a syenite dike hosted

by ultramafic rocks, east of the A zone. The dike strikes parallel to S_2 and is weakly foliated parallel to S_2 . It is one of only three syenite dikes found in the Cheminis mine area.

Sandstone-hosted mineralization

Sandstone-hosted mineralization consists of fibrous quartz-carbonate veins surrounded by an alteration haloe of ankerite, sericite, arsenopyrite and pyrite. The host least-altered sandstone consists of fine-grained chlorite, biotite, sericite, carbonate, quartz, albite and rutile. Alteration haloes surrounding the veins are marked by increases in ankerite abundance, the disappearance of chlorite, the coarsening of sericite, and the addition of up to 5% arsenopyrite and pyrite. Arsenopyrite is the most abundant sulphide mineral followed by pyrite and then disseminated chalcopyrite, tetrahedrite and gersdorffite. Chalcopyrite occupies fractures in arsenopyrite and is also present as small irregular grains along arsenopyrite grain boundaries. The mineralogical changes from unmineralized to mineralized sandstones correlate with variable mass gains in CaO and CO₂, and strong mass gains in S, As, W, and Sb (Fig. 9c). Only one gold grain was found. It has a gold to silver ratio of 5, and it fills intersecting fractures or cleavage planes within arsenopyrite.

Interpretation and Discussion

Observations and data presented in this study provide new insights on the origin of the turbiditic sandstone, the structural evolution of the LLCDZ, and the formation of gold deposits along the LLCDZ, as discussed below.

Origin of turbiditic sandstone

From Larder Lake to Virginiatown, the LLCZ occurs along a band of mafic and ultramafic volcanic rocks, which is bounded to the north and south by turbiditic sandstone. The origin of the turbiditic sandstone is controversial (Ayer et al. 2005). Hyde (1980) interpreted these rocks as resedimented submarine fan turbidites equivalent in age to the Timiskaming fluvial conglomerate and sandstone. Alternatively, Jensen and Langford (1985), Corfu et al. (1991), and Mueller et al. (1994) suggested that the turbidite sandstones represent an older sedimentary sequence similar to the Porcupine assemblage in the Timmins area, where it is unconformably overlain by the Timiskaming Group.

The whole rock geochemistry of the turbiditic sandstone reflects the origin of the rock. On a TiO_2 versus Zr binary diagram, turbiditic sandstone is intermediate in composition to the syenite and the mafic and ultramafic volcanic rocks. It shows similar but weaker light REE enrichment, and the same but less pronounced negative anomalies in Nb, Ta, P, Ti, Sc, as those of the syenite dike. This suggests that the turbiditic sandstone is derived from the erosion of Timiskaming alkaline rocks and older mafic and ultramafic volcanic rocks similar to the Larder Lake Group. This interpretation is consistent with the detrital zircon age of these rocks. Samples of turbiditic sandstone, which were collected south of Highway #66 near the main entrance of the Cheminis mine (detailed outcrop map in Fig. 4 of Wilkinson et al. 1999) yielded Timiskaming ages of 2674 ± 3.7 Ma and 2684.9 ± 1.9 Ma (Ayer et al. 2002). Thus, as argued by Ayer et al. (2005), the turbiditic sandstone represents a deeper marine facies of the Timiskaming assemblage as initially proposed by Hyde (1980) rather than an older Porcupine-like sedimentary succession as later suggested by Jensen and Langford (1985), Corfu et al. (1991), and Mueller et al. (1994).

Structural evolution of the LLCDZ

The LLCDZ formed as deformation during the D₂ event became localized along the contact between the Timiskaming assemblage and the Larder Lake Group, west of Larder Lake, and along the band of Larder Lake volcanic rocks, east of Larder Lake. During the D₂ deformation, the Timiskaming rocks and Larder Lake volcanic rocks were tilted, imbricated, and overprinted by F₂ folds and axial planar, regional, S₂ cleavage. With continued shortening during D₂, the simple shear component of the deformation was partitioned within the Larder Lake volcanic rocks and along their contact with the Timiskaming rocks, resulting in the formation of the pronounced S₂ cleavage and steeply-plunging L₂ lineation that characterize the LLCDZ. The D₂ deformation is constrained in age between 2672 ± 2 Ma, the U-Pb zircon crystallization age of the syenitic Murdoch Creek pluton, and 2665 ± 4 Ma, the U-Pb age of syn-S₃ titanite (Wilkinson et al. 1999).

In the Cheminis mine area, the LLCDZ and Larder Lake volcanic rocks are bounded on both sides by Timiskaming turbiditic sandstone with the same south-younging direction but opposite S₂-bedding angular relationships. This geometry can be explained by fault repetition of the south-younging limb of a doubly-plunging F₂ fold. F₂ folds plunge both to the east and west (Fig. 5), therefore it is plausible that a regional F₂ fold, now dismembered by faults, had a doubly-plunging fold axis. The reversal in the plunge of the fold could have been caused either by overprinting of the fold by a younger fold generation, or by rotation of its fold axis during stretching parallel to the steeply-plunging L₂ lineation. In either case, its axial planar S₂ cleavage would have changed in relative angular orientation from anticlockwise, to parallel, to clockwise to bedding along the same fold limb. If the south-younging limb of the fold were then to be

repeated by limb- or bedding-parallel faults, then the relative orientation of S_2 to bedding would change across the LLCDZ without reversal in the younging direction, as observed in the Cheminis mine area. This would further explain the imbrication of the Timiskaming turbiditic sandstone and Larder Lake volcanic rocks, prior to the partitioning of the deformation along the band of Larder Lake volcanic rocks. The early map-scale F_1 folds in the western part of the study area (Fig. 5) could be coeval with the imbrication of the turbiditic sandstone and volcanic rocks (Pilote et al. 2014), or they may have formed during an older D_1 deformation event associated with extensional faulting during formation of the Timiskaming basin (Bleeker 2012).

The shallowly-dipping, non-auriferous, quartz-carbonate veins, which overprint S_2 , mark a hiatus between the D_2 and D_3 deformations. The veins are typically straight and undeformed except where folded by Z-shaped F_3 folds. Thus, they were emplaced after the D_2 deformation and before the late D_3 dextral transcurrent reactivation of the LLCDZ.

Wilkinson et al. (1999) proposed a different structural interpretation for the LLCDZ. They interpreted the bends or changes in trend along the LLCDZ as early features predating the D_2 deformation. Thus, they suggested that north-south shortening during D_2 caused sinistral transpression along NE-trending segments, dextral transpression along SE-trending segments, and only contraction (or compression) along E-trending segments. The slip directions along the NE- and SE-trending segments were interpreted to be sub-horizontal and therefore oblique to the stretching lineation as modelled and observed in triclinic transpression zones with a dominant transcurrent component (Sanderson and Marchini 1984; Robin and Cruden 1993; Lin et al. 1998). However, along all segments of the LLCDZ from the Young-Davidson mine in Matachewan (Zhang et al. 2014), to Kirkland Lake (Thompson et al. 1950), to the McBean mine in Dobie (Ispolatov et al. 2005), the Cheminis mine in Larder Lake (this study), and the Kerr-

Addison mine in Virginiatown (Smith et al. 1993), the slip direction is south-side-up parallel to the stretching lineation. It is not oblique or perpendicular to the stretching lineation, as observed in triclinic transpression zones or transcurrent transpression zones, respectively (Robin and Cruden 1993; Lin et al. 1998). Furthermore, the attitude of the S_2 cleavage mimics the changes in orientation of the LLCDZ. E-trending segments are neither offset along the sinistral NE-trending segments nor are they offset along the dextral SE-trending segments. This suggests that the bends along the LLCDZ do not represent early pre-D2 feature as proposed by Wilkinson et al. (1999). They are late features that either formed during F_3 folding or during north-trending open folding, as suggested by Ispolatov et al. (2005, 2008) for the Dobie area, west of Larder Lake. The LLCDZ has a more pronounced S_2 cleavage and L_2 stretching lineation than the turbiditic sandstones surrounding it, suggesting that it underwent shortening and extension perpendicular and parallel to its boundaries, respectively. Thus, the LLCDZ is transpressional, but it is not a transcurrent transpression zone or triclinic transpression zone with a dominant transcurrent character as proposed by Wilkinson et al. (1999). Rather, it formed as a reverse, south-side-up, shear zone, which underwent shear during shortening perpendicular to its boundaries and stretching parallel to the slip direction.

Gold deposits along the LLCDZ

The three styles of gold mineralization at the Cheminis mine reflect the composition of the host rocks. Mineral chemistries differ amongst host rocks, but strongly altered rocks have the same simple mineralogy regardless of their initial composition. All consists of white mica (sericite or fuchsite), carbonate (ankerite or ferroan dolomite), chlorite, quartz, and albite (Table 2), suggesting high hydrothermal fluid to rock ratios (Kishida and Kerrich 1987). All underwent

significant mass gains in CO_2 , S, K_2O , Ba, As, W (Fig. 9). Thus, the alteration and mineralization of the host rocks, and the formation of the three styles of gold mineralization, occurred under the same hydrothermal fluid system.

The relative chronology between mineralization and deformation can be determined from textural relationships. In fuchsite-carbonate ultramafic rock, gold-rich drill core intervals are characterized by strongly deformed quartz-carbonate veins, which are folded and transposed parallel to S_2 . This suggests that mineralization was emplaced either pre- or syn- S_2 cleavage. In Fe-rich tholeiitic basalts, S_2 wraps around pyrite grains that are surrounded by strain shadows filled by alteration minerals, whereas other pyrite grains truncate S_2 and contain inclusions of sericite and ankerite. Pyrite growth therefore both predates and postdates the formation of the foliation and the crystallization of other alteration minerals. Because gold is associated with pyrite as inclusions and as discrete grains along pyrite grain boundaries (Fig. 11d,e), mineralization is therefore interpreted as a prolonged event that spanned the development of the S_2 cleavage along the LLCZ (Lafrance 2015).

The Kerr-Addison mine is located 6 km to the east and along strike of the Cheminis mine (Fig. 2). It is hosted by the same altered komatiite and Fe-rich tholeiitic basalt. Hydrothermal alteration is characterized by major gains in CO_2 , variable gains in Na_2O , K_2O , CaO, and strong additions in metals including As, Sb, W (Kishida and Kerrich 1987). These mass changes occurred during the alteration of komatiite to green fuchsite-carbonate schist (carbonate ore) and during the alteration of Fe-rich tholeiitic basalt to a bleached and foliated carbonate-chlorite-quartz-sericite-albite rock (flow ore), containing gold associated with disseminated pyrite (Thompson 1941; Kishida and Kerrich 1987; Smith et al. 1993). These mineralogical and geochemical changes are very similar to those observed at the Cheminis mine, and thus strongly

suggest that the two deposits have the same mineralization and alteration history. The Kerr-Addison mine has two other ore types: graphitic ore and albitite dike ore (Smith et al. 1993). The graphitic ore is hosted in sheared graphitic metasedimentary horizons occurring in Fe-rich tholeiitic basalt. It also occurs along the late, post-ore, Kerr Fault as entrained fragments of mineralized flow ore and broken quartz veins surrounded by a graphitic fault matrix (Smith et al. 1993). The albitite dike ore is hosted by a swarm of mafic albitite dikes. Gold is associated with pyrite, which is present along fractures and in quartz-carbonate veins that cut across the dikes (Smith et al. 1993). The dikes are locally strongly deformed, but they also overprint the foliation in the green fuchsite-carbonate schist and contain mineralized vein fragments of quartz-pyrite and carbonate-pyrite, suggesting that they were emplaced as intramineral dikes during deformation and mineralization (Smith et al. 1993). The swarm of albitite dikes is thought to be associated with a deeper parent intrusive body that provided magmatic hydrothermal fluids to the ore-forming hydrothermal system (Smith et al. 1993).

On an extended trace element diagram normalized to primitive mantle, syenite dikes at Cheminis show strong enrichment in the more incompatible trace elements and pronounced negative Nb, Ta, P, Ti, Sc anomalies (Fig. 12a). Their trace element patterns overlap with those of syenite dikes from Kirkland Lake (Fig. 12a) and albitite dikes from Kerr Addison (Fig. 12b). Because the dikes at Cheminis are similar in composition to the albitite dikes at Kerr Addison, their presence suggests that hydrothermal magmatic fluids may also have contributed to the formation of the Cheminis deposit. The dikes have a weak foliation which is parallel to the S_2 cleavage in the surrounding incompetent ultramafic rocks, suggesting that they were emplaced either before or during the formation of the cleavage. If the dikes were emplaced during the formation of the cleavage, then magmatic hydrothermal fluids may have contributed to the

auriferous hydrothermal fluid system because textural relationships indicate that gold was emplaced during the formation of the cleavage. However, as only three syenite dikes have been mapped in the Cheminis mine area, magmatic hydrothermal fluids, if any, were only a minor component of the hydrothermal fluid system.

Hamilton and Hodgson (1984), Hodgson and Hamilton (1989), Hodgson et al. (1991), and Smith et al. (1993) suggested that gold at Kerr Addison was emplaced along the LLCDZ during the development of a late cleavage. Although they refer to this cleavage as S_2 , it overprints an older cleavage, and it therefore correlates with the overprinting S_3 cleavage at Cheminis. This is later during the evolution of the LLCDZ than reported here for the Cheminis mine, where gold is interpreted to have been emplaced during the development of the S_2 cleavage. Thus, if their interpretation is correct, then gold could have been deposited over a period spanning both D_2 and D_3 deformation events. However, from east to west across Ontario, all deposits plunge parallel to the L_2 stretching lineation, including the McBean, Anoki, Upper Canada, as well as the Cheminis and Kerr Addison – Chesterville deposits (Wilkinson et al. 1999; Ispolatov et al. 2008), suggesting that mineralization was emplaced early during the D_2 evolution of the LLCDZ. Further evidence of the early timing of gold mineralization is the presence at Kerr Addison of komatiites with undeformed spinifex textures that are completely altered to fuchsite and carbonate (Kishida and Kerrich 1987; Smith et al. 1993). An example of the transition from the latter (Fig. 13a) to fuchsite-carbonate schist (Fig. 13b) can be seen in drill core. This suggests that hydrothermal alteration occurred early and locally predated ductile deformation along the LLCDZ.

Early hydrothermal alteration and coeval emplacement of gold during crustal extension and deposition of the Timiskaming Group in an extensional basin has been postulated by

Cameron (1993) and more recently invoked by Bleeker (2012). The LLCDZ may have originated as an early syn-depositional normal fault constraining the Timiskaming basin to the south (Dimroth et al. 1982; Cameron 1993; Mueller et al. 1994; Bleeker 2012). During crustal extension, upwelling followed by decompression and partial melting of the mantle produced alkaline magmas that were emplaced in the basin as flows, volcanoclastic rocks and intrusions (Cameron 1993). The alkaline magma migrated along the basin-bounding LLCDZ, which may also have acted as a conduit for the upward migration of gold-bearing hydrothermal fluids (Cameron 1993, Bleeker 2012). Early hydrothermal alteration of the spinifex-textured komatiite and the formation of F_1 folds possibly occurred during this early extensional event. A link between crustal extension, the deposition of clastic sediments, and gold mineralization has also been suggested for the Red Lake greenstone belt in the western Superior Province (Calvert et al. 2004). During subsequent basin inversion and shortening, the LLCDZ was reactivated as a reverse fault (Hodgson and Hamilton 1989; Cameron 1993; Bleeker 2012). The movement along the fault is generally thought to be south-over-north in the Ontario segments of the fault (Thompson et al. 1950; Hodgson et al. 1991; Cameron 1993; Smith et al. 1993; Ispolatov et al. 2005, 2008; Bleeker 2012; Zhang et al. 2014; this study) and north-over-south in the Quebec segments of the fault (Camiré and Burg 1993; Daigneault et al. 2002). The gold deposits along the LLCDZ, including the Cheminis and Kerr Addison – Chesterville deposits, formed during this contractional D_2 event as hydrothermal fluids were channelized and migrated upward along the fault (Card et al. 1989; Hodgson and Hamilton 1989; Cameron 1993; this study). The LLCDZ was subsequently reactivated as a late dextral transcurrent fault (Hamilton and Hodgson 1984; Toogood and Hodgson 1985; Robert 1989; Smith et al. 1993; this study) but with few

exceptions (Neumayr et al. 2000, 2007; Bédeaux et al. 2014), this last movement on the fault does not appear to have contributed significantly to its gold endowment.

Conclusions

From Larder Lake to the Ontario-Quebec provincial border, the LLCDZ formed by partitioning of the D₂ deformation along a band of mafic and ultramafic rocks of the Larder Lake Group, following regional folding of the Timiskaming turbidites (< ca.2674 Ma) and imbrication of these rocks with the older Larder Lake volcanic rocks (ca. 2705 Ma). The D₂ deformation (ca. 2672 Ma – ca. 2665 Ma) produced the strong S₂ foliation and L₂ stretching lineation that characterize the LLCDZ.

Gold was emplaced during D₂ south-side-up ductile shear along the LLCDZ. The styles of gold mineralization at the Cheminis mine differ according to their host rocks, but they have similar alteration mineralogy (sericite/fuchsite, carbonate, chlorite, albite, pyrite, arsenopyrite) and underwent similar mass gains in CO₂, S, K₂O, As, W ± CaO, Na₂O, Sb, suggesting that they interacted with the same hydrothermal fluids. Hydrothermal alteration minerals have higher Mg number in the mineralized host rocks than in their least altered precursors. Thus, detailed mapping of alteration mineral assemblages and their mineral chemistries could be an important vectoring tool to mineralization. Mineral exploration should focus on segments of the LLCDZ with pronounced S₂ and L₂ fabrics and well developed south-side-up shear sense indicators, because these segments underwent higher strain and were likely more permeable to upward migrating hydrothermal fluids.

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Draft

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Figure captions

Figure 1: Simplified geological map of the Southern Abitibi subprovince modified after Poulsen et al. (2000) showing the location of gold deposits along the Destor-Porcupine deformation zone (DPDZ) and the Larder Lake - Cadillac deformation zone (LLCDZ).

Figure 2: Simplified geological map of Larder Lake - Cadillac deformation zone from Kirkland Lake to the Ontario – Quebec provincial border. Modified after Ispolatov et al. (2008).

Figure 3: (a) Ternary Al-Mg-Fe_{tot}+Ti diagram of Jensen and Pyke (1982). (b) Binary TiO₂ vs Zr diagram. B, basalt; A, andesite; D, dacite; R, rhyolite; FeB, High-Fe tholeiite basalt; MgB, High-Mg tholeiite basalt.

Figure 4: Trace element plots of syenite dike and sandstone at Cheminis, using normalizing chondrite and primitive mantle values of Sun and McDonough (1989): (a) Rare earth element plot normalized to chondrite. (b) Extended trace element plot normalized to primitive mantle

Figure 5: Geological map of the Cadillac – Larder Lake deformation zone from the Fernland mine to the Cheminis mine. UTM coordinates in NAD83. Contours on stereonet plots are in multiple of random distribution. Number of measurements on the bottom right hand side of stereonet plots.

Figure 6: Photomicrographs of shear sense indicators along the Cadillac – Larder Lake deformation zone. (a) Carbonate in strain shadows around pyrite grains (indicated by arrows) and anticlockwise change in orientation of a S-fabric defined by chlorite, suggesting south-side-up movement. Natural light. (b) Steady-state S-fabric in a quartz vein suggesting south-side-up movement. Crossed polars.

Figure 7: Geological map of the ore zones at the Cheminis mine area. UTM coordinates in NAD83.

Figure 8: Underground, 275 feet, 525 feet, 865 feet map level plans of the Cheminis mine.

Figure 9: Mass change histograms of: (a) Flow ore mineralized Fe-rich tholeiitic basalt relative to least altered precursor. (b) Carbonate ore fuchsite-carbonate ultramafic rock relative to talc-chlorite ultramafic rock. (c) Mineralized sandstone relative to least altered precursor.

Figure 10: CaO /Al₂O₃, CO₂/Al₂O₃, MgO/Al₂O₃ versus K₂O/Al₂O₃ binary molar diagrams for flow ore, mineralized Fe-rich tholeiitic basalt

Figure 11: Flow ore mineralized Fe-rich tholeiitic basalt along the Cadillac-Larder Lake deformation zone: (a) Drill core photograph of pyrite grains overgrowing S₂ foliation (scale bar in mm). (b) Natural light photomicrograph of S₂ foliation truncated by a pyrite grain with quartz-filled strain shadows. (c) Crossed polar photomicrograph of a pyrite grain with carbonate inclusions and quartz-carbonate in strain shadow. (d) Backscattered

scanning electron microscope image of a pyrite grain with gold inclusions. (e) Backscattered scanning electron microscope image of a gold grain along pyrite grain boundary. (f) Backscattered scanning electron microscope image of gold and covellite in a late fracture cutting across pyrite grain.

Figure 12: Extended trace element diagrams normalized to primitive mantle comparing the trace element composition of syenite dikes at Cheminis to that of (a) syenite porphyry and mafic syenite at Kirkland Lake after Ispolatov et al. (2008), and (b) albitite dike at the Kerr-Addison mine after Smith et al. (1993).

Figure 13: Drill core photographs of (a) altered, undeformed, spinifex-textured komatiite cut by a quartz-carbonate vein with fuchsitic selvages at the Kerr Addison mine, and (b) altered, deformed, fuchsite-carbonate ultramafic rock at the Kerr Addison mine. The drill cores are from collections of the Ontario Geological Survey regional office in Kirkland Lake.

Table 1: Representative whole rock geochemical analyses

Table 2: Representative SEM mineral analyses

Figure 1: Lafrance

Figure 2: Lafrance

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Figure 4: Lafrance

Figure 5: Lafrance

Figure 6: Lafrance

Legend**Timiskaming Group**
Syenite dike

Turbidites
Larder Lake Group
Fe-rich tholeiite basalt

Green fuchsite-carbonate ultramafic rocks

Grey albite-chlorite ultramafic rocks

Black talc chlorite ultramafic rocks

Flow ore

Bedding: upright, top unknown

S2, S3

L2, L3

Drill hole

Bedding form surface
Figure 7: Lafrance

Figure 8: Lafrance

Figure 9: Lafrance

Figure 10: Lafrance

Figure 11: Lafrance

Figure 12: Lafrance

Figure 13: Lafrance

Table 1: Concentration data for major elements (wt.%) and trace elements (ppm) for representative samples

Sample No.		U-Unmin	U-310	U-321	2004-3	N-35-63	N-8-33	N-8-6	N-11-17	05-KL-9B	N-8-17.5	N-12-25
Rock Type		S-LA	S-M	S-M	SYE	GreenU	GrayU	BlackU	B-LA	B-M	B-M	B-M
Major oxides Method												
SiO ₂	FUS-ICP	57.59	49.39	53.05	52.26	31.98	32.67	41.27	53.08	52.92	22.92	23.09
Al ₂ O ₃	FUS-ICP	16.18	16.43	11.10	12.44	5.14	5.36	6.09	13.72	10.99	13.1	14.53
Fe ₂ O ₃	FUS-ICP	9.31	10.43	8.21	7.16	8.32	8.74	9.66	16.69	8.37	16.44	13.7
MnO	FUS-ICP	0.05	0.04	0.09	0.13	0.18	0.22	0.091	0.15	0.29	0.248	0.291
MgO	FUS-ICP	4.18	2.01	3.20	4.24	15.77	15.37	24.79	4.06	3.23	5.42	5.69
CaO	FUS-ICP	1.15	2.13	4.49	5.63	10.58	13.69	4.47	1.64	5.41	10.59	12.74
Na ₂ O	FUS-ICP	2.66	2.32	2.97	6.27	0.25	0.02	0.05	2.94	5.11	0.85	1.26
K ₂ O	FUS-ICP	2.77	3.59	1.19	2.67	1.27	0.03	0.03	0.70	2.08	3.65	3.99
TiO ₂	FUS-ICP	0.88	1.00	0.65	0.49	0.30	0.31	0.333	2.13	1.51	1.771	2.145
P ₂ O ₅	FUS-ICP	0.17	0.15	0.02	0.28	< 0.01	0.02	0.03	0.20	0.03	< 0.01	< 0.01
LOI	FUS-ICP	3.78	7.90	11.13	9.10	25.95	21.51	9.45	4.21	10.52	15.13	17.06
Total		98.72	95.39	96.09	100.70	99.76	97.94	96.25	99.51	100.50	90.08	94.48
CO ₂	COUL	1.20	3.09	7.41	7.94	27.2	21.70	7.03	7.44	11.9	17.5	20.4
S	TD-ICP	0.10	6.45	3.38	2.58	0.94	0.15	0.134	0.84	1.04	8.42	5.29
Trace elements Method												
V	FUS-ICP	200	204	139	101	111	108	117	483	373	435	497
Sr	FUS-ICP	100	164	217	207	306	91	86	81	63	144	130
Ba	FUS-ICP	355	512	175	115	330	8	10	158	109	648	553
Co	FUS-MS	36	43	31	14	64	107	98	79	23	47	54
Ga	FUS-MS	22	21	13	14	7	7	6	21	15	23	24
Ge	FUS-MS	1.9	1.7	1.5	0.8	0.8	0.6	1.5	1.1	0.9	0.9	0.9
Rb	FUS-MS	54	110	37	2	35	< 1	< 1	23	17	65	61
Y	FUS-MS	24.4	22.1	14.1	17.1	6.6	7.4	7.0	39.5	38.2	36.1	37.8
Zr	FUS-MS	139	130	95	154	18	18	17	111	111	101	127
Nb	FUS-MS	6.1	6.0	4.1	7.7	0.8	0.8	0.7	5.6	5.6	4.6	5.8
Sb	FUS-MS	12.1	42.7	55.4	2.4	60.7	23.8	2.7	41.6	2.9	21.7	12.8
Cs	FUS-MS	2.3	3.3	1.1	0.2	1.2	0.2	0.5	0.4	0.3	1.4	1.0
Bi	FUS-MS	0.3	1.2	0.9	< 0.1	< 0.1	< 0.1	0.1	0.1	< 0.1	0.4	0.2
La	FUS-MS	16.50	15.60	12.80	33.70	0.89	0.51	1.12	6.72	5.81	5.18	6.35
Ce	FUS-MS	36.50	35.50	27.80	71.50	2.51	1.59	2.91	18.50	16.10	14.50	17.1
Pr	FUS-MS	4.71	4.65	3.56	8.99	0.41	0.29	0.46	2.96	2.59	2.38	2.8
Nd	FUS-MS	17.30	17.30	12.90	31.50	2.12	1.70	2.27	14.10	12.40	12.30	13
Sm	FUS-MS	3.95	3.82	2.80	6.08	0.68	0.59	0.69	4.33	3.83	3.68	3.87
Eu	FUS-MS	1.12	1.12	0.86	1.64	0.29	0.30	0.19	1.26	0.89	1.08	1.25
Gd	FUS-MS	4.16	4.01	2.87	5.23	0.95	0.94	1.03	6.10	5.06	5.10	5.37
Tb	FUS-MS	0.73	0.70	0.49	0.70	0.19	0.19	0.20	1.18	0.99	1.01	1.05
Dy	FUS-MS	4.32	4.06	2.84	3.27	1.20	1.22	1.32	7.45	6.34	6.61	6.87
Ho	FUS-MS	0.83	0.80	0.52	0.56	0.24	0.25	0.26	1.49	1.28	1.36	1.39
Er	FUS-MS	2.38	2.34	1.51	1.52	0.69	0.71	0.75	4.45	3.76	4.01	4.12
Tm	FUS-MS	0.35	0.35	0.24	0.22	0.10	0.10	0.12	0.67	0.57	0.60	0.63
Yb	FUS-MS	2.21	2.28	1.51	1.38	0.67	0.64	0.74	4.32	3.56	4.01	4.06
Lu	FUS-MS	0.34	0.35	0.22	0.21	0.10	0.10	0.11	0.68	0.54	0.64	0.647

Notes: Total Fe as Fe₂O₃; LOI, loss on ignition; S-LA, least altered sandstone; S-M, mineralized sandstone; SYE, syenite dike; GreenU, fuchsite-carbonate ultramafic rock; GrayU, chlorite-carbonate ultramafic rock; BlackU, talc-chlorite ultramafic rock; B-LA, least altered tholeiitic basalt; B-M, mineralized tholeiitic basalt; FUS-ICP, lithium metaborate & tetraborate fusion ICP; FUS-MS, lithium metaborate & tetraborate fusion ICP-MS; TD-ICP, total four acid digestion ICP; INAA, instrumental neutron activation analysis; AR-MS, aqua regia digestion ICP-MS

Table 1 (continued)

Sample No.		U-Unmin	U-310	U-321	2004-3	N-35-63	N-8-33	N-8-6	N-11-17	05-KL-9B	N-8-17.5	N-12-25
Rock Type		S-LA	S-M	S-M	SYE	GreenU	GrayU	BlackU	B-LA	B-M	B-M	B-M
Trace elements Method												
Hf	FUS-MS	3.6	3.6	2.5	3.4	0.6	0.5	0.5	3.3	2.8	3.0	3.7
Ta	FUS-MS	0.42	0.40	0.24	0.32	0.02	0.04	0.03	0.29	0.22	0.25	0.30
Th	FUS-MS	3.00	2.57	1.98	5.30	0.08	< 0.05	0.11	0.59	0.57	0.51	0.60
W	FUS-MS	2.6	10.0	7.8	6.6	6.9	5.7	1.7	3.8	92.3	98.2	189.0
Ti	FUS-MS	0.26	0.22	0.08	< 0.05	0.23	< 0.05	< 0.05	0.14	0.07	0.14	0.09
U	FUS-MS	0.84	0.69	0.54	1.51	0.04	0.06	0.06	0.17	0.18	0.16	0.16
Cd	TD-ICP	1.4	2.7	2.2	2.0	2.1	2.2	0.8	1.0	0.8	3.0	2.2
Ni	TD-ICP	101	167	163	39	728	1140	1200	294	45	47	173
Cu	TD-ICP	55	43	111	44	68	58	93	112	101	276	165
Zn	TD-ICP	79	32	31	28	47	54	49	106	36	52	45
Ag	TD-ICP	< 0.5	0.5	0.9	1.1	< 0.5	< 0.5	< 0.5	< 0.5	< 0.5	1.9	0.7
Au	INAA	0.06	1.53	11.00	0.53	1.81	0.05	0.08	0.03	0.23	10.20	1.69
As	INAA	332	16,900	26,500	1,600	1,280	509	20	160	132	643	483
Cr	INAA	256	276	189	64	1,490	1,920	2,220	18	13	24	55
Sc	INAA	27.6	28.8	16.1	10.6	17.6	18.3	20.2	45.6	31.6	43.2	46.9
Sb	INAA	3.3	8.1	9.2	< 0.2	57.5	20.3	1.6	48.0	1.4	9.7	6.9
Au	AR-MS			> 10	0.56	1.66				0.33	> 10	1.60
Ag	AR-MS			0.85	0.54	0.10				0.14	1.57	0.34
As	AR-MS			> 10000	1,660	1,180				134	617	438
W	AR-MS			0.2	0.1	0.1				0.5	2.5	1.0
Zn	AR-MS			26.7	33.1	40.2				41.2	61.0	50.7
Cu	AR-MS			120.0	43.6	62.1				101.0	266.0	158.0
Pb	AR-MS			5.28	5.41	2.70				1.72	7.74	2.9
Bi	AR-MS			0.16	0.09	< 0.02				< 0.02	0.17	0.03
Te	AR-MS			0.07	3.20	< 0.02				0.24	0.52	0.16
Mo	AR-MS			1.88	0.08	< 0.01				0.07	0.14	0.14
Sb	AR-MS			40.80	1.51	45.50				0.59	6.04	3.82
Sn	AR-MS			< 0.05	0.32	< 0.05				0.09	0.08	0.1
Se	AR-MS			1.1	0.9	0.3				0.3	3.0	1.6

Table 2: Average mineral compositions

Rock Type	S-M	GreenU	GrayU	BlackU	B-LA	B-M
Carbonate	Ankerite (2)	Ferroan dolomite (9)	Ferroan dolomite (9)	Ferroan dolomite (7)	Ankerite (7)	Ankerite (14)
Fe/Fe+Mg	0.3	0.11	0.13	0.20	0.55	0.31
Mg#	0.67	0.88	0.86	0.78	0.41	0.68
Chlorite	(2)		(4)	(3)	(5)	(4)
Fe/Fe+Mg	0.50		0.17	0.30	0.74	0.49
Mg#	0.50		0.83	0.70	0.26	0.51
Atomic% Al	10.27		7.41	7.59	10.01	9.00
White mica	Sericite (5)	Fuchsite (8)			Sericite (3)	Sericite (11)
Fe/Fe+Mg	0.39	0.15			0.55	0.34
Mg#	0.61	0.85			0.45	0.66
Al IV	0.88	0.90			0.96	0.97
Al VI	1.84	1.80			1.82	1.80
Magnesite-Siderite		(16)	(10)			
Fe/Fe+Mg		0.19	0.19			
Mg#		0.81	0.80			

Notes: Number of analyses in parenthesis; S-M, mineralized sandstone; GreenU, fuchsite-carbonate ultramafic rock; GrayU, chlorite-carbonate ultramafic rock; BlackU, talc-chlorite ultramafic rock; B-LA, least altered tholeiitic basalt; B-M, mineralized tholeiitic basalt

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