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Tittel

Notat om den videre malmleting i Bidjovagge
Geological setting of the Bidjovagge deposit and related gold-copper deposits in the northern part of the Baltic shield.

Forfatter	Dato	Bedrift
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Råstofftype	**Emneord**	
Malm/metall	Au Cu	

Sammendrag

Rapporten inneholder et notat og en publikasjon om Bidjovagge gruvers geologi samt to canadiske publikasjoner om tilsvarende geologisk problematikk.

I notatet anbefaler prof. Arne Bjørlykke at det bores et dypt hull på ca 700m for å fastslå de geologiske forholdene mot dypet av forekomsten.

Notat om den videre malmleting i Bidjovagge.

Av professor Arne Bjørlykke

Mulighetene for å finne nye dagnære malmlinser i Bidjovagge-antiklinalen begynner å bli uttømt. Men det er ennå enkelte områder som må undersøkes videre (Se mitt tidligere notat fra juni 89). Arbeidet til Oddleiv Olesen ved NGU med en regional tolkning av Kautokeinoområdet er nå kommet godt i gang. Jeg regner med at vi kan dra nytte av resultatene fra dette prosjektet allerede i feltsesongen 1991. Jeg vil derfor komme tilbake med et notat om den regionale prospekteringen i april/mai.

Jeg vil i dette notatet ta opp mulighetene for malmlinser mot dypet i Bidjovagge. Jeg vil ikke gå i detail angående lønnsomhet ved underjordsdrift, da det ikke er mitt fagområde. De dagnære malmlinser som hittil er funnet varierer sterkt i gullgehalt, men de rikeste linsene, for eksempel K-malmen og C-Au burde kunne tåle økte utgifter til oppfaring og undersøkelsesboring.

Våre undersøkelser hittil har vist at gullmalmen i Bidjovagge er knyttet til skjærsoner og at de har mange likhetstrekk med de mer berømte arkeiske forekomstene i Australia og Canada (se vedlagte publikasjon av Bjørlykke et al. 1991). Erfaringer fra disse malmenen viser at de ofte har en stor vertikal utbredelse, ofte mer enn 2 000 meter (Se vedlagte publikasjon av Sibson et al. 1988). Malmdannelsen er knyttet til overgangen fra duktil til sprø deformasjon. I den duktile sonen vil en ha konstant lithostatisk trykk mens det i den sprø sonen vil være lavere trykk, tilnærmet hydrostatisk fluid trykk. I overgangsonen vil det veksle mellom hydrostatisk og lithostatisk trykk. Det er spesielt karbonatdannelsen som er trykkavhengig og karbonatisering er som regel relatert til gullmalmdannelsen. Hvorfor vet vi idag ikke eksakt.

Gullmalmer kan også endre karakter mot dypet. Dette er kjent fra andre gullmalmer, foreksempel Lac Shortt mine i Chibougamou hvor gullet i den øvre delen opptrer i en silifisert sone sammen med pyritt. Mot dypet opptrer gull og pyritt i en syenitt, sannsynligvis av metasomatisk opprinnelse. Dette har blitt tatt til inntekt for en magmatisk opprinnelse for gullet i området.

Den sprø oppsprekningen som karakteriserer koppermalmen i Bidjovagge, men som også finnes i partier av gullmalmen, indikere at det som hittil er produsert, er fra den øvre del av et malmsystemet. Ut i fra en ren skjærsone-modell så er det derfor ingen grunn til at Bidjovagge forekosten ikke skal fortsette mot dypet. Det er fortsatt usikkert om malmløsningene i Bidjovagge var av magmatisk eller metamorf opprinnelse. Opptreden av lyse porfyriske dioritter som ganger nær malmlinsene kan indikere en magmatisk dannelse. En annen indikator for en magmatisk opprinnelse er konsentrasjonen av tellurider i K-malmen. D- og O-isotop sammensetningen plotter i et området som er felles for magmatiske og metamorfe løsninger. Vi håper å kunne få et bedre svar på dette spørsmålet når fluid inclusion undersøkelsene er avsluttet.

Et problem i Bidjovagge er at malmen er knyttet til separate linser. Avstanden mellom linsene varierer fra noen få meter til flere hundre meter i horisontalplanet. En må forvente en tilsvarende geometri i vertikal- som i horisontalplan. Det betyr at en ikke kan forvente seg en kontinuerlig malm mot dypet og at en eventuel dyp malm kan bli vanskelig å finne. Den uregelmessige optreden av skjærlinser ser ut til å være primær-altså relatert til selv skjærsone dannelsen. Senere bevegelser av kaledonsk alder er dominerende steile vertikale NE-SW forkastninger langs. De er dannet ved isostatiske bevegelser for å kompensere lasten av dekkene i NW. Større horisontale forskyninger relatert til innskyning av de kaledonske dekkene er ikke observert i Bidjovagge.

Det andre problemet er den tilsynelatende relasjon mellom mineraliseringene i Bidjovagge og grafittskiferen. Det er mulig at denne relasjonen fører til at mineraliseringene blir bundet til et begrenset vertikalt nivå.

Mineraliseringene i Bidjovagge ligger i kulminasjonen på en N-S antiform. De stratigrafisk underste nivåene består av dolomitt og skifre med lagerganger av diabas (Se vedlagte figurer). Vi vet lite om bergartene mellom dolomittene og det underliggende basement. Sannsynligvis består sekvensen vesentlig av sandsteiner, men bedre kjennskap til de underliggende bergarter vil være av stor betydning for vurdering av malmpotensialet i Bidjovagge og for den regionale malmleting.

Siste sommers kartlegging og boring i Dæl`ljadas området gav en del ny informasjon. Området ligger langs kanten av grønnsteinsbeltet og forkastningsonen representer en første orden skjærsone. Sonen som er intrudert av diabaser,granodioritter og granitter har gjennomgått amfibolitt-fasies metamorfose. Det ble funnet relativt lave gehalter av kopperkis og amfibolittene viste forhøyet radioaktivitet. Det er mulig at vi i Dæl`ljadas har et dypere snitt i en tilsvarende skjærsone som i Bidjovaggeantiklinalen. Hvis dette er tilfelle så skulle vi forvente en hyppigere frekvens av intrusjoner mot dypet i Bidjovagge-antiklinalen. På lignende måte som i Lac Short kan mineraliseringen mot dypet ta andre former, for eksempel kan mineraliseringen opptre innen sure eller mafiske intrusjoner.

Forslag til videre undersøkelser

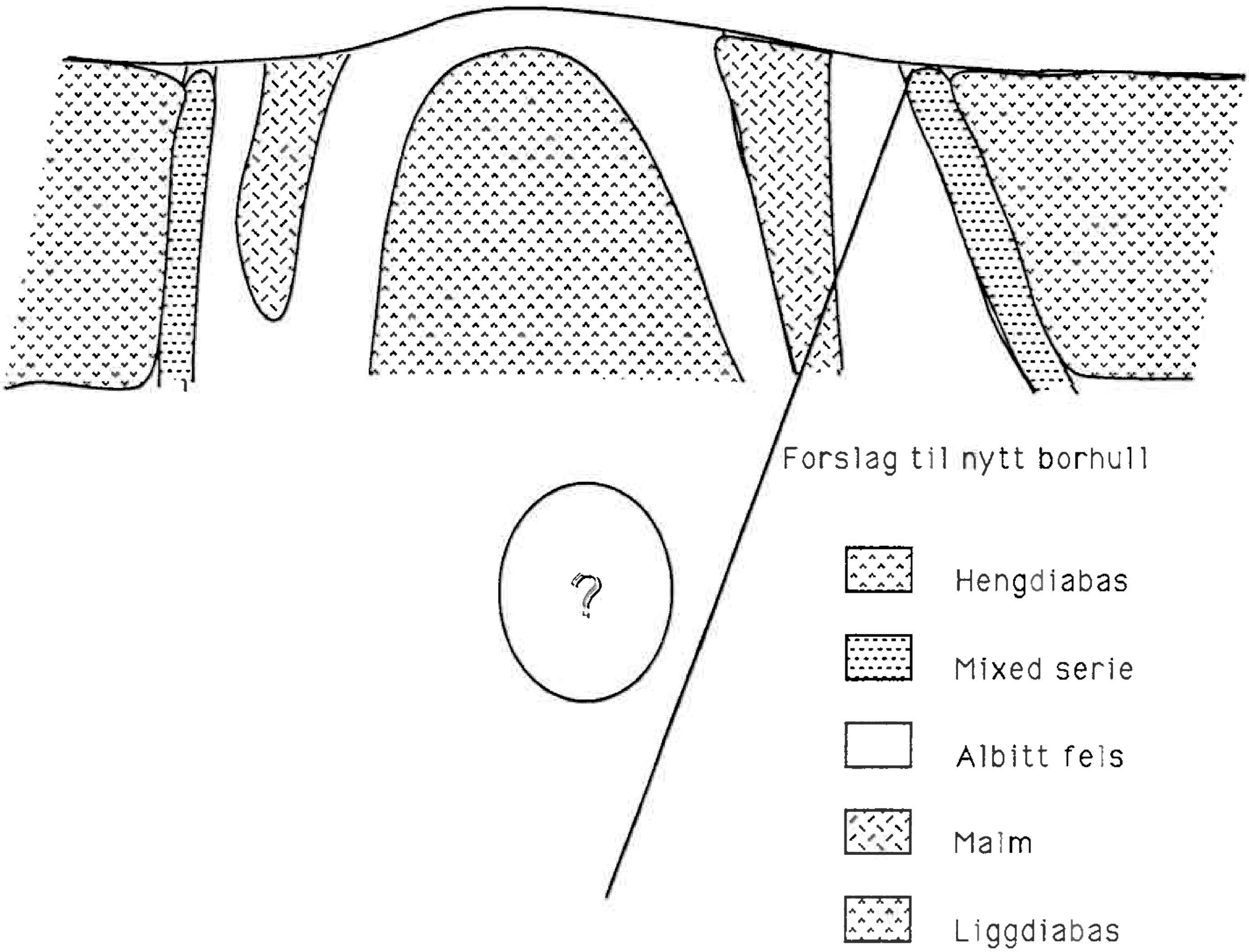
Muligheten for malm på et dypere nivå bør undersøkes nærmere i Bidjovagge. Det bør i første omgang bore ett dypt hull gjennom ligg-diabasen og ned i de underliggende sedimenter. Hullet bør settes på ved en av de viktige malmsonene. Jeg vil foreslå K / B området på grunn av innholdet av tellurider i K-malmen. Lengden på hullet bør være ca 600 til 700 meter. Det bør settes på ved ca 850 N og ca 550 -570 E (øst om kanten av dagbruddet) med fall 80 grader mot W.

Fluid inclusion undersøkelsene viser lovende resultater, men det er ennå for tidlig å si om de direkte kan være til nytte ved lokalisering av dypmalmer. Prøver fra et dyphull vil kunne gi oss viktige data i dette arbeidet.

Ottawa den 8/1-1991

Arne Bjørlykke
Professor

Profil 850 N



Forslag til nytt borhull
?
Hengdiabas
Mixed serie
Albitt fels
Malm
Liggdiabas

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GEOLOGICAL SETTING OF THE BIDJOVAGGE DEPOSIT AND RELATED GOLD-COPPER DEPOSITS IN THE NORTHERN PART OF THE BALTIC SHIELD.

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ABSTRACT

The Bidjovagge gold- copper deposit is the only producing gold mine in Norway. The annual production is approximately 350 000 tonnes of ore containing 4-5 g/t Au and 1 % Cu. The deposit occurs in the Lower Proterozoic Kautokeino Greenstone Belt in Finnmark and within the Baltic - Bothnian megashear zone.

The deposit consists of several ore bodies along a N-S trending anticline over a distance of 2.5 km. They are all hosted by albitic fels, representing strongly tuffite and carbonaceous shale.The albitization is related to the first generation of diabase sills that were intruded during an early compression of the sedimentary sequence.The Svecokarelian metamorphism, which increases from east to west in the area, approached upper greenschist facies in the Bidjovagge mine. A N-S- to NNW-SSE-trending shear zone was then developed, first with dextral strike- slip movements .The gold and copper mineralization occurs in carbonate, actinolite and quartz veins, and these are related to later sinistral strike - slip movements in the same shear zone .The carbonatization of the diabase is related to the same event. The deformation terminated with a late compressional phase which also folded the mineralized veins.U-Pb isotopic analyses of davidites from the ore assemblages give an age of approximately 1.85 Ga.

Gold deposits of similar geological setting and genesis are in production at Pahtevara near Kiruna in Sweden and at Saattopora near Kittilä in Finland.There are many similarities between these Proterozoic deposits and the more famous and larger Archean gold lode deposits. The main differences are the more extensive Na -metasomatism and the higher Cu-content .

INTRODUCTION

During the last 10 years several gold deposits and zones of mineralization have been found in Early Proterozoic rift-related volcanic belts in the northern part of the Baltic Shield. Some of them were known as copper deposits and copper minerals are commonly an important part of the paragenesis, but anomalous contents of uranium, scandium and REE are also common in these deposits.

The Early Proterozoic Kautokeino greenstone belt rift in Finnmark, northern Norway (Fig.1), contains the Bidjovagge gold-copper deposit which is the best example of this deposit type (Bjørlykke et al., 1987).The Bidjovagge gold-copper deposit is at present the only important gold producing mine in Norway. The mine produced copper in the first period of mining from 1970 to 1975. Au-rich mineralization was then discovered close to the Cu-rich orebodies and in 1985 the mine was reopened by Outokumpu Oy. Annual production is about 350 000 tonnes of ore with approximately 4 - 5 g/t Au and 1 % Cu. The known ore reserves are small.

Several other ore deposits and prospects occur in the Early Proterozoic greenstone belts in the northern parts of Norway, Sweden and Finland. Three gold-copper mines are now in production , Bidjovagge, Sattopora in the Kittila area of northern Finland and Pahtohavare near Kiruna in Northern Sweden. Gold mineralization in Proterozoic volcanic belts is also known to occur in Norway, at Ringvassøy near Tromsø and in the Rombak area near Narvik, and in Finland, near Kuusamo in Northeastern Finland (Pankka and Vanhanen, 1989) and at Pahtavaara in Finnish Lapland (Korkiakoski, 1989).

Several Cu deposits and Cu-U deposits occur in the same geological sequence (Gaal, 1990). Best known are the Viscaria Cu deposit (Godin, 1976) and Kopperåsen U-Cu deposit (Adamak,1975) in Sweden , the Pahtavouma Cu-Zn-U deposit in the Kittila area of northern Finland (Inkinen, 1979) and the Repparfjord,Kåfjord and Raipas deposits in northern Norway (Bjørlykke et al, 1985, Sandstad 1986).

The genesis of the Cu ± Au deposits on the northern part of the Baltic Shield has been controversial for a long time. Because of their stratabound character these deposits have been classified as exhalative sedimentary by Hollander (1979) and Inkinen (1979). Gaal (1990) suggested that they belong to the Besshi-type. An epigenetic genesis of the Bidjovagge deposit has been suggested by Gjelsvik (1958) , Padget (1959), Bjørlykke et al. (1987), Bjørlykke et al. (1990) and Nilsen & Bjørlykke (1990). Most of the Cu deposits are associated with strong sodium metasomatism, but the style of mineralization and the metal composition differ among groups of deposits. This may also reflect genetic differences.

The gold-copper deposits , which all are hosted by a very fine-grained albitic rock (albitic felsite), have been called Bidjovagge-type deposits (Gaal 1990). This paper will therefore give a short description of the Bidjovagge deposit and the geological setting for similar deposits in the Baltic Shield.

Regional geology

The gold-copper deposits in the northern part of the Svecokarelian / Baltic Shield are hosted by Early Proterozoic supracrustal rocks, mainly in the Kittela greenstone group, the Kiruna greenstone group and the Cas`kejas greenstone group (Kautokeino volcanic belt). The Archean basement (Fig. 1) consists mainly of tonalitic gneiss. Late Archean volcanic rocks (greenstone belts) are well developed in the Kittila area and occur also in the Komagfjord area.

The Early Proterozoic supracrustal rocks can be divided into four major units: 1.a basal clastic unit with quartzite and arkose; 2 a mixed unit with sedimentary rocks and mafic volcanic rocks; 3. a unit with felsic volcanic rocks; 4. an upper clastic unit with quartzite and arkose. The two first units were deposited in an extensional tectonic regime and we can distinguish between a platform and a rift environment. The time of deposition is poorly constrained but the units are older than a large layered intrusion dated at 2.45 Ga and younger than the Kiruna Porphyry Group at 1.93 Ga.

1. The basal clastic unit varies in thickness from a few metres in the platform environment to more than 1 km in the rift. The best exposed rift environment is represented by the Saltvann Group in the Komagfjord window, which contains sedimentary rocks deposited from alluvial fans and braided rivers (Pharaoh et al.,1983). In the platform environment the sandstones are pink , white or green (fuchsite stained). They are thick bedded, medium- to fine-grained and well sorted (Siedlecka, 1985). In both environments the upper part consists of schist and locally dolomite.

2. The mixed unit (Kiruna Greenstone Group, Kittila Greenstone Group, Cas`kejas Group and the Nussir Group) comprises dolomite, black shale, tholeiitic basalt and tuffite. The platform environment is represented mainly by plateau basalts with a quiet magnetic pattern. The rift environment is characterized by highly magnetic gabbroic sills and dykes that are easily identifiable on magnetic maps. The magnetic interpretation map of the Nordkalott region (Henkel et al., 1986) shows clearly that part of the volcanic sequence in Kittela is related to the same complex rift structure as at Kautokeino and Kiruna. Both the sills and the volcanic rocks are tholeiitic in composition. (Pharaoh and Pearce, 1984; Pharaoh et al.,1987). The Kiruna Group was deposited from 2200 Ma to 1930 Ma (Skiöld, 1986,table 1) .

3. Felsic volcanic rocks occur locally above the mixed unit. The best example is at Kiruna where a thick sequence containing syenitic and quartz-bearing porphyritic rocks

has been deposited. They yield 1.91-1.86 Ga zircon U-Pb ages (Skiöld and Cliff, 1984; Welin, 1987). The Raufjell Suite in the Komagfjord window (Pharaoh et al., 1983) are probably of the same age.

4. The uppermost unit consists of a thick sequence of fluvial sandstone (Kumpu, Vakko and Caravarri Formation).

The geological evolution of the northern part of the Baltic Shield has recently been compiled by Pharaoh and Brewer (1990) . They suggested that the Kola and Karelian continents collided at 1900 Ma and that this led to the formation of the granulite belt (Fig. 1) , followed by an accretion of the "Skjellefte Arc" at 1880 Ma. They also suggested that the accretion of the " Skjellefte Arc " resulted in the development of a large scale dextral shear zone along the Raahe - Ladoga line. According to Skiold (1988), there were two intrusive periods during the Svekokarelian orogeny. The first was between 1880 and 1900 Ma; the foliated " old " granites belong to this generation. The second period resulted in monzonitic intrusions emplaced between 1860 and 1870 Ma. Padget (1959) described the occurrences of leucodiabases in Svecokarelian rocks. He interpreted them as metasomatic rocks formed late in the orogenic event. It is interesting to see that the occurrences of leucodiabase correspond very well with the areas in which gold has later been discovered.

Berthelsen and Marker (1986) described a rather complex kinematic evolution of the northern part of the Baltic Shield in four stages: 1. Collisional to early post collisional stage with dextral movements on N-S megashears. 2. Dextral shear along NW-SE structures (Raahe-Ladoga). 3. Sinistral movements along N-S megashears. 4 Sinistral movements along NW-SE megashears.

GEOLOGY OF THE BIDJOVAGGE DEPOSIT

The Bidjovagge deposit occurs in a north-south trending volcanic belt between domes that contain older Archean gneiss and amphibolite. In the Bidjovagge area the lowest stratigraphic unit consists of dolomite and black shale. This is followed by a sequence of tuffite which grades into mafic volcanic rocks to the south, and which is succeeded by marine shale and carbonate. These, in turn, gradually pass upwards into fluvial sandstone (Bjørlykke et al., 1987). The lower part of the sequence has been intruded by gabbroic or diabase sills, and the sedimentary rocks on the upper side of the sills are strongly metasomatized to a "cherty looking" albitic felsite. This first generation of sills is chemically very similar to the overlying mafic volcanic rocks and they are probably related to the same volcanic episode. Later non-foliated porphyritic diorite occurs locally in structures related to shear zones. In the Kautokeino area, south of Bidjovagge, komatiite also occurs in the lower part of the sequence (Olsen & Nilsen, 1985).

The main structure of the Bidjovagge area is a north-south trending antiform, which can be followed along an axial length of 8.5 km (Figure 2). The lower parts of the Cas´kejas Formation, containing the ore bearing units, are exposed in this anticlinorium. The general stratigraphy of the Cas´kejas Formation is shown in Figure 3. The lowermost beds outcropping in the Bidjovagge area consist mainly of dolomite, and were probably deposited on tidal flats similar to the less deformed beds in Kvenangen , north of Bidjovagge (Vik , 1985). These are followed by argillite, commonly carbonaceous and typically altered to albitic felsite in the mining area. The uppermost beds consist of tuffite and amphibolite (Bjørlykke et al., 1987). The Cas´kejas Formation has been intruded by at least two generations of diabase sills, mainly in its lower part. The thickest and the most continuous sill occurs at the boundary between the carbonate and the argillite. The albitic felsite occurs mainly in direct contact with, and lies above, this main sill and the felsite alteration is interpreted to be related to the diabase intrusion.

The Bidjovagge area is intersected by several north-south trending faults, probably related to the N- S Baltic-Bothnia megashear (Berthelsen and Marker, 1986).Detailed mapping has revealed a complex zone with dextral and sinistral strike slip, as well as reverse and normal faults. Slickensides show two directions of movement, one sinistral strike-slip and one reverse. The ore lenses are related to this shear zone and lenses of diorite occur together with the ore. Unaltered diorite has phenocrysts of albite and biotite. The biotite is locally altered to chlorite and in contact with the orebody the diorite becomes white, and contains albite phenocrysts in an albitic matrix.

Ore geology

The Bidjovagge mine encompasses 10 orebodies that are distributed for a strike length of 2.5 km along the Bidjovagge anticline. There are large variations in their gold content and recent exploration has resulted in the discovery of several high-grade gold zones. For example, a 50 x 50 x 4 m zone along the hanging wall contact of the D ore body, grading 26 g/t Au and 0.1 % Cu, has been described by Söderholm & Nixon (1988). All economic ore is hosted by albitic felsites, but graphitic felsite is also commonly mineralized close to its contact with albitic felsite and some gold mineralization is also found in albitized diabase sills.

The most prominent alteration is the first generation of albitization. It can be followed for several kilometres stratigraphically above the main footwall diabase and its thickness is usually between 5 and 50 metres. The contact between albitic felsite and graphitic schist commonly crosscuts primary bedding, and directly adjacent to the footwall diabase, the graphitic schists are both albitized and oxidized and a gradational bleaching of the graphitic felsite . This is interpreted as being the result of oxidation fronts, which are related to hydrothermal alteration (Bjørlykke et al., 1987). Further away from the diabase, the graphitic schist becomes gradually less albitized. The Na_2O content in schist increase from 3 - 4 % to 7 - 8 % in the most altered zones.

The albitic fels consists of very fine grained pure albite (An 0-2) with grain size mainly less than 0.01 mm. Small rutile and calcite grains are commonly present. More schistose varieties contain biotite, sericite or greenish muscovite with high Cr and V contents and some quartz. Carbonate (calcite and/or dolomite) is common and is disseminated or in thin veins and bands. Scattered grains or blastic lenses less than 0.1 to 1 mm occur frequently, and these contain one or more of the minerals calcite, albite (An 0-8),quartz, pyrite, rutile, sphene, davidite, tourmaline, biotite, muscovite or amphibole (Hagen 1982; Mathiesen, 1970 ; Bjørlykke et al., 1987; Nilsen & Bjørlykke, 1990).

In the hanging wall , which contains tuffite, carbonate and shale, the albitization is followed by scapolitization. Both albitization and scapolitization are quite common near gabbroic intrusions in greenstone belts on the Baltic Shield and may be caused by highly saline brines or evaporites in the sedimentary sequence (Tuisku, 1983)

The footwall diabase is commonly altered to a coarse-grained albite-amphibolite rock. This occurred during the first albitization. Later, and related to the ore-forming event , carbonatization and pyritization of the diabase took place. Biotite alteration and fine-grained albitite were developed along shear zones. In the footwall fault between the main diabase and the albitite is a zone containing green sericite. In the ore zone actinolite and iron-rich amphibole have commonly replaced earlier albite.

Chlorite ± hematite seems to be late in the alteration process and possibly related to weathering during development of the Cambrian peneplain.

The ore mineralization can be divided into two main types, but the ore bodies are usually a mixture of these types :

- Copper-ore type(2-5 % Cu and less than 1-2 ppm Au) : This ore consists of coarse grained carbonate (ankerite and calcite) veins and dykes that contain albite, actinolite, quartz; these also contain moderately abundant sulfides (chalcopyrite and pyrite), minor tellurides and native gold.

- Gold-ore type (Au 5 - 20 ppm , Cu 0,1 - 0,5 %) : This consists of small fracture veins in microbrecciated albitic felsite, that contain quartz , actinolite/ Fe-amphibolite, pyrite, pyrrhotite and minor chalcopyrite, apatite, green muscovite, tellurides, davidite and gold. A telluride-rich orebody (K-orebody) was recently discovered near the B orebody. This ore represents a subtype of the gold ore-type. The main part of the gold occurs as calaverite and as native gold associated with altaite.

The copper type occurs in late central and/or oblique shear veins that typically are moderately undeformed but which, in some places, are boudinaged or gently folded.These veins are someplaces as much as 4 to 5 metres thick. The gold type occurs mainly in the shear fabric and in earlier structures than the copper type indicating an age relationship between this two ore-types.

Typically a good correlation exists between gold content and radioactivity (Bjørlykke et al., 1987), the latter being mainly related to the occurrence of davidite, and indicates that the gold and davidite are cogenetic. Together with davidite, uraninite occurs locally and gold has been found within a grain of uraninite (H. Åsen, pers. com. 1990). Davidite from Bidjovagge has been dated by U-Pb and Sm - Nd methods and ages of 1885 ± 18 Ma and 1886 ± 88 Ma respectively, have been obtained (Bjørlykke et al., 1990).

Geological evolution of the Bidjovagge district

The main mafic volcanism of the Kautokeino greenstone belt took place around 2.0-2.1 Ga (Krill et al., 1985; Olsen and Nilsen,1985).The albitic felsites at Bidjovagge were preferentially formed by intrusion of diabase sills into the shallow marine sequence, which probably contained highly saline water (early extensional rift phase). Folding to produce the N-S trending antiformal structure took place during the peak of metamorphism (approximately 1900 Ma), which reached lower to medium grade in the amphibolitic tuff.

The Bidjovagge antiformal structure is cut by shear zones trending N-S to NNW-SSE (Fig 2). Several generations of displacements, consisting of both dextral and sinistral movements and reverse faulting, are observed in the shear zones. The early generation of shear is represented by mylonite zones at the contacts of some albitite felsites and as ductile shear bands in the amphibolitic tuffites. Dextral movements along NW-SE trending shear zones indicate a possible tensional phase. Further movements, probably sinistral, along the N-S trending shear zones led to local microbrecciation of the albitic felses (early brittle stage of the deformation) hosting the Au-rich ore type. Porphyroblasts of davidite, formed in an early stage of mineralization associated with gold, give an age of 1885 ± 18 Ma (Bjørlykke et al., 1990). Intrusion of albite diorite dykes, further brecciation, and formation of the Cu-rich ore type and coarse albite-carbonate veins, were accompanied by sinistral shear movements and reversel block faulting (compressional deformation phases).

The observed close age relationship between the gold and copper ores is supported by lead isotope determinations from the copper ore. The Pb/Pb data are interpreted as a two stage isochron with an initial age of 1876 ± 15 Ma (Bjørlykke et al., 1990). The shear zone intersections were commonly favourable locations for the precipitation of the ore-forming solutions. The main ore bodies are preferentially located at the ends of the discontinous graphitic schist lenses where the shear cross-cuts the oxidation front between the graphitic and albitic felses. The albitic felses are more intensely brecciated, probably due to the great difference in rheological behavior relative to the graphitic schists.

Smaller faults and N-S to NE-SW fracture zones have in part extensive chloritization and associated hematite, carbonates and native copper. These features are probably inherited from the surface alteration under the Caledonian peneplane.

REGIONAL CORRELATIONS

Gold deposits in similar geological settings as the Bidjovagge deposits occur also in the Kiruna and Kittila districts (Fig. 1).

Pahtohavare deposit

The Pahtohavare copper-gold deposit, which is located 9 km south of the Viscaria Cu deposit in Kiruna, was found during an intensive exploration program from 1984 to 1988 and production started in 1990 (Carlson and Johansson, 1990). Viscaria AB (Outokumpu) is planning to produce 150 000 tonnes of ore per year at a grade of 1 g/t Au and 2.9 % Cu from an open pit mine. Reserves at Pahtohavare are estimated to be 5.4 million tonnes of ore with 1.28 g/t Au and 2.18 % Cu in 5 different lenses (Carlson and Johansson, 1990).

The area is geologically complex. Archean basement rocks consists of banded gneiss and diorite. Above the basement follows a supracrustal sequence of Early Proterozoic age starting with a basal metasandstone and this is overlain by the Kiruna Group, which includes metasedimentary rocks, in part graphitic schists, and mafic agglomerate, lapilli tuff, banded tuff , gabbro and gabbroic sills. The latter are probably related to the volcanic rocks. The uppermost units comprise the Kurravaara conglomerate and the Kiruna porphyry group.

The Pahtohavare deposit is located within the Kiruna Greenstone Group and five different types of ore and mineralization have been distinguished in the area (Carlson and Johansson, 1990) :

1. In the Archean basement quartz diorite with skarn zones carry disseminated chalcopyrite and bornite as well as traces of gold.
2. Gabbroic stocks and sills contain veins carrying disseminated chalcopyrite and pyrrhotite.
3. Hydrothermal alteration in gabbroic stocks consosits of zones of carbonatization, scapolitization, silicification and biotite alteration that are accompanied by disseminated chalcopyrite , gold and pyrite.
4. Exhalative chert and albitite. Veins of massive chalcopyrite and minor associated native gold occur close to the intrusive centers of the gabbroic bodies. In some areas one can observe oxidation of the graphitic schist similar to the front described by Bjørlykke et al. (1987) from Bidjovagge.
5. Layered and disseminated chalcopyrite mineralization in metatuffites and metasedimentary rocks, similar to the mineralization at Viscaria (Godin, 1976), are also observed at Pahtohavare.

Saattopora deposit

The Saattopora gold-copper deposit in the Kittila district, which was earlier known as a copper deposit, is also similar to the Bidjovagge deposit. The geological reserves at Saattopora are estimated at 1.2 million tonnes grading 4.4 g/t Au and 0.3% Cu.

Production started in 1989 at a mining rate of about 350 000 tonnes/year from an open pit mine (Wyllie, 1989).

Saattopora is located along an east - west tectonic structure (Cirka line), together with the Pahtavuoma Cu-Zn-U deposit (Inkinen, 1979)and the Sirka Cu deposit. An early Proterozoic sequence with quartzite and schist in its basal part is located above the Archean granitic gneiss basement. Stratigraphically above are amphibolite, greenstone with hypabyssal albite diabase and a mixed sedimentary group that includes schist (partly graphitic), tuffite and phyllite (Inkinen, 1979) . Conglomerate and quartzite of the Kumpu Formation occurs highest in the stratigraphic sequence. The supracrustal sequence has been intruded by ultramafic rocks (talc-carbonate chlorite schists) in the eastern part of the area, monzonite to the south and west,and syenite and granite to the west.. The metamorphic grade increases from low in the center to medium and high along the margin of the supracrustal synform

The Saattopora deposit occurs at the contact between micaschist to the south and metavolcanic rocks to the north. The contact zone, comprising graphitic schist, tuffite and dolomite, has been intruded by an ultramafic body and adjacent to this body the sedimentary rocks have been altered to albitic fels. There are two ore bodies, one located on the northern (A) and the other on the southern (B) side of the ultramafic body (Wyllie, 1989). Both ore bodies are within the albitic fels and the gold occurs in quartz-carbonate veins , which probably are related to the Cirka line structure.

DISCUSSION

The Bidjovagge-type gold-copper deposit is a complex ore-type and it is too early in our study of deposits of this type to suggest a detailed genetic model for their formation. We will therefore discuss some observations that we think are significant , in the context of present views of shear-zone-related deposits.

First of all , the tectonic style whereby greenstone belts surrounding domal areas of basement rocks and later granitic intrusions, is very similar to that described for many Archean areas, although these Early Proterozoic supracrustal sequences consist mainly of sedimentary rocks and the amounts of volcanic and ultramafic rocks in the sequences are less .

In the northern part of the Baltic shield we can distinguish between a rift phase of supracrustal rocks and a platform phase (Fig 4), as has been done for the Archean Yilgarn block of Western Australia (Groves et al., 1987). For example, in Finnmark, where the Cas`kejas greenstone represents the rift phase and the Suoluvuobmi Formation represents the platform phase, within the rift there are thicker sequences of fluvial sedimentary rocks , which wedge out toward the platform environment. This early sedimentation phase is poorly constrained by age determinations, but it must have taken place between 2.45 Ga and 2.1 Ga In the overlying volcanic sequence the rift environment is reflected by highly magnetic diabase (gabbroic) dikes and sills.

The main metamorphic event is related to the continent-continent collision to the east which lead to the formation of the granulite belt at approximately 1.9 Ga (Fig. 4). This phase resulted in dextral movement along north-south shear zones (Berthelsen & Marker, 1986). At approximately 1.88 Ga the Skjellefte arc collided with the Archean basement to the east (Pharaoh & Brewer, 1990). This later event was probably the cause of dextral movements along the Raahe-Ladoga line (Pharaoh & Brewer 1990) and the sinistral movements along the north-south Bothnia-Baltica megashear zone described by Berthelsen & Marker (1986).The shear zones are probably a reactivation of older rift structures, such as in the Kautokeino rift (Nilsen & Bjørlykke, 1990)

The ages of the davidites at 1885 ± 18 Ma shows that the ore mineralization was formed during or soon after peak metamorphism and at approximately the same time as the older granites at 1.88 - 1.9 Ga (Skiöld, 1987), and probably during the sinistral movements along the Bothnia-Baltica megashear zone. The fluvial sedimentation in the Baltic Shield occurred shortly after the main orogenic event and before the intrusion of post-orogenic granites, and reflects rapid crustal uplift. The locations of the gold deposits near the boundary of amphibolite facies rocks supports the interpretation that emplacement of the gold occurred before or during the uplift.

As shown on Fig. 1, only minor parts of the greenstone belts of the Baltic Shield are composed of low grade metamorphic rocks and major gold deposits may have been eroded away. This may be one important factor as to why the Proterozoic of the Baltic Shield contains less known gold reserves than Archean greenstone belts in Canada and Australia.

The spatial relationship between gold deposits and magnetite-enriched intrusions has also been described for Archean deposits (Groves et al., 1987), but in the Bidjovagge type.most of the gold is deposited in the more brittle albite alteration zone within the contact aureoles of the mafic intrusions.A direct relationship between pyritization of iron-rich minerals and gold deposition is therefore not obvious on a local scale. Gravimetric data indicate that the Bidjovagge and the Pahtohavare deposits occur near positive anomalies, and therefore near the centers of the intrusions.

In a recent paper on fluids and shear zones, Newton (1990) pointed out that deep shear zones are characterized by the occurrence of alkaline granite, carbonatite - lamprophyre - syenite complexes, and various types of metasomatism including regional carbonatization, fenitization, granitization and , probably, extreme depletion of large-ion lithophile elements. Most of these features have also been described from the shear zones in the Baltic Shield.

The relationship between large scale shear zones and these deposits was first pointed out by Mikkola and Vuorela (1977) and more detailed work (Ward et al., 1988; Nilsen and Bjørlykke, 1990) has supported their view. The shear zones have participated in several movements, but structures related to reverse faulting and sinistral strike-slip movements were probably most important for the ore-forming process at Bidjovagge (Nilsen & Bjørlykke, 1990) Also, in the Baltic Shield, the gold deposits occur on secondary structures related to the megashear zones in a similar setting as that descibed by Groves et al.(1987) for the Yilgarn Craton.

The role of magmatism in the genesis of these gold-copper deposits is still unclear. Emplacement of small porphyritic diorite dykes was temporally close to the mineralization event in the structural history of the area (Ward et al. 1988) and the age of the davidite at Bidjovagge corresponds, within the error limits, to both of the two synorogenic intrusion events dated by Skiöld (1988).The Bidjovagge-type has some similarities to that of Archean age in the Chibougamau area, particulare to that at the Lac Shortt mine (Guha et al., 1990 ; Quirion, 1990) where a magmatic source for the syenite alteration and the ore has been suggested.

The mineralization is, on regional and local scales, related to sodium- and carbonate-metasomatism. This is demonstrated on a regional scale by the occurrences of albitite (albitic fels) and leucodiabase (Padget ,1959). The sodium metasomatism is not related to a single phase in the history of the Svecokarelian orogen, but occurs in almost every stage in its evolution. This may indicate a sodium source within the supracrustal sequence. The arkosic red bed sequence which grades into tidal dolomites in the lower clastic unit may have contained evaporites. Another indicator of a probable evaporitic source of sodium is the widespread occurrence of scapolite (Serdyuchenko, 1975). Veins with anhydrite and gypsum occur in the copper-rich C-orebody.

Another interesting feature in these orebodies is the differentiation between Cu and Au. Copper mineralization occurs commonly in the rift environment, in most cases without significant associated gold contents, both in medium and low grade rocks. Some of the deposits may have been formed significantly earlier than the gold ore. Examples of this early ore-forming event are provided by the red-bed type copper deposit at Repparfjord and the Viscaria deposit in Sweden, which have many similarities to the Kupferschiefer-type deposits.

Different sources may thus be indicated for Cu-ore and Au-ore in the Bidjovagge deposit, and they may have been deposited from different ore-forming solutions. The Au-ore has many similarities to other shear-zone-hosted deposits and may have a deep crustal or mantle source. The Cu-ore may represent metamorphic remobilization of copper from red-bed mineralization originally deposited in an underlying evaporitic environment.

Acknowledgments

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Figure captions:

Figure 1. Geologic map of the north-western part of the Baltic Shield. Modified from Krill et al. (1988). Major shear zones and the location of Bidjovagge (B), Pahthavare (P) and Saattopora (S) are indicated. The numbers along the border of the map are degrees of longitude and latitude.

Figure 2. Geologic map of the Bidjovagge mine. The coordinates are given in metres.

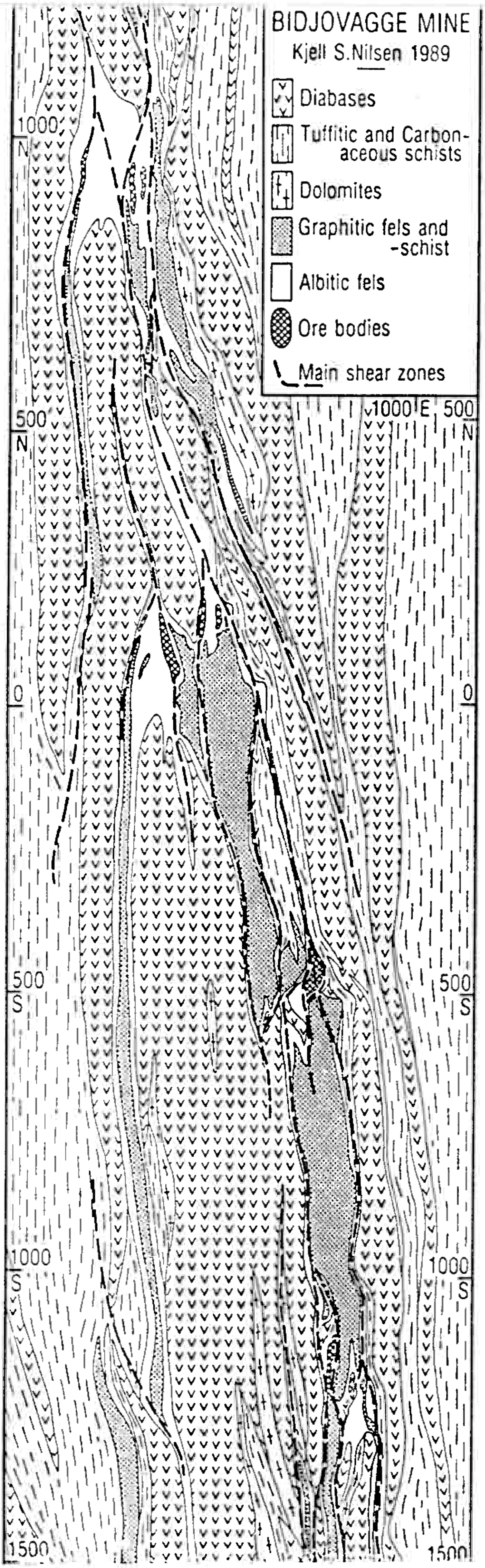
Figure 3. General stratigraphy of the Cas'kejas Formation in the Bidjovagge area.

Figure 4. Simplified tectonic synthesis of the early Proterozoic development of the Baltic Shield, modified from Pharaoh & Brewer (1990). Key to labels : BBZ: Bothnia-Baltic zone; BF: Bergslagen Field; JC: Jormua ophiolitic complexes ; KB: Karasjokk Belt; LGC: Lapland Granulite Complex; PG: Petchenga Group; SD: Svecofennian Domain; SF: Skjellefte Field.

Legend :

- Caledonian rocks
- Granitic intrusions
- Low grade Proterozoic supracrustal rocks
- Medium-high grade Proterozoic supracrustal rocks
- Granulite belt
- Archean rocks- mainly gneisses

Fig 1



BIDJOVAGGE MINE
Kjell S.Nilsen 1989
Diabases
Tuffitic and Carbon-aceous schists
Dolomites
Graphitic fels and -schist
Albitic fels
Ore bodies
Main shear zones
1000 N
500 N
0
500 S
1000 S
1500
1000 E
500 N
0
500 S
1000 S
1500

Fig 2.

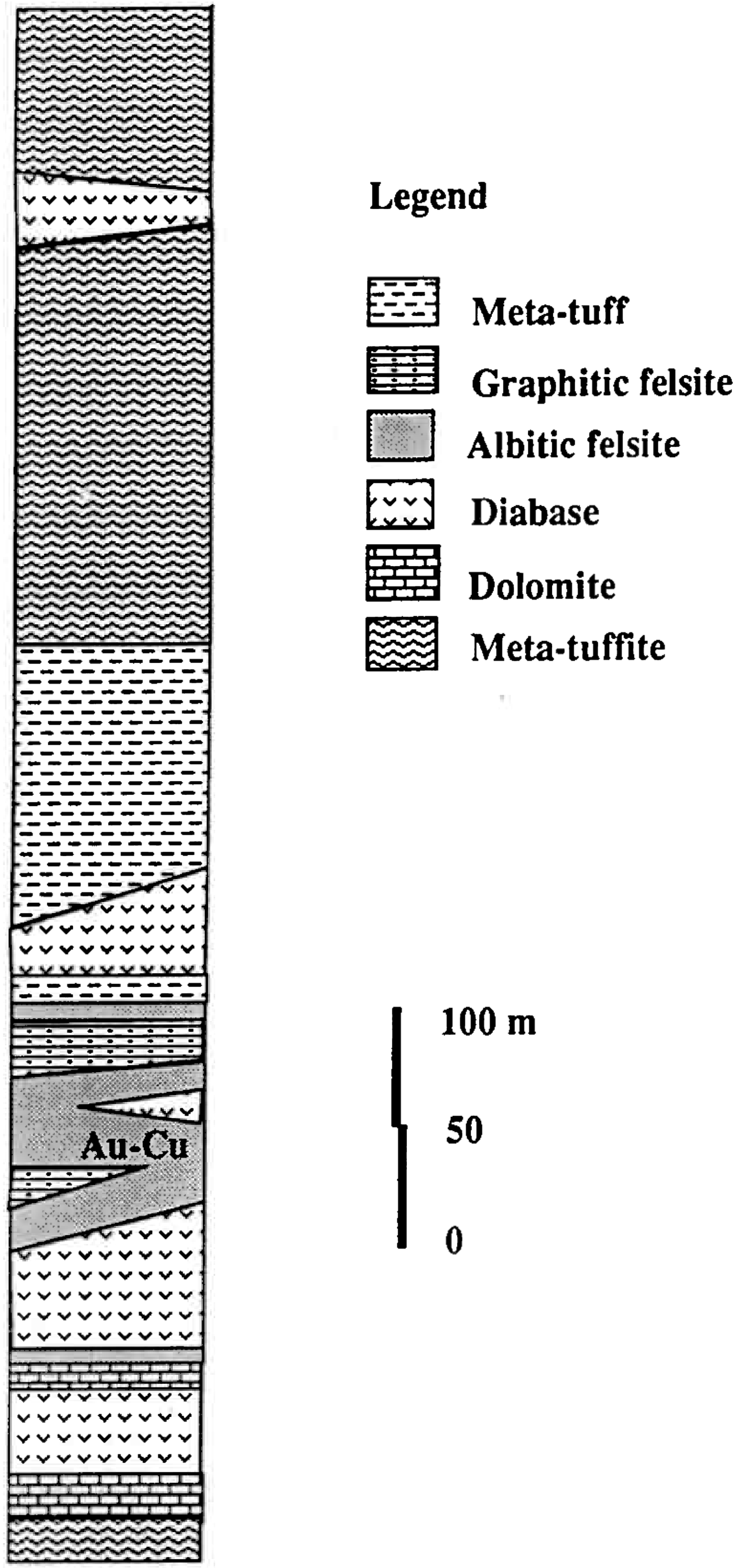

Au-Cu
Legend
Meta-tuff
Graphitic felsite
Albitic felsite
Diabase
Dolomite
Meta-tuffite
100 m
50
0

Fig 3.

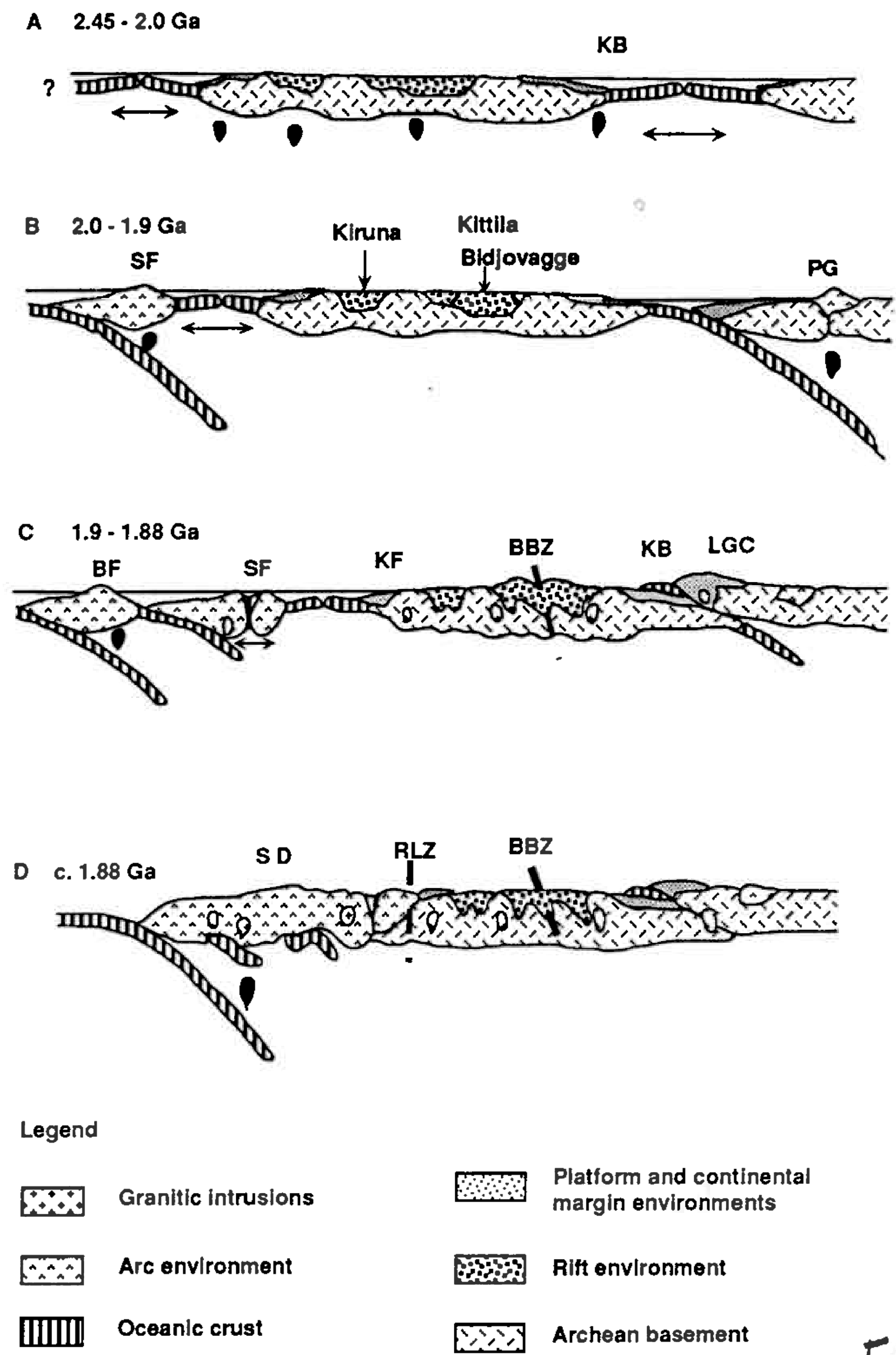

A 2.45 - 2.0 Ga
KB
?
B 2.0 - 1.9 Ga
SF
Kiruna
Kittila
Bidjovagge
PG
C 1.9 - 1.88 Ga
BF
SF
KF
BBZ
KB
LGC
D c. 1.88 Ga
S D
RLZ
BBZ
Legend
Granitic intrusions
Arc environment
Oceanic crust
Platform and continental margin environments
Rift environment
Archean basement

Fig 4.

High-angle reverse faults, fluid-pressure cycling, and mesothermal gold-quartz deposits

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ABSTRACT

Many mesothermal gold-quartz deposits are localized along high-angle reverse or reverse-oblique shear zones within greenstone belt terrains. Characteristically, these fault-hosted vein deposits exhibit a mixed "brittle-ductile" style of deformation (discrete shears and vein fractures as well as a schistose shear-zone fabric) developed under greenschist facies metamorphic conditions. Many of the vein systems are of considerable vertical extent (>2 km); they include steeply dipping *fault veins* (lenticular veins subparallel to the shear-zone schistosity) and, in some cases, associated *flats* (subhorizontal extensional veins). Textures of both vein sets record histories of incremental deposition. We infer that the vein sets developed near the roofs of active metamorphic/magmatic systems and represent the roots of brittle, high-angle reverse fault systems extending upward through the seismogenic regime.

Friction theory and field relations suggest that the high-angle reverse faults acted as *valves*, promoting cyclic fluctuations in fluid pressure from supralithostatic to hydrostatic values. Because of their unfavorable orientation in the prevailing stress field, reactivation of the faults could only occur when fluid pressure exceeded the lithostatic load. Seismogenic fault failure then created fracture permeability within the rupture zone, allowing sudden draining of the geopressured reservoir at depth. Incremental opening of flats is attributed to the prefailure stage of supralithostatic fluid pressures; deposition within fault veins is attributed to the immediate postfailure discharge phase. Hydrothermal self-sealing leads to reaccumulation of fluid pressure and a repetition of the cycle. Mutual crosscutting relations between the two vein sets are a natural consequence of the cyclicity of the process. Abrupt fluid-pressure fluctuations from this *fault-valve* behavior of reverse faults seem likely to be integral to the mineralizing process at this structural level.

INTRODUCTION

Mesothermal gold-quartz vein systems account for a significant fraction of world gold production and form the principal gold resource in several countries. In Canada, for example, they account for over 50% of current production. Most of the systems are associated with Archean greenstone belts in shield areas, but younger deposits are also known in similar settings; e.g., the Cretaceous Mother Lode system of California (Knopf, 1929). The lodes typically occupy steeply dipping shear zones and appear to have developed syntectonically in horizontal compressional or transpressional regimes. This is in distinct contrast to epithermal gold-quartz mineralization, which tends to be associated with extensional or transtensional fault systems. Furthermore, the prevalence of steep veins in regions undergoing horizontal contraction presents a profound structural paradox.

In this paper we demonstrate how this paradox may be resolved by considering the special role of high-angle reverse faults (and perhaps other faults unfavorably oriented for slip) as fluid-pressure-activated *valves*. In the structural environment of mesothermal gold-quartz deposits, large cyclic fluctuations in fluid pressure must accompany intermittent shear failure on master faults that transect the seismogenic regime. The abrupt drops in fluid pressure engendered by faulting episodes seem likely to play an important role in gold-quartz deposition. Previous workers have considered the progressive evolution of fault-hosted lode deposits in terms of *quasi-static* stress fields (e.g., Guha et al., 1983). Here we emphasize that understanding the mineralization process in this structural setting requires an appreciation of the *dynamic* stress change accompanying each seismic faulting episode, and its effect on the fluid regime.

MESOTHERMAL GOLD-QUARTZ VEIN SYSTEMS

Lode-gold vein systems in Archean granite-greenstone belts have been intensively studied in Canada, and the following summary of their characteristics is based largely on this data set.

Host Rocks

The vein systems are hosted dominantly by mafic-ultramafic volcanic sequences of tholeiitic affinity, with some intermediate calc-alkalic volcanic and volcaniclastic rocks. Associated clastic sedimentary rocks (commonly turbidite sequences) and felsic intrusions are usually subordinate. Generally, the host rocks have been regionally metamorphosed under low- to mid-greenschist facies conditions, but amphibolite facies assemblages occur in some areas. Intense carbonate alteration is common in the vicinity of the deposits. Many of the deposits within the Abitibi greenstone belt of the Superior Province occur in the vicinity of major east-west–trending fault zones, known locally as breaks, that have complex deformation histories. Two of the most prominent are the Porcupine-Destor and Kirkland Lake–Cadillac breaks. Recent tectonic analyses suggest that these structures began either as major normal-slip growth faults (Dimroth et al., 1983) or as part of a wrench-fault system (Hubert et al., 1984). However, in both interpretations the youngest deformation involves north-south crustal shortening and thickening with the breaks predominantly reactivated as reverse faults, sometimes with a minor strike-slip component. The development of the gold-quartz vein systems appears to be related to this last phase of reactivation, postdating most of the late alkaline magmatic activity within the orogenic belt. On the basis of isotope studies, Kerrich (1986) suggested that the mineralizing fluids were likely derived from late tectonic devolatilization accompanying metamorphism at depth, with perhaps some magmatic contribution.

Structural Characteristics of Lodes

The gold-quartz vein systems are usually hosted by steeply dipping shear zones exhibiting mixed brittle-ductile (discontinuous-continuous) behavior (Poulsen, 1986). These shear zones may coincide with, or form subsidiary structures to, the major breaks referred to above. They are characterized by an intense L-S tectonite fabric (schistosity with stretching lineation), locally disrupted by discrete shears and vein fractures, and were mostly formed under greenschist metamorphic conditions. The shear zones generally dip between 50° and 80°, with a stretch lineation raking at high angles in the steeply inclined schistosity. Fabric orientation and kinematic indicators associated with the shear zones in many of the larger deposits invariably show them to result from high-angle reverse or reverse-oblique shearing (Roberts, 1987). Included in this category are the major lode-gold deposits at Yellowknife (Henderson and Brown, 1965), Val d'Or (Robert et al. 1983), Kirkland Lake (Watson and Kerrich, 1983), Hollinger-McIntyre (Wood et al., 1986), and Red Lake (Sanborn-Barrie, 1987). Where established, the total reverse separation across the hosting shear zones is low, typically about a few hundred metres.

In considering the structure of the vein systems, we refer extensively to the Sigma Mine at Val d'Or in Quebec, where the relations between different vein components and the vein textures typify those exhibited by many of the major deposits. Detailed structural studies in this mine (Robert et al., 1983) have shown that motion across the steeply dipping (~70°) host shear zones involved nearly pure reverse dip-slip. A notable feature at Sigma and elsewhere is the great vertical extent of the mineralized veins, which in several mines exceeds 2 km.

The quartz-tourmaline vein system at Sigma Mine (schematically represented in Fig. 1) consists of two dominant vein sets: steep *fault veins* subparallel to the walls of the shear zones and *flats* (subhorizontal veins) (Robert and Brown, 1986). Textures in both vein sets are locally diagnostic of open-space filling and record episodic histories of deposition. The fault-hosted veins include hydrothermal wall-rock breccias and lenticular veins that are bounded or juxtaposed by slickenlined slip surfaces. Robert and Brown (1986) showed the flats to be extension veins that opened vertically and were incrementally filled. Flats have been mapped that extend laterally for tens of metres from fault veins, but they may persist to greater distances as barren fractures. Variable deformation of the fault veins (folding, local boudinage, etc.) makes it clear that they developed while the shear-zone complex was active. Flat veins can locally be traced as continuous structures into fault veins, but in other places members of the different vein sets crosscut each other. It is significant, however, that there is no consistent crosscutting relation between flats and fault veins. This argues strongly for a broad contemporaneity of the vein sets and a cyclic developmental sequence.

The flats clearly represent hydraulic extension fractures formed perpendicular to the least principal compressive stress and held open during vein-filling episodes by fluid pressures in excess of the lithostatic load (i.e., $P > \sigma_3$) (Robert and Brown, 1986). Their orientation, coupled with the reverse sense of motion across the shear zones and the existence of some conjugate reverse-sense shear zones, constrains the driving stress field to be as shown in Figure 1. The fault veins may then be interpreted as lenticular cavity fills formed mostly by local overriding during slip episodes on discrete shears (Guha et al., 1983). Similar structural configurations have been reported from other lode-gold deposits (e.g., Kerrich and Allison, 1978; Watson and Kerrich, 1983), though in some cases early-formed extensional veins have taken on a sigmoidal appearance from subsequent deformation. The preservation of flats in approximately their original form and attitude at the Sigma Mine results from the extreme strain localization in the shear zones; the wall rocks (andesitic metavolcanic rocks intruded by a porphyritic diorite) hosting the flats remain largely undeformed.

Metamorphic Environment

Estimates of the temperatures of hydrothermal deposition in the vein systems from fluid-inclusion studies (~300–400 °C; Robert and Kelly, 1987) agree well with those obtained from oxygen isotope analyses of wall-rock alteration (270–450 °C; Kerrich, 1986), and are broadly consistent with the general greenschist assemblages of the hosting shear zones. At Sigma Mine, alteration of metamorphic assemblages around the veins suggests that hydrothermal activity was either coeval with or slightly postdated the shear-zone metamorphism (Robert and Brown, 1986). Depths of vein formation are not well constrained, but pressures in the range of 2 to 4 kbar seem most probable for the gold-quartz lodes of the Abitibi greenstone belt (Brown and Lamb, 1986; Kerrich, 1986; Robert and Kelly, 1987), corresponding to depths of perhaps 7–14 m.

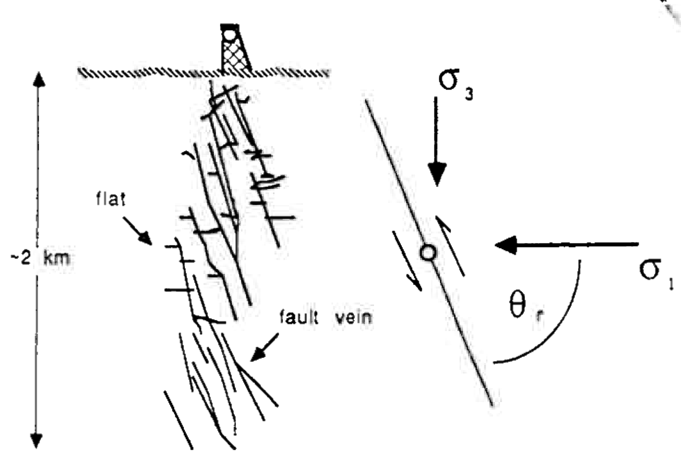

Figure 1. Schematic representation of vein structure within shear-zone-hosted, mesothermal gold-quartz deposit (loosely based on Sigma Mine, Quebec; after Robert et al., 1983), shown in relation to inferred stress field (principal compressive stresses, $\sigma_1 > \sigma_2 > \sigma_3$) and fault reactivation angle, θ_r.

Inferred Tectonic Setting

Background seismicity in deforming continental crust appears to extend to depths corresponding to the onset of greenschist facies metamorphic conditions, typically around 10–15 km (Sibson, 1983). Activity may, however, be restricted to shallower levels in regions of particularly high heat flow and geothermal activity. Larger earthquake ruptures that transect the upper crust tend to nucleate toward the bottom of this background seismogenic zone. Whereas these ruptures extend mostly laterally and upward, it also seems likely that they may penetrate some distance downward beneath the zone of background activity. Deformation around the base of the seismogenic zone is thus likely to involve complex transitional behavior, with aseismic shearing and mylonitization punctuated intermittently by loss of continuity during the propagation of large ruptures. From the greenschist assemblages and the mixed continuous-discontinuous character of the host shear zones, we infer that many mesothermal gold-quartz vein systems represent the exhumed *roots* of high-angle reverse fault zones near the base of the continental seismogenic regime (Fig. 2). Given the tectonic history of regional shortening and crustal thickening and the isotopic evidence on the origin of the mineralizing fluids, it also seems probable that the shear zones lie within the *roofs* of actively prograding metamorphic piles. As cogently argued by Cox et al. (1986), this structural level within the crust likely acts as an impermeable cap to the lithostatic fluid-pressure regime associated with regional metamorphism. Any associated magmatic activity will only enhance the capacity of the metamorphic system for generating high fluid pressures.

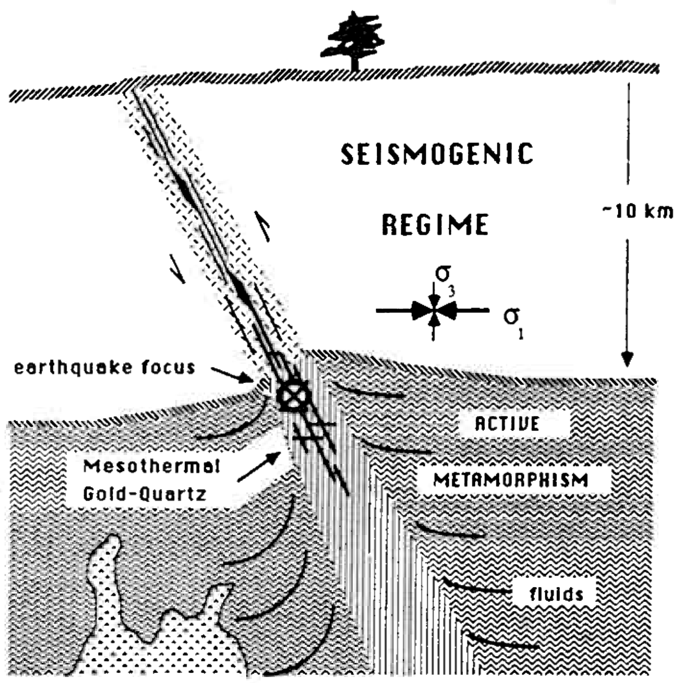

Figure 2. Synoptic diagram (not to scale) of inferred tectonic setting for mesothermal gold-quartz vein system in relation to continental seismogenic regime.

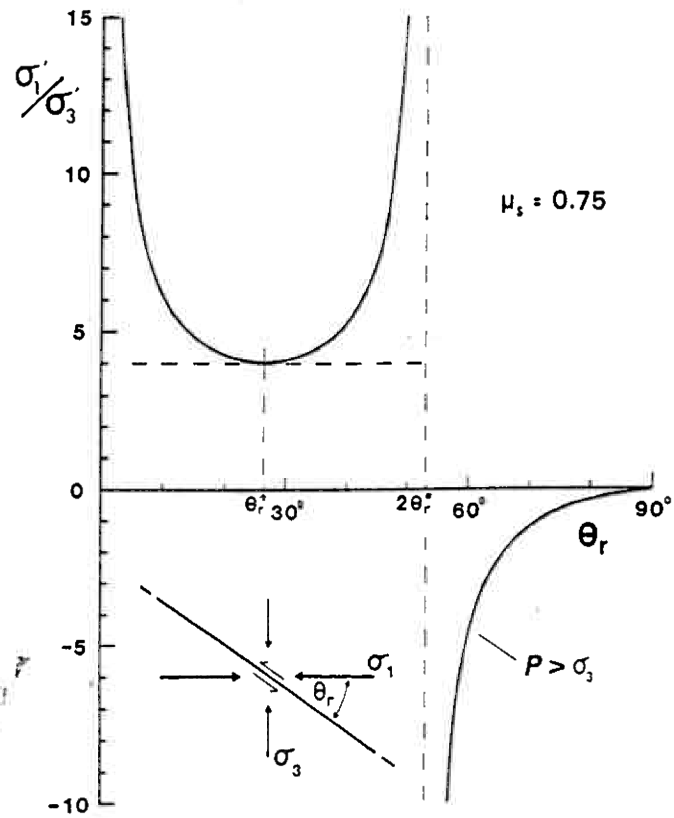

Figure 3. Stress ratio (σ_1'/σ_3') required for reactivation of cohesionless fault plotted against angle of reactivation (θ_r) for static coefficient of rock friction, $\mu_s = 0.75$ (after Sibson, 1985).

ORIGIN AND REACTIVATION OF HIGH-ANGLE REVERSE FAULTS

High-angle reverse faults are anomalous structures with reference to Anderson's (1905) classical theory for the inception of faults in intact crust. In a stress field with principal compressive stresses $\sigma_1 > \sigma_2 > \sigma_3$, the expectation from Coulomb failure theory is that faults should form along planes that contain the σ_2 axis and are oriented at 25°–30° to the maximum principal compression, σ_1. Thus, on the reasonable premise that stress trajectories near the surface of the earth will tend to be either subhorizontal or subvertical, thrust faults with dips well under 45° should develop in regions of horizontal compression. It seems likely, therefore, that most high-angle reverse faults have developed by the reactivation of existing structures. Common situations where this may occur are (1) reactivation of old normal faults when an originally extended region is shortened, (2) transpression across an originally vertical strike-slip fault when the stress field becomes misaligned with respect to the fault system, and (3) continuing reactivation of an initially thrust-sense shear zone that steepens progressively as the wall rocks absorb some of the regional shortening strain. Evolutionary models for the shear zones hosting the Canadian lode-gold deposits generally invoke one of these tectonic settings.

In saturated crust with fluid pressure, P, failure is governed by the effective principal stresses (Hubbert and Rubey, 1959):

$$\sigma_1' = (\sigma_1 - P) > \sigma_2' = (\sigma_2 - P) > \sigma_3' = (\sigma_3 - P). \quad (1)$$

The frictional shear strength of an existing fault may then be approximated by a Coulomb-type criterion:

$$\tau = C + \mu_s\sigma_n' = C + \mu_s(\sigma_n - P), \quad (2)$$

where τ and σ_n are, respectively, the shear and normal stresses on the fault surface; C is the cohesive strength of the fault that might arise from hydrothermal cementation processes; and μ_s is the static coefficient of rock friction, which typically has a value of about 0.75 (Byerlee, 1978). When the cohesive strength is zero, the expression reduces to the equivalent of Amonton's Law:

$$\tau = \mu_s\sigma_n' = \mu_s(\sigma_n - P). \quad (3)$$

For a fault whose pole lies in the σ_1/σ_3 plane, this failure criterion may be rewritten as

$$\sigma_1'/\sigma_3' = (1 + \mu_s \cot\theta_r)/(1 - \mu_s \tan\theta_r), \quad (4)$$

where θ_r is the angle of reactivation as defined in Figure 3 (Sibson, 1985). This plot illustrates the relative ease of frictional reactivation for cohesionless faults of different orientation. Sliding occurs under the minimum positive stress ratio at the optimum reactivation angle, $\theta_r^* = 0.5 \tan^{-1}(1/\mu_s)$, which is about 27° for a typical coefficient of rock friction, $\mu_s = 0.75$. At other than the optimum angle, greater stress ratios are needed for reactivation. Where existing faults are very unfavorably oriented, abnormal fluid pressures are needed to keep the differential stress level for reactivation below that which would cause shear failure of the surrounding intact rock.

Of particular relevance to the discussion here is that for $\theta_r > 2\theta_r^*$, a *necessary* condition for reactivation is $\sigma_3' < 0$ or $P > \sigma_3$. Thus, in a regime of horizontal maximum compressive stress (as in Fig. 1), reactivation of high-angle faults with dips greater than $2\theta_r^*$ (typically ~54°) requires the fluid pressure to exceed the lithostatic load, as is demonstrated by the presence of the flats at Sigma Mine and at other lode-gold deposits. When the misaligned faults have a cohesive strength, the requirement of abnormal fluid pressures for reactivation becomes even greater. In these circumstances, fault reactivation *must* occur at rather low values of shear stress, probably less than about 100 bar, at most (Sibson, 1981, 1985).

FAULT-VALVE MODEL

Fault reactivation under supralithostatic fluid pressures likely leads to nearly total relief of shear stress, and to a marked reduction in the component of normal stress acting across the faults. Thus, in the low deviatoric stress state following failure, geopressured fluids will tend to drain rapidly along the steep fluid-pressure gradient coincident with the fracture-permeable rupture zone. In this manner, high-angle reverse faults may act as fluid-activated *valves* promoting large cyclic fluctuations in fluid pressure (cf. Sibson, 1981). Given the observed structural relations, the inferred stress field, and the tectonic setting of mesothermal gold-quartz lodes hosted in shear zones (Figs. 1 and 2), the following structural cycle is envisaged in relation to major seismic failure episodes on a steep existing fault system (see Fig. 4).

1. *Prefailure*. Fluid pressure below the seismogenic zone, which forms the impermeable roof to a geopressured fluid reservoir at depth, builds up to supralithostatic values, allowing flats to open up in the compressive stress regime. The fault remains sealed at this stage.

2. *Seismogenic Fault Failure*. Once fluid pressure exceeds the lithostatic load, accumulating shear stress may trigger failure on the fault near the base of the seismogenic zone in accordance with the reactivation criterion (equation 2 or 3), nucleating an earthquake rupture that propagates largely updip to the surface and creates extensive fracture permeability (Fig. 2). Shear stress along the fault is substantially reduced, and so the deviatoric stress state becomes very low.

3. *Postfailure Discharge*. Under this low deviatoric stress, drainage of the geopressured fluid reservoir at depth occurs along the rupture zone and associated subsidiary fractures. The abrupt drop in fluid pressure toward hydrostatic values triggers mineral deposition in the fracture network at the base of the seismogenic zone and in minor, lenticular "rideovers" along the rupture. Comparison with modern ruptures (see below) suggests that the discharge is likely to decrease progressively through the aftershock phase, perhaps over a period of several months.

4. *Self-Sealing*. Hydrothermal deposition during discharge seals fractures and destroys permeability along the rupture zone, perhaps preventing a total pressure drop to hydrostatic values.

5. *Repetition of Cycle*. Fluid pressure and shear stress rebuild, and the cycle repeats.

On the assumption that the slip increment accompanying each failure episode is about 1 m (a reasonable value for an ~M 7 rupture transecting the upper crust), a total displacement of a few hundred metres across the fault system would involve 10^2–10^3 episodes of fluid-pressure accumulation and discharge.

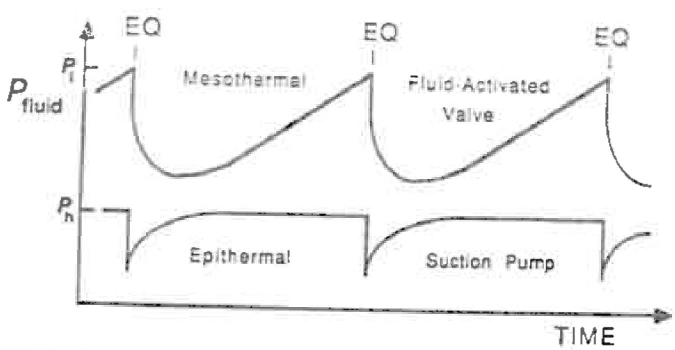

Figure 4. Inferred fluid-pressure fluctuations induced by *valve* and *pump* mechanisms in relation to successive earthquake rupturing episodes (EQ). P_h and P_l represent hydrostatic and lithostatic fluid pressures, respectively. Top curve relates to mesothermal environment of gold-quartz lodes at base of seismogenic regime; bottom curve relates to pressure regime in vicinity of dilational fault jog.

Seismological Considerations

The existence of high fluid pressures and low deviatoric stress in the area of rupture nucleation has wide-ranging implications for seismic source theory and for potential precursory behavior before large ruptures in comparable tectonic settings. Recognition of rupture triggering after fluid-pressure accumulation also raises several important questions. What is the degree of coupling between the buildup of fluid pressure and shear stress at the base of the seismogenic zone? How commonplace is *fault-valve* behavior? Is there any evidence that it occurs today? One possible example is the massive surface discharge following the M 7.6 Kern County, California, earthquake of 1952. This major event, involving left-reverse oblique slip on a steeply (~65°) dipping fault (Gutenberg, 1955), was followed by the effusion of ~10^7 m^3 of water in the vicinity of the rupture trace over a period of two months (Briggs and Troxell, 1955; Sibson, 1981).

MINERAL DEPOSITION FROM FLUID-PRESSURE FLUCTUATIONS

The difference between lithostatic and hydrostatic fluid pressures increases downward at ~170 bar/km, and so the potential pressure fluctuations caused by this mechanism at around 10 km depth are large, even if the pressure drop is only partial. Abrupt reductions in fluid pressure may have a dramatic effect on the aqueous solubility of quartz (Walther and Helgeson, 1977) and are likely to play a major role in mineral precipitation (Helgeson and Lichtner, 1987).

Fluid-inclusion studies from the gold-quartz veins have shown that the mineralizing fluid was CO_2-bearing and of low salinity (Robert and Kelly, 1987). Induced boiling or phase separation of CO_2 accompanying sudden drops in fluid pressure thus seems likely to be an important mechanism for rapid precipitation (cf. Spooner et al., 1987). Supporting evidence includes the widespread carbonatization around the vein systems and the presence of carbonates as an important gangue constituent.

DISCUSSION

From the tectonic setting, structural relations, and vein textures of mesothermal gold-quartz lodes hosted in shear zones, we have developed a mechanical model in which high-angle reverse faults act as fluid-activated valves on geopressured reservoirs, causing large cyclic fluctuations in fluid pressure that promote mineral deposition. Previous workers have recognized the need for fluid-pressure fluctuations in the genesis of these deposits, but have generally found it necessary to invoke radical switches in static stress regimes to account for the different vein sets (Kerrich and Allison, 1978; Rigg and Helmstaedt, 1981; Kerrich, 1986). The *fault-valve* model must be regarded as preliminary; careful consideration of different lode-gold deposits may make it possible to build a more comprehensive picture of the linked fault-vein systems and the process of mineralization at different crustal levels. For example, the presence or absence of flats adjacent to fault veins may depend on the particular structural level exposed by erosion and mining.

High-angle reverse faults capping regions of active crustal shortening and metamorphism provide the optimum setting for *fault-valve* behavior, capable of giving rise to the largest fluctuations in fluid pressure. However, other unfavorably oriented fault structures may also induce fluid-pressure fluctuations of lesser extent. A vertical strike-slip fault at a high angle to σ_1 in a transpressive regime would be one example. In all such circumstances, one would expect a combination of discharge fault veins and prefailure extension veins in the σ_1/σ_2 plane.

One may compare the anticipated effects of the *fault-valve* model (Fig. 4) with the *suction pump* effect of earthquake ruptures within the seismogenic regime (Sibson, 1987). In this latter situation, slip transfer across dilational fault jogs during rupture propagation causes abrupt drops in fluid pressure below ambient (hydrostatic?) values, triggering boiling and mineral deposition in the epithermal environment. The important point is that although the *pump* and *valve* mechanisms differ in detail, both depend on *dynamic* effects associated with earthquake faulting. The observed structural relations are inconsistent with a static stress state.

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JOUR / DAY 5B

GEOLOGIE DE LA MINE D'OR LAC SHORTT

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INTRODUCTION

La mine Lac Shortt, une division de Minnova Inc., exploite un gisement d'or à Waswanipi (Québec). La mine est localisée à 120 km au sud-ouest de Chibougamau soit à 450 km au nord de Montréal. Une route de gravier de 12 km de long relie la propriété à la route provinciale 113.

La plupart des indices aurifères de la région montrent une association spatiale avec des failles d'orientation E à NE (Brisson et Guha 1989b). A la mine Lac Shortt, la zone minéralisée principale est formée d'une mylonite située dans l'éponte inférieure d'une faille régionale orientée ENE appelée faille Lac Shortt. Toutefois, le gisement constitue un cas particulier de par sa filiation avec un complexe alcalin à carbonatite et syénite.

HISTORIQUE ET PRODUCTION

L'histoire de la découverte du gisement Lac Shortt remonte à 1950 où McWatters Gold Mines Limited, à la recherche de nickel, a foré l'anomalie magnétique située dans l'éponte inférieure du gisement. Par la suite, diverses compagnies se sont succédées dans l'exploration de la structure pour son potentiel aurifère. En novembre 1984, La Corporation Falconbridge Copper débute l'exploitation souterraine du dépôt. La mine est depuis 1986 une division de Minnova Inc., le successeur de Corporation Falconbridge Copper et la production est de l'ordre de 1000 t par jour.

Le gisement est connu jusqu'au niveau 800 et actuellement, 2 zones sont en exploitation. La Zone Principale est constituée d'une mylonite à fragments de syénite dans une matrice carbonatée et fénitisée avec pyrite et hématite et la Veine Sud qui lui est subsidiaire, comprend des veines de quartz injectées dans un cisaillement mineur. La méthode d'exploitation dans les niveaux supérieurs est par chantiers longs trous tandis que sous le niveau 500, on utilise la méthode AVOCA. Le gisement a fourni à ce jour 1,882,845 t de minerai à une teneur moyenne de 4,93 g/t Au. Les réserves au 1er janvier 1990 étaient de 885,219 t à une teneur de 4.82 g/t Au et la production totale de la mine Lac Shortt à ce moment était de 270,255 oz tr (9,266 kg) d'or (Coulombe, 1990).

GEOLOGIE LOCALE

Cadre lithologique

La stratigraphie de la région du lac Shortt comprend les Formations d'Obatogamau et du Ruisseau Dalime, lesquelles sont corrélées avec les roches de la base du Groupe de Roy reconnues dans le district de Chibougamau (Sharma et Gobeil, 1987). Sur la propriété Lac Shortt, les failles Lamarck, Lac Shortt (Mica Vert Principal) et le Mica Vert du Sud marquent les frontières de 4 secteurs caractérisés par des ensembles lithologiques distincts (Fig. 5B.1).

Secteur 1

Le secteur 1 comprend des roches du Complexe des Chutes de l'Esturgeon (Lamothe, 1983). Les pyroxénites et les gabbros différenciés sont associés à des niveaux mineurs de volcanoclastites et de basalte. Cette unité est tronquée au sud par la faille Lamarck qui est définie par une bande de schiste à chlorite, calcite et séricite.

Secteur 2

La séquence de tuf à lapillis-blocs homogène qui repose au sud de la faille Lamarck constitue l'éponte supérieure de la faille Lac Shortt. De composition intermédiaire, le tuf appartenant à la Formation du Ruisseau Dalime, est à fragments lithiques volcaniques dans une matrice feldspathique et chloriteuse. On observe une déformation progressive de la roche

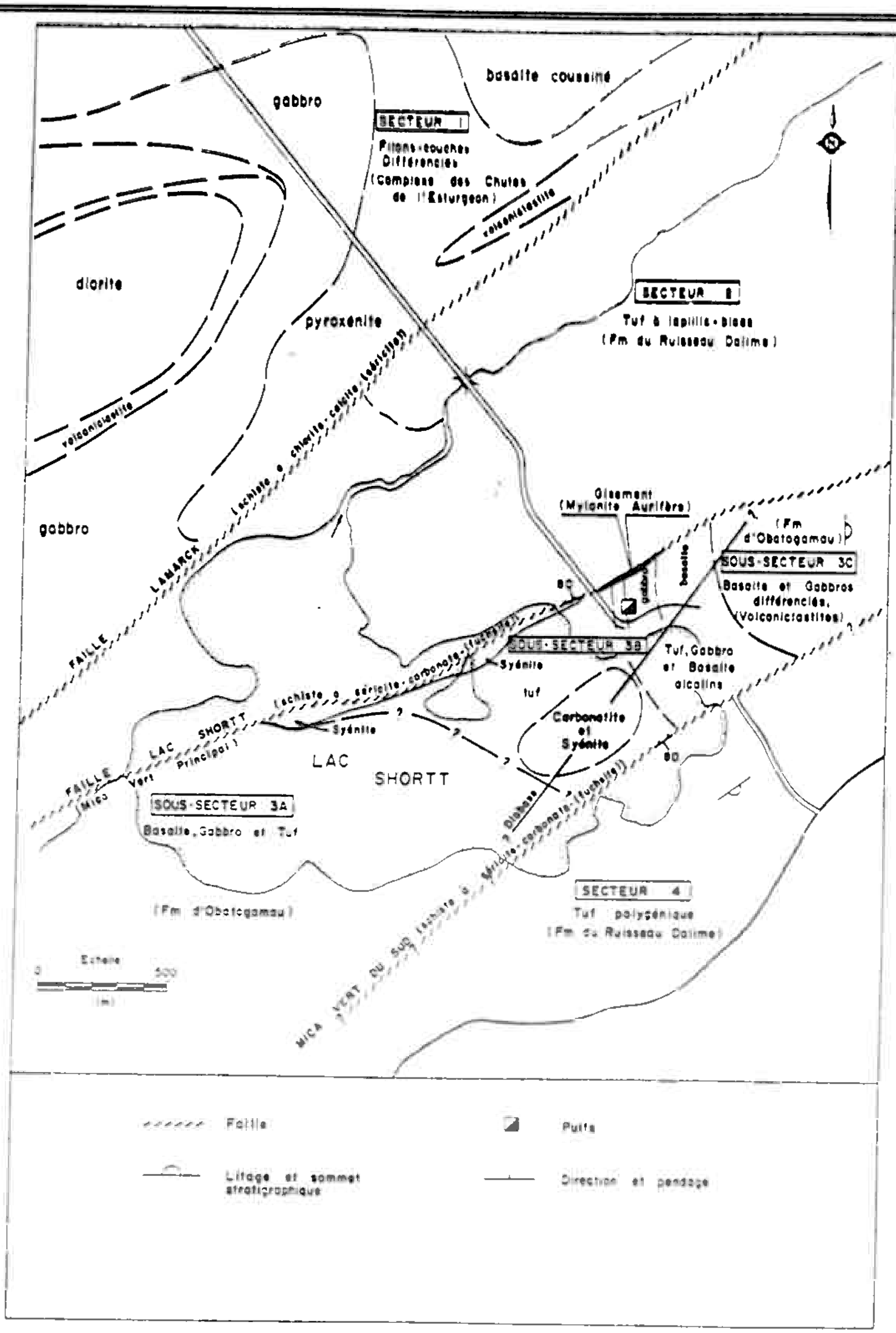

Figure 5B.1. Géologie de la propriété Lac Shortt. Les failles Lamarck, Lac Shortt et le Mica Vert du Sud divisent le territoire en 4 secteurs distincts.

pyroclastique en s'approchant des deux failles et dans le cas de la faille Lac Shortt, la texture de la roche varie de protomylonitique à ultramylonitique au coeur du Mica Vert Principal.

Secteur 3

Le secteur 3 est directement associé à la minéralisation aurifère. Les sous-secteurs 3A, 3B et 3C constituent respectivement les portions ouest, centre et est du territoire situé entre les 2 failles à mica vert.

Sous-secteur 3A

Le sous-secteur 3A est peu connu puisqu'il est en partie recouvert par les eaux du lac Shortt. Il comprend des roches volcaniques mafiques incluant des gabbros, des basaltes et des tufs.

Sous-secteur 3B

Le sous-secteur 3B hôte de la minéralisation aurifère et de l'intrusion de carbonatite, est formé d'ouest en est de tuf intermédiaire, de gabbro magnétique et de basalte. La masse gabbroïque qui constitue le protolite du gisement, est grossièrement orientée nord-sud avec un pendage vertical à fort vers l'est.

Les roches du sous-secteur 3B sont injectées de dykes de syénite et de carbonatite et elles sont fénitisées (altération sodi-potassique). On retrouve des syénites dans la masse principale de carbonatite dont elles constituent près de 20%, sous la forme de dykes satellites, au sein de zones altérées de façon intense (ultrafénites) et sous l'aspect de clastes dans la mylonite aurifère de la Zone Principale. Les syénites et ultrafénites du Lac Shortt sont hématisées ce qui leur confère une couleur rouge.

Sous-secteur 3C

Le sous-secteur 3C est formé d'une série de coulées basaltiques coussinées ou massives et de gabbro associés. On y retrouve, en plus, des intercalations mineures de volcanoclastite. Les roches de l'est du secteur 3 sont généralement peu altérées et elles présentent une polarité vers l'est.

Secteur 4

Le secteur 4 est séparé du domaine de la mine (secteur 3) par une seconde faille à mica vert, le Mica Vert du Sud, qui possède une orientation de 055°E avec un pendage de près de 80° vers le sud. La structure est anastomosée et matérialisée par un schiste à séricite et dolomite avec des traces de fuchsite. Au sud de la propriété, on observe une unité de tuf polygénique de la Formation du Ruisseau Dalime particulière de par son contenu en fragments accessoires de chert et de sulfure (pyrite). La dimension des débris varie des cendres aux blocs métriques de tri très pauvre. Le modèle de genèse proposé pour ce tuf évoque la destruction de matériel volcanique felsique par mécanisme explosif (Brisson, 1989a). Des critères sédimentaires et la taille des fragments indiquent un milieu de déposition proximal et une polarité vers le sud.

Géochimie

Au total, 48 échantillons de roches représentatives des différents secteurs de la propriété ont été analysés pour les éléments majeurs, le CO2 et le zirconium (Tableau 1). Le graphique alcalis versus silice indique que les roches analysées sont sous alcalines à l'exception des lithologies associées à la carbonatite (sous-secteur 3B), lesquelles se situent dans le champ alcalin (Fig. 5B.2a).

Roches sub-alcalines

Les tufs de la Formation du Ruisseau Dalime échantillonnés dans les secteurs 2 et 4 se situent dans le champ calco- alcalin du diagramme AFM (Fig. 5B.2b). Pour sa part, le Complexe des Chutes de l'Esturgeon (secteur 1) présente une signature tholéiitique particulièrement appauvrie en alcalis. Les roches sub-alcalines situées entre les 2 failles à mica vert (sous-secteurs 3A et 3C) entrent également dans le champ tholéiitique suggérant une filiation avec la Formation d'Obatogamau. La présence dans l'éponte inférieure du gisement de lave à gloméroccristaux de plagioclase typique de l'Obatogamau appuie cette conclusion.

Roches alcalines

Les tufs, les gabbros et les basaltes alcalins du sous-secteur 3B sont caractérisés par un rapport $Na_2O/K_2O > 1$ (Fig. 5B.3). Les syénites et ultrafénites du lac Shortt ont un contenu en alcalis de l'ordre de 12% avec un rapport K_2O/Na_2O élevé de près de 13 ce qui les distingue de la syénite du lac Opawica dont le contenu moyen en alcalis est d'environ 10% et le ratio K_2O/Na_2O de 0.30 (Fig. 5B.3). Les syénites associées à

Tableau 1: Analyses géochimiques des roches volcaniques provenant des différents secteurs de la propriété Lac Shortt; analyses par spectrographie cd (éléments majeurs), par fluorescence x (Zr) et par colorimétrie (C02).

SECTEURS	ECHANTILLON	SiO2 %	TiO2 %	Al2O3 %	FeO tot. %	MnO %	MgO %	CaO %	Na2O %	K2O %	P2O5 %	CO2 %	Zr ppm
1 COMPLEXE DES CHUTES DE L'ESTURGEON	106-1	46.10	1.42	11.50	14.50	0.13	4.37	5.79	3.86	0.05	0.16	8.48	125
	106-2	47.90	0.53	3.97	13.70	0.23	14.10	15.20	0.61	0.07	0.19	0.47	36
	107-2	43.80	0.54	3.21	10.10	0.15	14.00	12.00	0.04	0.12	0.32	9.44	31
	107-3	49.00	1.78	12.30	13.30	0.33	4.63	7.76	3.04	0.03	0.33	5.61	122
	S-21	50.30	0.61	3.58	11.40	0.21	15.10	17.40	0.72	0.07	0.05	0.12	32
	S-22	51.70	1.96	12.90	12.40	0.28	3.72	8.17	3.34	0.58	0.28	1.60	133
	S-23	43.60	1.57	13.90	19.60	0.22	6.13	7.29	2.91	0.26	0.15	0.08	49
	S-24	46.30	0.55	13.60	10.00	0.18	10.30	11.70	1.20	0.23	0.17	0.39	43
2 TUF (FM DU RUISSEAU DALIME)	111-3	45.60	0.68	11.40	9.43	0.18	7.93	8.39	4.15	< 0.01	0.19	11.21	66
	113-2	55.90	0.74	16.40	7.87	0.12	5.75	4.80	4.22	1.85	0.33	0.13	115
	2166	51.00	1.02	15.60	11.90	0.14	3.07	4.28	2.79	1.53	0.39	2.89	121
	3141	55.30	0.73	15.10	6.93	0.12	2.54	6.08	2.55	1.99	0.35	4.20	147
3 3A OUEST DU GISEMENT	103-1	51.90	1.90	13.50	12.20	0.24	4.01	7.54	2.15	0.15	0.23	1.53	125
	103-2	49.60	2.17	13.00	14.80	0.24	3.60	8.11	3.32	0.50	0.41	1.23	172
	111-1	54.50	0.80	12.60	8.55	0.19	3.95	6.04	3.98	1.05	0.21	4.02	98
	111-2	46.20	2.68	12.20	16.30	0.29	4.51	5.45	3.52	0.48	0.10	3.75	86
3B EPONTE INFERIEURE DU GISEMENT	05	55.60	0.48	14.10	5.08	0.13	1.86	7.53	5.17	3.25	0.17	2.95	< 1
	11-AE	48.85	0.93	15.80	10.39	0.17	3.61	5.36	5.80	2.62	0.15	3.12	25
	11-01	45.80	1.55	12.40	18.00	0.26	3.18	4.80	5.98	1.72	0.16	3.91	70
	12-08	42.80	0.58	14.40	10.90	0.19	7.74	6.83	4.35	0.16	0.12	6.83	< 1
	29	46.10	1.52	12.30	16.40	0.25	3.52	5.51	5.11	1.49	0.39	4.08	97
	30	55.00	0.47	15.70	5.25	0.12	2.37	4.87	7.81	2.19	0.39	3.83	119
	3149-1	48.40	1.21	14.10	13.50	0.22	5.39	9.54	4.07	0.37	0.26	0.91	71
	3153-1	44.70	1.60	14.00	15.70	0.22	6.00	6.63	4.36	1.55	0.43	1.28	58
	3235-8	50.40	0.58	14.50	7.14	0.25	3.96	7.30	5.70	3.65	0.44	4.79	75
3C EST DU GISEMENT	3149-2	46.30	0.56	14.30	11.40	0.19	9.06	11.10	1.48	0.42	0.12	2.51	23
	3149-3	47.30	0.54	14.30	11.90	0.21	7.60	11.30	0.55	0.15	0.17	0.33	35
	3149-4	45.80	0.46	18.20	9.43	0.16	8.56	12.40	1.35	0.15	0.12	0.30	27
	3149-5	50.60	0.83	14.70	12.50	0.19	6.95	9.21	0.79	0.01	0.13	1.02	53
	3149-6	44.00	0.76	13.20	10.10	0.22	3.51	13.10	1.56	0.10	0.23	5.11	50
	3149-7	56.20	0.60	16.30	6.47	0.05	3.23	3.71	4.55	0.46	0.23	0.44	120
	3149-8	50.20	1.04	13.70	14.30	0.24	5.07	7.85	2.30	0.12	0.29	0.06	37
4 TUF POLYGENIQUE (FM DU RUISSEAU DALIME)	3147-4	57.30	0.48	13.30	5.38	0.21	2.22	7.28	3.53	1.43	0.32	5.02	119
	3153-3	62.10	0.28	15.50	4.80	0.09	1.23	2.80	3.27	2.58	0.21	2.71	122
SYENITES INTRUSIVES (SOUS-SECTEUR 3B)	03	51.90	0.88	12.70	3.30	0.21	1.62	3.63	2.20	11.80	0.12	5.52	1470
	07	50.00	0.15	11.50	7.37	0.15	0.20	6.95	2.54	10.90	< 0.01	5.40	3803
	15	47.70	0.93	11.10	10.30	0.32	2.08	5.64	2.53	8.75	0.55	3.92	395
	09	53.80	0.12	21.60	2.37	0.07	0.40	2.95	0.65	12.70	0.01	2.15	2529
	3142-1	48.30	0.11	15.40	5.06	0.24	0.38	9.89	0.72	11.20	0.23	5.29	1872
	3235-3	51.50	0.20	17.30	3.20	0.16	0.87	7.24	1.86	9.19	0.19	6.31	1459
FENITES (SOUS-SECTEUR 3B)	11-BD	53.00	0.48	15.70	5.31	0.17	1.30	5.19	5.20	4.39	0.38	6.24	34
	11-C	55.50	0.13	19.60	2.33	0.05	0.29	2.58	2.10	12.00	0.10	2.04	1695
	3175-AG	44.35	0.48	12.70	5.68	15.68	3.48	11.95	2.05	7.90	0.37	8.65	134
	3175-BF	46.25	0.52	15.25	6.70	0.25	2.54	8.44	1.96	9.12	0.47	6.05	517
	3175-CE	52.40	0.24	20.40	4.19	0.12	1.63	5.05	1.70	11.15	0.09	3.41	1370
	3175-D	53.40	0.08	23.30	1.20	0.10	0.29	5.00	0.63	12.10	< 0.01	3.79	357
MYLONITE AURIFERE (SOUS-SECTEUR 3B)	20	37.00	1.40	10.60	13.80	0.34	3.57	7.85	0.25	3.34	0.17	11.20	51
	3166-2	37.90	1.43	10.40	15.50	0.32	3.08	8.14	1.59	5.48	0.19	15.48	84

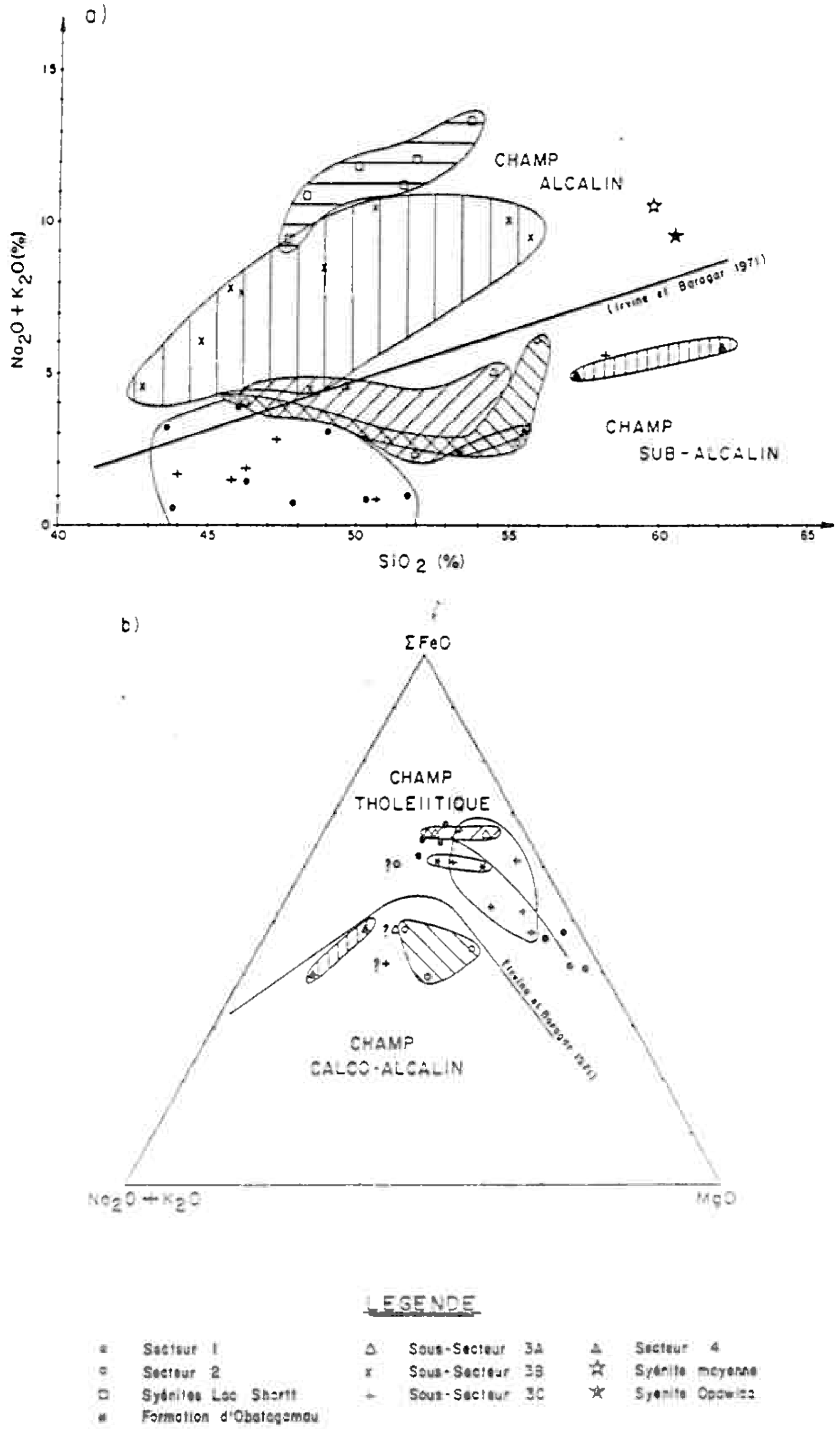

Figure 5B.2. Géochimie des roches de la mine Lac Shortt a) Diagramme alcalis versus Si02 et b) Diagramme AFM. Echantillons de la Formation d'Obatogamau tirés de Ludden et al., 1984; Syénite moyenne d'après Le Maître, 1976; Syénite Opawica d'après une moyenne de 21 syénites tiré de Averill et al., 1989.

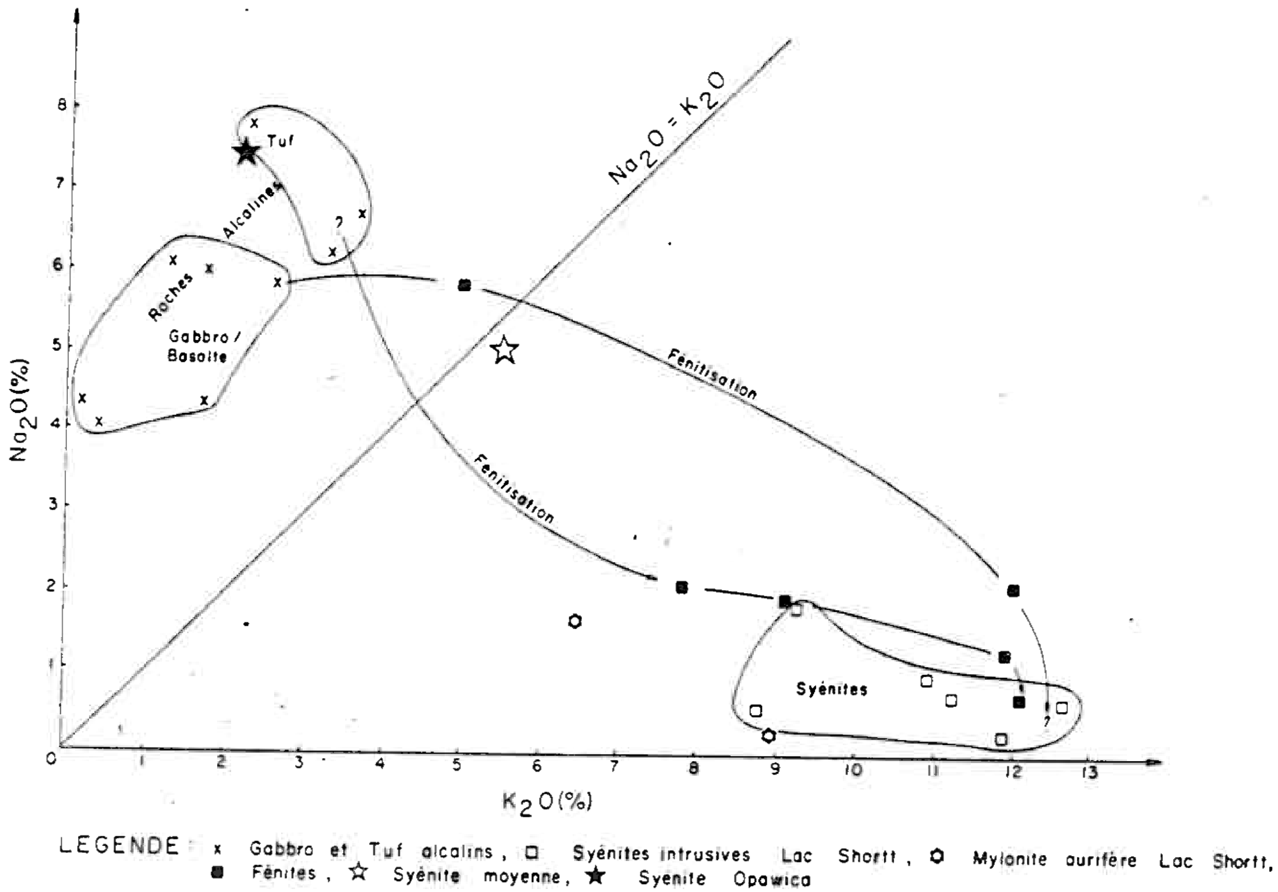

Figure 5B.3. Diagramme Na20 versus K20: caractérisation de la fénitisation à la mine Lac Shortt. Syénite moyenne selon Le Maitre, 1976; Syénite Opawica d'après une moyenne de 21 syénites tiré de Averill et al., 1989.

la faille Lac Shortt sont géochimiquement anormales par rapport à la moyenne des syénites qui selon Le Maitre (1976), présente un contenu en alcalis de 10% avec des quantités sensiblement égales de Na_2O et K_2O (ratio K_2O/Na_2O _ 1). Sur la propriété lac Shortt, le contenu particulièrement élevé en K_2O des syénites est imputable au phénomène de fénitisation qui a accompagné la mise en place de la carbonatite.

Minéralisation aurifère

Le coeur de la Zone Principale est constitué d'une mylonite à fragments de syénite boudinés et cataclasés dans une matrice carbonatée, hématisée, fénitisée et pyritisée. Le gisement a été formé par remplacement dans une zone de déformation fragile-ductile, du gabbro alcalin magnétique injecté de syénite. Le minerai renferme des quantités variables de microcline (35-75%), dolomite (10-50%) et pyrite (1-15%) avec de la séricite, hématite et fluorite mineures (Cormier et al., 1984). La teneur en or du minerai est proportionnelle à la quantité de sulfure puisque le métal précieux est majoritairement sous forme de grains de la taille de 8 microns en moyenne, accolés à la pyrite ou localisés dans des micro-fractures à l'intérieur de cette dernière. La pyrite aurifère s'est développée au détriment de la magnétite primaire contenue dans la roche gabbroïque hôte de la minéralisation (Morasse 1988, page 87). Dans la Veine Sud, le quartz est stérile et l'or est associé à la pyrite dans le halo d'altération.

Dans l'ensemble, le gisement a une forme tabulaire continue avec une dimension verticale indéterminée (Fig. 5B.4). La zone minéralisée a une épaisseur moyenne de 5.5 m dans les niveaux supérieurs avec une largeur de près de 300 m tandis qu'en profondeur, sa largeur diminue et elle s'épaissit par anastomose. Vue en section, la mylonite aurifère tend à se détacher graduellement du Mica Vert Principal sous le niveau 500. En plan, le gisement est bordé à l'ouest par une unité de tuf tandis qu'à l'est, le contact avec le basalte marque la limite de la zone minéralisée qui vient se buter sur la faille Lac Shortt (Fig. 5B.5). Au dessus du niveau 500, la largeur du gisement est contrôlée par la présence, dans l'éponte inférieure, d'une intrusion gabbroïque divisée en 2 branches principales (Fig. 5B.5a). La branche ouest du gabbro se pince sous le niveau 500 ce qui explique la diminution de la largeur du gisement en profondeur.

Altérations

Dans les tufs intermédiaires qui constituent les secteurs 2 et 4 on observe, surimposée à l'assemblage métamorphique à séricite calcite épidote, une augmentation progressive du contenu en séricite et carbonate en s'approchant des failles à mica vert. Dans l'éponte inférieure de la faille Lac Shortt (sous- secteur 3B) les structures aurifères ont un halo d'altération visible qui varie de 10 cm dans les structures secondaires à 40 m d'épais dans la Zone Principale. Les types d'altération associés à la minéralisation sont la carbonatisation, la fénitisation et la pyritisation. La silicification est une altération spécifique à la Veine Sud.

Deux sections de roches de l'éponte inférieure dont l'altération potassique varie graduellement de la roche mafique fraîche en apparence à la syénite (ultrafénite) ont été échantillonnées et analysées (Tableau 1). L'altération est marquée par un enrichissement en SiO_2, Al_2O_3 et K_2O et par une diminution en TiO_2, MgO, CaO, Na_2O et Fe0 tot. Le graphique Na_2O versus K_2O permet de visualiser l'effet de la fénitisation du gabbro et du tuf alcalin (Fig. 5B.3). Avec l'augmentation de l'altération potassique, les roches mafiques sodiques dont le ratio Na_2O/K_2O est initialement supérieur à 1

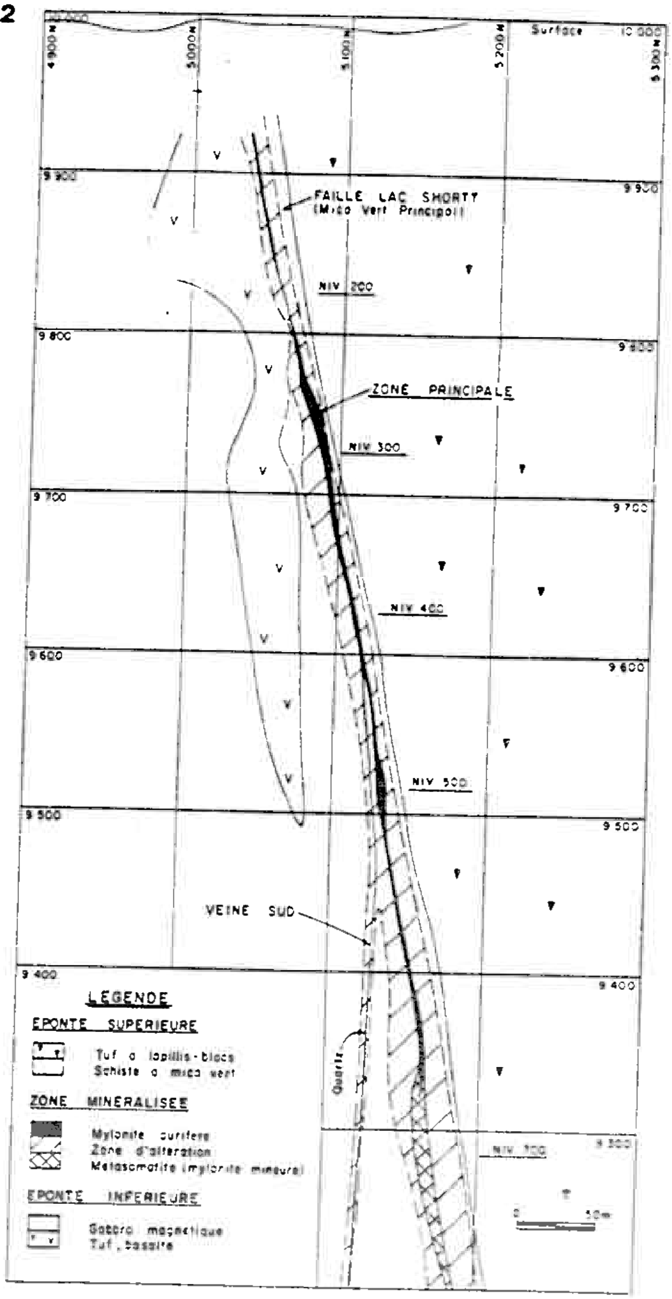

Figure 5B.4 Section transverse dans le gisement Lac Shortt. Zone Principale: section 5325E; Veine Sud: section composite 5350 à 5400 E.

voient ce rapport diminuer jusqu'à ce que la quantité de K_2O devienne supérieure à la quantité de Na_2O. A[1] la limite, les ultrafénites atteignent une composition et un aspect identiques à ceux des syénites échantillonnées au coeur de la masse principale de carbonatite ou dans les dykes satellites. L'étude de la fénitisation des roches au lac Shortt par ajout de potassium est toutefois incomplète puisqu'elle ne rend pas compte du transfert du sodium qui accompagne généralement ce

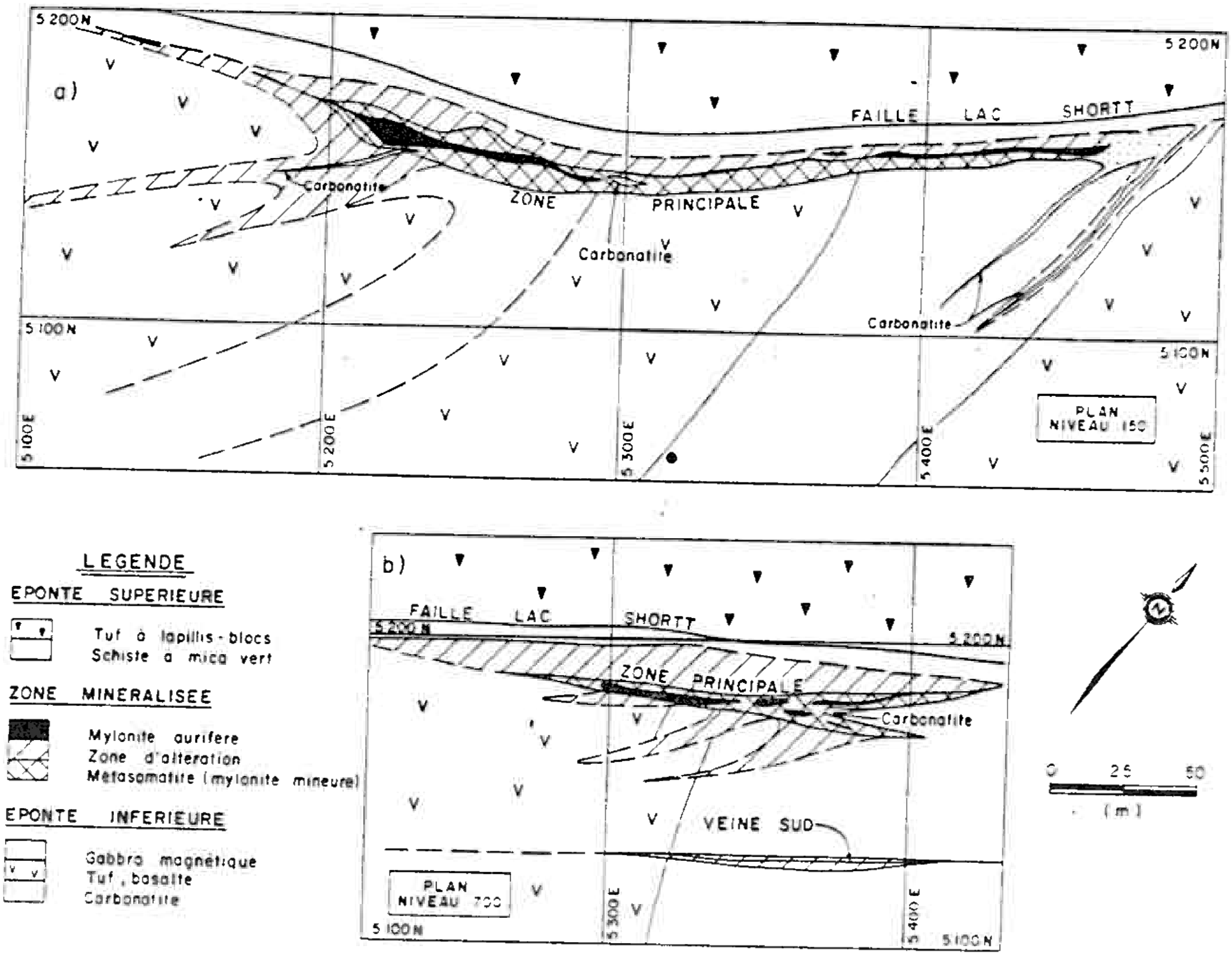

Figure 5B.5. Géologie des niveaux a) 150 et b) 700 à la mine Lac Shortt.

phénomène. L'albitisation des cristaux de plagioclase dans le gabbro magnétique hôte de la minéralisation (Morasse, 1988) suggère que l'affinité alcaline sodique des mafites du sous-secteur 3B est imputable à un processus métasomatique plutôt que primaire. L'albitisation à grande échelle est associée à un ajout d'amphibole sodique, la magnésio-arfvedsonite (Morasse, 1988) en bordure de la carbonatite. On retrouve, surimposées à ce métasomatisme sodique, des plages locales d'altération potassique jumelées à une hématisation visible ayant amené la roche mafique vers une composition syénitique anormalement élevée en K_2O.

La restriction de la fénitisation aux roches de l'éponte inférieure du Mica Vert Principal, son association spatiale avec la mylonite aurifère et le fait que les syénites ou fénites observées dans les épontes de la Zone Principale sont aurifères indiquent que l'intrusion alcaline est la source la plus probable des fluides minéralisants. Dans ce con texte, les zones de cisaillement agissent à la façon de soupapes en favorisant la migration, par pulsations, des fluides hydrothermaux (Sibson et al., 1975). La mylonite aurifère constitue la zone de déformation vers laquelle les solutions hydrothermales réactives riches en H_2O, CO_2 et alcalis issus de la carbonatite auraient été canalisées.

Le protolithe gabbroïque de la mylonite a servi de piège structural de par sa compétence, et de piese chimique par réduction de la magnétite et précipitation de l'or avec la pyrite.

Géologie structurale

La faille Lac Shortt, le Mica Vert du Sud et la mylonite de la Zone Principale constituent les accidents structuraux majeurs associés à la minéralisation aurifère. Les deux failles à mica vert résultent de la déformation ductile des tufs de composition intermédiaire situés de part et d'autre des structures et leur similarité suggère qu'elles font partie d'une même dynamique globale. Dans le cas de la Zone Principale, le protolithe est un gabbro compétent injecté de syénite qui a plutôt réagi à la déformation de façon fragile ductile en développant une texture mylonitique.

Les éléments structuraux secondaires comprennent la Veine Sud, les micro-fractures de tension stériles remplies d'albite et de quartz concentrées dans la zone minéralisée et les failles subsidiaires. Ces dernières, orientées N à NNE font un angle de 0 à 45° avec la Zone Principale sur laquelle elles viennent se greffer (Fig. 5B.5).

La Zone Principale présente une texture interne irrégulière et une nature anastomosée autant à l'échelle de la mine qu'en microscopie. Plusieurs familles de foliations se sont développées dans la matrice de la mylonite avec une orientation qui varie de ENE à franchement EW avec un pendage de 60° à 80° vers le nord.

On observe généralement peu de quartz libre à la mine Lac Shortt puisque la silice à été littéralement pompée lors de la formation des syénites métasomatiques. L'introduction de silice dans la Veine Sud est probablement postérieure au pic de la fénitisation et dans ce contexte, l'analyse de cette structure, même si elle permet de contourner la complexité de la Zone Principale, ne s'applique qu'aux derniers stades de la déformation.

La Veine Sud s'inscrit dans un réseau de failles de cisaillement conjuguées disposées en échelon, orientées à 045°/85° et 225°/85°. L'enveloppe d'altération de la Veine Sud a une largeur moyenne de 5 m et elle est sub-verticale (Fig. 5B 4). La partie interne de la structure est injectée de veines de quartz cisaillées également disposées en échelon. Une foliation sub-verticale orientée à 217° et des linéations d'étirement plongeant à environ 60° vers le SW suggèrent un mouvement oblique inverse avec une composante dextre sous l'influence d'une contrainte principale maximum sub-horizontale orientée NW-SE (307°). Cette dynamique, lorsqu'étendue à l'ensemble de la mine, implique un mouvement du même type au sein des failles à mica vert. Dans le modèle proposé à la figure 5B.6, la Zone Principale est sub-parallèle au champ d'aplatissement maximal de l'ellipsoïde de la déformée ce qui est en accord avec le boudinage des dykes de syénite et la présence de veinules de tension subhorizontales dans la mylonite. Dans ce contexte, les failles subsidiaires sont assimilées à des Riedels senestres de type antithétique.

La cinématique inverse dextre proposée est compatible avec les mouvements les plus récents du Mica Vert Principal déduits des fabriques CS, des plis d'entraînement et des gradins de stries (Boisvert, 1986 et Morasse, 1988). Cependant, des mouvements secondaires contradictoires, de type normal-senestre ont également été reconnus (Boisvert, ibid.) mettant en évidence la complexité de l'histoire de la déformation dans la faille Lac Shortt.

CARBONATITE

La carbonatite est localisée dans l'éponte inférieure de la faille Lac Shortt, à environ 250 m au SW de la mine (Fig. 5B 1). D'une superficie approximative de 300 m x 500 m, la carbonatite, de type sovite, est majoritairement constituée de calcite à laquelle est affiliée environ 20% de syénite. Les roches avoisinantes ont subi une fénitisation typique des roches associées aux carbonatites. Elles sont, de plus, recoupées par de nombreux dykes de carbonatite et de syénite dont la structure très irrégulière témoigne d'une mise en place explosive sous pression de fluide.

Des signes de déformation tels des mâcles courbes dans la calcite, une texture mylonitique développée en bordure nord de l'intrusion et la déformation à divers degrés des dykes de carbonatite et de syénite dans la mylonite aurifère témoignent d'une mise en place syntectonique du complexe alcalin. Des critères texturaux tels la présence de roches hybrides et de dykes bimodals à carbonatite et syénite indiquent que les 2 lithologies sont comagmatiques. Des profils de terres rares comparables dans leur enrichissement particulier en terres rares intermédiaires (Nd à Tb) appuient cette conclusion (tableau 2 et Fig. 5B.7).

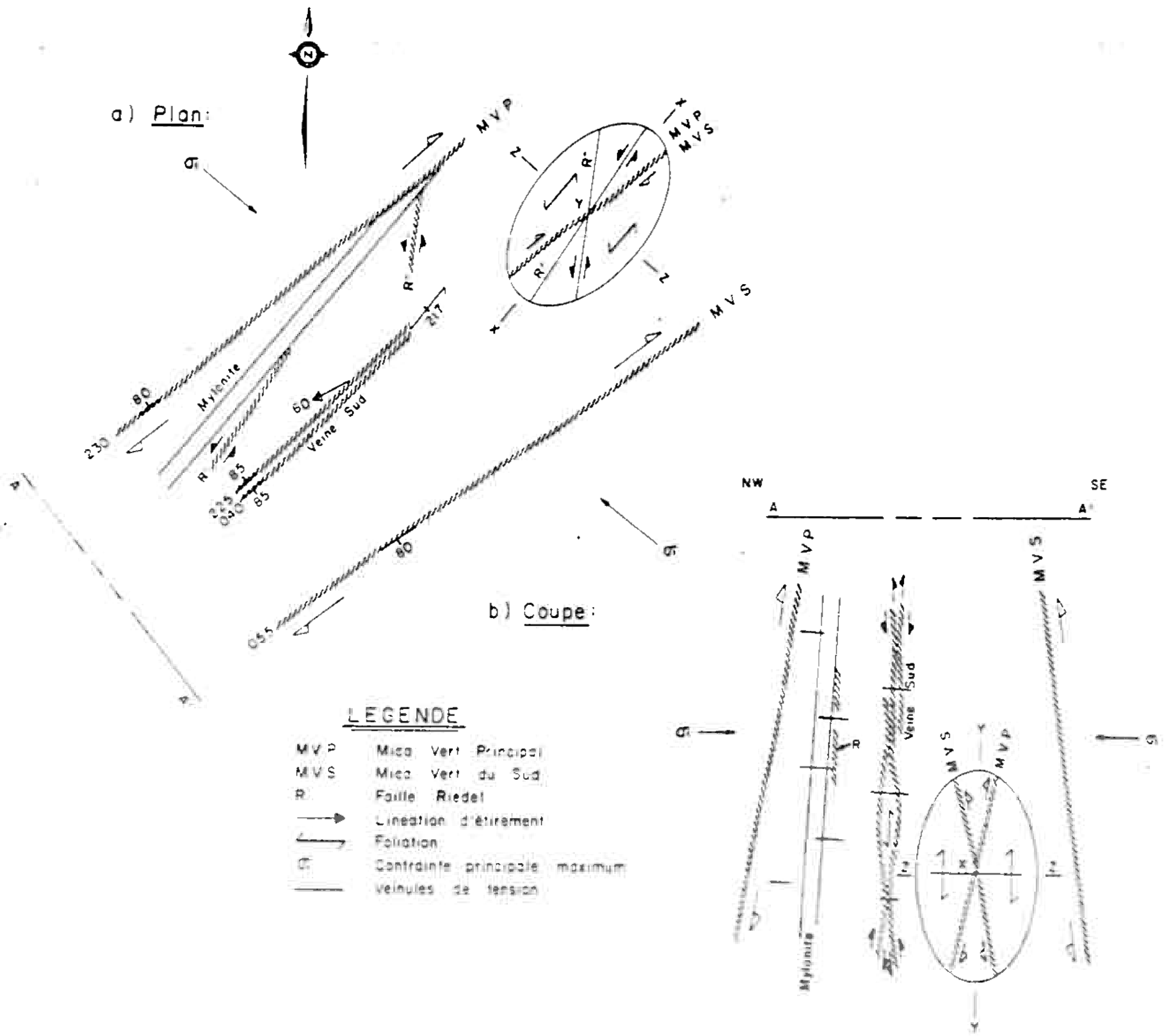

Figure 5B.6. Analyse des éléments structuraux à la mine Lac Shortt.

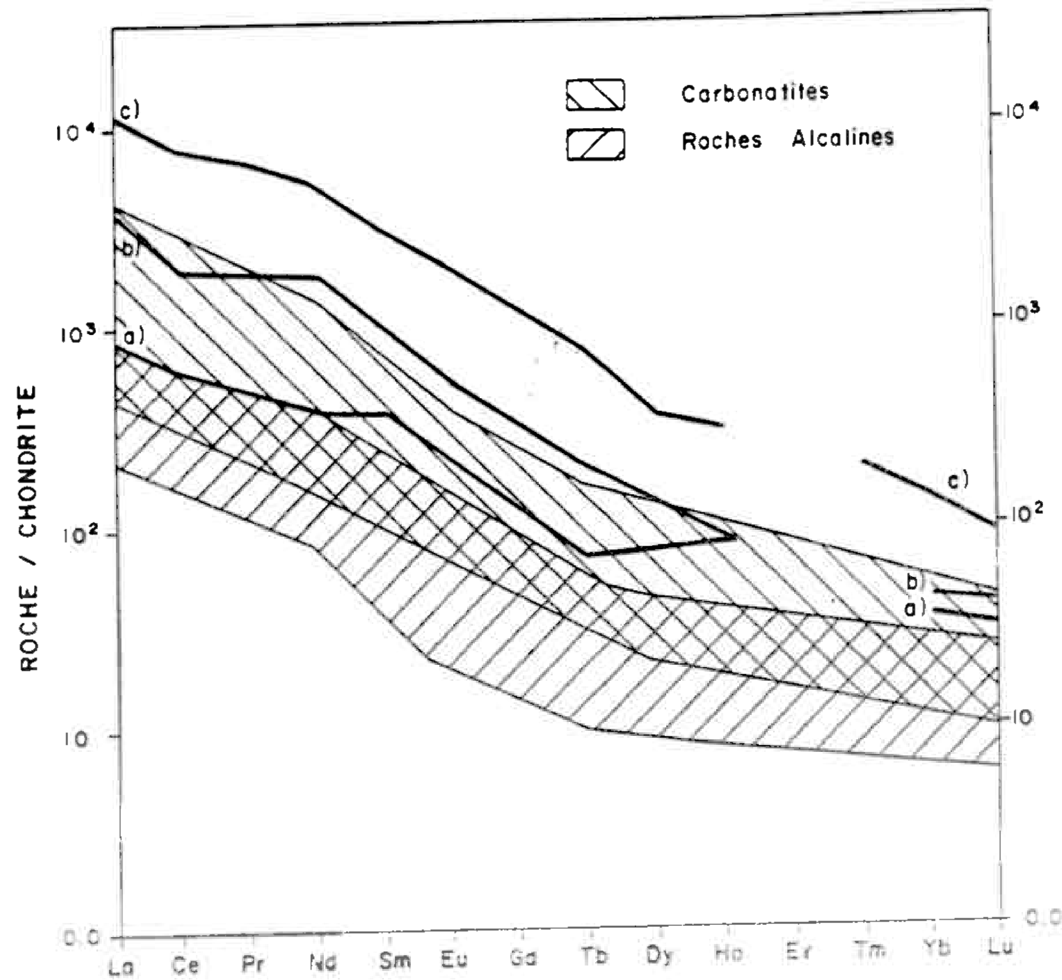

Figure 5B.7 Patrons des terres rares moyens pour la syénite (a), la carbonatite (b) et sol résiduel (c) du Lac Shortt. Normalisations d'après un composé de 9 chondrites (Haskin et Frey, 1968). Champs des carbonatites et des roches alcalines d'après Moller et al., 1980.

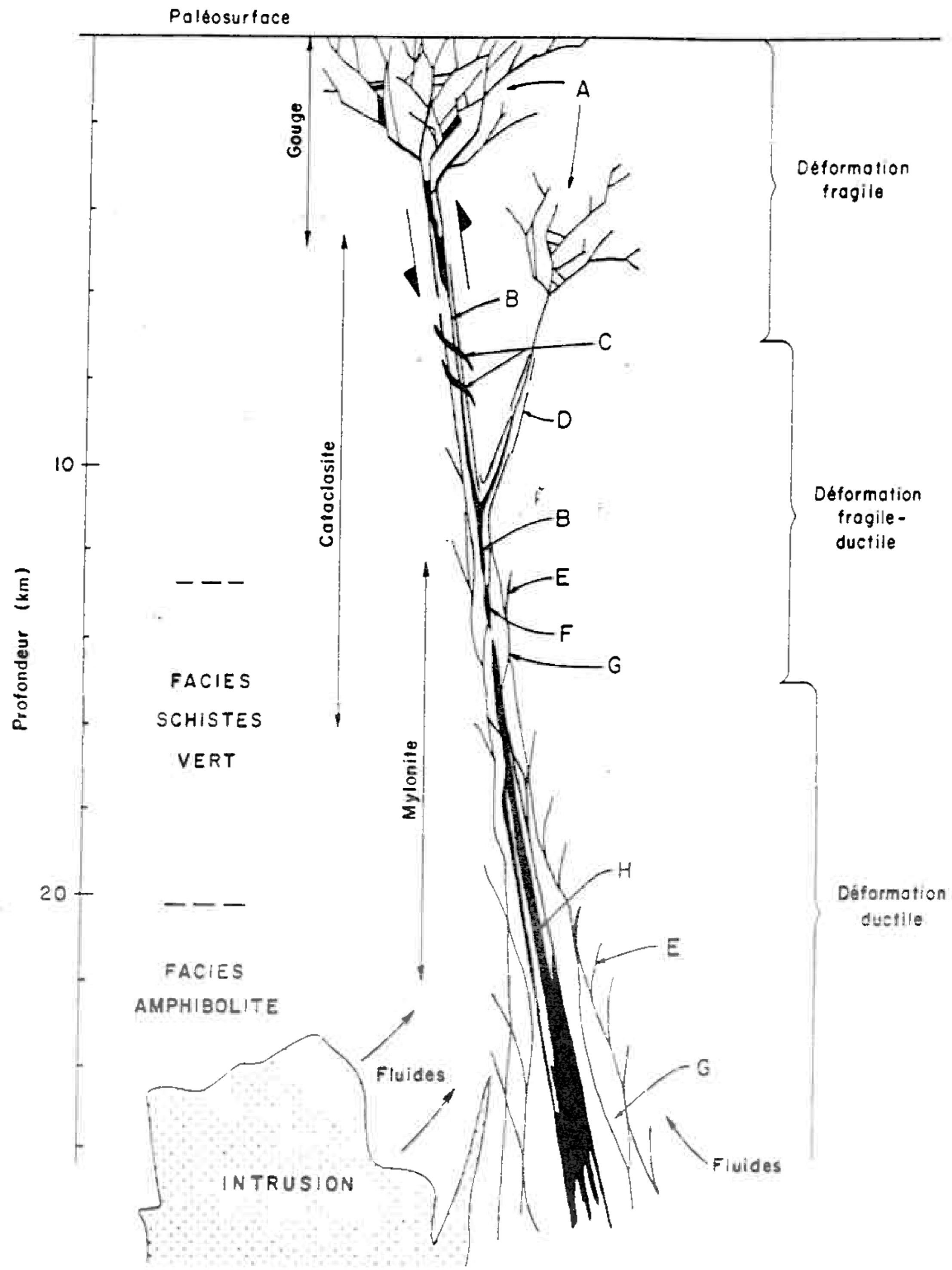

Figure 5B.8. Représentation schématique idéalisée d'une zone de cisaillement aurifère. Veines bréchiques (A), veine de cisaillement (B), veines de tension (C), faille Riedel (D), foliation (E), boudinage (F), anastomose (G), veine de remplacement (H), (modifié de Sibson et al., 1988).

Tableau 2: Concentration en éléments traces et terres rares de la syénite, de la carbonatite et du sol résiduel du lac Shortt; analyses par activation neutronique.

	SYENITE	CARBONATITE				SOL RESIDUEL	
		DYKES		INTRUSIF PRINCIPAL		REGOLITE	SAPROLITE
(ppm)	7995	7762	7768	7539	7578	01-29	01-36
Ba	1035.00	1316.00	572.00	1724.00	2114.00	> 20 000	> 20 000
Be	-	-	-	-	-	7.70	8.30
U	338.70	-	-	11.20	22.20	130.00	81.00
Th	87.00	-	-	37.70	52.50	129.00	215.00
Zr	6267.00	664.00	525.00	1608.00	1097.00	274.[illegible]	175.00
Sc	0.58	0.80	0.43	0.72	0.79	10.00	1.90
La	286.78	987.02	855.83	1098.38	1701.92	4690.00	4990.00
Ce	531.00	1470.00	1403.00	1530.00	1965.00	8660.00	9280.00
Pr	-	-	-	-	-	950.00	950.00
Nd	229.00	914.00	975.00	1120.00	961.00	3680.00	4170.00
Sm	66.44	132.30	138.00	175.74	158.63	624.00	704.00
Eu	15.00	28.00	30.30	39.30	34.30	170.00	178.00
Tb	3.20	8.00	8.10	9.36	7.33	42.00	40.00
Dy	-	-	-	-	-	162.00	153.00
Ho	5.50	6.00	6.00	-	-	25.00	23.00
Tm	-	-	-	-	-	7.60	6.10
Yb	8.40	9.10	9.90	6.30	8.40	38.00	34.00
Lu	1.55	1.31	1.61	1.14	1.44	3.90	3.50
Y	-	-	-	-	-	493.00	465.00

D'autres travaux sur ce problème sont présentement en cours.

La carbonatite du Lac Shortt est constituée à 90% de calcite rose. Les minéraux accessoires sont les micas, la magnétite, le feldspath potassique, l'apatite, l'aegyrine et l'amphibole bleue. Les minéraux traces identifiés sont le zircon, la fluorite, la barytite, la célestite, la magno-columbite (Prud'homme, communication personnelle) et les minéraux de terres rares soient la monazite, un phosphate associé à l'apatite et la bastnaésite, un carbonate exsuolvé par la calcite lors de son refroidissement. La teneur en lanthanides de la carbonatite est de l'ordre de 0.5% (Tableau 2). Avec un contenu en lanthane de près de 3000 x chondrite et un rapport La/Lu d'environ 1300, la carbonatite du Lac Shortt présente le patron de terres rares fortement penté typique des roches alcalines (Fig. 5B.7).

La météorisation de la carbonatite au Tertiaire a causé l'effondrement du complexe alcalin et la création d'une fosse de mort-terrain d'au moins 200 m de profondeur sous le lac Shortt. Un trou de sondage foré dans la fosse a intersecté un sol résiduel de 13.7 m d'épais à l'interface entre le drift glaciaire et une syénite karstique (Quirion, 1989). Le sol résiduel est une brèche de phosphorite à francolite zonée avec un saprolite (ferricrète) à la base et une latérite (silicrète) au sommet. Le paléosol renferme en moyenne 1.5% de terres rares + Y. Les minéraux de terres rares identifiés sont la monazite et la francolite, une apatite secondaire. Le paléosol a une teneur en lanthane correspondant à plus de 10000 x chondrite et un rapport La/Lu de près de 1300. Le patron des terres rares normalisé du sol résiduel témoigne de l'enrichissement supergène de la roche-mère, ici la carbonatite, par un facteur de l'ordre de 3. Le sol résiduel du lac Shortt se caractérise par un contenu avantageux en terres rares intermédiaires, principalement l'europium, lesquelles constituent près de 5% des terres rares en présence.

DISCUSSION

La déposition de matériel pyroclastique proximal au sud de la mine Lac Shortt est liée à l'édification de centres volcaniques calco-alcalins (Formation du Ruisseau Dalime) au dessus de laves tholéiitiques (Formation d'Obatogamau). Dans cet environnement, la faille Lac Shortt et le Mica Vert du Sud ont pu constituer des failles de croissance comparables aux structures anciennes décrites par Dimroth et al. (1983) au sud de la ceinture de l'Abitibi. Selon cette optique, les failles à mica vert inverses à fort pendage qui sont aujourd'hui le siège d'une carbonatation et d'une foliation intenses témoignent d'une évolution structurale plus complexe. L'ultime mouvement oblique inverse-dextre que l'on a déduit des fabriques observées dans la faille Lac Shortt ne serait que le reflet de la phase de compression finale de l'orogenèse kénoréenne.

CONCLUSION

Le gisement Lac Shortt se situe dans la portion fragile ductile d'un cisaillement aurifère typique dont le modèle a été développé par Sibson (1977) et Sibson et al. (1988) (Fig. 5B.8). La Zone Principale a été formée par mylonitisation et remplacement du gabbro magnétique compétent localisé dans l'éponte inférieure de la faille Lac Shortt. La minéralisation est associée à la fénitisation qui a accompagné la mise en place de la carbonatite au sein d'un domaine volcanique tectoniquement actif contrôlé par les deux failles à mica vert. L'augmentation des effets de la déformation ductile liée à la migration des fluides minéralisants par diffusion en profondeur est responsable de l'épaississement et de la dilution de la zone d'altération par anastomose sous le niveau 500.

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