GEOLOGICAL SURVEY OF CANADA

OPEN FILE 2158

Litho-tectonic framework and associated mineralization of the eastern extremity of the Abitibi Greenstone Belt, Quebec (Field Trip 3)

edited by

J. Guha
E.H. Chown
R. Daigneault

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Energy, Mines and Resources Canada / Énergie, Mines et Ressources Canada

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GEOLOGICAL SURVEY OF CANADA

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LITHO-TECTONIC FRAMEWORK AND ASSOCIATED MINERALIZATION OF THE EASTERN EXTREMITY OF THE ABITIBI GREENSTONE BELT [FIELD TRIP 3]

EDITED BY

J. GUHA, E.H. CHOWN and R. DAIGNEAULT

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8TH IAGOD SYMPOSIUM

FIELD TRIP GUIDEBOOK

8th IAGOD SYMPOSIUM

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8^{e} SYMPOSIUM DE l'IAGOD
8TH IAGOD SYMPOSIUM
EXCURSION #3

CADRE LITHOTECTONIQUE ET MINERALISATION ASSOCIÉE DE L'EXTREMITE EST DE LA ZONES DES ROCHES VERTES DE L'ABITIBI (QUEBEC).

LITHOTECTONIC FRAMEWORK AND ASSOCIATED MINERALIZATION OF THE EASTERN EXTREMITY OF THE ABITIBI GREEN STONE BELT, QUEBEC.

Lundi 6 août/ Monday Aug. 6

Ottawa-Chibougamau (par Autobus/by bus)
Départ/Departure 7:00 a.m. ("Commons" - Carleton University) (Voir la carte/see enclosed map)

Jour/Day 1 (7/08)

Cadre régional / regional setting

Obatogamau, Waconichi, Gilman and Chibougamau
Formations; Brêches de Queylus/Queylus breccias, Roberge sill, section Mine Lemoine/Lemoine mine section; Complexe du lac Doré/Dore Lake Complex.

Jour/Day 2 (8/08)

AM Copper Rand/Portage Mines
PM Norbeau and Taché deposits

Jour/Day 3 (9/08)

Caopatina Segment
(volcanologie, sedimentologie, structure et hydrothermalisme-syn-volcanique, minéralisation aurifère/gold mineralisation)

Jour/Day 4 (10/08)

AM Mine Joe Mann mine + surface
PM Autour/around mine Opemisca - surface

Jour/Day 5 (11/08)

AM Lac Shortt Mine + surface
PM départ pour Ottawa - arrivé vers minuit
departure for Ottawa - arriving around midnight

DE/FROM 6/08 - 11/08 addresse/address
Hotel Chibougamau
473, 3e rue
Chibougamau, Qc
G8P 1N6
Tél 418-748-2669
Fax 418-748-2107

COLLABORATEURS / CONTRIBUTORS

AVANT-PROPOS

Cette excursion met l'emphase sur l'environnement géologique des segments de Chibougamau et de Caopatina- Desmaraisville. Ces segments appartiennent tous deux à la portion nord de la Sous-province de l'Abitibi et s'étendent du Front de Grenville, à l'est, jusqu'aux environs de la mine d'or du Lac Shortt (150 km à l;ouest), un région qui a été la cible d'une forte exploration minérale au cours des quatre dernières années. Les objectifs de l'excursion sont de présenter une vue synthétique de la stratigraphie volcanique (de conditions variant d'eaux profondes à subaériennes), des filons-couches mafiques à ultramafiques et des plutons (tonalite à carbonatite), des structures (synvolcaniques à grenvilliennes) et de leurs contrôles sur la minéralisation. Le cadre régional donnera une perspective dans laquelle les différents gîtes de la région pourront être placés.

Les livrets-guides d'excursion précédents (Allard et al. 1979; Guha and Gobeil, 1984) ont davantage porté sur l'environnement géologique du segment de Chibougamau proprement dit. Ce livret-guide, bien qu'inspiré en partie des ouvrages précédents, tentera d'apporter une dimension nouvelle en considérant la géologie régionale en relation avec des modèles tectoniques globaux, qui tenteront d'expliquer à la fois les processus de formation des roches supracrustales et intrusives ainsi que les mécanismes de la déformation régionale. Deux milieux de déposition contrastant, une plaine de basalte au sud, et un arc volcanique au nord, contrôlent la minéralisation syngénétique. Ces deux domaines semblent toutefois liés par une histoire plutonique et structurale commune. De plus, une longue histoire protérozoïque marque l'ensemble de la région et culmine vers l'orogenèse grenvillienne au Protérozoïque inférieur.

Ce livret-guide a été structurée de façom à présenter une mise à jour de la géologie régionale suivie de la description des arrêts. La première section représente un condensé provenant de plusieurs auteurs. Les points de vue peuvent varier sur un sujet donné ce qui devrait générer des discussions stimulantes entre les participants et les différents auteurs présents.

REMERCIEMENTS

La préparation de cette excursion vient d'un effort collectif de plusieurs personnes et organismes. Mentionnons les contributions de Minnova Inc., Meston Lake Resources, Westminer Canada Ltd. du Ministère de l'Energie et des Ressources du Québec (contribution ## 90-5130-07) et de l'Université du Québec à Chicoutimi.

FOREWORD

The field trip will focus on both the Chibougamau segment and the Southern Caopatina segment of the Northern Abitibi Belt stretching from the Grenville Front to the Lac Shortt gold mine (150 kms to the west), an area which as been intensely explored over the last four years. The objective is to present an overall view of the volcanic stratigraphy (from deep water to subaerial environment), mafic to ultramafic sills and plutonic intrusions (tonalite to carbonatite), structures (synvolcanic to Grenvillian) and their various controls on ore formation. This leads to an examination of the exploration significance of various types of deposit with respect to their regional setting.

Previous guidebooks (Allard et al. 1979; Guha and Gobeil, 1984) have been concerned chiefly with the Chibougamau region. An attempt is made in the regional synthesis of this guidebook to consider the regional geology in the light of possible global tectonic settings, both at the time of formation of the supracrustal and intrusive rocks, and during regional deformation. Two contrasting depositional settings, a southern basalt plain and a northern volcanic arc, affect syngenetic or deposits in the two domains, whereas the entire region is united by a common deformation and intrusion pattern. In addition the area was marked by geological processes through out the Proterozoic and culminating in the Late Proterozoic Grenvillian Orogeny.

The guidebook has been structured to present an updated version of the geology of the area followed by stop/deposit descriptions. The latter section presents some divergent views and ideas which hopefully will generate stimulating discussions between the participants and the various authors.

ACKNOWLEDGMENTS

Preparation for this excursion has been a team effort, involving the efforts of Minnova Inc. Meston Lake Resources, Westminer Canada Ltd., the Ministère de l'Energie et des Ressources du Québec (contribution #90-5130-07), and the Université du Québec à Chicoutimi.

PART 1: GEOLOGICAL SETTING OF THE EASTERN EXTREMITY OF THE ABITIBI BELT

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Introduction

The Chibougamau-Caopatina region (CCR) is located in the northeastern extremity of the Abitibi Belt (Fig. 1). Differences in lithology, such as the occurrence of komatiites in the south, and the greater abundance of plutons and mafic to ultramafic sills in the north, as well as a difference in stratigraphic thickness, prompted Dimroth et al. (1982, 1984) to divide the Abitibi Belt into a northern, Internal Zone and southern, External Zone (Fig. 1). In the same manner, Ludden et al. (1986) made a distinction between the same two segments, naming them the Northern Volcanic Zone and the Southern Volcanic Zone. The (CCR) forms part of the Northern or Internal Zone which is bounded to the north by the Opatica Belt, and to the east by the Grenville Province. The contact zone between the two provinces, the Grenville Front (Wynne-Edwards,1972; Rivers and Chown, 1986), is marked by a disruption in the common East-West tectonic trend. The metamorphic grade, typically greenschist facies, reaches the amphibolite facies near the Grenville Front.

The CCR (Fig. 2) is the most extensively studied part of the northern Internal Zone. Initial mapping was carried out by Norman (1937), Retty and Norman (1938), Beach (1941a, b) and later continued by Gilbert, (1949, 1955). Duquette (1970) was the first to establish a coherent stratigraphy, composed of the older Roy Group and the younger, essentially epiclastic Opemisca Group. Allard (1976) subsequently redefined the volcanic formations of the Roy Group. Recent mapping by Tait and Chown (1987), Tait et al. (1987), Lauzière and Chown (1988), and Lauzière et al. (1989) led to the recognition of a sedimentary unit interstratified with the Roy Group volcanic rocks in the southern part of the CCR (Sharma et al., 1987). Detailed mapping the Québec Ministry of Energy and Resources has revealed a need to reevaluate the stratigraphy in certain areas of the CCR. Recently, Mueller et al. (1989) described the evolution of this region in terms of volcanic cyclicity, sedimentation, and plutonic emplacement history, and Daigneault and Allard (1990) described the structural setting, and presented detailed overview of the formations and groups in the Chibougamau segment.

Volcano-Sedimentary Supracrustal Sequence (General Aspects)

Archean volcano-sedimentary sequences are generally characterized by volcanic cycles and inter-stratified sedimentary units (Windley, 1986) and the CCR is no exception (Allard et al., 1979; 1985; Allard and Gobeil, 1984; Dimroth et al., 1985, Mueller, 1986). Recent advances in the study of both modern and ancient environments enables a direct comparison to be made between ancient sequences and those from modern tectonic regimes. As a preliminary step, the significance of the volcanic and sedimentary phases of the supracrustal suite will be considered based on studies from the CCR, as well as other Archean and modern areas.

Volcanic phase

Ideal volcanic cycles are represented by an extensive komatiitic and/or tholeiitic base and an upper andesitic-dacitic-rhyolitic part (Goodwin, 1982, Dimroth et al., 1982, Jensen, 1985). The former has been interpreted as a subaqueous basalt plain, composed of massive, pillowed, and brecciated basalts and comagmatic sills (Dimroth et al., 1982, Mueller et al., 1989). The felsic part of the cycle represents a volcanic edifice (Ayres, 1982, Easton, 1984) of highly variable dimension (from 0.2-2.5 km thick / 5-20 km in diameter to 10-14km thick/80-120 km in diameter; Goodwin, 1982; Mueller et al., 1989). The felsic lava flows, pyroclastic flows and their reworked counterparts are predominantly the products of subaqueous deposition. Hydrothermal alteration of these centres is a common feature in the CCR (Guha, 1984). Preservation of subaerial deposits

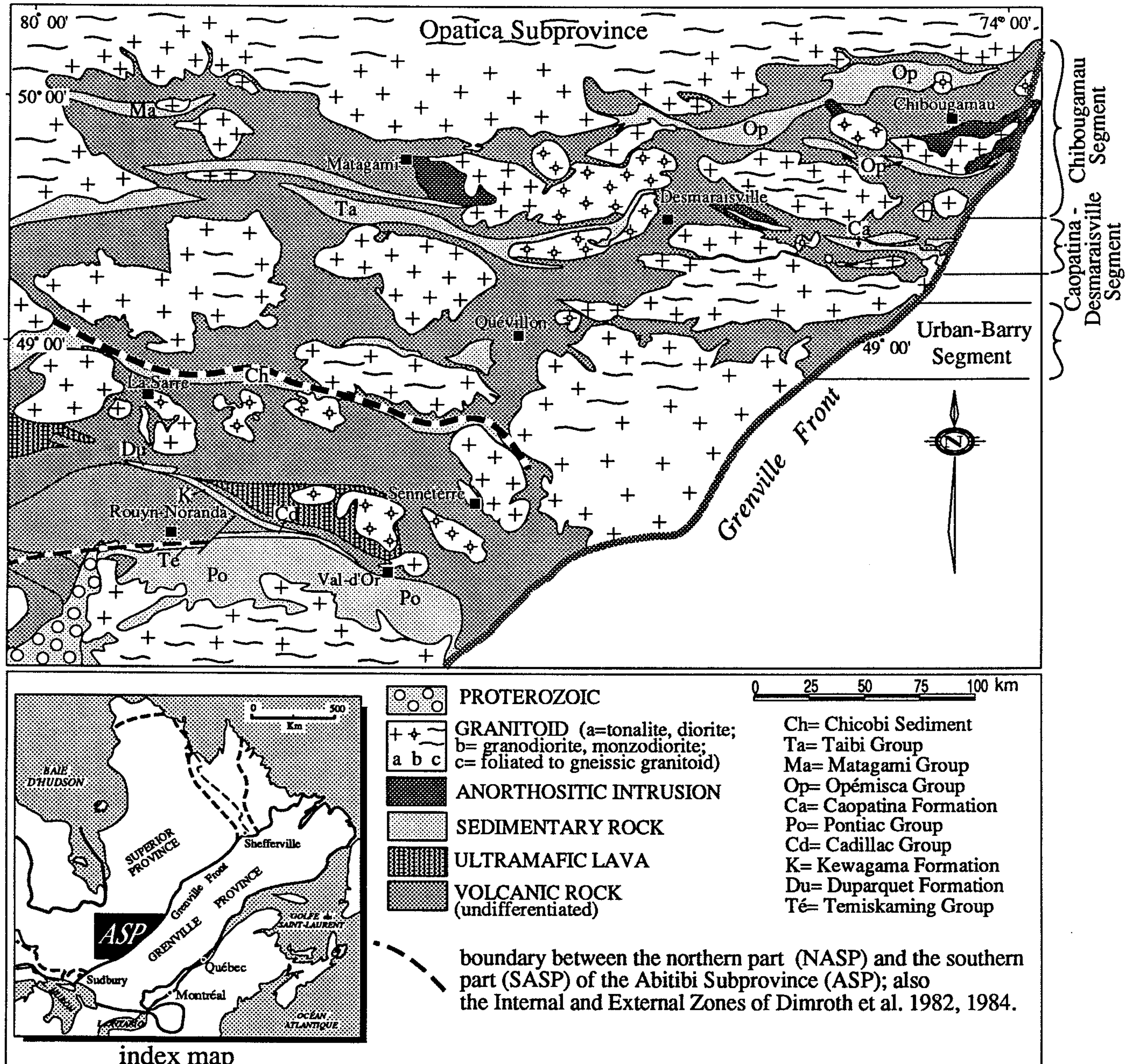


Figure 1. General Geology of the Abitibi Sub-Province and location of the studied area.

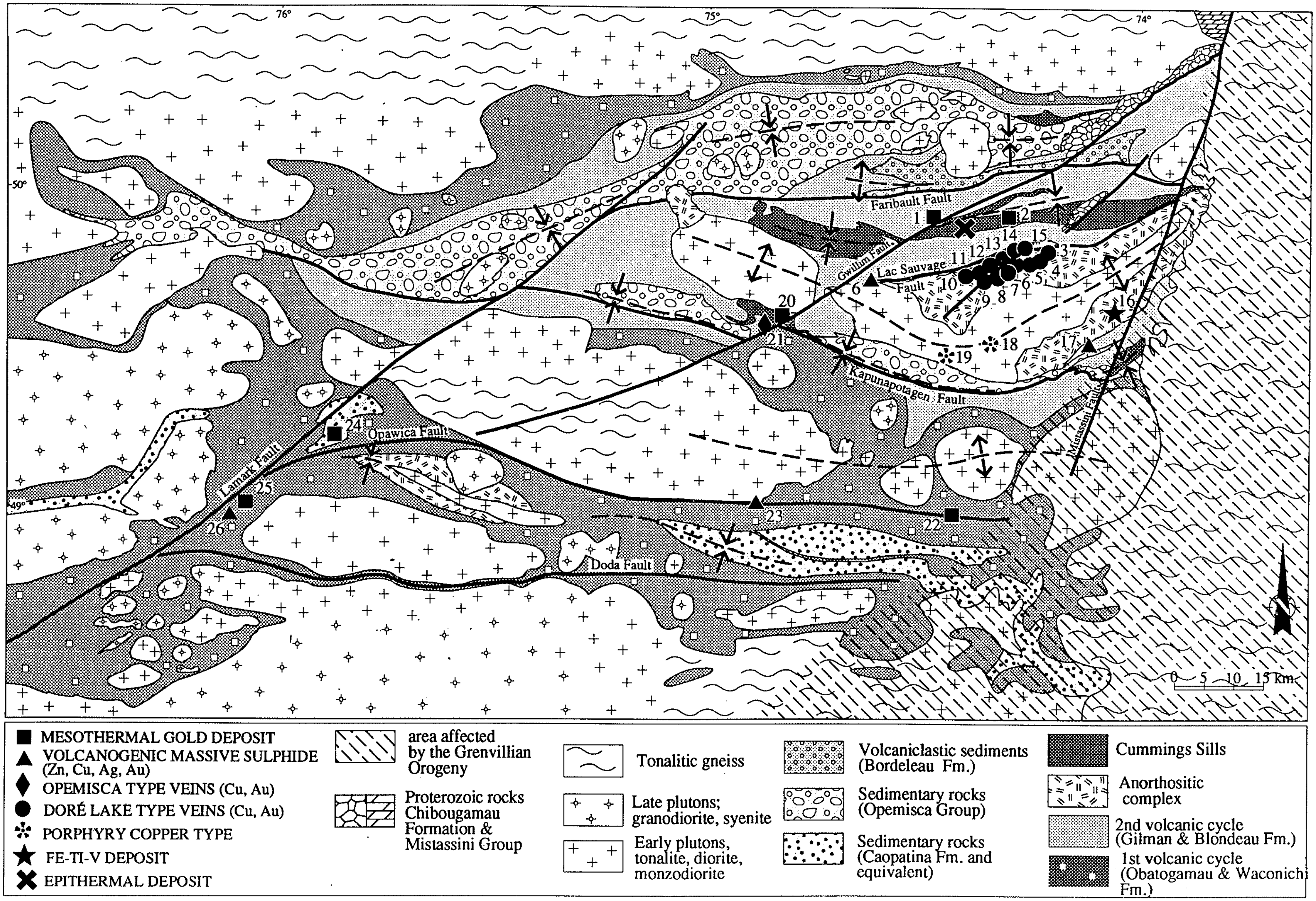


Figure 2. General Geology and Ore Deposits of the Chibougamau-Caopatina Region.

around these centres is negligible due to a high rate of erosion. Evidence of emergence of these edifices is found in the adjacent sedimentary basins (Mueller and Dimroth, 1987) and synvolcanic plutons, which represent the core complexes of the edifices (Tarney et al., 1976, Dimroth et al.; 1985; Chown and Mueller, in press).

Sedimentary Phase:

Emergence is the result of continued volcanic construction by pyroclastic deposits, effusive lava, and thick intrusive sills (Ayres, 1982, Staudigel and Schmincke, 1984), as well as by plutonic emplacement and uplift after volcanic activity ceased (Mueller et al., 1989). Erosive processes then came into effect, so that abundant volcanic detritus was shed into the adjacent elongated basins bordering these islands. Some of these volcanic centres were eroded down to their plutonic roots, as documented by the abundance of tonalitic/dioritic clasts in the conglomerates (Tarney et al., 1976; Mueller and Dimroth, 1984, 1987; Chown and Mueller, in press).

Chibougamau-Caopatina Region (CCR)

The CCR is subdivided in a southern Caopatina segment (SCS) and a northern Chibougamau segment (NCS) based on the recognition of sedimentary formations at different stratigraphic levels (Fig. 3). Due to the complex stratigraphic relationships inherent in all volcano- sedimentary terranes several columns are employed (Fig. 4). In the SCS (Fig. 4a), an essentially volcaniclastic sedimentary formation is interdigitated with the uppermost portion of cycle 1 volcanic rocks. In contrast, the NCS features an extensive sedimentary unit that formed after two volcanic cycles (Fig. 4b, c, d, e). The volcanic cycles and sedimentary units can be related to three paleogeographic evolutionary phases (Table 1) as follows: (1) formation of an early extensive subaqueous basaltic plain, (2) formation of central volcanic complexes, their gradual emergence and emplacement of synvolcanic plutons, and (3) denudation of the volcanic centres with deposition of the volcano-plutonic detritus in the adjacent basins.

Southern Caopatina Segment

The SCS is characterized by an extensive primitive basalt plain with small interspersed volcanic centres (Fig. 2, 4a) and capped by an early sedimentary sequence. The deep marine basalt plain is composed of masive, pillowed and brecciated basalt. A local increase in brecciation and vesicles is tentatively employed to indicate shoaling of shield volcanoes (Jones, 1969; Wells et al, 1979).

Small, time-equivalent felsic volcanoes appear to be omnipresent. Some are clearly intercalated in the mafic volcanic sequence, whereas others are superposed on the mafic volcanics. The volcaniclastic sediments of the first sedimentary unit (Fig. 4a, c) feature clastic material deposited by sediment gravity flows and minor intercalations of fine-to coarse- grained tuff. These tuffs document ongoing volcanic activity. Contacts between the basalt are conformable and gradational. Erosional contacts were not observed, but this may be an artifact of outcrop exposure.

Stratigraphy of the Caopatina Segment

The Obatogamau Formation, a tholeiitic basalt unit, is commonly porphyritic. This 3 to 5 km thick unit is very extensive and has been recognized laterally from the Grenville Front to the east (Daigneault, 1986; Hébert, 1980) to the Miquelon area to the west (Gauthier, 1986), a distance over 150 km; and it is common to both the Caopatina and Chibougamau segments. The basalts are tholeiitic in composition (Midra, 1989), and are characterized by abundant phenocrysts of plagioclase distributed throughout certain flows, but commonly restricted to phenocryst-rich zones generally occurring near the top of flows.

Local felsic volcanic units, such as the Lac des Vents Complex in the east, the Wachigabau Member and the Ruisseau Dalime Formation (Sharma and Gobeil, 1987) in the west, are intercalated with or superposed on the Obatogamau Formation. The Lac des Vents Complex (Fig. 3, 4a, b), a 2 to 2.5 km thick sequence, is representative of one of these edifices. The mafic-felsic centre is composed of five felsic units interstratified with basalt flows and gabbro sills (Mueller et al., 1989). The initial three units have the following components: (1) massive to brecciated felsic (dacitic) lava flows, (2) primary and reworked pyroclastic debris, (3) pelagic sediments, and (4) volcaniclastic sediments deposited by turbidity currents. The upper unit documents the destruction of the edifice and is characterized by framework-supported conglomerates followed by epiclastic volcaniclastic sediments. The former is interpreted as a slope deposit and the latter formed in a high-energy, volcanic shelf or nearshore setting (Mueller and Chown, 1989). Renewed volcanic island subsidence and cessation of volcanic activity caused submergence.

Table 1. Paleogeography and paleotectonic evolution of the Abitibi Belt (Mueller et al., 1989).

PALEOGEOGRAPHY AND PALEOTECTONIC EVOLUTION

Paleogeographic Phases	INTERNAL ZONE *N* (Chibougamau)	INTERNAL ZONE *S* (Caopatina)	EXTERNAL ZONE (Rouyn-Noranda)
Late Evolutionary Stage	3b Uplift and erosion of volcanic edifices down to plutonic roots; development of fault-bounded sedimentary basins containing plutonic detritus 2b Emergence and growth of felsic volcanic islands, substantial plutonic accretion and rise of tonalitic gneiss 1b Renewed basaltic volcanism		3 Uplift and erosion of volcanic edifices rarely down to subvolcanic plutons; development of fault-bounded sedimentary basins 2 Emergence and growth of felsic volcanic islands, plutonic accretion 1 Extensive submarine lava plain composed of little-differentiated komatiitic, low-K tholeiitic, and calc-alkaline sequences
Early Evolutionary Stage	1a Local shoaling of small felsic complexes	3a Uplift and erosion of volcanic edifices; development of fault-bounded sedimentary basins.Plutonic detritus is rare. 2a Emergence and growth of felsic volcanic islands, plutonic accretion and rise of tonalitic gneiss	
	1a Extensive (100 x 100 km), low-K tholeiitic basalt plain		

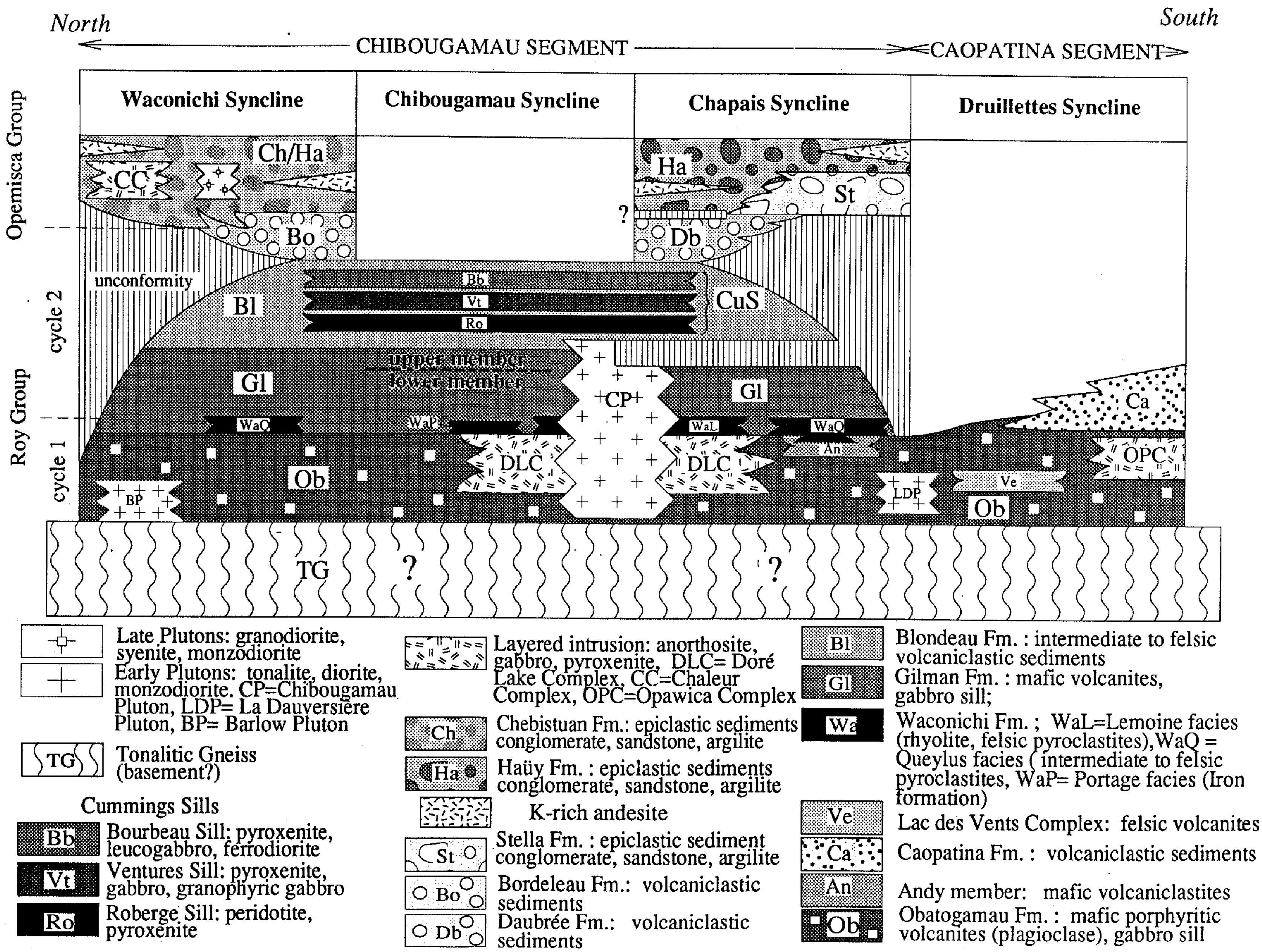


Figure 3. Generalized tectonostratigraphic relationships in the Chibougamau- Caopatina Region with respect to the position in synclines. Modified from Daigneault and Allard (1990), Dimroth et al. (1984) and Mueller et al. 1989.

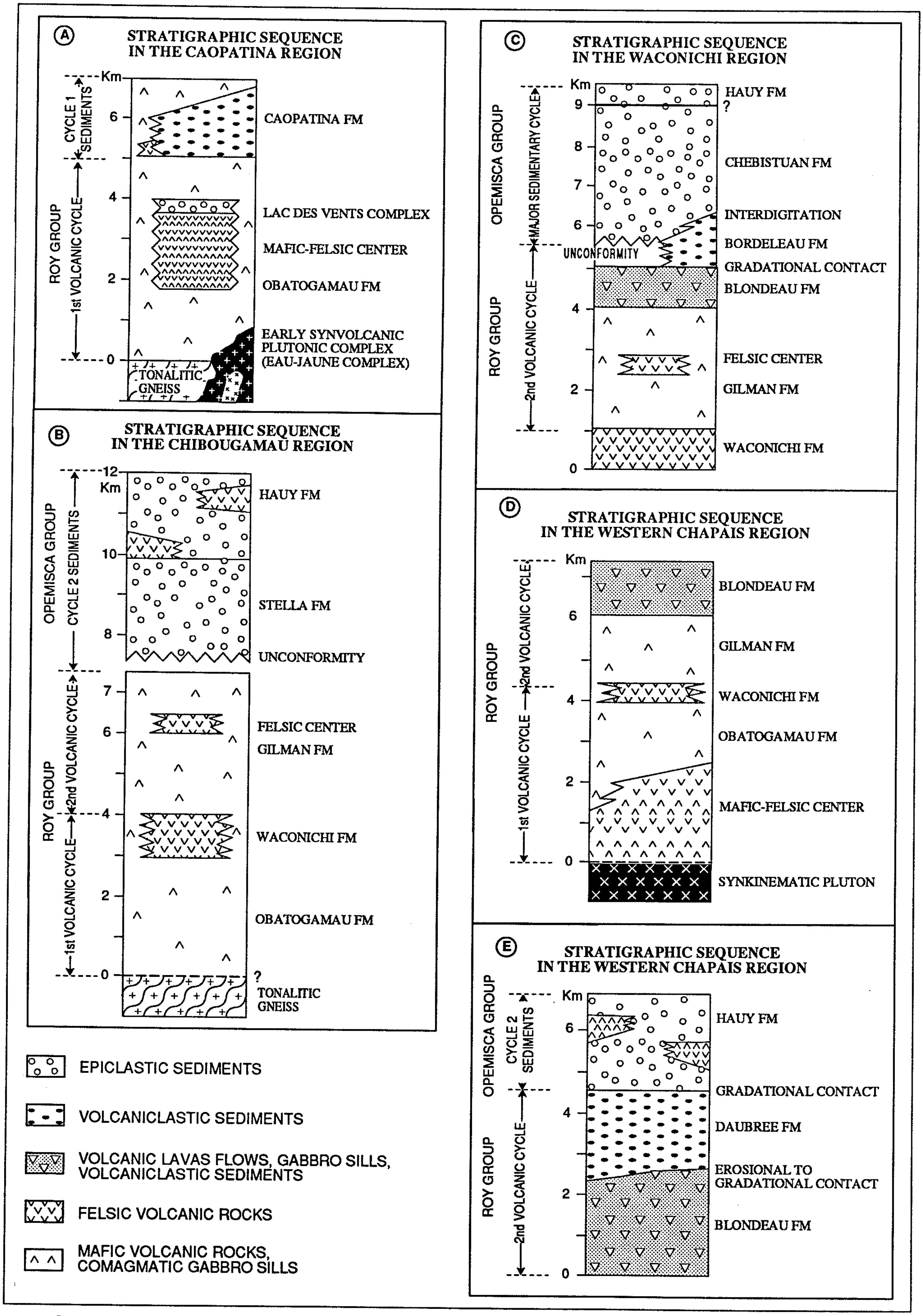


Figure 4. Stratigraphic sections Chibougamau-Caopatina Region (Mueller et al., 1989).

The Caopatina Formation represents a 1 to 2 km thick volcaniclastic sedimentary unit, locally interstratified with ash flow tuffs. Conglomerate, sandstone and argillite are the prevalent lithological components deposited in this sedimentary basin. Thick sequences of thin-to- medium-bedded volcaniclastic sandstone display the grading typical of turbidity current deposits, whereas argillite/shale indicate normal pelagic deposition typified by settling out of particles from the water column. Associated massive, stratified and graded bedded conglomerates suggest deposition from high-density turbidity currents or laminar debris flows (Lowe, 1982). Sedimentary features of these deposits suggest moderate to deep water depths, definitely below storm wave base (>200m). This unit, interstratified with pillowed basalts, indicates a marine setting. Local occurrence of tuff deposits corroborates ongoing volcanic activity. An early phase of extension must have occurred in order to accomodate a significant sedimentary sequence.

Layered Intrusions

The Opawica River Complex (Maybin, 1976) intrudes the Obatogamau Formation near its base. It consists of a Lower Anorthosite Zone (3500 m), composed of gabbroic anorthosite, gabbro and anorthositic gabbro, overlain by Gabbro-Ferropyroxenite Zone (900m). The rocks of the complex are coarse-grained and display a well-preserved cumulus texture and rhythmic layering despite the effects of regional metamorphism. Plagioclase phenocrysts are larger and more abundant in Obatogamau basalts near the complex (Midra, 1989) suggesting a close affinity between the two. This affinity is also suggested by their similar chemistry.

The Chutes de l'Esturgeon Complex (Lamothe, 1983) is 1300m- thick and is composed of peridotite at the base grading to quartz gabbro at the top. This layered intrusion also has a tholeiitic affinity (Quirion, this paper).

Northern Chibougamau Segment (NCS)

The NCS consists of two volcanic cycles (Fig. 4b) followed by a terminal sedimentary unit. Cycle 1 volcanic rocks in the NCS are similar to those described in the SCS and will not be dealt with in further detail. Recent U-Pb zircon age dating of felsic cycle 1 at 2730 $\pm$ 2 Ma constrains the age of volcanic cycle 1 (Mortensen, in prep). Cycle 2 volcanic rocks show that a primitive plain with overlapping shield volcanoes developed (Allard and Gobeil, 1984) and central volcanic edifices evolved upon them. The felsic part of cycle 2 is characterized by volcanic activity centred around the synvolcanic Chibougamau pluton (2718 $\pm$ 2 Ma, Krogh, 1982), which Dimroth et al. (1985) described as Chibougamau Island. This centre was the locus of paroxysmal volcanism and felsic to intermediate lava flows (Fig. 5a). Volcanic activity evidently occurred in the surrounding basins where marine volcaniclastic deposits prevail (Archer, 1984). Some areas to the north and west of the major centre experienced local volcanism. Mafic to ultramafic sills up to 1.2km thick, the youngest of which has been dated at 2717$\pm$1 Ma (Mortensen, in prep), are related to synvolcanic pluton emplacement. The coeval emplacement of felsic plutons and mafic sills high in the stratigraphic succession contributed considerably to the uplift of this central volcanic complex (Mueller et al., 1989).

The unconformably overlying sedimentary cycle largely derived its principally detritus from cycle 2 rocks and synvolcanic plutons (Fig. 5b). This sedimentary group can be divided into an initial epiclastic formation where the detritus originates from the cycle 2 volcanic rocks, sills, and synvolcanic plutons (Mueller and Dimroth, 1984, 1987). A second and terminal epoch is characterized by shoshonitic- dominated volcanism in a marine (Mueller, 1986) and terrestrial environment, interstratified with coarse clastic sediments (Mueller and Dimroth, 1987). An overview of the different facies and facies associations in the northern and southern basin of the NCS is given in figure 6.

Stratigraphy of the Chibougamau Segment

The stratigraphy of the Chibougamau segment (Fig. 3, 7 and 8) is divided into two groups; the Roy Group dominated by volcanic rocks, and the Opemisca Group dominated by sedimentary rocks (Allard et al. 1985). The older Roy Group is made up of five formations composed of volcanic rock of various compositions. The Opemisca Group is divided into three formations composed of conglomerates, sandstone and argillite. The contact between the Opemisca Group and underlying Roy Group is locally unconformable, but in most cases, is marked by E-W trending longitudinal faults (Daigneault and Allard, 1987).

The Obatogamau Formation (3-4 km) thick represents the base of the volcano- sedimentary succession and marks is the most important volcanic

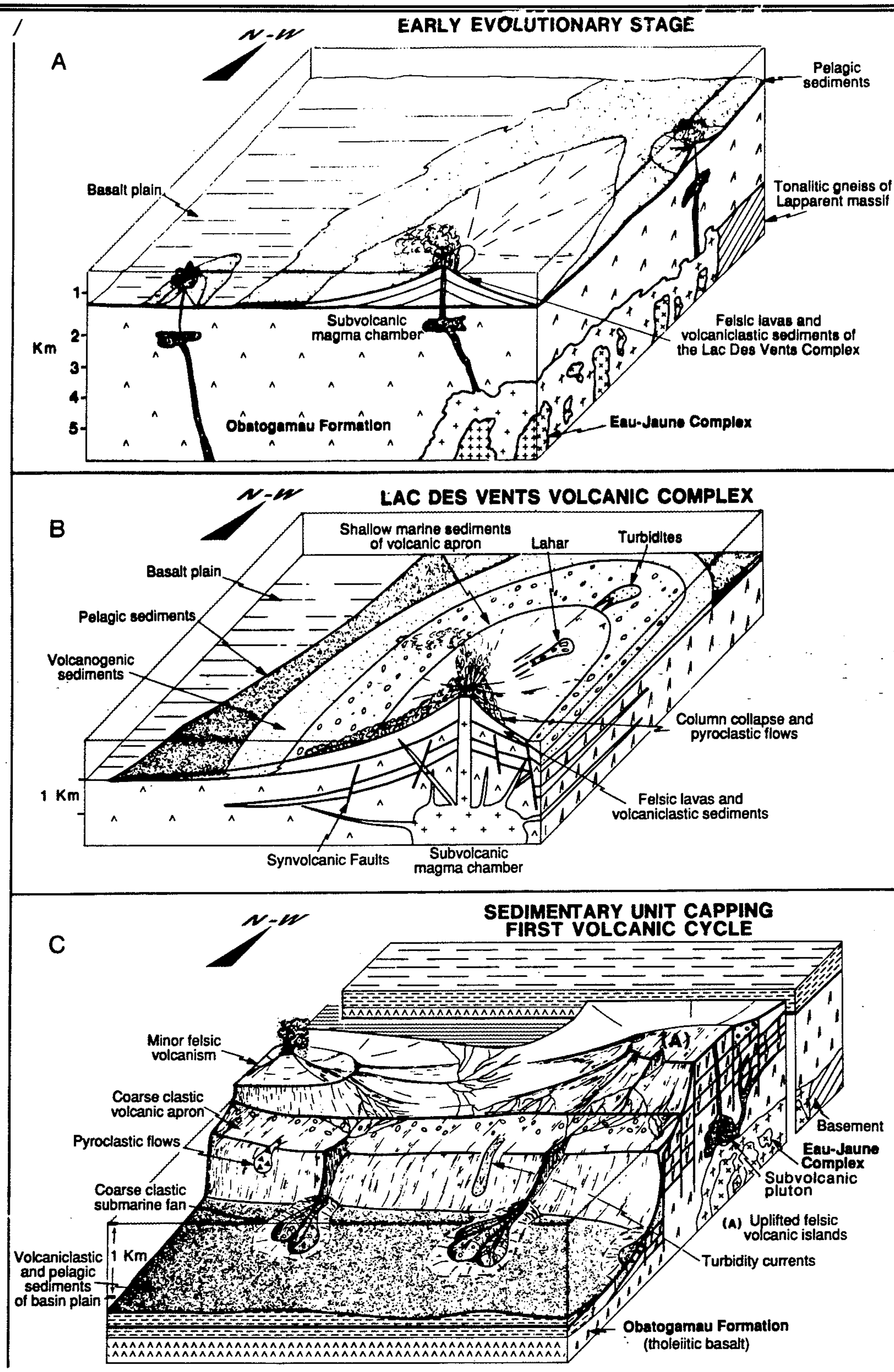
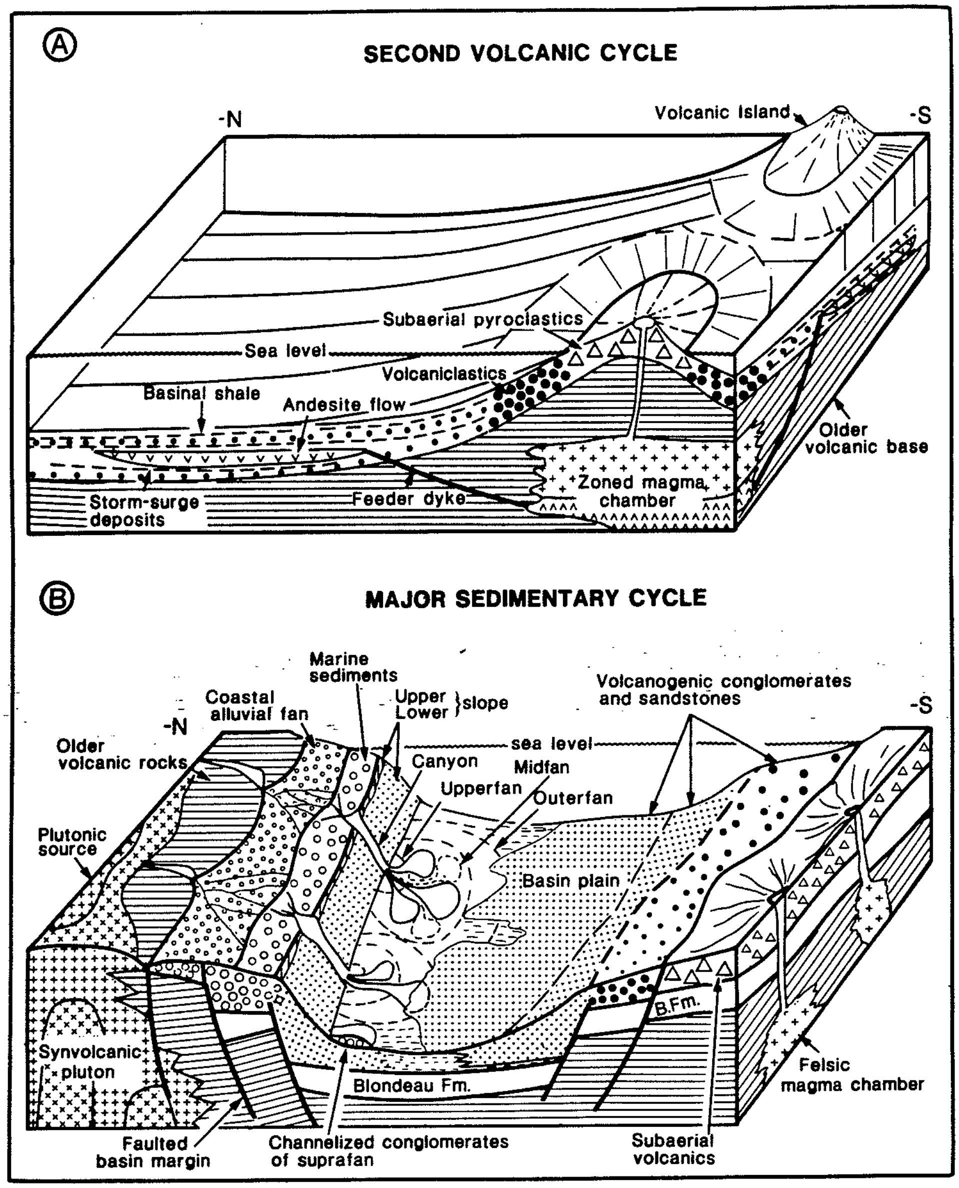


Figure 5. Evolution of the First Volcanic Cycle (Mueller et al., 1989).



Modified from Mueller and Dimroth 1984

Figure 6. Evolution of the Second Volcanic Cycle (Mueller et al., 1989).

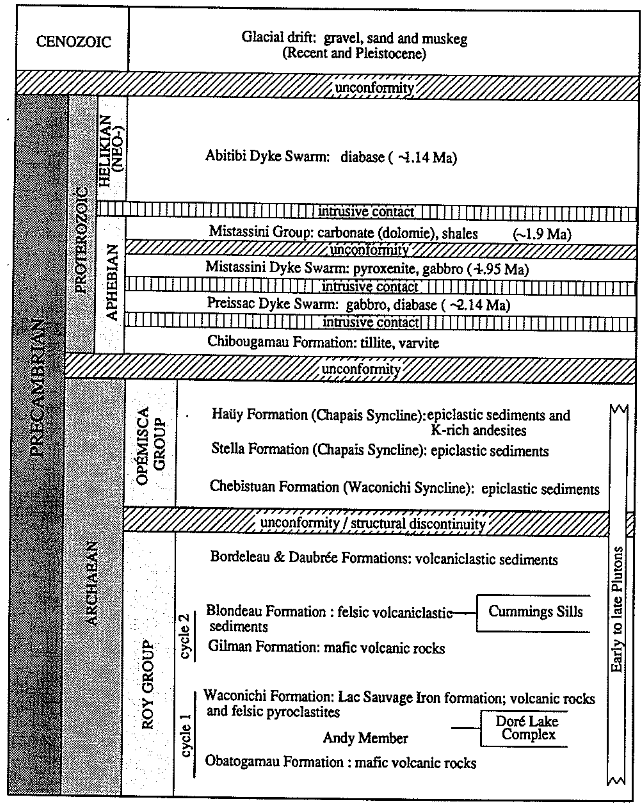


Figure 7. Stratigraphy of the Chibougamau Segment.

LEGEND FOR FIGURE 8:

1 (Ob) = Obatogamau Fm. (mafic porphyritic volcanite); 2 (Wa) = Waconichi Fm., WaL= Lemoine type (rhyolite, felsic pyroclastite), WaQ = Queylus type (intermediate to felsic pyroclastite), WaP= Portage type (Iron Formation); 3 (Gl) = Gilman Fm. (mafic volcanite); 4 (Bl) = Blondeau Fm. (intermediate to felsic volcaniclastite); 5 (Bo) = Bordeleau Fm. (felsic volcaniclastite); 6 (St) = Stella Fm. (sedimentary rocks); 7 (Ha) = Hauy Fm. (sedimentary rocks and K andesitic lava); 8 (Ch) = Chebistuan Fm. (sedimentary rocks); 9 (DLC1) = Dore Lake Complex, anorthositic Zone (anorthosite, gabbro); 10 (DLC2) = Dore Lake Complex, layered Zone (gabbro, magnetitite); 11 (DLC3) = Dore Lake Complex, granophyric Zone and Upper border Zone (gabbro, anorthosite); 12 (CuS) = Cummings Sills (gabbro, pyroxenite, peridotite); 13 (TG) = Tonalitic Gneiss; 14 = Early plutons (tonalite, diorite, monzodiorite); 15 = Late Plutons (granodiorite, syenite); 16 (Cb) = Chibougamau Fm., Proterozoic (sedimentary rocks); 17 (Al) = Albanel Fm., proterozoic (carbonates); 18 = diabase dyke, Proterozoic; 19= younging direction; 20= anticline; 21= syncline; 22 = high angle reverse fault; 23= NE trending fault; 24= NNE trending Grenvillian fault.

For the plutons:

BP= Barlow Pluton, ChP= Chevrillon Pluton, FP= France Pluton, CP= Chibougamau Pluton, EJC= Eau Jaune Complex, MP= Muscocho Pluton, VP= Verneuil Pluton, LDP= La Dauversiere Pluton, BvP= Boisvert Pluton.

For the structural elements:

BF= Barlow Fault, FF= Faribault Fault, AF= Antoinette Fault, LSF= Lac Sauvage Fault, GdF= Goudreau Fault, KF= Kapunapotagen Fault, GlF= Gwillim Fault, MF= McKenzie Fault, MF= Mistassini Fault, WF= Winchester

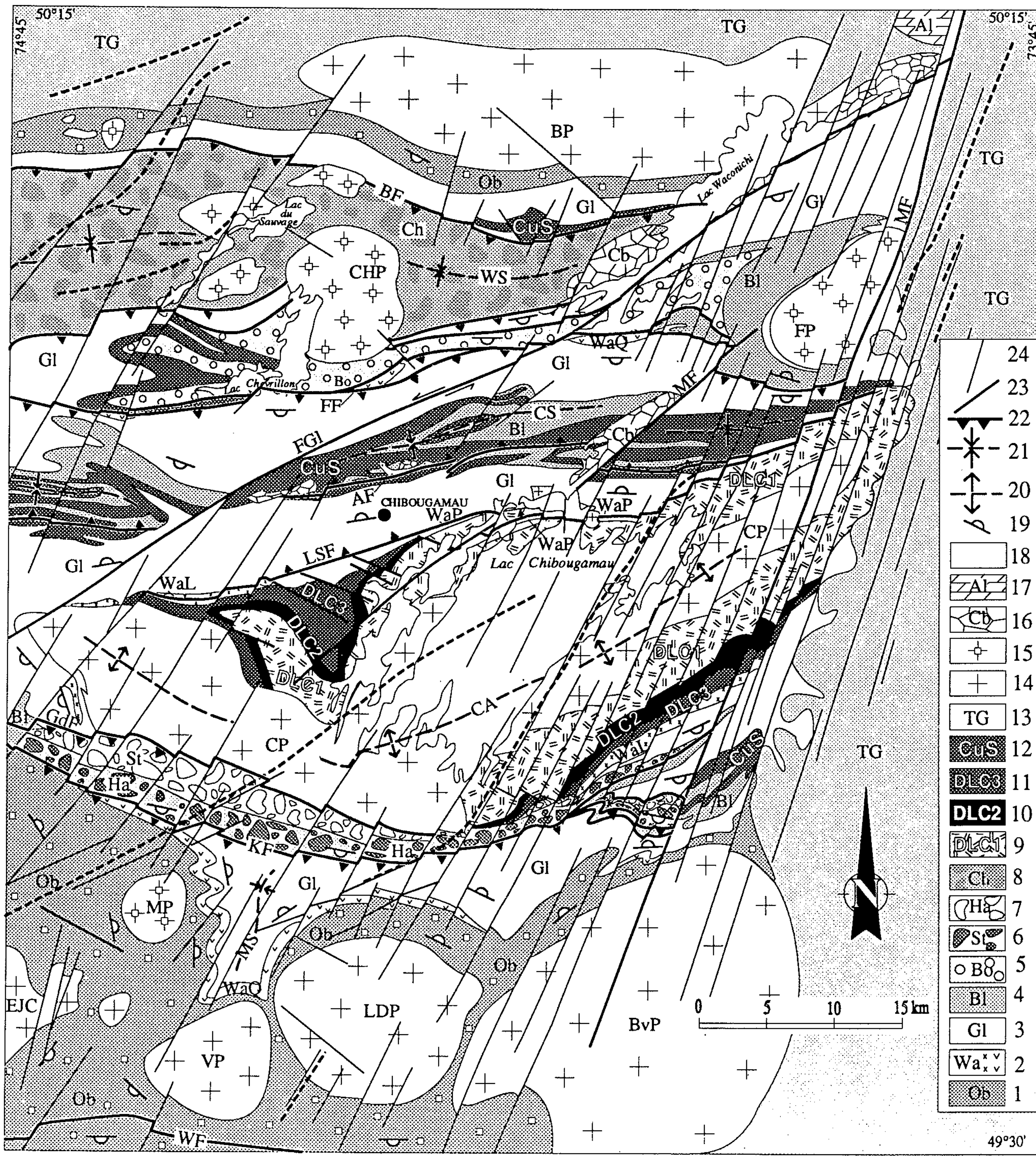


Figure 8. Detailed structural map of the Chibougamau area (adapted from Daigneault and Allard, 1990).

episode of the Roy Group. Centimetric megacrysts of plagioclase make up 3 to 20% of the pillowed to massive basalts. This stratigraphic unit systematically occurs at the base of the supracrustal assemblage. Felsic and intermediate tuffs and breccia constitute a very small portion of the unit.

The Waconichi Formation (800m-thick, volumetrically less important, is composed of dacitic and rhyodacitic rocks which mark the end of the basaltic volcanism of the first cycle. The Waconichi Formation is divided into three principal facies. The Lemoine facies is composed of rhyolitic porphyry and rhyolite, the Queylus facies consists of an intermediate to felsic pyroclastic assemblage, and the Portage facies is an exhalite typified by the lac Sauvage Iron Formation (Henry and Allard, 1979). The Lemoine facies is economically the most important since it contains the Lemoine mine on the south limb of the Chibougamau anticline and an important prospect on the north limb. The dominant rhyolitic porphyry of the Lemoine facies is interpreted to be subvolcanic intrusion. The presence of rhyolitic volcanoclastites and minor lobed rhyolite is also characteristic of this facies. Synvolcanogenic mineralization is characterized by well-laminated massive sulphide lenses rich in sphalerite and chalcopyrite.

The Gilman Formation (3-4 km thick) marks a return to mafic volcanism. The lavas, commonly very similar to those of the Obatogamau Formation (Ludden et al., 1984), are generally aphyric. Near Chibougamau, the unit can be locally divided into two members, the lower Gilman, characterized by basaltic flows, and the upper slightly more evolved Gilman characterized by andesitic to dacitic flows. Small lenses of intermediate volcaniclastites occupy the contact zone between the two members. This more evolved member representing less than 5% of the Gilman Formation has only been recognized at this one locality.

The Blondeau Formation (2-3 km thick) is composed of volcaniclastic rocks of varying composition. The principal rock types are felsic and cherty tuff, volcanogenic sandstone, graphitic shale and rhyodacitic flows. Some comformable disseminated and massive sulphide lenses are present in the pile. Mafic variolitic flows and hyaloclastite are also present in this unit. The Blondeau Formation hosts the three mafic sills of the Cummings Complex.

The Bordeleau Formation, and its equivalent the Daubrée Formation, are composed of a monotonous sequence of felsic to intermediate volcaniclastic rocks consisting of volcanogenic sandstone, argillite and minor conglomerate that were probably derived from the Blondeau Formation. The Bordeleau formation represents a transition to the sedimentary rocks of the Opemisca Group (Dimroth et al., 1985).

The Opemisca Group is composed of three formations. The time- equivalent Stella and Chebistuan formations (Fig. 3 and 8) are epiclastic sedimentary units derived from volcanic cycles one and two. Conglomerate, sandstone, argillite and shale are the major constituents of these formations. Abundant granitoid fragments in the conglomerates are principally derived from the Chibougamau Pluton, an unroofed synvolcanic intrusion. The different sedimentary facies of these formations are summarized in Table 2 and indicate that depositional environments ranges from deep marine, to slope, to shallow marine, to fluvial.

The younger Haüy Formation is characterized by a late shoshonitic- type volcanism (Picard and Piboule, 1986) forming massive and brecciated andesite/basalt flows. The depositional environment, as documented by the intercalated sedimentary facies (Tables 3, 4), varies from fluvial to marine (Piché, 1984, Mueller and Dimroth, 1987).

Doré Lake Complex

The Doré Lake Complex is an anorthositic layered intrusion that originated from the differentiation of a tholeiitic magma similar to the one that produced the basalts of the Obatogamau Formation. It was emplaced at the base of the Waconichi Formation. A schematic map summarizing the principal subdivision of the Doré Lake Complex is shown in figure 9 and the detailed division of the complex is presented in Table 5. Of the three major subdivisions of the complex, the Lower Series is the thickest. The most important part of the Lower Series is the Anorthositic Zone, although a minor magnetitite zone occurs locally at the base of the intrusion and a peridotite zone occurs in the northwestern part of the intrusion (Fig. 9).

The Anorthositic Zone is essentially composed up of 2500 to 3600 m of gabbro and anorthosite and contains most of copper-gold deposits. It is characterized by very coarse (up to 30 cms in places cumulus bytownite crystals, replaced by albite speckled with zoisite. The intercumulus magnesium-rich pyroxene is replaced by chlorite and locally by actinolite. An

Table 2. Overview of Sedimentary facies associations, Waconichi Syncline (Mueller et al., 1989).

AN OVERVIEW OF THE SEDIMENTARY FACIES ASSOCIATION IN THE NORTHEASTERN WACONICHI SYNCLINORIUM (NCS)

FACIES	LITHOLOGY	MODE OF DEPOSITION	LOCUS
Black Shale Facies Association	Black shale, argillite, volcaniclastic sandstone, intraclast beds	Suspension sediments Storm surge deposits	Basin plain, 200-300 m water depths
Submarine Fan Facies Association	Channeled conglomerates, graded bedded sandstone, argillite	Sediment gravity flows (high and low density turbidity currents)	Upper fan, midfan, outer fan
Transitional or Slope Facies Association	Graded bedded sandstone, rare conglomerate beds, argillite	High and low density turbidity currents	Upper and lower slope
Marine-Terrestrial Conglomerate-Sandstone Facies Association	Prevalence of matrix-supported conglomerates hummocky cross-stratified sandstone, graded bedded sandstone	Sediment gravity flows Storm surge	Shallow marine shelf
	Framework-supported conglomerates, massive to laminated sandstone	Traction currents	Terrestrial ? alluvial fan?

Table 3. Non-marine sediments of the Chapais Syncline (Mueller et al., 1989).

NON-MARINE SEDIMENTS IN CHAPAIS SYNCLINE (NCS)

	FACIES	CHARACTERISTICS	LOCUS
Conglomerate-Sandstone Facies Sequence, 1a (> 50% conglomerate)	Gms, Gp, Gt, Sm, Sh, St, Sp	Amalgamation of conglomerate beds, poorly developed fining-upward cycles	Coastal alluvial fan (fan-delta)
Conglomerate-Sandstone Facies Sequence, 1b (< 50% conglomerate)	Gm, Gt, Sh, St, Sp, Sr, Fl	Excellent fining-upward cycles, dominance of sandstone beds	Alluvial fan (distal portion) grading into braid delta

Table 4. Marine sediments of the Chapais Syncline (Mueller et al., 1989).

SHALLOW MARINE SEDIMENTS IN CHAPAIS SYNCLINE (NCS)			
	LITHOLOGY	PROCESS	LOCUS
Heterolithic Facies (2.1)	Thin beds of laminated argillite, ripple cross laminae, graded beds and laminated sandstones	Fair weather suspension sediment, weak current motion, storm surge deposits	Proximal offshore
Sandstone-Argillite Facies (2.2)	Wavy, lenticular sand-stone with intercalated argillite beds	Fair weather, high wave surge and suspension deposits during tranquil periods	Lower shoreface
	Plane-bedded sandstone graded bedded sandstone capped by argillite	Storm surge and storm abatement deposition	Lower shoreface
	Planar crossbeds with clay drapes on fore-sets, cosets of trough crossbeds with clay drapes on foresets and between sets	Fair weather deposition wave-induced structures	Lower shoreface
Fine-to Very Coarse-Grained Sandstone Facies (2.3)	Parallel laminated and graded bedded sandstones rich in rip-up clasts from facies 2.1	Storm surge	Lower shoreface
	Clay draped trough crossbeds	Fair weather wave-induced structures	Lower shoreface
	Plane-bedded to low-angle crossbedded sandstone, trough crossbeds with super-posed ripples, absence of argillite laminae	Fair weather high wave surge	Middle shoreface

Table 5. Stratigraphy of the Doré Lake Complex (Daigneault and Allard, 1990)

	South flank			North flank				
				western segment			eastern segment	
	ZONES	SUB-ZONES	LITHOLOGY	ZONES	SUB-ZONES	LITHOLOGY	ZONES	LITHOLOGY
UPPER SERIES	BORDER GRANOPHYRE		Gabbro Gabbroic Anorthosite	BORDER GRANOPHYRE		Gabbro Diabase Gabbroic anorthosite	BORDER	Ferropyroxenite Anorthositic gabbro
LAYERED SERIES				FERRODIORITE S.1.	F_3 F_2 F_1	Ferrodiorite (s.s.) Ferrogabbro Ferropyroxenite	FERRODUNITE* S. 1.	Ferrodunite (s.s.) Magnetitic ferrodunite Ferroperidotite Ferropyroxenite
	LAYERED ZONES P_3	*quartz* *apatite and ilmenite* *ilmenite and ferrosilicates*	Ferrropyroxenite Ferrogabbro	LAYERED ZONES NP_3		Quartz-ferropyroxenite		
	A_2		Gabbroic anorthosite					
	P_2		Magnetitite Ferropyroxenite Ferrogabbro	NP_2		Apatite-ilménite-ferropyroxenite		
	A_1		Gabbroic Anorthosite					
	P_1		Magnetitite Ferropyroxenite Ferrogabbro	NP_1		Magnetitic ferrogabbro		
LOWER SERIES	ANORTHOSITE s.l.	*Magnetite*	Gabbro Anorthosic gabbro Noritic anorthosite Anorthosite (s.s.) Ferrogabbro Magnetitic ferropyroxenite	ANORTHOSITE S.l.	*Peridotite*	Gabbro Anorthositic gabbro Noritic anorthosite Anorthosite (s.s.)	ANORTHOSITE S.l.	Gabbro Anorthositic gabbro Noritic anorthosite Anorthosite (s.s.)

* unlayered rocks outcropping on Mont du Sorcier and lac Robert area

intrusive breccia related to the mafic phases of the Chibougamau commonly occupies the lower contact of the Anorthositic Zone. At the top of the sequence, the gabbro and anorthosite change abruptly to magnetite-rich rocks of the Layered Series.

The Layered Series is well developed on the south flank and in the north-west part of the complex. On the south flank, five zones are distinguished, three rich in oxide phases (P1, P2 and P3) and two others similar to the Anorthositic Zone (A1 and A2; see Table 5). At the base, the 30-to-90 m thick P1 Zone contains economic vanadium values in the magnetitite horizon. The P2 Zone (9-60 m thick) resembles the P1 Zone except for its lower vanadium values. A ferrogabbro and a magnetitite characterize its upper part. The P3 Zone is composed of a dark green ferropyroxenite interlayered with thin ferrogabbro horizons.

The Upper Series includes the Granophyre Zone and the Upper Border Zone. The Granophyre Zone is found in the south flank and on the northwest part of the complex in juxtaposition with rhyolitic rocks of the Waconichi Formation. The rocks of the Granophyre Zone ressemble leucocratic tonalites. On the south flank the thickness of the Granophyre Zone varies from 150 to 900 m. The Upper Border Zone forms thin discontinuous lenses in contact with the volcanic rocks of the Roy Group overlying the complex. Where the Upper Zone is absent, the granophyric zone is in direct contact with the volcanic rocks. The Upper Zone consists of fine-grained gabbro (diabase), ferropyroxenite rich in magnetite, anorthosite and gabbro. Enclaves of Upper Border gabbro are found in the Granophyre Zone and are considered to be evidence of a chilled zone. In several localities, there are blocks of rhyolitic porphyry which represent remnants of rocks overlying the intrusion that have been incompletely melted.

The Granophyre Zone and the Layered Zone are present only where the complex is in contact with the rhyolitic rocks of the Waconichi Formation. Where the complex is in contact with mafic rocks on the north east flank, the Granophyre and Layered Zones are absent and the upper contact of the complex is characterized by the presence of massive magnetite-rich ferropyroxenite and ferrodunite typified by the presence of olivine.

The parent magma of the Doré Lake Complex is a tholeiitic basalt of Obatogamau type containing over 10% plagioclase phenocrysts. The main pulse, corresponding to a huge magma surge, produced the main body of the complex (Fig. 9). A chilled zone (Upper Border Zone) was formed at the border of the magma chamber in contact with the host rock. Coarse plagioclase megacrysts suggest very slow crystallization. The accumulation of plagioclase produced the Anorthositic Zone. The accumulation process may have been accelerated by expulsion of liquid during compaction. Heat loss at the top of the complex was responsible for melting the rhyolitic rocks of the Waconichi Formation. The interaction between two magmas, one rich in iron, the other rich in silica, radically transformed conditions in the magma chamber. The input of silica contributed to the formation of plagioclase and iron-rich pyroxene (two saturated minerals). The formation of abundant primary magnetite is favoured by a change in oxygen fugacity in zones P1 and P2. The conditions changed in zone A2 and became radically different in zone P3 where the magnetite was replaced by ferroaugite and ferrohypersthene. After deposition of the P3, the input of more siliceous magma gave rise to the formation of the granophyre. Fragments of Upper Border Zone gabbro were broken off and fell into the viscous magma.

Cummings Complex

The Blondeau Formation is intruded by a trio of differentiated sills called the Cummings Complex (Duquette 1976). These sills represent an excellent marker horizon because the petrography, texture, and chemistry of each sill is characteristic and their order of superposition is identical throughout the district: the Bourbeau sill at the top, the Ventures sill in the centre and the Roberge sill at the base. They have been mapped from the Grenville Front westward for a distance of 160 km. The total thickness for the three sills is approximately 1200 metres, each sill varying in thickness between 200 and 1000 metres. These sills have been recognized in the core of the three major synclines of the Chibougamau Segment (Fig. 8). They have been described on the north limb of the Waconichi Syncline (Caty, 1978; Boudreault, 1977), from both limbs of the Chibougamau Syncline (Duquette, 1976) from the Chapais Syncline in two places, around the Chapais area copper mines (Duquette, in Allard et al., 1972) and near the Grenville Front area (Allard 1981).

The Roberge sill is composed of ultramafic rocks varying in composition from dunite to pyroxenite and wehrlite. The olivine is transformed to serpentine and magnetite and the pyroxene is uralitised. The Ventures sill includes a black pyroxenite at the base, a foliated gabbro, a coarse poikilitic gabbro named "Ventures Gabbro" and locally, small lenses of granophyric gabbro. The more evolved Bourbeau, is composed of a thin pyroxenite at the base overlain by a leucogabbro and a large ferrodiorite and quartz ferrogabbro unit.

Chaleur Lake Complex

The Chaleur Lake Complex (Durocher, 1985) intrudes rocks of the Opemisca Group (Gobeil, pers. comm), and is intruded by the Opemisca pluton. The complex, although strongly deformed, appears to be composed of leucogabbro overlain by a layered zone (gabbro, pyroxenite and anorthosite) overlain in turn by a gabbroic anorthosite containing lenses of peridotite and dunite, particularly near the contact with the Layered Zone.

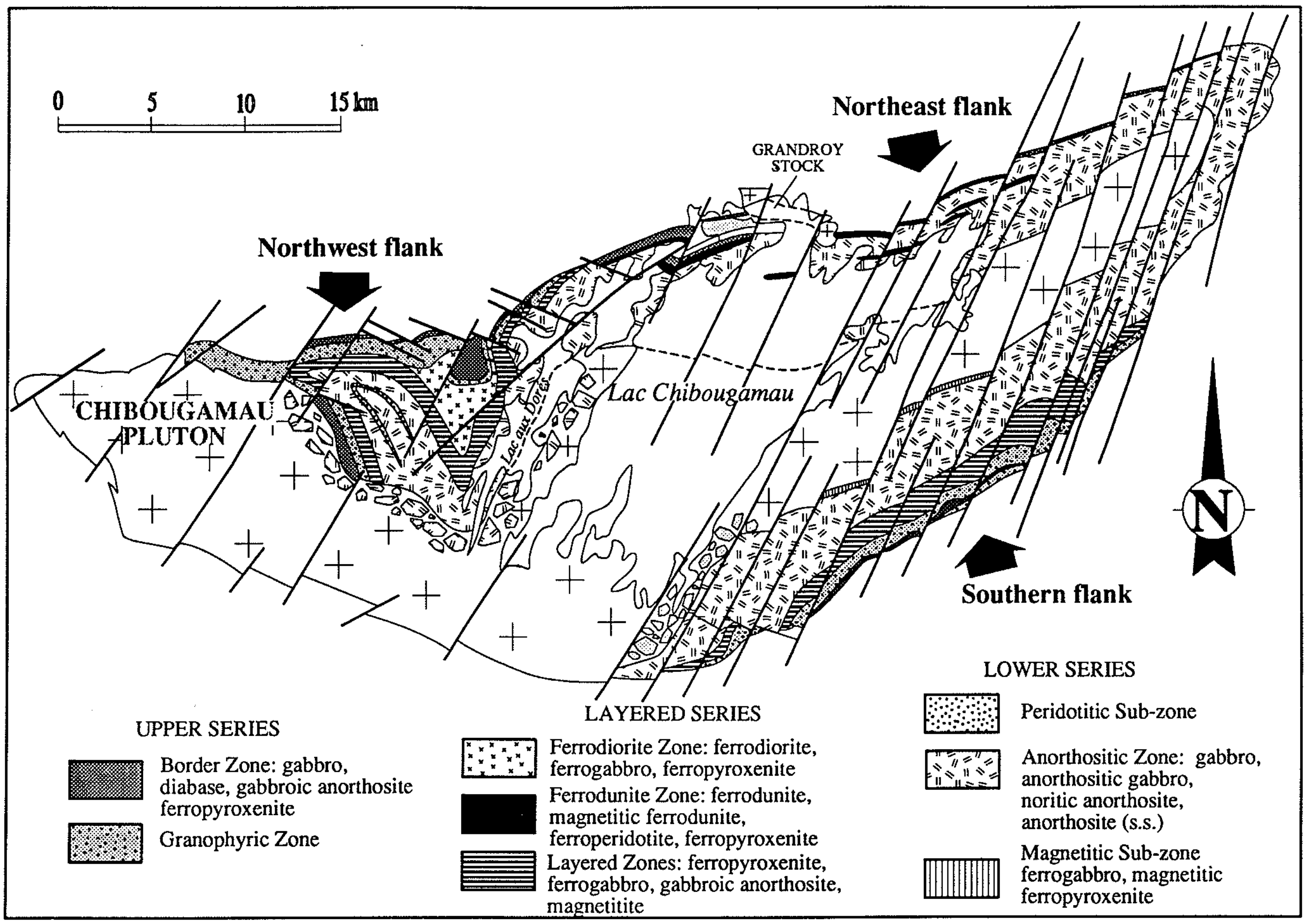


Figure 9. Schematic map of the Doré Lake Complex (adapted from Daigneault & Allard, 1990).

STRUCTURAL GEOLOGY

The Chibougamau area is part of an Archean greenstone belt representing a large synclinorium of volcanic and sedimentary rocks enclosed within tonalitic gneisses. Within this basin, several E-W trending regional folds are responsible for the vertical disposition of the strata. Synclinal structures, with younger sediments enclosed within the core, are well outlined and are clearly associated with a regional axial plane schistosity. Anticlines, on the other hand, either form domes with a core occupied by early tonalitic to dioritic plutons or are destroyed by E-W trending longitudinal faults. In the latter case, the presence of an anticline is deduced by reconstitution of stratigraphic sequences dismembered by reverse movements along E-W trending tectonic zones.

The Kenoran Orogeny, the cause of the major deformation in the area, accounts for the large folds and the regional schistosity. The age of deformation in the Abitibi belt is considered to be 2695-2700 Ma (Corfu et al., 1989) corresponding to the Shebandowan event (Stott et al.,1989) of the Kenoran Orogeny. The actual age of deformation in the Northern Abitibi Belt is poorly constrained, although ages of syntectonic plutons (Table 6) appear to agree with those from the Southern Zone. Model Pb ages for a number of E-W-trending-shear-hosted mineral deposits in the Chibougamau district (Thorpe et al., 1984) also fall in this range.

Folding

Four important folding events are distinguished (D_1, D_2, D_3 and D_4). Three of these events are Archean, the fourth one is Grenvillian and limited to the eastern border of the Belt where it is in contact with the Grenville Orogen. The three Archean deformation phases can be considered as part of a deformation continuum and not as chronologically separate and distinct phases. Figure 8 illustrates the main structures of the area.

An early phase of deformation (D1) produced broad N-S folds lacking schistosity. In the southern portion of the area, an important regional N-S striking fold known as the Muscocho Syncline has been identified (Daigneault and Allard, 1983). This structure developed early in the tectonic history as suggested by the inversion of structural facing in axial zones of the folds associated with the regional schistosity, and by the fact that the regional foliation clearly cuts across both limbs of this early flexure. The early deformation may be the precursor of the regional deformation; in this phase flexures are caused by the simple sinking of the crust under its own weight. The location and attitude of the folds may be controlled by paleogeographic factors responsible for the thickness of the volcanic prism (i.e. volcanic centres).

The regional deformation (D2) is responsible for the large E-W folds). Six major structures of regional extent control the attitude of the rocks of the belt (Fig. 2, 10) as follows from north to south; 1) the Waconichi Syncline, 2) the Chibougamau Syncline, 3) the Chibougamau Anticline, 4) the Chapais Syncline, 5) the La Dauversière Anticline, and 6) the Druillettes Syncline.

The superposition of both north- south and east-west systems produces the framework of the regional dome and basin interference pattern (Fig. 8). The climax of regional deformation generated a regional schistosity, vertical stretching lineations and tightening of folds with the rotation of their axes toward the vertical.

The regional schistosity is a flow schistosity generally well developed throughout the whole area. It is an axial-plane structure and it contains a subvertical stretching lineation. The principal tectonic fabric is locally disturbed in the vicinity of early plutons showing an interaction between a north-south horizontal shortening and a local stress field generated or deviated by granitoid masses. A greater level of deformation (L-S and L type tectonites) is observed in areas adjacent to the early plutons when compared to the rest of the region. A concentric schistosity trajectory around the plutons act as an obstacle to the regional E-W trajectories, thus producing small interaction zones characterized by strong vertical extension. These relations suggest an interference between a regional stress field producing a north-south horizontal shortening, and local stress fields controlled or deviated by granitoid plutons acting as resistant nuclei.

The third event (D3) includes several phenomena superposed on elements generated during events 2 and 1. It is not necessarily a distinct phase since it could be the continuation of the regional deformation (D2). These manifestations tend to be more developed in anisotropic zones. Their main characteristic is the presence of a northeast-trending crenulation cleavage with a subvertical dip. Where this cleavage becomes more developed, asymmetric Z- folds are locally observed. These folds are generally open and their plunge is moderate to abrupt with a tendency to be coaxial with the stretching lineation.

Faults and shear zones

Four fault systems can be distinguished on the basis of direction and structural signature. These include an East-West, Southeast, Northeast and North-Northeast system. The last fault system like the D4 deformation, is related to the Grenvillian Orogeny.

The East-West trending faults belong to a system of large regional breaks found throughout the Abitibi Belt. Consisting of ductile shear zones varying in thickness from 100 to 1000 metres, those faults represent the oldest in the area and the most important in terms of stratigraphy and metallogeny. These faults, showing evidence of reverse movement, represent the final stage of the regional deformation. Notably in some cases, they are responsible for repetition of portions of the stratigraphic sequence and, in other cases, truncate regional folds, juxtaposing different structural domains (Daigneault and Allard, 1987). Locally, reverse

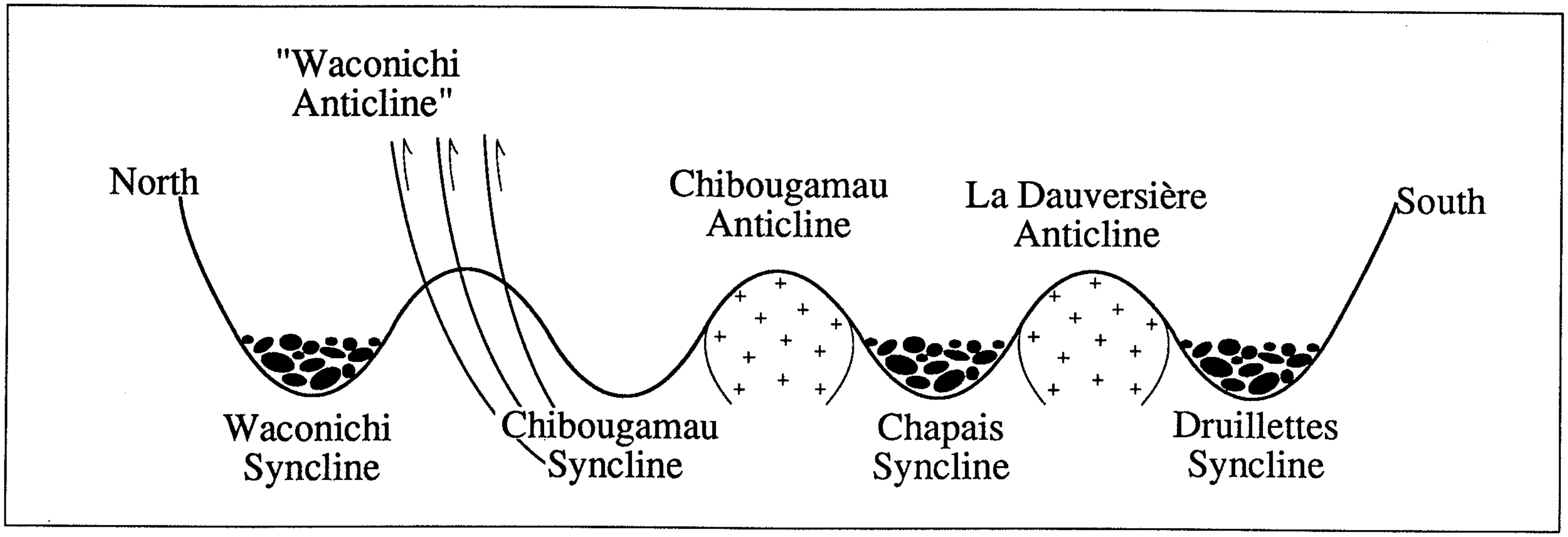


Figure 10. Schematic cross section of the Chibougamau-Caopatina Region (adapted from Daigneault & Allard, 1990). The Chibougamau area consist of one large synclinorium constituted by 4 major synclines. Three of them contain younger sediment in their core. Two anticlines are occupied by early plutons while the other one is destroyed by a series of E-W trending longitudinal faults

movement along the faults places unusual stratigraphic units in contact.

The Kapunapotagen Fault (Fig. 11) is well outlined by the contact between the sedimentary Opemisca Group and the volcanic Roy Group. This fault is recognized for a length of 90 km, and it occupies the axial plane of the Chapais syncline The structural cross-section in Figure 11 shows the position of the fault in the hinge of the syncline. The presence of strong vertical stretching lineations in the fault zone is suggestive of a vertical movement. A south over north sense of movement is indicated.

The Faribault Fault (Fig. 12) separates a sedimentary basin to the north from volcanic asssemblage to the south. It is characterized by important shear zones concentrated in the volcanic domain. The sedimentary domain is more or less unaffected by the deformation. Strong subvertical stretching lineations in the volcanic domain, and their relative absence in the sedimentary domain, is suggestive of a vertical movement of south over north.

The Doda Fault (Fig. 2) is located in the Druillettes Syncline (Lauzière et al., 1989). It consist of a large E-W trending high strain zone slightly different from the E-W discontinuities belonging to the northern Chibougamau segment due to its prominent subhorizontal stretching lineation. This deformation zone seems to have had a complex strain history as indicated by the variability in attitude of the stretching lineation and contradictory kinematic indicators. The influence of the Grenvillian Orogeny on this fault remains to be evaluated.

The Lac Sauvage Fault (Fig. 13) includes several E-W trending shear zones, part of a large anastomosing deformation corridor. The high strain zone locally reaches a width of 400 m. The presence of a down-dip stretching lineation favours a vertical component of movement. Kinematic indicators are suggestive of a north over south movement which explains the repetition of slices of the Doré Lake Complex and other stratigraphic units.

The southeast-trending shear zone system represents in some cases large corridors reaching 300 m in width. However, their extent is limited, varying from 2 to 5 km. Most occur in the Doré Lake Complex between the Chibougamau Pluton and the lac Sauvage Fault. Generally, these structures have a steep dip (60-80°) to the southwest. They host most of the copper-gold deposits of the Chibougamau mining camp, and they are interpreted to be subsidiary faults associated with the Lac Sauvage fault.

The northeast-trending faults are younger than the E-W trending system. Their principal representative is the Gwillim Fault which may be traced for 150 km. The latest movements on this fault post- date the deposition of the Lower Proterozoic Chibougamau Formation producing mylonite in these rocks. The horizontal offset deduced from a series of marker horizons indicates a sinistral component of a few kilometres. The McKenzie and Lamarck Faults are also members of this system. The Lamarck fault, which is 200 km long, shows a similar sinistral offset, whereas the McKenzie Fault has a dextral movement.

PLUTONIC ROCKS

The compilation map of the plutons of the region (Fig. 14) gives an appreciation of the spatial distribution of the intrusive bodies, which, combined with an estimation of their probable time of emplacement, allows an evaluation of their influence in the structural evolution of the belt. A recent preliminary classification of the granitic rocks of the region (Racicot et al., 1984) has been modified and extrapolated for the entire northern Abitibi (Chown and Mueller, in press). The model is based on detailed mapping and petrologic studies in the northeast coupled with reconnaissance mapping and petrographic examination of plutons farther west. Further information has been gleaned from the relatively sparse literature available. All the plutons in the Northern Zone of the Abitibi Subprovince are I-type (Chappel and White, 1974). Further classification is difficult without extensive geochemical studies, but at present relatively few intrusions have been the subject of recent geochemical studies including rare earths, although some studies are in progress.

The available age dates (Table 6) show a clear division into two distinct periods ca. 2718 and 2700-2695 Ma corresponding to a period of volcanic activity and to the main Abitibi deformation respectively. The plutonic rocks may be divided into three categories in terms of tectonic age, synvolcanic (or pretectonic) plutons, syntectonic plutons and post-tectonic plutons. The plutons of various ages fall into different petrographic families, although there is considerable overlap in the suites. Tonalitic gneiss, which may be remnants of an older "basement", or may be part of the

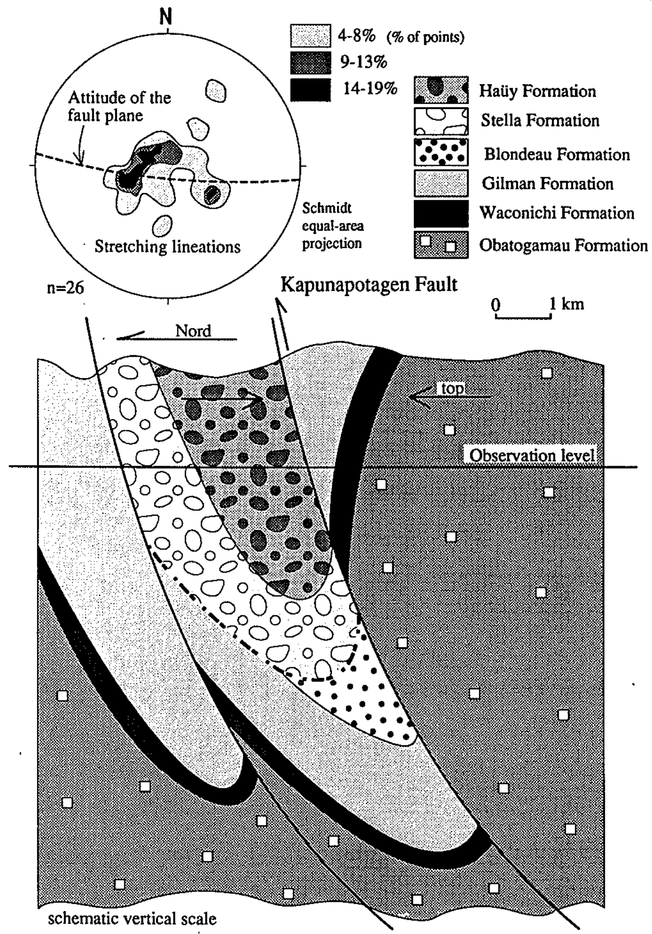


Figure 11. Schematic cross-section across the Kapunapotagen Fault in the northern Chibougamau segment. The fault is responsable for the juxtaposition of the sedimentary domain of the Opemisca Group and the volcanic domain of the Roy Group. The subvertical stretching lineation is indicative of the vertical component of movement.

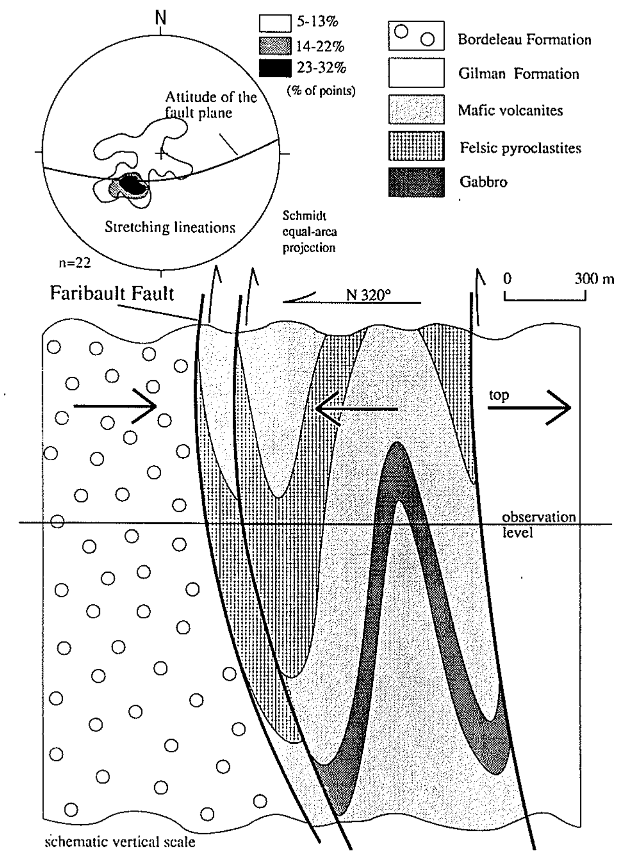


Figure 12. Schematic cross-section across the Faribault Fault in the northern Chibougamau segment. The fault creates an anormal contact between the volcaniclastic sediment of the Bordeleau Formation and the volcanic rocks of the Roy Group.

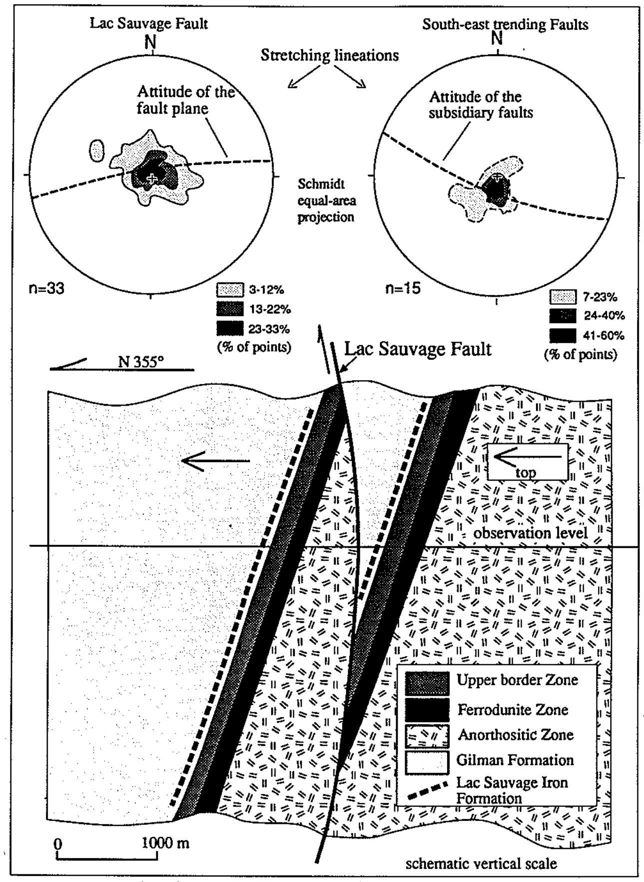


Figure 13. Schematic cross-section across the Lac Sauvage Fault in the northern Chibougamau segment. The fault is responsable for the repetition of the sequence "anorthosite zone - ferrodunite zone - Border zone-Lac Sauvage Iron Formation - Gilman".

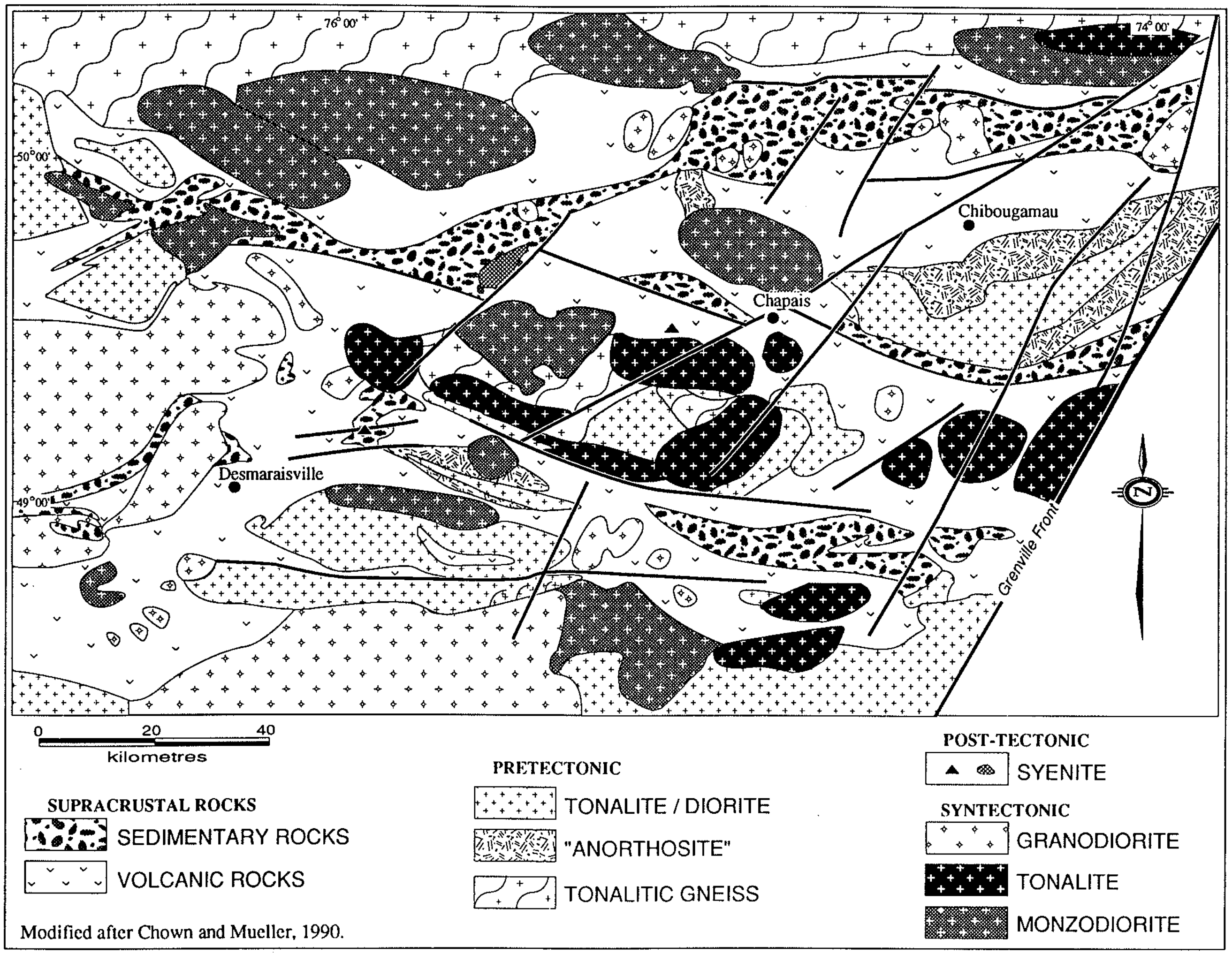


Figure 14. Distribution of felsic plutons; Chibougamau-Caopatina Region.

Table 6. Age dates of plutons of the Northern Abitibi Belt.

PLUTON	AGE (Ma)	METHOD	SOURCE	PETROGRAPHIC SUITE
WASWANIPI	2616 ± 19	Pb-Pb	1	GRANODIORITE
MUSCOCHO	>2698	U-Pb	2	
FRANQUET	2692 ± 4	U-Pb	3	
OLGA	2693 ± 3	U-Pb	2	TONALITE/GRANODIORITE
RENAULT	2718 ± 12	Pb-Pb	1	
RENAUD	2700 ± 2	U-Pb	2	
ABITIBI	2690 ± 4	U-Pb	2	
PALMAROLLE	2696 ± 1	U-Pb	2	
DAUVERSIERE	2720 ± 2	U-Pb	2	
BARLOW	2695 ± 3	Pb-Pb	1	MONZODIORITE
OPEMISCA	2695 ± 8	Pb-Pb	1	
OPEMISCA	2697 ± 2	U-Pb	3	
RADIORE	2715-2721	U-Pb	2	TONALITE/DIORITE
CHIBOUGAMAU	2718 ± 2	U-Pb	5	
TASCHERAU	2718 ± 2	U-Pb	3	
DETOUR	2722 ± 3	U-Pb	4	
QFP	2718 ± 2	U-Pb	2	
LAPPARENT	2708 ± 12	U-Pb	2	TONALITIC GNEISS

1) Gariepy and Allègre 1985; 2) Mortensen in prep; 3) Frarey and Krogh, 1986; 4) Marmont and Corfu, 1988; 5) Krogh, 1982.

plutonic suite, has been identified in a number of places within the belt and along the northern border.

The synvolcanic plutonic suite includes tonalite/diorite related in part to the genesis of the volcanic sequence, but intrusive into it. Syntectonic plutons were intruded during the late stages of regional deformation, and their emplacement was controlled by it. Three overlapping suites of syntectonic intrusions are recognized; a monzodiorite suite, a tonalite/granodiorite suite, and a granodiorite suite. The genesis of these suites of plutons must ultimately be related to the deformational history. Post-tectonic plutons are generally controlled by late tectonic structures cutting across the regional deformation and are, in part, deformed by these structures. The post-tectonic suite includes both a granodiorite suite and a less voluminous syenite/carbonatite suite. Absolute age determinations (Table 6) give a broad framework for the intrusive episodes, but insufficient studies are available as yet to detail the succession of intrusive episodes from monzodiorite to tonalite/granodiorite to granodiorite which appears evident from field relationships, nor to separate the syntectonic from the post-tectonic granodiorites.

All the plutonic rocks were intruded into the supracrustal rocks, creating contact metamorphic aureoles and engulfing large quantities of inclusions, and many of the rock compositions of successive suites are similar. Classification, therefore, depends on an accurate evaluation of the structural relationships. Many polyphase intrusions are laden with cognate inclusions, particularly of the earlier mafic phases, which may easily be confused with inclusions of mafic volcanic rocks.

Synvolcanic Plutons

The synvolcanic plutons have been resolved into five large masses scattered throughout the belt (Fig. 15). Some are clearly separate individual plutons, but others are an agglomeration of plutonic masses including numerous adjacent elongate slivers resulting from the principal deformation. The plutons often occur in the cores of anticlines, and regional structures swirl around them as they form resistant knots in the deformation pattern. Strong L-fabrics with steeply plunging lineations are developed in the intensely deformed zone around

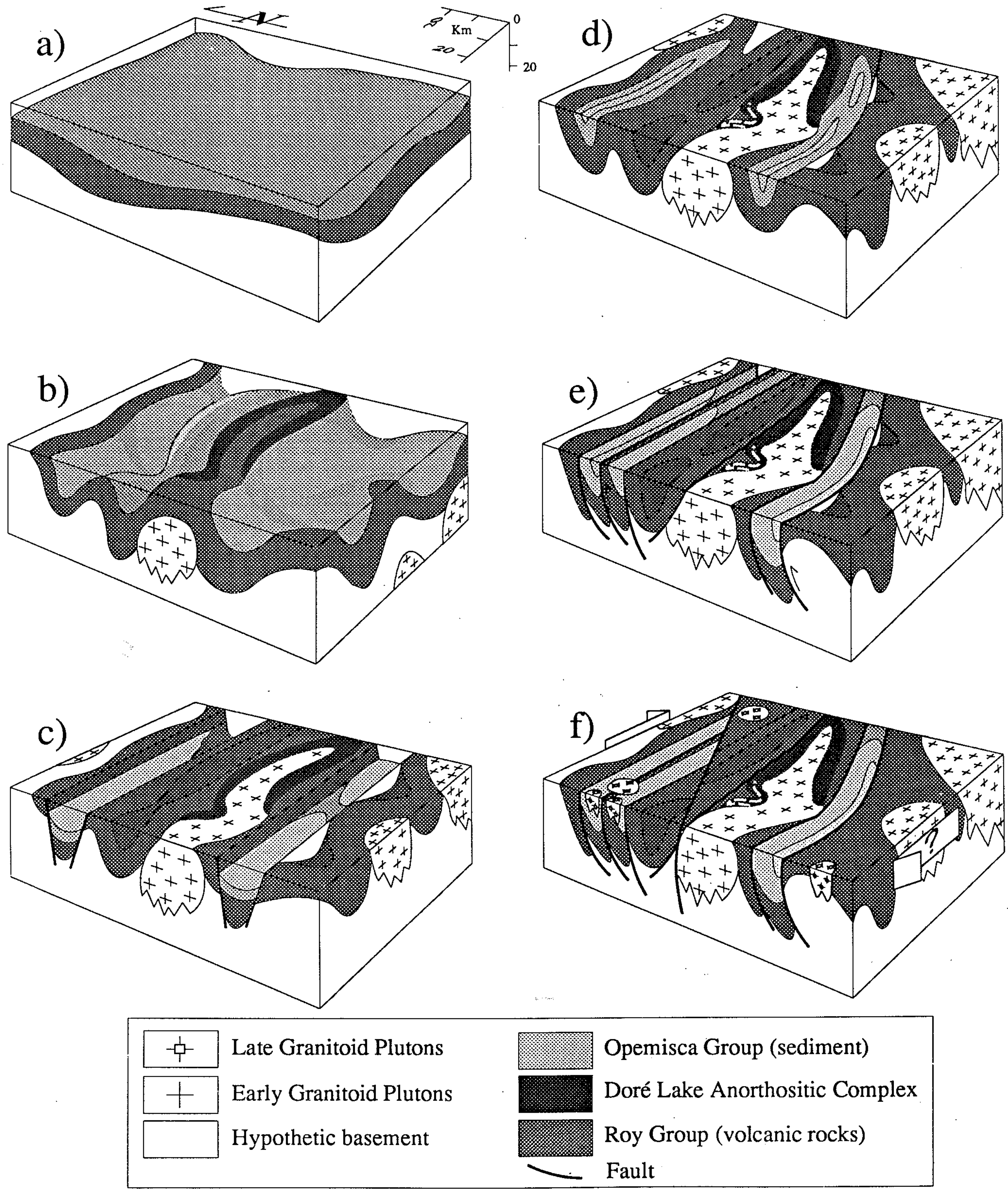


Figure 15. Structural evolution of the Chibougamau Region

such plutons (Daigneault and Allard, 1990). Syntectonic plutons have used the contacts of many of these complexes as an intrusion corridor, adding to the complexity of the plutonic mass and commonly causing confusion in the geologic evaluation of the plutons.

Synvolcanic plutons are thought to be largely subvolcanic cauldron subsidence complexes probably developing as a series of annular or eccentric intrusions. They are expected to be relatively flat bodies (Hansen et al., 1988; Bateman, 1984). The Aulneau batholith in the western Superior Province, which is petrologically and structurally very similar, has been shown by gravity analysis, to form a shallow saucer 7 km thick (Brisbin and Green, 1980). The present elongate form of most of the plutons is in part the result of deformation, and in part the result of surface expression. Regional deformation patterns are observed to swing around the batholiths and many of their margins are wide fault zones, although some flattening of these large masses is possible. Within the batholiths, the rocks have deformed along discrete mylonite zones transposing dykes and elongating enclaves to produce a banded rock. Apart from these zones the rocks show only local development of regional schistosity.

The plutons are polyphase intrusions, generally ranging from dioritic early phases to tonalitic and leucotonalitic, and occasionally granodioritic later phases. A composite model batholith may be envisaged from three better studied batholiths (Chibougamau, Eau Jaune and Lichen). Distribution of the phases is typically irregular, in part because the present exposure is the result of tectonism, but more importantly because the pluton did not have an originally symmetrical zonal arrangement. Contacts between phases of the same pluton may be gradual or irregular and are characterized by zones of cognate inclusions (Racicot, 1981, 1982). Septa of volcanic country rock also mark the partition between successive phases (Tait et al., 1988), which commonly become mylonite zones during regional deformation. All stages of plutonic activity are characterized by a vast array of chiefly porphyritic, dykes intruding both earlier phases of the pluton and occurring widely through the adjacent country rock. Unlike later tonalites, few of the dykes are pegmatitic. The intrusions commonly contain foundered blocks and xenoliths of country rock.

The irregular, polyphase nature of the plutons, the asymmetric distribution of these phases, and the vast array of dykes coupled with their close relation to the volcanic sequence (Jolly, 1980; Ludden et al., 1984) lead to the conclusion that these were intruded at shallow depths. The Chibougamau Pluton was eroded before deformation (Mueller et al., 1989) and the presence of plutonic clasts in sediments elsewhere in the Northern Zone suggests that other early plutons have also been uplifted and eroded. This places a probable limit of 5-6 km depth for the intrusion (Chown and Mueller, in prep.). The near surface expression of porphyry copper (Guha et al., 1984) and epithermal (Guha et al., 1988) mineralization associated with the Chibougamau Pluton is characteristic of an even shallower depth of intrusion. Estimates of depth of intrusion of 2-4 km based on barometric studies (Feng and Kerrich, 1989) for plutons of the Southern Abitibi appear to confirm this.

All the complexes must have had an original contact metamorphic aureole, which should have been reasonably extensive, judging by the volume of the intrusion. Few of these aureoles are more than partly preserved, as they have been either retrograded to greenschist facies or obliterated by a regional amphibolite facies metamorphism. The effects of regional metamorphism are variable within the plutons. The more mafic phases are particularly affected, with hornblende and biotite being retrograded to chlorite, and plagioclase sericitized and saussuritized. The relatively dry plutonic rocks are generally only affected near their contacts.

Tonalitic gneiss is a coarse-grained orthogneiss with a strong foliation and relatively little compositional layering. The gneiss does not contain xenoliths of supracrustal rocks, in contrast to all the other plutonic rocks. The gneiss is intruded by a regular succession of dykes related to synvolcanic plutons (Chown and Mueller, in press). These dykes are in turn cut by the regional deformation structures and dykes related to the syntectonic plutons. Although it appears to be structurally older than all other plutonic rocks, the gneiss may represent early ductile deformation in deeper parts of synvolcanic plutons. The close relation between the gneiss and syntectonic intrusions indicates that the emplacement of the gneiss at its present structural level took place during regional deformation.

Syntectonic Plutons

The syntectonic plutons are nearly all elongate parallel to the predominant tectonic fabric of the belt. Most occur at the discontinuity between crystalline rocks (synvolcanic plutons or mafic stratiform complexes) and the more anisotropic supracrustal rocks. The intrusions, tabular in shape and up to 30 kilometres in length, appear as lenses or ellipses on the map due to of the dominance of vertical structures in the belt. Some, however, intrude as sheets or phacoliths exploiting the subhorizontal top of earlier intrusions, and these may have extremely irregular shapes as a result of the erosion of shallow-dipping tabular bodies. A relatively small number of syntectonic plutons occur as isolated intrusions wholly within the supracrustal rocks, commonly along major structures.

Syntectonic plutons have narrow amphibolite grade contact metamorphic aureoles imprinted on the regional greenschist metamorphism. Their close association with regional deformation suggests the probable level of intrusion was 6 to 8 km, the level of formation of the greenschist facies.

The magmatic flow foliation within syntectonic plutons, defined by oriented tabular minerals and flattened xenoliths, follows the outline of the body and is commonly subparallel to regional deformation structural patterns. Most plutons have a wide marginal zone of ductile deformation, producing rocks with a mylonitic fabric and pronounced flattening of xenoliths. The width of this ductile deformation zone varies from intrusion to intrusion. In general, it is much wider in the tonalites than in the monzodiorite or granodiorite intrusions.

The three petrographic suites of syntectonic plutons, monzodiorite, tonalite/granodiorite and granodiorite appear to occur in this order in individual areas (Benn et al., 1989; Chown and Mueller, in press), although the available absolute age determinations (Table 6) do not show a clearcut division. Inclusion of the granodiorite suite as late syntectonic intrusions is a departure from earlier classifications (Racicot et al., 1984). Recent work (Benn et al., 1989; Midra et al., in press; Moukhsil, personal communication) demonstrates that larger granodiorite intrusions were also intruded under regional tectonic controls, and were affected by the last phase of strike-slip movements during emplacement.

The principal rock type of the monzodiorite suite is a coarse-grained monzodiorite to quartz monzodiorite ranging to granodiorite with varying quartz content, and is a consistently mafic-rich rock with an M of 15 to 22. Hornblende dominates over biotite as the principal mafic mineral, and pyroxene is common. Hornblende, and less commonly pyroxene, form primary layers. A wide variety of cognate mafic phases occur as an inclusion-rich border zone enclosed in a leucogabbro to tonalite matrix in many of these intrusions. The mafic early phases are also present as amphibolite dykes intruding both the adjacent crystalline rocks, and rarely the supracrustal suite. Similar suites of rocks have been described from elsewhere in the Abitibi Belt and adjacent belts (Stern, 1989; Sutcliffe et al., 1989). The plutons rose along the faulted margin of a synvolcanic pluton, spreading laterally along the roof of the pluton and projecting up into the overlying supracrustal rocks, possibly ballooning slightly.

Tonalitic intrusions show a multi-stage evolution from melatonalite at the border to granodiorite near the centre. The border phases have been intruded by successive dykes documenting the petrologic evolution of the pluton. Ductile deformation, affecting most of the early phases, is probably related to the emplacement of the pluton. It is in turn cut by láter ductile deformation zones parallel to regional deformation. The structures of a typical tonalite pluton are almost identical to those of the monzodiorite plutons.

Granodiorite plutons are the most numerous intrusions in the region. Two distinct types of pluton are observed, large sheet-like bodies, generally intruding along a contact, and discrete small stocks, commonly in clusters of two to four plutons. Mineralogic and textural similarity among groups of small stocks suggests they are derived from larger underlying plutonic masses. The larger plutons usually demonstrate syntectonic structural characteristics, with the development of magmatic and ductile fabrics particularly related to the latest regional strike-slip movements. The smaller stocks, circular in plan from several hundreds of metres to 10 km in diameter, display concentric structures, and are interpreted as cylindrical structures. Granodiorite bodies in the Opatica Belt to the north are observed to contain inclusions of both tonalite and monzodiorite (Benn et al., 1989).

Post-Tectonic Plutons

At least three small leucogranodiorite to syenite stocks with associated carbonatite and lamprophyre dykes have been observed associated with northeast trending faults (Bédard, 1987; Morasse., 1988; Proulx, in press). The leucogranodiorite is zoned inward to syenite (Bedard, 1987) and cut by many aplite dykes of granodiorite composition. Later pyribole dykes show more alkalic tendencies. Biotite sovite dykes with narrow zones of fenitization occur in close association. Olivine-biotite lamprophyres are the latest phase of the suite. Some porphyritic granodiorites were possibly intruded during this later period of deformation. Lauzière (1989) has demonstrated the close relationship between a small granodioritic stock and the late northeast-trending fault system. The association of small plutons with late northeast-trending faults may indicate that they are part of an extensive period of late tectonic activity.

TECTONIC MODELS

Based on similarities with other studied areas tentative models may be proposed relating the origin of the volcano-sedimentary supracrustal sequence, the deformation and the sequence of intrusions.

Modern analogues of the supracrustal and associated plutonic suite described above are found in the Western Pacific, the Andaman Sea marginal basin and the western coast of South America. Plate tectonic movement at convergent plate boundaries and intra-arc settings are responsible for this type of succession and volcano-sedimentary configuration in modern-day environments. Plutonic activity associated with volcanic accumulation produced polyphase batholiths composed of relatively primitive tonalite/diorite on a fairly regular spacing across the belt. Some of these occur within evolved volcanic complexes, but most intrude the basalt plain. This synvolcanic tonalite/diorite suite appears to be consistent with a derivation from one or more subducting oceanic plates. The petrographic composition and tectonic setting of these intrusions, when compared to other areas (Beakhouse et al. 1989 Type 1; Martin, 1986) supports this interpretation.

The principal driving mechanism in the Archean has not been established with any certainty, but it is generally believed that a modified form of plate tectonics operated at this time (Windley, 1986, Sleep and Windley, 1981, Tarney et al. 1976). The Northern Abitibi Zone , composed of an initial volcanic cycle (2720-2730 Ma) and a second cycle restricted to the Chibougamau Segment (2705-2720 Ma) are both characterized by terminal sedimentary cycles which indicate an extensional phase. Mueller et al. (1989) consider this type of succession to represent an incipient island arc that evolved in certain sections into a mature arc, as documented by the late-stage shoshonitic volcanism. The presence of volcanics of the same age (2718 Ma) in the Southern Abitibi (Corfu et al., 1989), suggests the existence of a volcanic arc along the south margin of the Northern Zone, with a broad back-arc area to the north where a shallow-dipping subducting plate generated tonalite/diorite magma over a relatively wide area.

The initial phase of deformation formed either by gravitational subsidence or by the earliest effects of the N-S horizontal shortening (Fig. 15B). The broad N-S trending folds developed before regional foliation and may have originated at the same time as juvenile E-W folds. Their origin may be related to a topographic depression in the volcanic terrane or to a distinct phase of deformation. These early flexures did not produce an axial plane foliation but the combination of the N-S and E-W systems generated the framework of a regional interference pattern of domes and basins dominated by the large E-W synclines.

Juvenile synclines formed from marginal grabens filled with sediment derived from the eroding volcanic edifice (Fig. 15C). N-S horizontal shortening followed this early extensional phase and developed the dominant, pervasive regional E-W foliation. The isotropic synvolcanic plutons interfered with the N-S horizontal shortening (Fig. 15D) participating in the upturning of the enclosing strata to the vertical, and creating a wrap-around pattern in the regional schistosity. These structural aureoles are characterized by the development of strongly linear tectonites.

The climax of the regional deformation is responsible for the tightening of the large E-W folds and the steepening of their axial planes parallel to the vertical stretching lineation. Finally, the last episode of the regional deformation (Fig. 15E) led to the development of the long E-W reverse faults and zones of intense shearing. Many of these major faults appear to be reactivated early normal faults.

The most evident result of the regional deformation is the important N-S horizontal shortening which created the dominant E-W foliation and regional isoclinal folds. The origin of the N-S shortening may be related to plate convergence and possibly subduction. Accretion of a series of volcanic arcs onto other arcs in a north-to-south progression across the Superior Province (Card, 1989), by a northerly dipping flat-plate subduction as initially inferred by Dimroth et al. (1983) probably caused N-S compression resulting in E-W trending folds.

The persistent vertical regional schistosity and down-dip stretching lineation is the major constraint to tectonic evolution models for the region. It is well documented in Phanerozoic orogenic belts that the most efficient horizontal shortening is produced by low angle thrust faults and nappes. These processes are postulated for the Himalayas (Gansser, 1964), the Western Cordillera (Coney, 1989) and Eastern Appalachian Belt of North America (St-Julien and Hubert, 1976; Williams and Hatcher, 1982). These examples are clearly cases of sub-horizontal strain zones associated with a stretching lineation parallel to the transport direction. The low energy budget of this process is an attractive alternative to the high energy processes required for a vertical response to deformation by the supracrustal pile. To apply the thrust model to the Abitibi Belt, one can propose that the regional east-west high angle reverse faults are, in fact, subhorizontal faults, rendered subvertical by continued horizontal shortening.

The large E-W longitudinal faults represent the final stage of the regional deformation. The structural elements observed in these large faults suggest high angle thrusting. In spite of their great lateral persistence within the Chibougamau area, the transport was never great enough to obliterate the stratigraphic relationships and the area is part of an contiguous domain, and not a series of nappes. The coherent stratigraphic sequence suggests that the much of Chibougamau Segment behaved in a coherent manner, and a thrust fault model must consider the whole Chibougamau Segment to be one terrane.

Examination of the major E-W faults in conjunction with the southeast faults, which display strong subhorizontal stretching lineations, clearly indicates a dextral movement sense (Daigneault and Archambault, 1990). This suggests a late dextral movement superimposed on the reverse movement recognized in the majority of these faults (Fig. 15F).

Syntectonic plutons consist of largely tabular bodies intruding around synvolcanic plutons and clearly controlled by the deformation. This took place during an extensional event following the major compression. Successive development of the intrusive suites, from monzodiorite to tonalite/granodiorite to granodiorite is observed, based on field relations. Plutons post-date the principal deformation and were affected by the latest phases of deformation. The monzodiorite and tonalite/granodiorite suites are intruded during, or slightly after, the main period of folding and development of regional schistosity. The granodiorite suite, on the other hand, appears to have intruded during the final period of dextral strike-slip faulting which concluded the major deformation. Small stocks of granodiorite within the supracrustal sequence are believed to be offshoots of deeper large granodiorite bodies. These rarely show unequivocal relations to the major structures, and most may well be late syntectonic in age. Thus, arc-arc collision and major regional deformation from 2700 to 2695 Ma (Corfu et al., 1989) successively produced monzodiorite, tonalite/granodiorite and granodiorite magmas as products of partial fusion of the lower crust and mantle, the former thickened by the deformation. The basic similarity between the syntectonic tonalite/granodiorite and granodiorite suite suggests derivation from much the same source, either from the mantle (Sutcliffe et al. 1989) or possibly by partial fusion of the base of the folded volcanic pile. Monzodiorite magmas were probably subduction related, from a deep crustal or mantle source (Stern et al., 1989; Bédard et al., 1989).

A small suite of syenites with associated carbonatites, as well as some granodiorites, occur along late northeast structures, suggesting that their emplacement took place during the last period of post-tectonic adjustment. These latest intrusions of porphyritic granodiorite and syenite/carbonatite are thought to be controlled by late tectonic faults, and are petrologically consistent with derivation from beneath a thickened crust.

PROTEROZOIC

The Chibougamau region underwent periodic activity during the Proterozoic. Two periods of sedimentation and four separate igneous (dyke) events are known. In addition, the Chibougamau region lies on the border of the Archean craton, and was the autochthonous foreland to the Grenville Orogen. Much of the district has been moderately to profoundly affected by this major deformational event.

Both sedimentary formations, the Chibougamau Formation and the Mistassini Group are relatively undisturbed, except near faults, and have undergone little more than low grade burial metamorphism. Similarly the dykes have suffered only local deuteric alteration and minor development of prehnite. Dykes are seldom more than 100 m thick, and typically have sharp glassy contacts. Most may be traced by their prominent positive linear anomalies on regional aeromagnetic maps. Dykes belonging to four swarms, the Preissac, Mistassini, Otish and Abitibi swarms, are present in the Chibougamau district.

Chibougamau Formation (2450-2490 Ma)

The Chibougamau Formation consists of up to 200 m of boulder to pebble conglomerate and sandstone lying unconformably on the folded Archean basement. The conglomerates are made up of fragments of granitic and metavolcanic rocks representing all the Archean lithologies, contained in a fine-grained angular matrix. The sedimentary structures and distribution of facies led Long (1974) to postulate a fluvial origin for the formation, and the presence of mixtites, and dropstones in banded argillites suggests a glacial origin. The formation is derived from the northwest. The Chibougamau Formation is tentatively correlated with the Gowganda Formation of the Huronian Supergroup, 400 km to the southwest (Young, 1970).

The Chibougamau Formation occurs in a series of outliers north and west of the town of Chibougamau. Clastic dykes undoubtedly related to the formation are found within the Archean basement up to 80 km west of the outliers (Chown and Gobeil, in press) strengthening Young's (1970) correlation and indicating that the present erosion surface is close to the post-Archean land surface.

Preissac Dyke Swarm (2140 Ma)

The Preissac dyke swarm consists of nearly north-south trending dykes cutting much of the eastern Abitibi belt (Fahrig et al., 1985). The quartz tholeiitic Gabbro Island dyke of the Chibougamau region (Chown, 1984) appears to be part of this swarm, but other prominent north-northeast dykes may belong to the younger Otish swarm (Chown and Archambault, 1987).

Mistassini Dyke Swarm (1960 Ma)

The Mistassini dyke swarm (Fahrig et al., 1986) is a northwest-trending dyke system which occurs in a spectacular fanning array just north of Chibougamau. The dykes range in composition from peridotite to olivine tholeiite. Only a few examples of this dyke swarm occur in Chibougamau, but include the pyroxenite dyke outcropping on islands in Lake Chibougamau and in some of the mines in the Doré Lake Complex. The other Mistassini dykes in the area are olivine tholeiites.

Mistassini Group (1950-1750 Ma)

The Mistassini Group consists of a thick dolomite sequence (Lower and Upper Albanel formations) composed of stromatolitic and argillaceous dolomite with interspersed black shale units. This is overlain by a mixed formation of quartzite, argillite and iron formation (Temiscamie Fm.). Clastic fluviatile formations (Papaskwasati and Cheno fms.) delineate the northern margin of the sedimentary basin. The group unconformably overlies the Archean basement, the unconformity being marked by a thick paleosol which has been largely replaced by carbonate during the diagenesis of the overlying dolomites (Chown and Caty, 1983). The southernmost extent of these sediments is immediately to the north of Chibougamau.

Otish Dyke Swarm (1730 Ma)

The prominent north-northeast trending swarm of olivine tholeiite dykes extending north from Senneterre over 900 km includes most of the dykes in the Chibougamau area. These dykes range from coarse-grained ophitic textured gabbros at the centre of dykes to fine-grained chilled phases at the contact. Large dykes have numerous apophyses and parallel dykes associated with them.

Abitibi Dyke Swarm (1140 Ma)

The Abitibi dyke swarm (Ernst et al., 1987) was intruded along fractures generated by crustal flexure during the early stages of the Grenville Orogeny (Ranalli and Ernst, 1986). Two extremely continuous dykes extend from Lake Superior and terminate in the Chibougamau region near the margin of the Grenville Orogen. These feldspar phyric quartz tholeiites are commonly multiple injection dykes.

Grenvillian Orogeny

The Grenville Orogen, a complex zone of late Proterozoic deformation, lies southeast of the Chibougamau region. Although existing mapping of the Archean Abitibi Belt suggests the latter effectively terminates against the Grenville Front just southeast of the town of Chibougamau, the results of recent reconnaissance mapping indicate that major units may be traced across the front into the Parautochthonous Belt of the Grenville. The Grenville Orogen is interpreted to be the result of a continent-continent collision (Rivers et al., 1989), in which the Grenville mobile belt was shortened against the Archean (Superior Province) foreland to the north. The Grenville Front is mapped locally as the limit of readily recognizable Archean formations and structure. Elsewhere, the Grenville Front is mapped as the northern limit of Grenville structure and metamorphism, which has resulted in some confusion concerning its location in the Chibougamau area.

It is important to realize that two major structural events took place in the Archean Parautochthon during the Grenville Orogeny. First, large nappes were thrust from southeast to northwest, and following the thickening of the crust produced by the first event, isostatic adjustment took place on a series of high-angle reverse faults.

The first structural event produced a pervasive southeast-plunging stretching lineation in all the rocks of the Parautochthonous Belt (Ciesielski and Ouellet, 1985), and a high temperature/high pressure regional metamorphism on the Proterozoic rocks bordering and within the parautochthon. Grenvillian metamorphic effects on the already metamorphosed Archean rocks are more difficult to characterize because of variable overprinting relations. The principal manifestation is the development of large garnet and hornblende porphyroblasts in the abundant metabasites, commonly overprinting a background greenschist facies assemblage, and occurring in conjunction with the development of younger structures (Ouellet, 1988). High pressure assemblages, in particular kyanite bearing assemblages, are noted in some Archean rocks, similar to the assemblages developed in Proterozoic rocks and in contrast to the typical medium pressure andalusite bearing assemblages of the Abitibi Belt (Jolly, 1978). Feathery hornblende porphyroblasts yielded Ar39/Ar40 plateau ages from 1105 Ma at the upper greenschist facies to 953 Ma in the amphibolite facies (Baker, 1980). Proterozoic diabase dykes also develop coronitic structures and grade into garnet amphibolites within the Parautochthonous Belt, although the metamorphic changes in these relatively dry rocks do not appear as far north as corresponding changes in the metabasites. The imprint of Grenvillian amphibolite grade metamorphism along the edge of the foreland is relatively clear, leaving a vast zone to the northwest where Grenville-related greenschist and lower grade metamorphism must have overprinted Archean metamorphism, but whose effects are not readily identified. N-E trending folds associated with an axial plane crenulation cleavage and the amphibolite facies metamorphism are common.

The isostatic readjustment phase of Grenvillian deformation resulted in a spectacular series of 010-020 trending reverse faults (Fig. 8). These faults have a combined sinistral and east-side-up movement. Faults in the south and east deformed in a dominantly ductile manner, whereas those farther west display brittle characteristics. Late brittle movements on the ductile faults are shown by the presence of pseudotachylite (Daigneault and Allard, 1990), which have produced a systematic sinistral throw of a few 100 to 1000 metres. These faults, in particular the Mistassini Fault, the most pronounced member of the system, dictate much of the configuration of the mapped location of the Grenville Front in the Chibougamau region. It is clear that some, if not all, Archean faults (eg Gwillim) have been reactivated during the Grenvillian Orogeny.

A major result of the Grenvillian Orogeny has been to uplift the Archean rocks of the Parautochthonous Belt so that the supracrustal rocks of the Abitibi Belt are in juxtaposition with tonalite, tonalite gneiss and small lenses of metabasite from deeper levels of the Archean crust. Clinopyroxene-garnet rocks are widespread within the Parautochthon, partly retrograded to garnet amphibolites. Thermobarometric analysis of these assemblages yields 800 C and 1000 mpa (Ouellet, 1988). It is suggested, but not yet proven, that these high pressure rocks are the product of Grenvillian, rather than Archean metamorphism.

METALLOGENY OF THE EASTERN EXTREMITY OF THE ABITIBI BELT

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Since 1954 the Chibougamau mining district has produced approximately roughly 1.2 million tonnes of copper, 115,000 kgs of gold, 650,000 kgs of silver, 115,000 kgs of zinc and 4000 kgs of lead. The district hosts several types of ore deposits and occurrences in a variety of settings (Fig. 2). The multitude of vein type mineralization makes the task of correlation harder, but examined within the geological framework, a pattern emerges. There have been a number of publications detailing different deposits and settings and in two publications there has been an attempt to sort out the different types of occurrences in relation to their mode of emplacement (Guha, 1984; Guha et al., 1988). This section will provide a short overview in order to visualize the geological position of some of the occurrences that will be visited and summarize the earlier classifications in the context of recent advances in the understanding of the tectono-stratigraphic evolution of the area.

Although precise dates of the mineralization/alteration are unavailable, the dating of some of the important lithologies makes it possible to correlate the mineralizing events with the evolutionary cycle of the deformed Archean belt. Mineral deposits can be divided into 6 broad categories (each having sub-categories), based on their characteristics, settings and the geological process(es) that played a role in their emplacement.

I-Mineralization related to the emplacement of mafic intrusions (oxides and sulphides of magmatic origin)

II-Sulphide deposits related to the synvolcanic period.

III-Mineralization related to high level plutonic activity and associated volcanic landforms

A. Porphyry-type and associated mineralization

B. Epithermal precious metal veins

IV-Archean mesothermal gold deposits spanning a period of major deformation and plutonic activity.

V-Deposits of uncertain age: Opemiska-type veins

VI-Mineralization with a major post-Archean influence:- Cu-Au veins of the Dore lake Complex.

This is not an attempt to classify all the mineralized occurrences of the camp nor a detailed description of the deposit types. A number of prospects are currently being studied, and their position in this scheme is an important element in their evaluation. In the following section each category will be examined briefly. Deposits within the Proterozoic Mistassini basin and metamorphosed equivalents of Archean mineralization in the Grenvillian Orogen have not been considered.

I Mineralization related to the emplacement of mafic intrusions (oxides and sulphides of magmatic origin)

These occurrences are related to the emplacement of the major mafic and ultramafic sills such as the Dore Lake Complex (DLC). The DLC has been described in detail by Allard (1976) and Allard and Daigneault (1990) and they indicate economic grades for vanadium in the P1 member of the DLC in Rinfret Township. Nearly 245,000 000 tonnes of ore with an average grade of 27.6 % Fe and 1.1% TiO_2 have been outlined from the Mont du Sorcier and Magnetite Bay area. Assessment of PGE associated with sulphides indicates that no enrichment has occurred, due probably to an early scavenging of the PGE from the magma at depth (Barnes and Couture, 1990). A Ni-Cu sulphide showing (Lantagnac Township) in a metapyroxenite intrusion, on the other hand, does contain PGE concentrations (Barnes and Couture, 1990).

II Sulphide deposits related to the synvolcanic period.

The development of incipient island-arcs (Mueller et al., 1989) with active felsic centres favoured development of hydrothermal systems that produce volcanogenic sulphide deposits. Evidence of synvolcanic alteration and mineralization is widespread. This includes silicification of large tracts of the Gilman Formation (Couture, 1986; Trudeau, 1981) as well as the hydrothermal alteration of the Lac des Vents complex described in this guide. The collapse of small volcanic edifices produced megabreccias overlying pyroclastic debris flows and tuffs accompanied by mineralization associated with synvolcanic faulting and sulphidized pillow rims (Bouchard, 1986). Discharge of fluids into the volcaniclastic rocks or into paleotopographic depressions gave rise to bedded sulphides, which can be seen in the upper units of all the cycles. Most of these accumulations are pyrrhotite and pyrite-rich with varying amounts of Cu, Zn, and precious metals.

Significanct deposits significance discovered so far are Lemoine (728,000 tonnes at an average of 4.2 % Cu, 9.6 % Zn, 4.5 g/t Au and 83.85 g/t Ag - mined), Lac Scott (680,000 tonnes at an average of 0.55 % Cu, 6.9 % Zn and 13.3 g/t Ag - reserves), Cooke 8-5 , Lac des Vents (100, 000 tonnes at 2 % Cu) and Coniagas (700, 000 tons. at an average of 10.7 % Zn, 1.0 % Pb and 0.3 oz/t Ag - mined) . The Waconichi Formation was considered a favourable felsic stratigraphic unit with the discovery of the Lemoine deposit and the Lac Scott prospect, based on the concept that volcanogenic massive sulphide deposits occur in clusters confined to specific horizons. Why should this felsic sequence be favoured? The probable answer to this is the development of a heat source capable of generating the hydrothermal system. The porphyritic rhyolite of the Waconichi Formation at Lemoine formed at 2728 $\pm$ 1.5 Ma and the DLC (Granophyre Zone) at 2728 $\pm$ 1 Ma (Mortensen, in prep.). This time frame indicates that the DLC was available to provide the heat to generate the hydrothermal sytem required for metalliferous sea-floor deposits. The presence of such a hydrothermal system can be envisaged from indications that a large volume of rock was affected by hydrothermal fluid-rock interaction. Gobeil (1980) observed Ca and Na depletion in rocks stratigraphically below and laterally more than a kilometre to either side of the Lemoine mine. Felsic centres such as this were instrumental in developing high heat flow areas which then acted as foci for fluid discharge. The nature of the Lemoine deposits (bedded/laminated sulphides and structures indicating flow on the slight slope of an edifice) points to more of a slope plus topographic low type of accumulation than a mound type. The supposition that DLC acted as the subvolcanic heat engine is indirectly supported by the fact that the "Obatogamau bridge rhyolite" (cf. Daigneault and Allard, 1990), a possible stratigraphic equivalent of the Waconichi Formation (dated by Mortensen as 2728 $\pm$ 1 Ma), away from the influence of DLC does not show this hydrothermal activity. Moreover, the hydrothermal activity in the Gilman Formation on the northern flank of the DLC, which shows silicification with large lateral extent, could well be due to a hydrothermal system kept active by the DLC. The preservation of the deposits in the Waconichi Formation is the result of a number of post-depositional factors such as the final position of the DLC, the intrusion of the Chibougamau Pluton, deformational events and erosion. The Lac des Vents Complex shows more explosive activity which favours a mound type evolution with repeated fluid surges and eruptive breccias giving way to lenticular deposits with the rapid build-up of the volcanic edifice. The emergence of the volcanic edifice to subaerial conditions did not favour the preservation of massive sulphide accumulations nor their formation.

III-Mineralization related to high level plutonic activity and associated volcanic landforms

A. Porphyry type and associated mineralization

The emplacement of tonalite/diorite synvolcanic plutons, thought to be generated by a subducting oceanic plate forming high level cauldron subsidence complexes, would be extremely favourable to the development of porphyry type mineralization. This is well illustrated by the Chibougamau pluton, where typical porphyry copper-type mineralization is exposed. The Queylus prospect is considered to be a typical porphyry showing on the basis of the large areal extent of its alteration, its hydrothermal breccia zones and its stockwork/ disseminated copper mineralization (Cimon, 1976; Bureau, 1980; Bureau et al., 1979; Racicot, 1980). The Devlin deposit which also belongs to the same mineralizing episode (Guha et al., 1984), is a flat-lying vein system, with distinctive breccia zones and a widespread selective alteration pattern (Bureau, 1980; Guha et al., 1984).

B. Epithermal precious metal veins

The development of a hydrothermal system and the type of deposit associated with felsic plutons is related to the physical evolution of the volcanic system. Archer (1984) and Dimroth et al. (1984) have shown that there was growth and subsequent uplift of volcanic islands during the deposition of the Blondeau Formation and that this volcanic island was probably underlain by plutonic rocks. In subaerial conditions such as these any porphyry-like hydrothermal system developing at deeper levels within or centred around subvolcanic felsic intrusions would manifest itself in the upper parts of the volcanic pile as epithermal Au-Ag-Pb-Zn-As veins controlled by synvolcanic faults; or intrusion- related faults generated by cauldron subsidence. This has been recognized in this mining district (Guha, 1984). One of the best example is the Berrigan prospect (Guha, 1984; Pilote, 1987; Guha et al., 1988) described in detail as one of the stops. Although affected by subsequent deformation, this epithermal mineralisation can be identified by its geometry which is controlled by synvolcanic faults; and by the associated breccia zones and vein textures indicating extensional regimes, although subsequent deformation has obliterated many of the latter criteria. The epithermal veins are best developed in the mafic to ultramafic sills due to the competence of the these units. The segmenting of these sills during subsidence makes the identification of the synvolcanic faults possible.

IV Archean mesothermal gold deposits spanning periods of major deformation and plutonic activity.

This category of deposits spans the range of deformation and felsic plutonic intrusion. In spite of the lack of precise age dating of the mineralization and alteration, the structural setting provides a clue to the late syntectonic emplacement of these deposits with respect to the major deformation of the Archean rocks. The major deformation period probably related to arc-arc collision also produced late syntectonic plutons. The hydrothermal alteration associated with the gold mineralization which is superimposed on regional greenschist facies metamorphic assemblages also demonstrates late emplacement.

In the Chibougamau segment the mesothermal deposits are best developed in the Bourbeau sill, a differentiated unit which shows certain similarities with the Golden Mile dolerite (Dubé, 1990; Dubé and Guha, 1989; Allard, 1982). In the Caopatina segment mesothermal gold occurrences are found along an E-W axis stretching for several kilometres from the Joe Mann Mine to the Bachelor Lake mine. Although specific features vary from one deposit to the other, there are some common traits Most of them are in E-W trending shear zones or conjugate NE and NW systems, and are part of, or subsidiary to, E-W trending regional shear zones. The geometry of the ore lenses is due to dominant vertical reverse movements, and carbonate alteration is ubiquitous. Several deposits and showings show a spatial relationship to regional NE- or-NW trending faults (Guha et al., 1988). The host unit is commonly a mafic intrusive and in some cases porphyritic felsic dykes are closely associated. Some occurrences are hosted by felsic plutons or stocks, such as Meston Lake (Dion et Guha, 1989), and showings in the Chibougamau pluton where this category of mineralization post-dates the greenschist facies metamorphic assemblage (Ouellet, 1986). Not all are vein-type deposits but some, such as the Tadd prospect in the Bourbeau sill (Dubé et al., 1987) and the Philibert prospect (Dion et Guha., 1989), are associated with disseminated pyrite in altered and sheared host rocks .

There are some marked differences in the composition of the ore bodies even though they are controlled by the same structural event and hosted by a similar lithology, as is the case of Cooke and the Norbeau mines hosted by the Bourbeau sill. The former is more Cu rich and has a poorly developed carbonate alteration, and the latter is gold-rich and has well developed carbonate alteration. Significantly different deposits which show a strong correlation with late tectonic porphyritic granodiorite and syenite/carbonatite intrusions such as Lac Shortt (Morasse, 1988; Morasse et al., 1988; Quirion, this volume) and the Bachelor Lake Mine (Lauzière, 1989) are observed. A marked hematization and potassic alteration is present in both Bachelor Lake and Lac Shortt mines and a sodic amphibole alteration is present in the latter (Morasse, 1988, Lauzière, 1989). Although not directly pertinent to this section, the carbonatite bodies may themselves be classed as a distinct group due to their rare earth and strategic metal potential since more and more carbonatite intrusions are being noted and evaluated (Bédard, 1988; Quirion, this volume; Prud'homme - M.Sc. thesis in progress).

The genesis of these deposits may be summarized as follows. The emplacement is evidently structurally controlled and the host rock plays a dual role. Firstly, a mechanical role is indicated for the more competent rocks such as gabbroic sills and felsic porphyries, and secondly a chemical role is indicated where fluid-rock interaction not only produced alteration assemblages but helped fix the gold (Dubé et al., 1987; Guha et al., 1988 and in press). The source of the fluid (with varying CO_2-H_2O contents Guha et al., in press), is not known precisely for all the deposits. Metamorphogenic fluid cannot be ruled even though the hydrothermal activity appears to post-date peak regional metamorphism in a majority of examples. A case may be made for a magmatically derived hydrothermal fluid for the Bachelor Lake and Lac Shortt deposits (Lauzière, 1989; Morasse, 1988; Morasse et al., 1988), and a reconnaissance oxygen-hydrogen isotope study (Tremblay, 1987) indicates a possible magmatic affiliation for the Cooke deposit based on only two analyses. A diffuse pyrite-gold "halo" type mineralization in the Lac Shortt deposit appears to be a hydrothermal event related to the syenite intrusion which precedes mylonitization, although its extent and the geometry is not completely known. It must also be emphasized that there has been surface water percolating through the rocks during the evolution of Proterozoic rocks (Guha, 1984; Guha and Kanwar, 1987), and care must be taken as to which generation of fluid inclusions is analyzed for the deuterium signature of the fluid.

V-Deposit of uncertain age: - Opemiska-type veins

The reason for placing the Opemiska veins in a category of their own is the fact that their structural position is still open to discussion. Derry and Folinsbee (1957) and Watkins and Riverin (1982) relate the structures to the Gwillim Fault, whereas Lavoie (1972) and Morin and Boisvert (this volume) demonstrate that the structures related to the Opemiska veins can be correlated with the major folding events. From the work of Daigneault et Allard (1990), and Dimroth et al. (1984) it is known that the major movement on the Gwillim Fault post-dates the regional E-W trending shear zones. These veins have higher copper versus gold content compared to mesothermal deposits of category V, and are hosted by the coarse gabbroic phase of the Ventures sill. A characteristic feature of the Opemiska veins is the presence of significant quantities of scheelite and molybdenite, which is rare or non-existant in the mesothermal type (category V) or the Cu-Au veins of the Dore Lake Complex. The main mineralization of the Opemiska vein is overprinted by late pitchblende-uraninite-molybdenite mineralization of an uncertain age (Guha, 1984).

VI-Mineralization with a major post-Archean influence:- Cu-Au veins of the Dore Lake Complex.

The Cu-Au deposits of the DLC, along with the Opemiska mines, gave the mining camps its start. Altogether 14 deposits have been discovered and the total production (up to 1989) has been 39.5 million tonnes at 1.57 % Cu and 1.95 g/t Au. At present three mines are still in operation.

Most of the deposits are located in NW trending shear zones, except Henderson and Portage which are in NE trending shear zones in the meta-anorthosite and gabbroic meta-anorthosite layer of the Dore Lake Complex. The host rocks are quartz-carbonate-sericite and/or chlorite schists produced by the shearing and alteration of the meta-anorthosite (Allard, 1976; Guha, 1984). The sulphide minerals are predominantly chalcopyrite, pyrite and pyrrhotite with minor amounts of sphalerite and galena. Gold occurs mainly as discrete grains associated with pyrite and chalcopyrite, in contrast with the free gold that is observed in the lode type deposits in the region.

Although they can be classed as structurally-controlled vein-type deposits, certain features make them somewhat different than the Archean mesothermal gold deposits found in the region. In contrast to the mesothermal deposits, the dilation zones are extremely large and the ore types vary from laminated sulphide schist to dilation type fillings with large blocks of rubble breccia at the margin of dilation zones (Guha and Koo, 1975; Guha et al., 1983). The geometry of the veins in the Henderson 2 Mine indicates that this orientation cannot be developed by a vertical reverse movement. On going work of Archambault and workers indicates superposition of structures of different episodes in the Copper Rand Mine. The ore-forming fluids were characteristic $CaCl_2$-NaCl rich brines with a coexisting methane rich fluid (Guha et al., 1979).

The following points summarize the facts and underline the problems pertaining to the genesis of these deposits. The ore fluids, as well as the character and attitude of the different ore lenses evolved with the structural deformation (Guha et al., 1983) indicating an emplacement synchronous with the shearing. The ore fluids have a strong "surface" water signature and the sericite and chlorite is also in isotopic equilibrium with these waters (Tremblay, 1987). A systematic homogeneous Proterozoic age is obtained for galenas from the ore deposits (Thorpe et al., 1981, 1984). This evidence led to the interpretation that the deposits in their present form are post Archean (Guha, 1984; Guha et al., 1988). The question remains as to whether there could have been a pre-existing deposit which has been remobilized. It is important to emphasize that a number of deposits that have undergone deformation and remobilization, whether lode-gold or volcanogenic massive sulphides, may be reconstructed by studying preserved primary features. Entry of later fluids in pre-existing gold deposits has been documented and the signature of the primary mineralizing fluid can still be seen (eg, Macassa Mine, Kerrich and Watson, 1985). Guha (1984) discussed the development of these deposits and the possibility of earlier events. It has been shown that the dykes (felsic to mafic) observed in these deposits can be related to the Chibougamau Pluton (Maillet, 1978). These dykes provided excellent competency contrast during deformation and large ore-filled dilation zones are observed at their contacts. It is possible that mineralized veins could have been emplaced during this time period. Ore formation has to be synchronous with the deformation of the pluton. Even if a late "reworking" of this early mineralization is possible, it is still necessary to add metal components because mineralization fluids derived from a felsic pluton would not produce zones with 1 % nickel that have been observed in certain mines. The nature and the origin of a pre-existing mineralization, if any, will remain conjectural. What is important is that a major shearing event produced the deposits in the DLC as they are now preserved.

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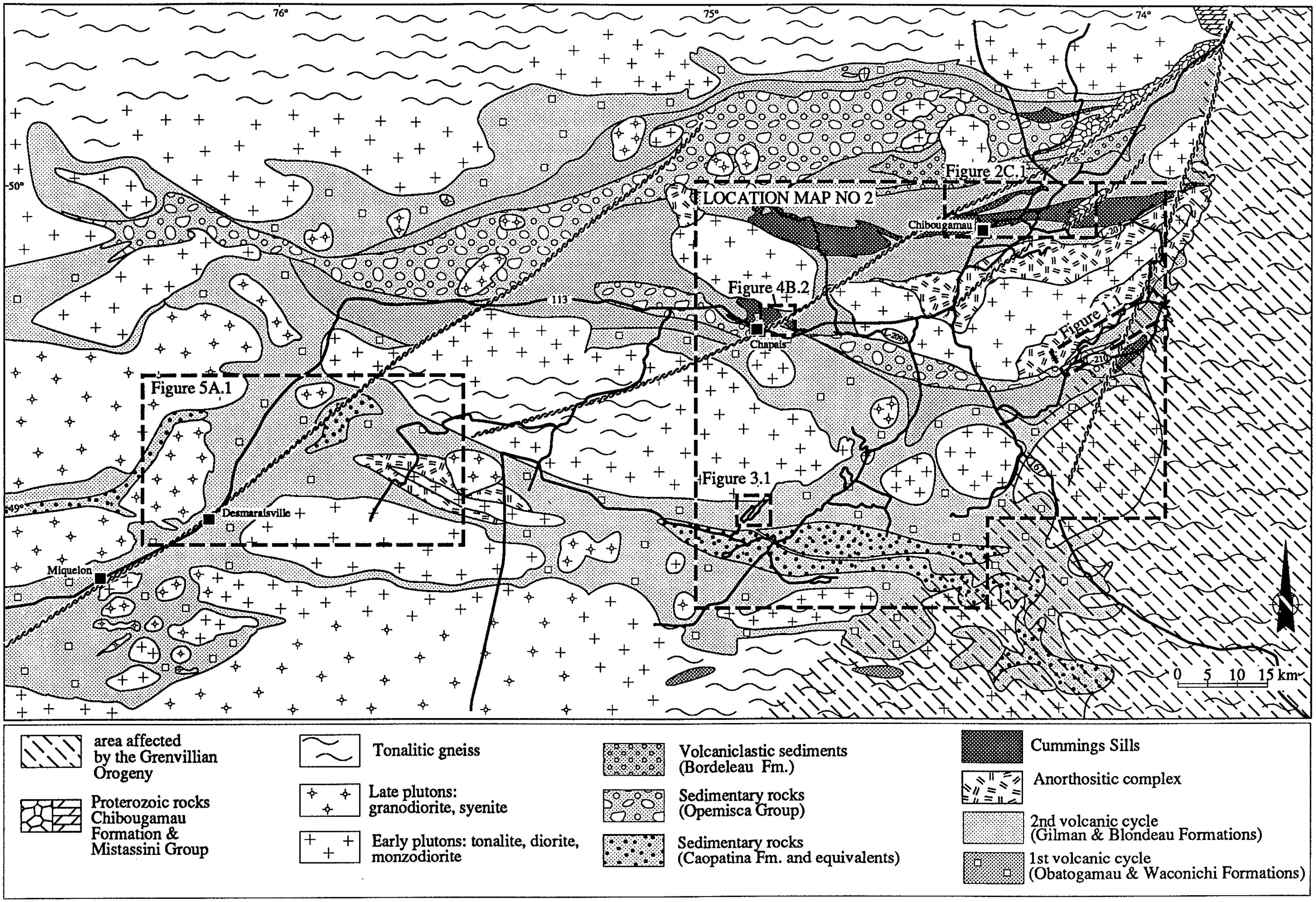
LOCATION MAP NO 1



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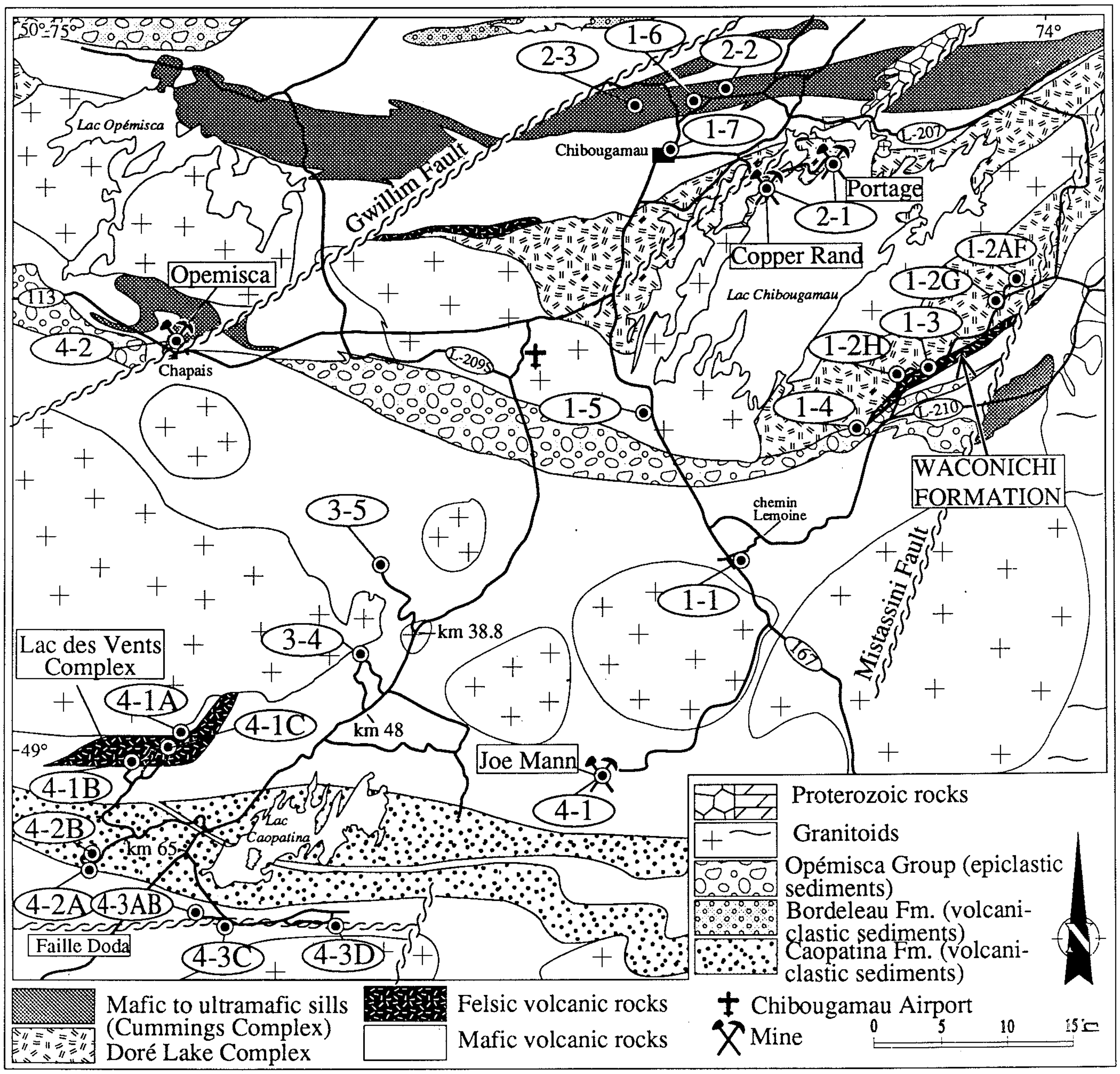

PART 2: STRATIGRAPHIC SETTING OF THE CHIBOUGAMAU RÉGION

Gilles O. Allard[1], Réal Daigneault[2], André Gobeil[3]

The first day outlines the main stratigraphic units of the Chibougamau segment. The visit will focus on the southern limb of the Chibougamau anticline, starting with the Obatogamau Formation at the base of the sequence. The excursion will then move to the Doré Lake Complex, where a section through the Layered Zone, Fe-Ti-V showings at its base and the transition to the granophyric zone will be examined. A visit to the rhyolitic rocks of the Waconichi Formation (representing the roof of the intrusion) will follow. The Waconichi Formation hosts a volcanogenic massive sulphide deposit (Lemoine Mine). The day will continue with a visit of the basal conglomerate of the Opemisca Group which uncomformably overlies the Roy Group. On the road back to Chibougamau, we will stop on the Queylus breccia showing, a mineralized zone with porphyry copper affinities. A good example of pillowed basalts of the Gilman Formation will be seen in Chibougamau. The day will end on Mount Bourbeau with the visit to the Aphebian Chibougamau Formation, a showing in the Bourbeau sill, and ultramafic rocks of the Roberge sill.

STOP 1.1 The Obatogamau Formation (adapted from Cimon in Allard et al. 1979)

The Obatogamau Formation, the oldest volcanic formation in the Chibougamau area, is the mafic portion of the first volcanic cycle of the Roy Group. It was first recognized on both flanks of the La Dauversière pluton. Its thickness varies between 2200 and 3000 metres. Plagioclase phenocrysts (now transformed into albite and clinozoïsite) in both the lavas and the comagmatic sills are characteristic of this formation. Basalt constitutes more than 80% of the Obatogamau Formation. Individual flows are only a few metres thick and are pillowed throughout. The phenocryst proportion peaks roughly halfway between the base and the top of the unit.

Location:

From Chibougamau, take highway 167 to the south and to kilometre 197.7 . Turn left (east) on the Campeau road. Stop 1.1A is located at this intersection on the western part of the road 167.

STOP 1.1 A

The first stop presents well developed pillow basalts with abundant plagioclase megacrysts.

STOP 1.1 B

For Stop 1.1B, continue east on the Campeau road for 500 metres. Outcrops are on the north side a few tens of metres in the bush. The outcrops at this stop are in the megacryst-rich portion of the formation. The crystals constitute roughly 10% of the rock and can be observed throughout the flows in the cores as well as in the rims of pillows. The megacrysts are idiomorphic or slightly rounded, chalky white, and in positive relief on

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the weathered surface. In narrow shear zones marked by an intense foliation, the feldspar megacryst have been greatly flattened and elongated. The ubiquitous nature of the plagioclase megacrysts make these basalts an important regional marker horizon.

STOP 1.1 C

For Stop 1.1C, proceed easterly 100 metres to an open area below a wooded hill. Cross the open area and follow a flagged trail 1 km to the outcrop. The outcrop consists of a sill that has been mapped for more than 2 kilometres along strike. In this locality, it is less than 100 metres thick. An outcrop of basalt flows lies less than 10 metres south of this outcrop. The stratigraphic top is to the north.

The sill is composed of distinct layers with by different proportions of plagioclase crystals. The lower two or three metres of the sill contain more than 80% cumulus plagioclase in a chloritic intercumulus matrix. The plagioclase crystals locally attain 10 centimetres in diameter and average approximately 2.5 centimetres. This layer is overlain by another layer composed of only 40% megacrysts. A poikilitic gabbro with a minor quartz content and less than 2% feldspar megacrysts occurs stratigraphically higher in the same sill.

STOP 1.2: The Doré Lake Complex

Two different sections, giving an overall view of the Doré Lake Complex, can be visited; the IGC section (Allard et al. 1984) and a section to the east designated the vanadium showing section (Figures 1.1 and 1.2). The latter is proposed for this excursion but the reader is referred to the description of the IGC section by Allard et al. (1984) to complete the description. The Layered Series of the Doré Lake Complex is certainly the most spectacular portion of this intrusion, and is one of the most interesting geological features of the Chibougamau region. The sequence is best developed on the southern limb of the Chibougamau anticline and is composed of a sequence of gabbroic, anorthositic and magnetitic rocks of a little less than 1000 m in thickness. It overlies the anorthositic zone composed of 3000m of anorthosite and coarse-grained gabbro. The Layered Series is overlain by 100m of granophyre followed by a thin gabbroic horizon constituting the Border Zone.

The Vanadium Showing Section, Layered Series of the Doré Lake Complex.

Several magmatic phenomena can be seen in the layered magnetite-bearing gabbros of this section. The vanadium showing (Fig. 1.2) and the transition between the Layered Series and the Granophyre Zone will be visited. However a continuous section through the Layered Series cannot be seen here, in contrast to the IGC section.

Location

To go to the vanadium showing section, take the Lemoine road (at km 200 from road 167) go east towards the L210 logging road, follow it up to a triangular intersection at km 21.3 and turn left (west) (Fig. 1.1). Go for 1.3 km and turn left (southwest). Follow this road for 4.8 km to the next three-way intersection. Go north on the road that passes just west of Laugon Lake and through a gravel pit (Fig. 1.2). From there go 1.3 km and park the vehicles before going down a steep slope. A 1.2 km walk to the end of the road leads to the first stop. The others are located on the road going back.

STOP 1.2A

Stop 1.2A is the main showing that was used to estimate the vanadium grades within the P1 zone (70 000 tonnes at 30% Fe, 0.5% V_2O_5, and 10% TiO_2. This outcrop is part of the P1 zone, at the base of the Layered Series. Farther south, outcrops of coarse-grained gabbro of the Anorthositic Zone can be seen. The well developed magmatic layering that characterizes the P1 zone is produced by alternating decimetric to centimetric gabbro horizons and magnetitite beds. A magmatic foliation results from the preferred orientation of plagioclase laths. Bedding is approximately N070° with a sub-vertical dip. The preferred orientation of plagioclase laths parallel to bedding is slightly enhanced by the regional deformation. Several magmatic phenomena related to convective currents can be observed, such as truncated beds and slump structures. Compaction induced structures such as flame structures can also be seen. Near the main showing a large gabbroic block of more than 3m in diameter is embedded in layers of the P1 zone. A number of injections of pegmatitic leucogabbro

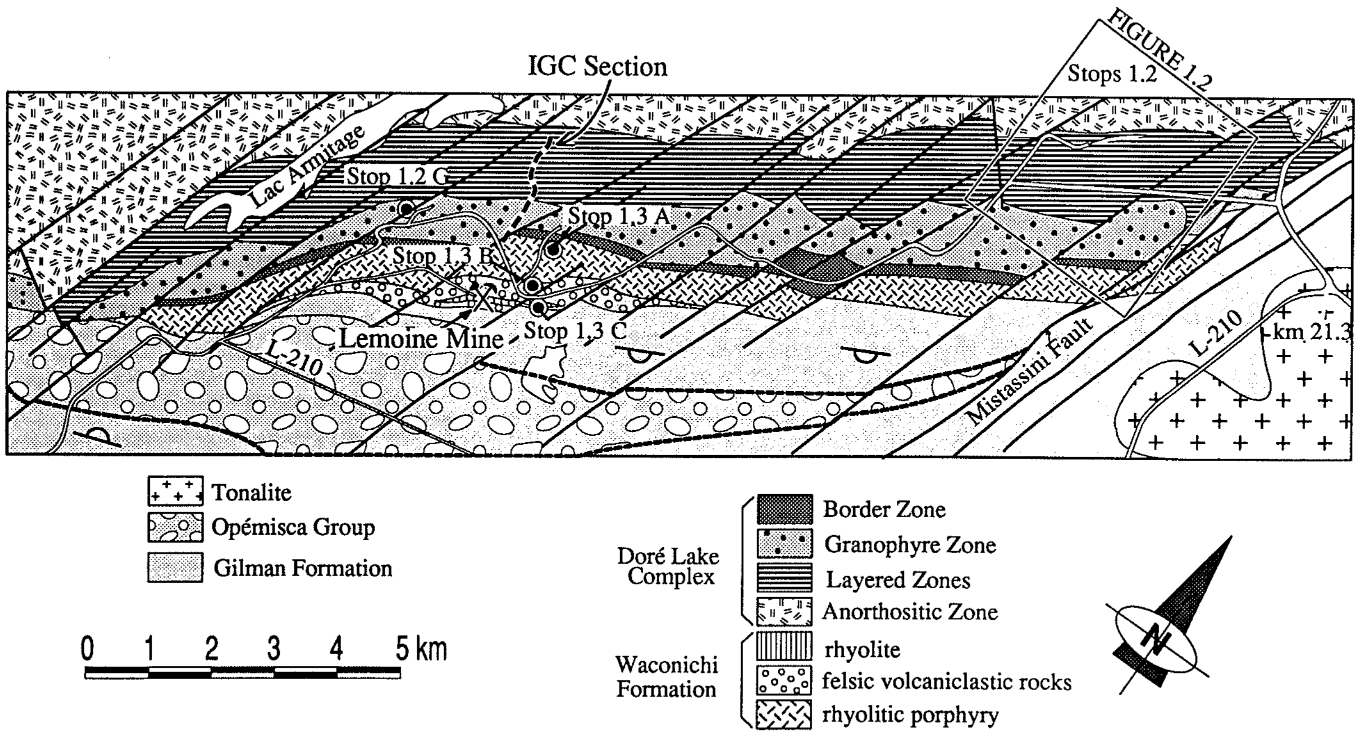

Figure 1.1 - Schematic geological map of the south limb of the Chibougamau anticline focussing on the position of the Layered Series of the Doré Lake Complex and on the Waconichi Formation. Note the location of Figure 1.2.

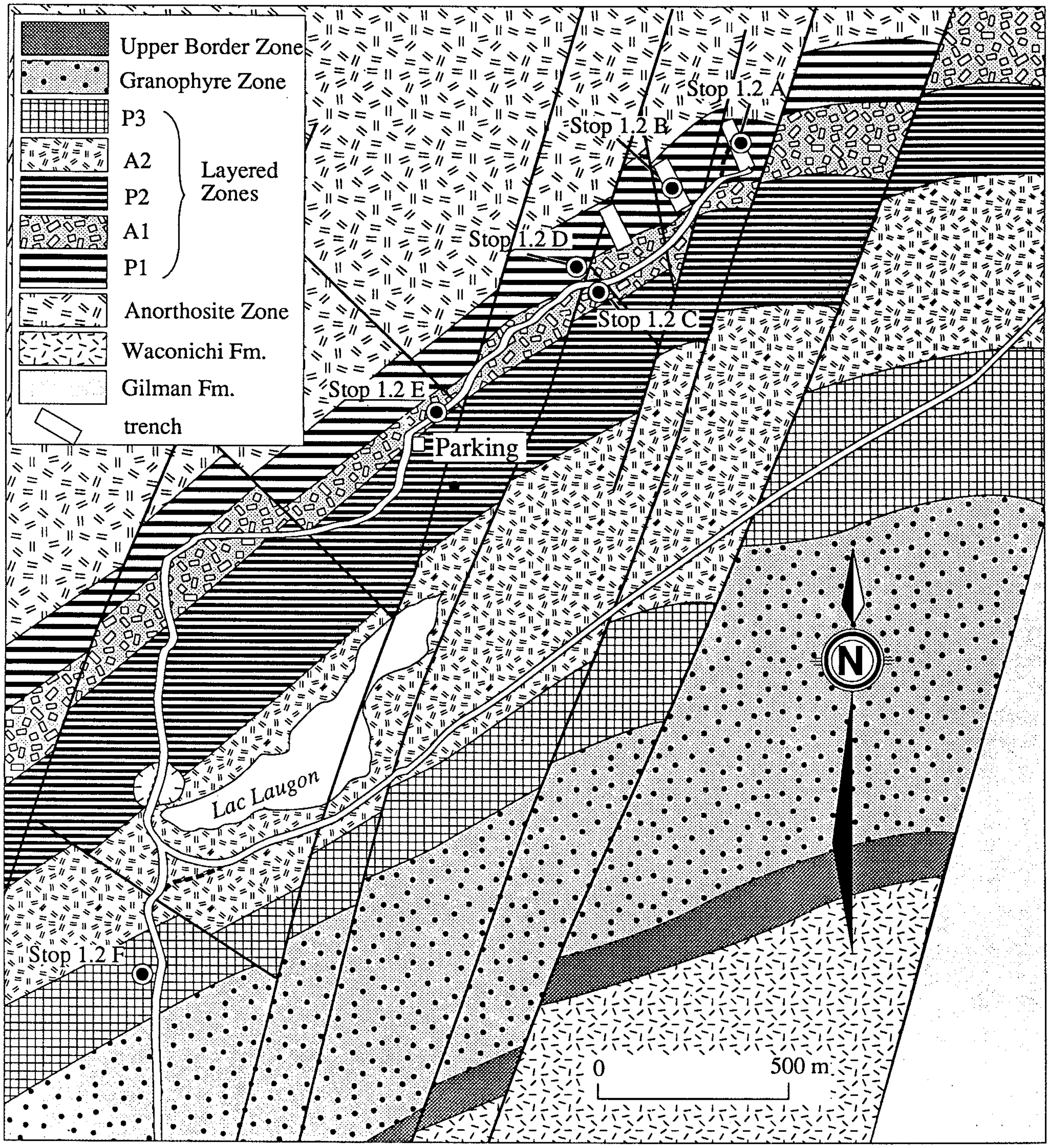

Figure 1.2 - Subdivisions of the Layered Series at the vanadium showing section.

disrupt the layering at several localities, probably increasing the total thickness of the Layered Series.

Stop 1.2B

Stop 1.2B is located 200m away, on the road back to the vehicles (Fig. 1.2). A magnificent fold can be seen within the P1 zone. The Z shaped fold was probably produced during the regional deformation episode by tightening of an originally undulating structure formed by a magmatic process. The numerous examples of slump structures seen on the previous outcrops testify to strata instability. In the hinge zone, the plagioclase crystals are aligned parallel to the axial plane of the fold. Note the necking of strata at the core of the fold.

STOP 1.2C

Stop 1.1C is on a small mound, immediately south of the road, 200m from the last outcrop. Stop 1.2C is located within the A1 zone that separates the P1 and P2 zones. The outcrop is of a gabbro with large plagioclase crystals. A cumulus-intercumulus texture is well developed, and the original intercumulus pyroxene is replaced by chlorite and actinolite. A poikilitic texture is locally present.

STOP 1.2D

Stop 1.2D is located approximately 100 m from stop 1.2C, near a small stream oriented N015°. The contact between rocks of the P1 zone with those of the A1 zone can be seen north of the road Near the stream, beds are progressively dragged towards a NNE direction. A N015° schistosity crosscuts the magmatic layering, suggesting the presence of a Grenvillian fault in the stream (Fig. 1.3). An apparent sinistral movement, consistent with the known movement along these faults, is deduced from the dragging of beds. The map in Fig. 1.1 illustrates the abundance of such faults in this area. Certain zones of this outcrop contain high concentrations of massive magnetite.

STOP 1.2E

Stop 1.2E is located on the slope leading to the vehicles. This outcrop is part of the A1 zone. It is a cumulate made up of plagioclase crystals up to 15 cm in diameter. The bedded rocks at the top of the slope are part of the P2 zone, which is similar to the P1 zone, but has a lower vanadium grades and a higher TiO_2 grade.

STOP 1.2F

Take the vehicles to stop 1.2F. It is located at the last three-way intersection, past Laugon Lake (Fig. 1.2), less than 100 m away from the road on the west side. The large outcrops on the slope belong to Zone P3, at the top of the Layered Series. Zone P3 is much thicker than zones P1 and P2 (180 m to 360 m). It is composed of dark green ferropyroxenite interlayered with thin ferrogabbro horizons. Bedding is well defined. The rock is mostly composed of ferrochlorite, ferrohastingsite and ferroactinolite replacing the original pyroxene, with minor quantities of ilmenite and epidote. There is no magnetite in this zone even though the rocks contain over 30% total iron. Iron is contained in the ferromagnesian minerals. One hundred metres to the south, the transition to the Granophyre Zone can be seen. Granophyre layers are intercalated with quartz-rich ferropyroxenites. The apatite-rich member is not seen in this section.

STOP 1.2G

Stop 1.2G will be visited after the outcrops of the Waconichi Formation (Fig. 1.1). From the triangular intersection south of Laugon Lake, go south for approximately 9 km. From there, a gravel pit will serve as a reference for the next stops. To visit stop 1.2G, go 3.2 km from the gravel pit and take the next road going north for 150 m, to the intersection. From there, go west on foot for less than 100 m. The outcrop zone is on the road. At this stop, which is the end of the Doré Lake Complex traverse, the Granophyre zone can be seen. Acid and basic (pyroxene-rich) granophyric horizons alternate to produce good layering. At this stop, one can see fine-grained gabbro blocks of the Border Zone floating in the Granophyre Zone.

STOP 1.3 The Waconichi Formation

The Lemoine-type Waconichi Formation is composed of rhyolite and porphyritic rhyodacite, felsic pyroclastite, and gabbro sills. In 1973, the discovery of the Lemoine Mine on the southern limb of the

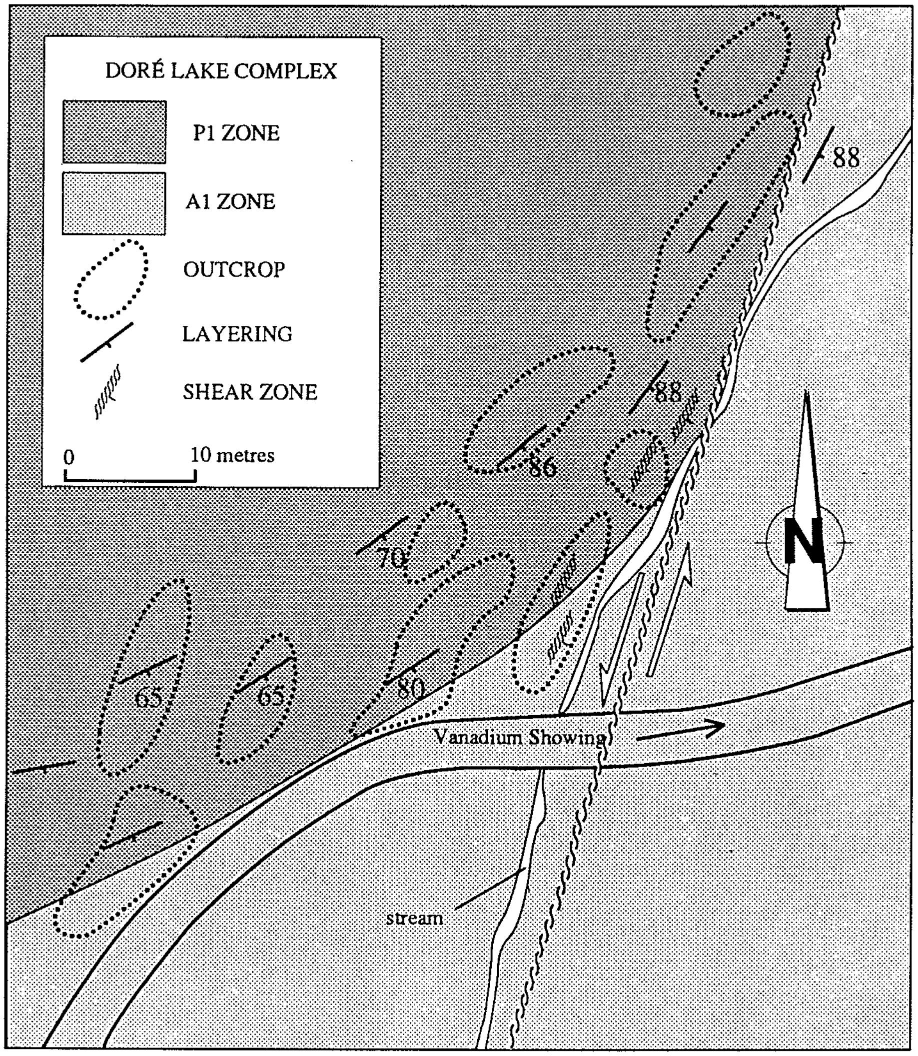

Figure 1.3 - Drag fold along a Grenvillian NNE fault affecting the layering of the Doré Lake Complex.

Chibougamau anticline (728 000 tonnes, 4.2% Cu, 9.6% Zn, 4.6 g/t Au and 83.8 g/t Ag), within the Waconichi Formation underlined the strategic importance of this formation. The later discovery (1975) of the Scott showing by SELCO on the northern limb of the anticline in Scott Township indicated the importance of this volcanic horizon as a metallotect in the Chibougamau region. In the Lemoine sector, the Waconichi Formation is approximately 1000 m thick. These rocks constitute the roof of the Doré Lake Complex and are overlain by mafic volcanic rocks of the Gilman Formation. The three stops of this part of the excursion represent a typical section through the Waconichi Formation (Fig. 1.1).

Location:

In the continuity of the field trip, start from the gravel pit and go 500 m past the lake on the northern side of the road. Turn right (north) at the intersection. The outcrops may also be reached directly from the Lemoine road. From the intersection of the Lemoine road and L-210, proceed for 1.8 km on the Lemoine road, then take a secondary road to the left (north) for 4.2 km.

STOP 1.3A

The first outcrop is 850 m farther to the right (east) away from the road. It is stratigraphically at the base of the Waconichi Formation and the contact with the Doré Lake Complex being less than 300 m. away to the north. More than 70% of this unit is composed of rhyolitic porphyry. It typically contains bluish quartz phenocrysts, consituting more than 20% of the rock. The plagioclase phenocrysts are milky white and are altered to white mica. Numerous enclaves of chlorite-rich rocks characterize the northern part of the outcrop. The angularity of the fragments suggests that they have been locally derived. The presence of mafic enclaves has been noted at several localities at the base of this rhyolitic porphyry unit.

STOP 1.3B

The second stop is 200 m away from the intersection, on the lakeside. The outcrop is composed of felsic pyroclastites of different sized fragments. Bedding is well defined in the quartz bearing crystal tuff. In the southern part of the outcrop, a white felsic tuff horizon outlines a metre-scale fold closure. Coarse volcaniclastics are found within the fold, angular fragments similar to the felsic tuff can be recognized. Bedded exhalite fragments are also present in this unit. Close examination of the fold and of the axial plane schistosity indicates that the regional schistosity crosscuts the northern limb of the fold in two places. This suggests that the fold could have been produced by slumping and later refolding during the N-S horizontal shortening, rather than being purely tectonic in origin. Irregular laminations (convolute bedding) also suggests the presence of slumping. All these features suggest that the coarse volcaniclastics were produced by a debris flow on the unstable slopes of a volcanic edifice.

STOP 1.3C

The third outcrop (Stop 1.3C) is located south of the main road near the gravel pit, in front of the lake. To get there return to the intersection and go 500 m beyond. The rocks at this outcrop are rhyolitic in composition and exibit closed, irregular-shaped structures, interpreted as lava lobes. They are ellipsoidal and up to 10 m in diameter. Other smaller ones are surrounded by a matrix which is probably recrystallized hyaloclastite. The borders of the lobes are highly vesicular. This unit is less than 30 m thick and is of limited lateral extent.

From this outcrop take the SE path for 100 m. Southward-facing pillowed basalts of the Obatogamau Formation constitute the next outcrop.

STOP 1.4 Basal conglomerate of the Stella Formation, Opemisca Group (adapted from Cimon _in_ Allard et al., 1979)

This stop in the basal conglomerate can be considered as one of the most important of the area since it has contributed much to the understanding of the regional stratigraphy and relative age of many formations. This outcrop shows an angular unconformity between the Roy Group and the Opemisca Group.

Location:

From the intersection of the Lemoine road and the L-210 road, proceed to the south for 900 m. Turn right (north) on a secondary road and continue 200 m up to a junction with another road, continue on this road for another 200 m. From there, follow the well marked trail for 1 km, keep left (Fig. 1.4).

The outcrop of the basal conglomerate lies on top of the Granophyre Zone of the Doré Lake Complex which outcrops just a few metres north of the conglomerate outcrop. The conglomerate consist of well rounded boulders varying from a few centimetres to more than one metre in diameter. The most common clasts are granophyre, quartz bearing gabbro, anorthosite, fine grained mafic rocks, both bedded and porphyritic felsic rocks, and quartz pebbles set in a coarse matrix of quartz, feldspar and rock fragments surrounded by sericite and chlorite. Negative relief of the rims of mafic clasts is due to the differential erosion of the softer altered outer portion. This is the result of weathering of the surficial rind of the clast before sedimentation. A stratigraphic hole drilled by the Ministère de l'Énergie et des Ressources du Québec under this outcrop cuts 70 metres of conglomerate (Cimon & Gobeil, 1978).

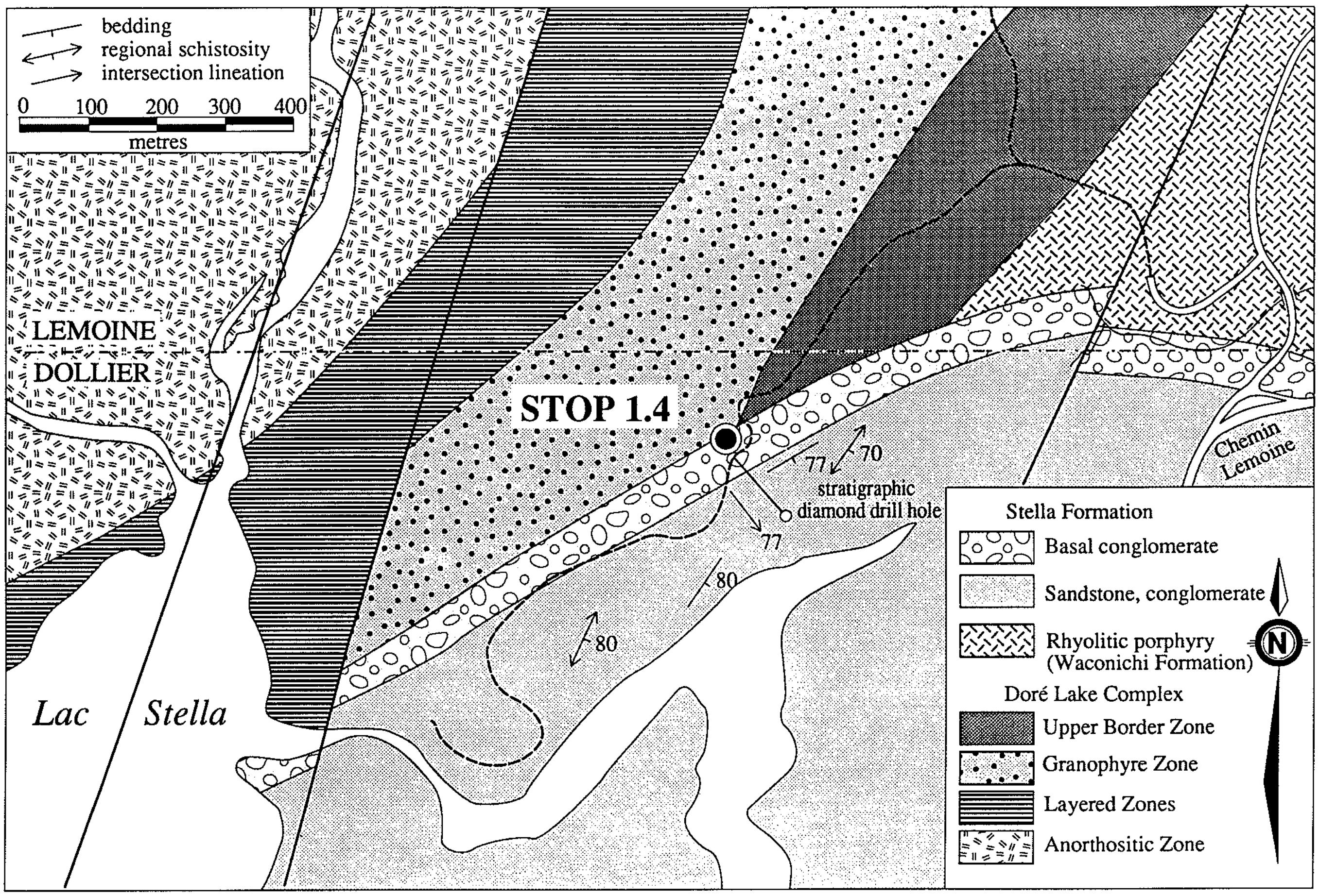

Figure 1.4 - General geology of STOP 1.4, showing an unconformity between the main units of the Roy Group and the basal conglomerate of the Stella Formation (modified from Cimon, 1977).

The upper part of the Stella Formation consists of feldspathic sandstone, with locally argillaceous and conglomeratic lenses. The Stella Formation has been folded into a tight syncline whose axial trace lies roughly 1.5 km to the south. In the vicinity of this outcrop, the beds of the Stella Formation have a northeast strike at a 20° angle to the rocks of the Doré Lake Complex. A good bedding-schistosity relationship can be seen in the well-bedded sandstone in the southwest part of the outcrop, suggesting the closure of a synclinal fold.

Stop 1.5 The Queylus breccias (adapted from Cimon in Allard et al. 1979)

The Queylus breccias bear striking similarities to porphyry copper type mineralization. A brecciated and highly-fractured zone affected by hydrothermal alteration and weakly mineralized with copper is located near the margin of the Chibougamau Pluton.

<u>Location</u>:

Take an old logging road to the west of the main highway at kilometre 208 of highway 167 (see Fig. 1.5).

The outcrop is in one of the tonalite units of the pluton. Hydrothermal alteration has almost completely

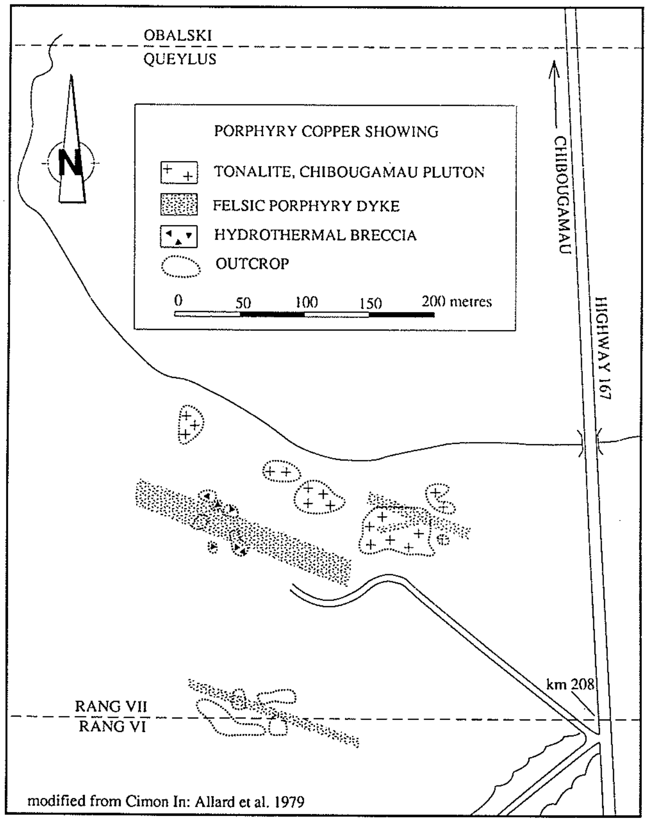

Figure 1.5 - Detailed geology of the "Porphyry Copper Showing" (modified from Cimon, In Allard et al., 1979).

obliterated all primary ferromagnesian minerals but the sub-porphyritic texture of the quartz is still evident. The alteration occurs over several square kilometres and affects both the tonalite and the crosscutting porphyritic dykes. The alteration is characterized by a red colour, and by the addition of potassium (increasing from 1.5% to 5%). The mineralogical changes include the appearance of microcline and sericite at the expense of plagioclase, epidote and biotite along fractures, rosettes of tourmaline and a pervasive hematitic impregnation. Carbonatization, which is locally very important, affects the primary ferromagnesian minerals.

Two types of breccia occur in the zone. The first is composed approximately angular pebble-sized fragments in a very fine-grained pulverized matrix. This breccia crops out over several hundred square metres, but its extent and form are unknown. Various types of fragment include tonalite identical to the country rocks, quartz feldspar porphyry identical to the adjacent dykes, magnetite-rich ultramafic rocks, and fragments of milky quartz. The matrix locally displays fine laminations.

The second type of breccia is composed solely of wallrock fragments cemented by tourmaline, quartz and pulverized rock. This type of breccia is located along the borders of the first type where it forms lenses several metres thick.

Chalcopyrite-pyrite mineralization in the form of veinlets occurs in a network of fractures associated with the formation of tourmaline, quartz and carbonate. Molybdenite is present locally in fractures, and finely disseminated in quartz veins. Magnetite is very abundant and is locally replaced by hematite.

Fluid pressure seems to have been the principal energy source for the formation of the breccias. The hydrothermal activity in this sector is a late event in the history of the pluton.

STOP 1.6 Radar dome showing (adapted from Gobeil in Allard et al., 1979 and from Pilote et al., 1984)

A series of different mineralization types their diverse relations with the host rock will be seen on this outcrop. All the mineralization occurs either within felsic to intermediate tuffs of the Blondeau Formation or in the ultramafic portion of the Bourbeau sill (Fig. 1.6). This stop also provides the opportunity to observe the Aphebian Chibougamau Formation interpreted as a fluvio-glacial deposit (Long, 1974).

Location

Take the gravel road north of the cemetery in the direction of Mount Bourbeau. At 0.3 km from the cemetery, keep right (east) at the intersection of the golf club road. Continue about 2.6 km to the radar dome road toward Mount Bourbeau. Take this road toward the summit up to an intersection.

STOP 1.6A

The first outcrop, to the southeast, is a conglomerate of the Chibougamau Formation. Rocks of this unit are undeformed and constitute an outlier overlying the Archean basement. The formation is made up of paraconglomerate with a variety of fragments, sandstone and rhythmites with dropstones. Further north down the road, cherty tuff (more or less silicified) of the Blondeau Formation crops out. Topographic relief on the pre-Proterozoic erosion surface is at least 2 m

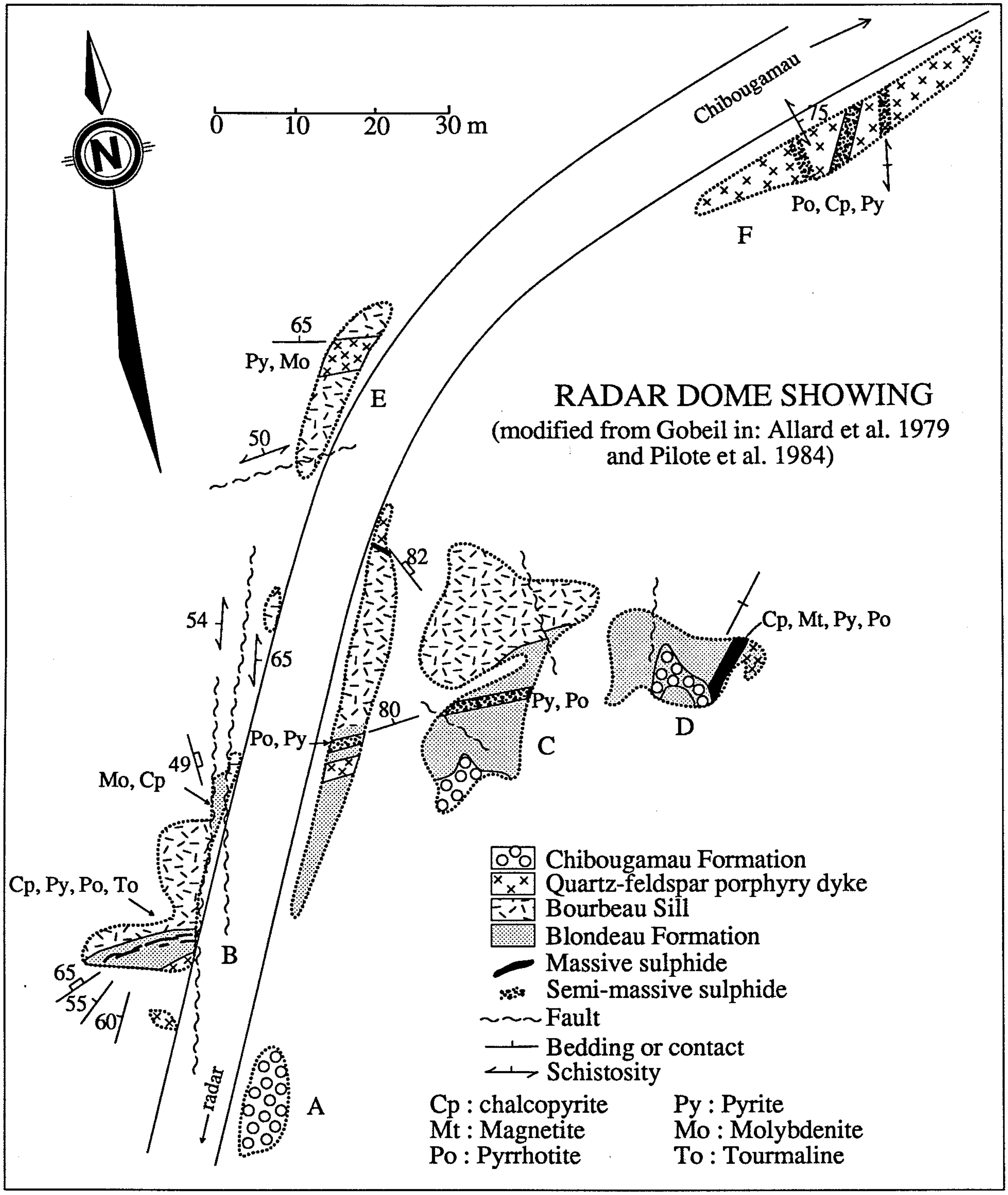

Figure 1.6 - Detailed geology of the "Radar Dome Showing". Letters A to F are referred to in the text (modified from Gobeil, In Allard et al. 1979).

here, although Smith (1960) records up to 9 m in this general vicinity.

STOP 1.6B

A massive sulphide vein (chalcopyrite, pyrite, pyrrhotite, with local tourmaline) is located just at the contact between the ultramafic portion of the Bourbeau Sill and the felsic tuffs. This vein is 40 to 60 cm wide, and branches into two segments east of this outcrop. To the north, the black pyroxenite is intruded by a quartz-feldspar porphyry dyke that contains traces of pyrite. NNE shear zones are common, and shear planes are locally coated with siderite. One of these shear zones is mineralized with molybdenite and chalcopyrite, with a quartz-carbonate gangue where it cuts the volcanic rocks.

STOP 1.6C

On the east side of the road pyrite, pyrrhotite and chalcopyrite mineralization may be seen becoming progressively more carbonatized and chloritized to the east before being truncated at an erosion surface. A fine-grained black sediment or possibly mafic lava containing chert fills fractures in the beds. A gabbroic zone at the base of the Bourbeau Sill is found 1 to 2 m north of the mineralization.

STOP 1.6 D

East of this zone a lens of chalcopyrite, magnetite, pyrite and pyrrhotite occurs, and is roughly concordant with the bedding but locally cross-cutting. The host sediment on the west side is highly chloritized. On the east side, a quartz-feldspar porphyry dyke containing chert fragments borders the lens. Note that the Aphebian conglomerate unconformably overlies both the volcanics and the sulphide lens, clearly indicating the mineralization to be Archean in age.

Stop 1.6E

Return to the west side of the road, and observe a quartz-feldspar- porphyry dyke identical to the previous stop, hosting molybdenite-pyrite mineralization.

Stop 1.6 F

Another similar porphyry dyke to the northeast contains several discontinuous veins mineralized with pyrrhotite, chalcopyrite and pyrite.

STOP 1.7 The Gilman Formation in Leblanc park (adapted from Gobeil in Allard et al., 1979)

In the town of Chibougamau, take 4th Street north, turn left on Normand Street, then right on Gendron Street, right again on McLean street and then left on the Leblanc Street to the park.

This series of outcrops sequentially numbered in figure 1.7 displays the typical rocks of the Gilman Formation as well as preserved volcanic structures (Fig. 1.7). The Gilman Formation is composed of pillowed basalt and andesite and gabbro sills. As well, several lenses of volcaniclastic (tuff and breccia) occur locally but mainly in the upper part.

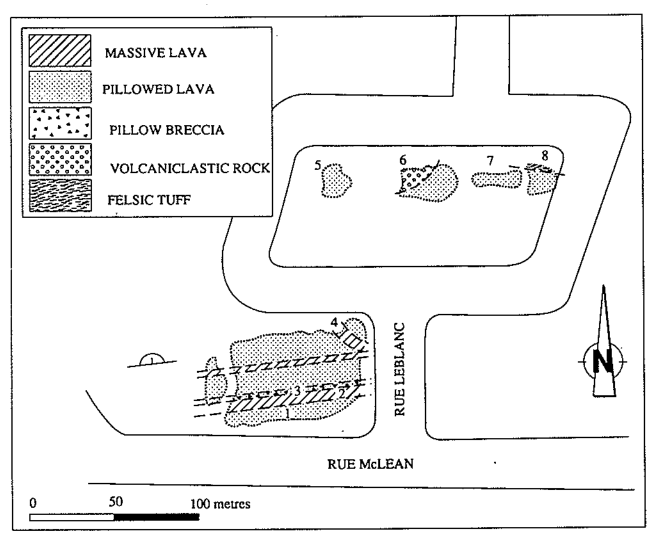

Figure 1.7 - Detailed geology of the Gilman Formation in Parc Leblanc. The numbers are referred to in the text (modified from Gobeil, In Allard et al., 1979).

1- The south part of the outcrop consist of one flow of pillowed lava. Some pillows are more than 2 m long, but the majority have an average diameter of 4 m. Amygdules are either concentrated around the chilled rim or evenly distributed throughout the pillows. Note that many pillows are attached together, and that smaller pillows or hyaloclastite fill the interstices at the junction of several pillows. Some radial cooling fractures are also seen.

2- Proceed north to a well-exposed massive lava flow about 6 m thick. This is a pale grey aphanitic andesite containing amygdules and small plagioclase phenocrysts. Pale andesites are characteristic of the upper part of the Gilman Formation.

3- At the top of this flow, a 2 m-thick bed of pillow breccia, containing fragments several centimetres in diameter occurs in a fine-grained hyaloclastic matrix.

4- The highest part of the outcrop is composed of pillow lavas. At the northeast extremity of the outcrop, a massive lava flow seems to be emplaced atan angle with the general trend of the pillow lava. The two contacts with pillow lavas appear gradual, with many pillows originating in the massive lava. Is it a massive flow on a slope gradually filling a depression with pillows forming on the front and flanks?

5- On this outcrop several pillows show well-developed concentric chill structures. Considerable pillow breccia is present.

6- A phenomenon similar to outcrop 4 occurred here. An irregular bed of coarse porphyritic volcanic pebbles fills a depression. The pebbles are moderately well-rounded, suggesting some transport has occurred.

7, 8- Enormous pillows with concentric chill fractures occurred. Intense silicification is observed principally at the junction of several pillows. Cherty tuff beds occur at the northern edge of the outcrop.

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COPPER RAND MINE
by
(Alain Blais)

HISTORY

The mining and exploration history of the Chibougamau area date back to I903 when copper ore was discoverd at Copper Point on Portage Island by Peter McKenzie. In I9I0, Captain H.C. Machin found a highly sheared outcrop hosting very good copper-gold mineralization near the shoreline of Dore Lake. This discovery led to a series of events which ended with the development of the Copper Rand and Portage mines 50 years later by Patino of Canada. Both mines were operated by Patino until I98I when Northgate Exploration acquired the North American assets of the Patino Group, including their operations in the Chibougamau area. In December I987, Northgate's operations in Chibougamau were sold to Western Mining. At the same time, a series of acquisitions throughout the country by Western Mining brought about the formation of Westminer Canada Ltd. of which Les Mines Chibougamau (Copper Rand and Portage mines) is a division.

The Copper Rand and Portage mines are located in the Dore Lake intrusive sill complex of which the main component is a highly altered and tectonized anorthosite. The ore is hosted by a series of en echelon lenses within a major shear zone which strikes north to west (Copper Rand Mine). Most of the more important sulphide mineralization is associated with W-to-NW-striking shear structures which are cut by the Lake Dore Fault (a regional fault trending NE and dipping 60 to 65 degrees towards the NW). To date, fourteen separate mines have been exploited at one time or another adjacent to the Dore Lake Fault. Although this fault's role in mineralization is not fully understood, it is certainly a major structural element.

MINE GEOLOGY

The ore deposits of the Copper Rand Mine lie east of the Dore Lake Fault within a shear zone trending N50W to N70W and dipping steeply southward at 60 to 75 degrees. The shear zone is 350 to 425 meters wide and extends for I.5 km. The rocks of the shear zone were originally part of the anorthosite transition zone and gabbro sequence (upper part of the Anorthosite Zone Fig. 2A.1). The alteration, consisting of sericite, carbonate, chlorite, and chloritoid, completely obliterates the original rock texture with a superimposed foliation. Mineralization is known to occur along the entire length of the shear zone, with the important mineralized lenses concentrated towards the footwall and hangingwall where dykes have been emplaced. Major sulphide minerals are pyrite, chalcopyrite, pyrrhotite, sphalerite and galena. Dykes are a major feature of the Copper Rand shear zone. They range from a few centimetres to over 20 metres in width and can be continuous up to 400 metres long. They are sinuous and branching with dips approximately paralleling the dip of the shear zone. They are good marker horizons for exploration purposes, considering that ore lenses are most often located at or near their contacts.

Principal rock types

The rocks present in the Copper Rand mine are transitional between meta-anorthosite to the south and metagabbro to the north. Megascopically, the meta-anorthosite is typically coarse-grained containing more than 80% altered light grey to white plagioclase phenocrysts. The interstitial areas are pale yellowish-green to dark green. Microscopically, the rock is composed of subhedral twinned andesine altered locally to zoisite, clinozoisite, sericite and oligoclase. The matrix consists of sericite, chlorite, oligoclase, epidote, clinozoisite and minor quartz and carbonate. Magnetite, sphene and leucoxene are present as accessories. The transition zone, which forms a complete gradation from meta-anorthosite to metagabbro, is defined by a gradual decrease in the amount of plagioclase (decrease in % of $Al2O3$) and an increase in the mafic minerals. Within the transition zone, there are zones rich in magnetite as well as a few lenses of coarse amphibolite. The metagabbro is coarse-grained, consisting essentially of equal amounts of altered plagioclase, amphibole and pyroxene. The alteration products consist of saussuritized andesine, actinolitic hornblende, minor pyroxene, magnetite, titaniferous magnetite, chlorite and apatite. The rocks within the shear zone have all been

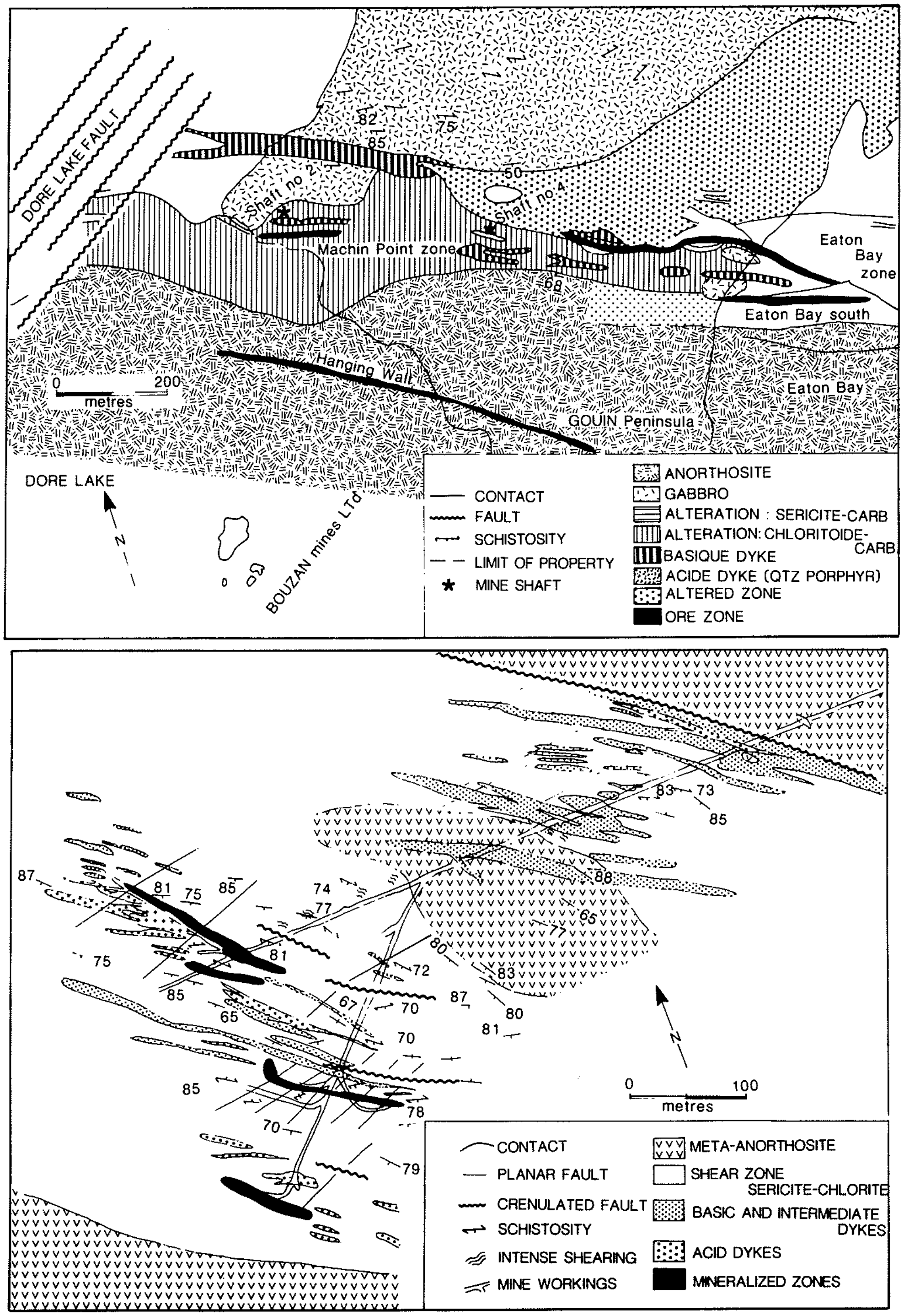

Figure 2A.1 Generalized level plans showing relative positions of the Copper Rand and Dore Lake Shear zones (top) and structural details of the Copper Rand shear zone (bottom). From Archambault et al., 1984.

converted to their highly altered, schistose equivalents. The alteration minerals, in most cases, have completely masked the original composition and texture of the anorthosite-gabbro transition sequence. In some isolated localities within the shear zone, the original porphyritic texture is preserved. This relict porphyritic texture consists of white to light green remnant or "ghost" phenocrysts in a dark green chlorite matrix. In areas of intense shearing, particularly in the hangingwall and footwall areas, the rocks are highly schistose with well developed "banding" and "flow structure". Sericite, carbonate, chlorite, quartz and chloritoid predominate. They occur in various combinations and their relative amounts, particularly of chlorite, determine the colour of the rock.

Dykes

The dykes, a major feature of the Copper Rand shear zone, occur in three distinct types based mainly on megascopic characteristics, as follows: (l) basic dykes (2) intermediate or grey dykes (the altered equivalent of the basic dykes), (3) acid dykes (quartz porphyry, quartz-feldspar porphyry). The dyke rocks consist almost entirely of alteration minerals and the names applied to them may be somewhat misleading. In fact, basic and intermediate or grey dykes are essentially of dioritic composition, and the intermediate or grey dykes exhibit an intense carbonatization. Since cross-cutting relationships are absent, the relative ages of the dykes are unknown. However, the basic dykes are assumed to be the oldest.

Basic dykes

The basic dykes are named for their dark green to black colour rather than their mineral composition. They are commonly dense, fine-grained, massive and structureless, but are locally weakly to moderately schistose. Microscopically, the "freshest" sample is composed mainly of chlorite, calcite, sericite and quartz. Accessory minerals include magnetite, ilmenite and leucoxene with minor apatite and pyrite. The presence of over 50% chlorite accounts for the dark colour. The rock is fairly homogeneous with many granular quartz and chlorite filled microfractures showing minor displacements.

Altered basic dykes (intermediate/grey dykes)

This type of dyke is a highly carbonatized version of the basic dyke, varying widely in colour light grey to brownish grey and in texture. The dyke is altered to siderite, with varying amounts of sericite, calcite, chloritoid, and quartz. Accessory minerals include magnetite, ilmenite and pyrite. Calcite stringers cut the schistosity, banding, and the chloritoid.

Quartz porphyry dykes

Megascopically, the rock is yellowish green to whitish grey with about l0% spherical to oval quartz eyes varying in size from less than 1 mm to over 50 mm and may be aggregates or single grains. Calcite and sericite, occurring either as fine bands or forming a mosaic with fine grained quartz, form the matrix. Fresh quartz porphyry also contains fine-to medium-grained chloritoid and chlorite. The chlorite is banded with the sericite giving the rock a distinct pale green colour. Minor disseminated and finely banded leucoxene occurs throughout.

Quartz feldspar porphyry dykes

Megascopically, this olive-greenish-grey rock is fine-to medium-grained, with minute to 50 mm, angular to oval phenocrysts. Some of the phenocrysts are shades of green and brown whereas others are light grey to white. The rock is relatively unsheared. Microscopically, the green and brown coloured phenocrysts are composed of a very fine to medium grained mosaic of carbonate and chlorite with minor calcite. Similarly, the matrix is composed mainly of granular calcite and chlorite.

Structure

The northwesterly-trending Copper Rand shear zone is the dominant structural feature in the mine area. It contains two major and two minor fault sets. The major sets strike northwest and northeast, while the minor sets strike N-S and E-W. All the faults appear to be younger than the shear zone, but faults, both older and younger than the chalcopyrite mineralization have been found. There are numerous northwesterly-trending faults throughout the mine. They dip for the most part steeply south, and generally transect the schistosity at small angles. Displacement along the northwesterly faults, while rarely observed, is generally not more than a few centimetres to several metres. The less abundant northeasterly- trending faults are also widely distributed. This set generally dips moderately to steeply to the northwest. They are locally found displacing the northwesterly set. The N-S fault set dips steeply west while the E-W set dips steeply to S.

Mineralization

While mineralization is known to occur along a length of 1200 m, it is found in abundance in four main areas, the Eaton Bay zone, the Machin Point zone, the Kerr-Addison zone and Hangingwall zone. In many areas, the dyke appears to have exerted a damming effect on the ore-forming fluids. Slight changes in the strike and dip of the major dykes also appear to have influenced the localization of the ore. The sulphide minerals in order of abundance are: chalcopyrite, pyrite, with minor pyrrhotite and sphalerite, and accessory galena. Magnetite occurs in lenses and veins, particularly in the Eaton Bay area, and in stringers and disseminated blebs throughout the remainder of the shear zone. Fine-grained specular hematite is locally present.

<u>Machin Point Zone</u>

The Machin Point ore zone dips approximately parallel to the shear zone at 87 degrees South, and rakes approximately 75 degrees West. Chalcopyrite, forms small blebs, pockets and lenses distributed throughout the medium-to coarse-grained siderite. This mineralization forms zones that reach a length of over 85m, but usually not more than 1 to 2 metres wide. Sericite, calcite, quartz and chloritoid occur throughout the zone. However, chlorite and chloritoid are the alteration minerals not closely associated with the ore.

<u>Eaton Bay Zone</u>

This ore zone also dips parallel to the shear zone. The dip is 75 degrees south at section 75E and changes to 65 south at section 85E. The rake is about 55 degrees west. Economically, this is the second most important of the four main zones. The ore minerals consist of chalcopyrite, with minor pyrite and magnetite and accessory sphalerite and pyrrhotite. The zone lies between a basic dyke to the north and a quartz-feldspar porphyry dyke to the south. The chalcopyrite occurs as stringers and sinuous lenses. Dark green chlorite is the main alteration mineral.

<u>Hangingwall Zone</u>

This zone parallels the south contact of the shear zone. The zone dips 55 degrees south and the rake averages 55-65 degrees west. Economically this is the most important of the four main zones. Chalcopyrite and pyrite are the most abundant sulphides. This zone is distinguished by its relatively larger content of pyrrhotite and sphalerite. Galena occurs in accessory amounts. The zone is also distinguished by its higher average gold grade and by the size of some of the ore lenses.

<u>Kerr Addison</u>

This zone is quite similar to, but smaller than the Hangingwall zone. It is located further east and has much the same attitude.

Type of veins mineralization

a) Sulphides (py-cpy) in stringers and/or semi-massive and/or patches with chlorite-quartz-occasionally minor pyrrhotite.
Walls : chloritic alteration
Dyke : basic or intermediate

b) Siderite vein (massive) with disseminated sulphides (py-cpy) and minor quartz injection.
Walls : chloritic alteration
Dyke : acid

c) Quartz vein with sulphides (py-cpy) locally minor pyrrhotite.
Walls : sericite and/or chlorite
Dyke : acid

d) Magnetite vein with disseminated sulphides (cpy-py-po) - chlorite.
Walls : chlorite
Dyke : basic and intermediate

e) Semi-massive to massive sulphide veins (cpy-py-po) with chlorite-quartz.
Walls : chlorite
Dyke : basic and intermediate

Gold in three types of pyrite

<u>a) Sugary pyrite</u>
Fairly well crystallized, usually associated with chlorite and/or black sericite. Occasionally with quartz vein.
Gold grade : less than 1 g/t

<u>b) Concretionary pyrite</u>
Very fine-grained pyrite, in massive patches within black sericite or quartz vein.

Grading usually between 2.7 to 4 g/t

c) Brittle pyrite
Very large pyrite crystal, intensively fractured and filled by quartz material. Brittle, large patches within quartz vein.
Grade : No limit. Visible gold frequently associated.
Gold also occurs in chalcopyrite ore as fine particles within chalcopyrite crystals and along their edges.

SHAFTS & ACCESS

The main production openings of the Copper Rand mine are the 1024 m No. 4 shaft from surface and 1035 m to the SW the 382m. No. 6 internal shaft between the 817m (2550 ft) and 1200m (3600 ft) levels. In the No. 6 shaft area between the 900m (2700 ft.) and 1200m (3600 ft) levels, a ramp provides alternate access and is used for moving trackless mining equipment between working places. At Portage, a 3-compartment rectangular shaft services the operation from surface to 1307m.

MINING METHODS

Initially the mining methods employed at Copper Rand and Portage were shrinkage stoping in the narrower ore lenses, long hole stoping in the wider lenses and hydraulic cut-and-fill mining. In the first two methods, very large voids were left while broken ore was being drawn from the stope. As both mines became deeper, increasing ground pressure caused spalling of the hangingwall waste rock which mixed with and diluted the ore. Also as stoping progressed, many leads of ore were encountered which, when followed, resulted in additional tonnages of ore. These methods did not permit follow-up of such leads in spite of their relatively low cost. As a consequence, both operations have been largely converted to mechanized hydraulic cut-and-fill mining. Mill tailings are used for the Copper Rand operation. A sand pit located 25 km from Portage provides backfill material for the Portage Island operation. All the development and most of stope drilling is done with hand-held pneumatic jackleg machines (I single and I twin-boom pneumatic jumbo drills are used in large stopes). The ore is removed either using slusher, cavo or electric LHD unit (scooptram, 84 m^3 and I.7 m^3).

PRODUCTION (1960 TO 1989 INCL.)		
	lb Cu	oz Au
Copper Rand 13,387,466 t @ 1.85% Cu, 2.3 g/t(.068 oz/t) Au	495 336 242	910 348
Portage 4,794,338 t @ 1.86% Cu, 2.4 g/t(.071 oz/t) Au	78 349 374	340 398
Other divisions 2,690,462 t @ 1.86% Cu, 0.7 g/t(.021 oz/t) Au	100 085 186	56 500
TOTAL 20,872,266 t @ 1.855% Cu, 2.1 g/t(.063 oz/t) Au	773 770 802	1 307 246
ACTUAL RESERVES (01/07/89)		
Copper Rand 4,550,000 t @ 1.71% Cu, 2.15 g/t(064 oz/t) Au	159 372 000	298 240
Portage 2,407,000 t @ 1.53% Cu, 3.23 g/t(.095 oz/t) Au	73 654 200	228 665
TOTAL 7,067,000 t @ 1.65% Cu, 2.51 g/t(.074 ox/t) Au	233 026 200	526 905

Present production is 2700 tpd (Copper Rand 1700 tpd, Portage 1000 tpd) at an average of 1.53% Cu, 4.1 g/t Au (.132 oz/t). The mill has a capacity of 3700 tpd.

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JOUR / DAY 2B

L'INDICE DU LAC BERRIGAN (TACHÉ), UN EXEMPLE DE MINÉRALISATION ÉPITHERMALE EN Au-Ag-Pb-Zn D' AGE ARCHÉEN

Pierre Pilote[1], Jayanta Guha[2]

INTRODUCTION

Cette excursion, portant sur la Zone Nord de l'indice Taché, permettra aux participants d'examiner un exemple présumé de minéralisation épithermale en Au-Ag-Pb- Zn d'âge archéen (Guha, 1984; Pilote, 1987; Guha et al., 1988). Plusieurs points sugggèrent qu'il s'agit d'un gîte dont la mise en place s'est faite antérieurement au développement du métamorphisme régional et de la déformation.

LOCALISATION

L'indice Taché se situe en bordure nord du lac Berrigan, à environ 6 kilomètres au nord-ouest de la ville de Chibougamau. On peut s'y rendre par le chemin du golf municipal au nord de la ville de Chibougamau.

CADRE GÉOLOGIQUE

L'indice Taché est situé au sein du filon-couche de Roberge qui, avec les filons-couches de Ventures et de Bourbeau, constituent le Complexe de Cummings. Cet ensemble de filons-couches est intrusif dans la Formation de Blondeau, constituée de sédiments et de volcanites felsiques. Des dykes de composition tonalitique à localement syénitique recoupent ces roches et sont vraisemblablement associées au stock du lac Line. Ce stock est interprété comme une intrusion tardi- à post-tectonique (Pilote et al., 1984). La trace axiale du synclinal de Chibougamau, d'orientation ENE, est située à environ un kilomètre au nord de l'indice Taché.

L'indice Taché comprend de deux zones minéralisées, appelées Zone Nord (ou Zone Taché ou principale) et Zone Sud (ou Zone Berrigan) se situant respectivement au nord et au nord-est du lac Berrigan. Ces zones sont encaisées dans les faciès ultramafiques du filon-couche de Roberge. La Zone Nord est économiquement la plus importante des deux.

Historique

Ces indices minéralisés ont été découverts en 1929 par deux prospecteurs, D. Berrigan et L. Larone. Cominco a pris ces terrains sous option en 1930 et a procédé aux premiers travaux d'exploration. En 1944, O'Leary Malartic Mines Ltd. les prend en possession et les laisse sous option à la compagne Noranda de 1947 à 1948. Taché Lake Mines Ltd achète ces terrains en 1951 et effectuera jusqu'en 1968 plus de 12 600 m de forage. En 1969, Canadian Merrill Ltd acquière la propriété en finançant le fonçage d'une galerie d'exploration de 70 m, inclinée de 10° et deux travers bancs de 30 et 37 m chacun (Bidgood, 1969). Elle effectue des forages jusqu'en 1977. En 1980, Francana Oil and Gas Ltd (maintenant nommé Sceptre Energy Ltd) prend à sa charge les actifs de Canadian Merrill Ltd, dont la propriété Taché qui est laissée sous option aux Mines Camchib. Cette dernière compagnie reprend en 1981 l'approfondissement de la galerie d'exploration originale (Fig. 1), effectue en 1982 une campagne de forage et en 1984 un nouveau calcul des réserves. Depuis (1986), Greenstone Resources Inc et Bitech Energy Resources Ltd, cette dernière pouvant acquérir 50% de cette propriété contre le financement des travaux d'exploration, travaillent sur cette proprieté. Ces deux compagnies ont entrepris depuis l'automne 1987 des campagnes de forage, des travaux de décapage et sont à évalu et la possibilité d'exploiter ce gîte par une opération à ciel ouvert (Anderson, 1988).

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Zone Nord

Cette zone se situe à environ 200 m au nord du lac Berrigan, dans les péridotites et dunites du filon-couche de Roberge (Fig. 1). Les minéralisations se composent de filonnets et de brèches contenant de la sphalérite, de la pyrrhotine, de la galène, de l'arsénopyrite et de la chalcopyrite avec plus ou moins de pyrite. Les teneurs en argent et en or sont significatives, de 107ppb Au à 8,3 g/t Au et de 14 g/t Ag à 380 g/t Ag (Pilote, 1987), Elles occupent des fractures orientées 010° à 040° et dont les pendages varient de subverticaux à 45° vers le NW. Le filon-couche de Roberge est pour sa part orienté ENE et son pendage est abrupt vers le nord. La faille du lac Antoinette (Pilote, 1987), située à une centaine de mètres au nord de l'entrée de la galerie d'exploration, explique la répétition de certaines unités de ce filon-couche.

En 1977, les réserves prouvées atteignaient 346 000 t à 7,5% Zn, 34,1 g/t Ag et 7,5 g/t Au. En 1984, une seconde estimation des réserves, faite par Les Mines Camchib, avait donné 840 730 tonnes métriques probables contenant 4,12% Zn et 2,4 g/t Au. En 1988, un nouveau calcul, comprenant les catégories probables et possibles, a donné 1,43 tonnes métriques à 3,31% Zn et 1,9 g/t Au. De ce chiffre, des réserves probables de 568 700 tonnes métriques contenant 4,06% Zn et 2,4 g/t Au sont comprises de la surface jusqu'au niveau 107 m (350 pieds) et sont considérées pour une exploitation à ciel ouvert (Anderson, 1988).

Près de l'entrée de la galerie d'exploration, les lithologies consistent en des péridotites et dunites, serpentinisées à divers degré, alors que vers le sud-est, on rencontre principalement des pyroxénites (Huang, 1976; Pilote, 1987). Les pyroxénites et les péridotites alternent de façon intense, depuis le pied de la colline contenant cette entrée jusqu'à la rive nord du lac Berrigan (Fig. 2B.1). La minéralisation est exposée sur le flanc de cette colline dont une grande partie a été décapée en 1987 par la cie Bitech Energy Ressources Ltd. Les meilleures expositions se trouvent à environ 85 m au sud de l'entrée de la galerie d'exploration, à l'intérieur d'une séquence de pyroxénite et de péridotites carbonatées et silicifiées à divers degré. Ces observations permettent de proposer l'hypothèse que la position stratigraphique occupée par les minéralisations de la Zone Nord se situe entre le milieu et le sommet de l'unité basale composant le filon-couche de Roberge (Poitras, 1984). Cette partie, contrairement à l'unité supérieure, présente un litage magmatique bien développée illustré par l'alternance de wehrlite-dunite-pyroxénite. Cette alterance brusque et variable des lithologies ultramafiques dans cette partie du filon-couche rend les corrélations latérales laborieuse. De plus, la Zone Nord est affectée par différents réseaux de failles d'orientation pricipalement nord-est. On peut ainsi comprendre les difficultés qu'ont rencontrées bon nombre de géologues à établir des corrélations stratigraphiques tant dans leurs travaux de cartographie que dans leurs campagnes de forage.

La pyroxénite encaissant les minéralisations est intensément bréchifiée le long d'un "couloir" légèrement ondulant de direction nord-est et d'une épaisseur de 3 à 5 mètres. Les filonnets minéralisés, de 3 mm à 4 cm d'épaisseur, ont comblé ces cassures, donnant ainsi en affleurement l'impression de se subdiviser en plusieurs embranchements qui se rejoignent ici et là. Ces observations témoignent également des pulsions successives des fluides minéralisateurs. Certains filonnets enveloppent complètement des fragments de pyroxénite de taille considérable (de 2 à 3 mètres de diamètre).

Les filonnets minéralisés consistent en un matériel bréchique qui se compose de fragments d'encaissant silicifiés, recoupés à l'occasion par de minces veinules de quartz et enrobés par des bandes de sulfures et de quartz d'épaisseurs variables, généralement inférieures à 2 cm. La silicification des fragments pyroxéniques et péridotitiques est localement si intense que ceux-ci ont parfois été identifiés dans le passé comme des fragments cherteux. La partie centrale des filonnets est marquée par une forte bréchification tandis que les bordures montrent plutôt un caractère rubané, démontré par la disposition préférentielle du quartz, de la pyrrhotine et de l'arsénopyrite. La pyrrhotine et la sphalérite représentent couramment plus de 75% de la composition des filonnets. Ordinairement, la sphalérite et le quartz se retrouvent en bordure de ces filonnets et la pyrrhotine dans la partie centrale. La pyrrhotine se retrouve également sous forme disséminée dans les fragments silicifiés et dans les épontes encaissantes.

Dans la partie bréchifiée des filons, tous les sulfures montrent des textures de déformation, de recristallisation et de rééquilibration. Ainsi, la pyrrhotine et la sphalérite exhibent une texture granoblastique. Tous les sulfures présentent de façon commune des points triples. Les sulfures au comportement plus fragile, tels l'arsénopyrite et la pyrite, ont fréquemment leurs fractures comblées par les autres sulfures voisins, soit

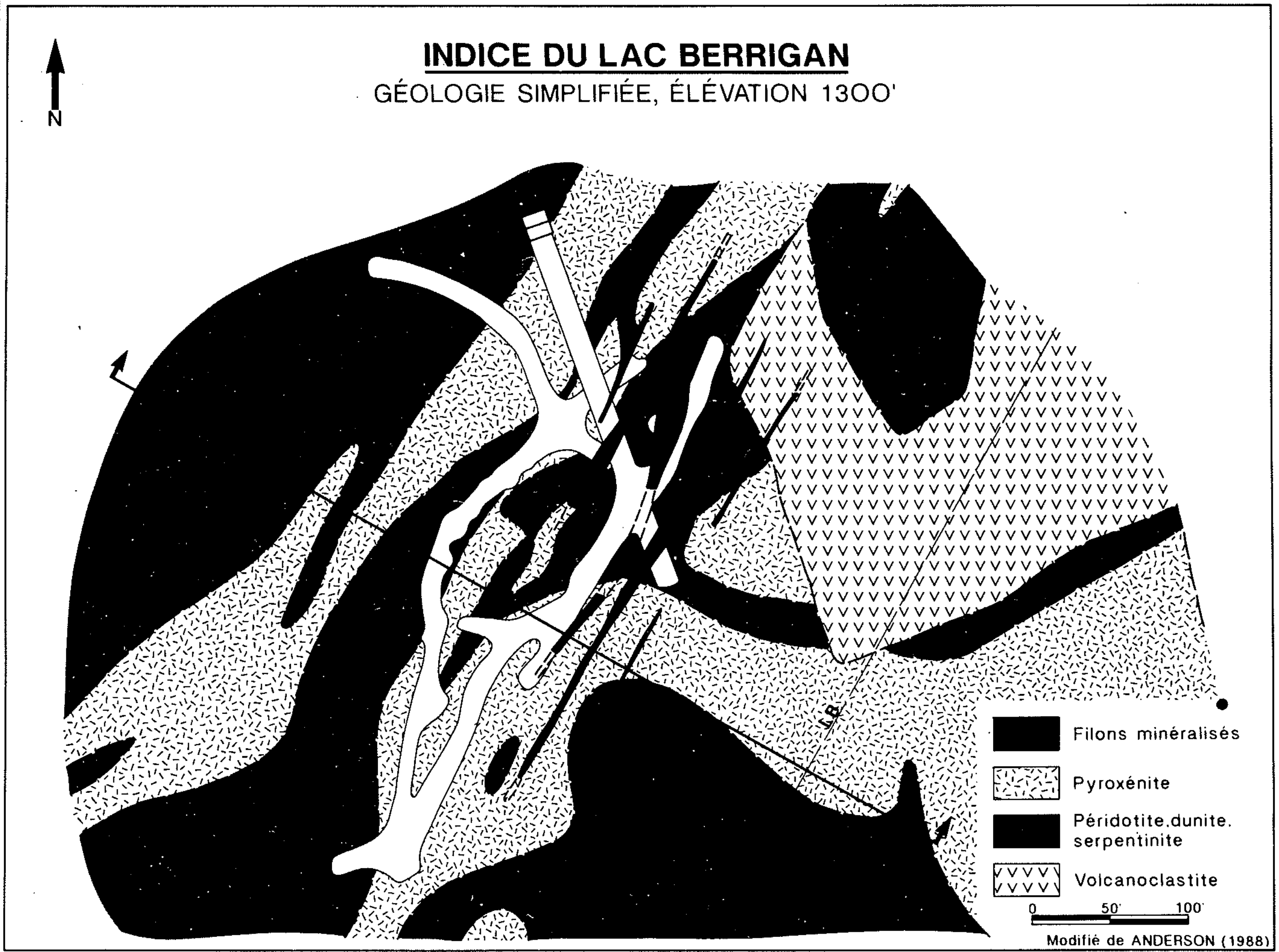

Figure 2B.1. Plan de niveau simplifiée (élévation 1300 pieds) de la Zone Mord de l'indice Taché, avec projection des contours de la galerie d'exploitation.

par la sphalérite et la pyrrhotine.

Les épontes sont fortement carbonatées. La silicification est cependant restreinte à la zone minéralisée et à une mince zone de bordure de moins de 0,5 mètre de large dans les roches encaissantes. A une vingtaine de mètres des filons, la pyroxénite est encore affectée par la carbonatation. L'ouralitisation des pyroxènes est totale mais des textures ignées reliques demeurent reconnaissables.

La séquence d'événements représentant la mise en place de la minéralisation peut se résumer de la façon suivante d'après Landry (1984), Pilote (1987) et Guha et al. (1988). Il y aurait d'abord eu altération en talc-carbonate des roches hôtes ultramafiques (autométamorphisme contemporain à la mise en place du filon-couche de Roberge) puis fracturation ponctuelle et silicification intense des lithologies, le long d'un couloir plus ou moins fortement fracturé. De faibles minéralisations en pyrrhotine et en sphalérite ont accompagné la silicification. Lors d'un événement minéralisateur majeur subséquent amenant la plus grande partie des sulfures, les roches subissent un second épisode de bréchification. De légères bréchifications et silicifications signalent la fin de cet événement minéralisateur. Ceci est démontré par la présence de veinules stériles de quartz et de carbonate de diverses orientations qui recoupent les filonnets minéralisés. Finalement, une partie de la bréchification observée dans la zone minéralisée, de même que les textures de déformation et de recristallisation présentes dans les assemblages sulfurés, témoingnent d'un événement dynamothermique tardif (kénoréen) et expliquent la déformation, la recristallisation et la remobilisation partielle des minéralisations.

Les failles nord-nord-est, dont certaines minéralisées, sont recoupées par des failles inverses d'orientation est-ouest à est-nord-est d'échelle régionale (Pilote, 1987); l'ampleur des rejets est cependant inconnue. Ces failles inverses produisent des dédoublements ou des contacts anormaux dans l'empilement stratigraphique. Toute la Zone Nord et les secteurs compris entre les lacs Berrigan et Larone ont été par la suite fortement affectés par des failles nord-nord-est de longues extensions, à décrochement apparent senestre et de pendage variant de subvertical à abrupt vers le sud-est. Ces failles apartiennent au réseau nord-nord-est (failles de type "R") de la zone de cisaillement de Gwillim, d'âge tardi-kénoréen et probablement même tardi-archéen (Dimroth et al., 1984 et 1985). Une telle corrélation permet d'expliquer les textures de déformation, de recristallisation et de remobilisation constatées dans les minéralisations sulfurées.

D'une manière sommaire, les diverses caractéristiques présentées par les minéralisations de la Zone Nord de l'indice Taché sont (Pilote, 1987):

1) une dispostion irrégulière de la zone bréchifiée, préférentiellement selon un couloir (ou cheminée),

2) la faible épaisseur en surface de la zone minéralisée [plus considérable en profondeur selon les travaux de sondage (Fig. 2B.2)],

3) la forme couramment anguleuse des fragments et la forte silicification associée,

4) la faible quantité de poussière de roche totalement broyée (masquée par la silicification?) dans la matrice de ces brèches,

5) caractère épisodique de la bréchification, de pré- jusqu'à post-minéralisation,

6) la présence de minéralisations subéconomiques à économiques.

Ces différents traits permettent, selon la nomenclature établie par Sillitoe (1985), de classifier la Zone Nord dans la catégorie des brèches phréatiques contenant des minéralisations de type épithermal. La roche encaissant ces minéralisations est par contre une lithologie peu commune pour ce type d'environnement, soit les ultramafites du filon-couche de Roberge. Cette hypothèse épithermale permet de relier l'existence du bassin d'effondrement avec un paléo-environnement en extension (Pilote, 1987; Guha et al., 1988).

Thorpe et al. (1984) ont procédé à l'analyse d'isotopes du plomb contenus dans la galène provenant des filonnets minéralisés de la Zone Nord. Les résultats ont donné un âge se situant autour de 2,72 Ga pour la mise en place des minéralisations. Ceci indique que la mise en place de ces filons est antérieure à l'orogenèse kénoréenne et au métamorphisme régional qui l'accompagne. Cet âge est aussi en accord avec d'autres résultats obtenus sur des gîtes de sulfures massifs d'âge volcanogène situés ailleurs dans le camp minier de Chibougamau et soutien ainsi l'hypothèse que les minéralisations de la Zone Nord du prospect Taché ont une mise en place précoce, vraisemblablement d'âge synvolcanique.

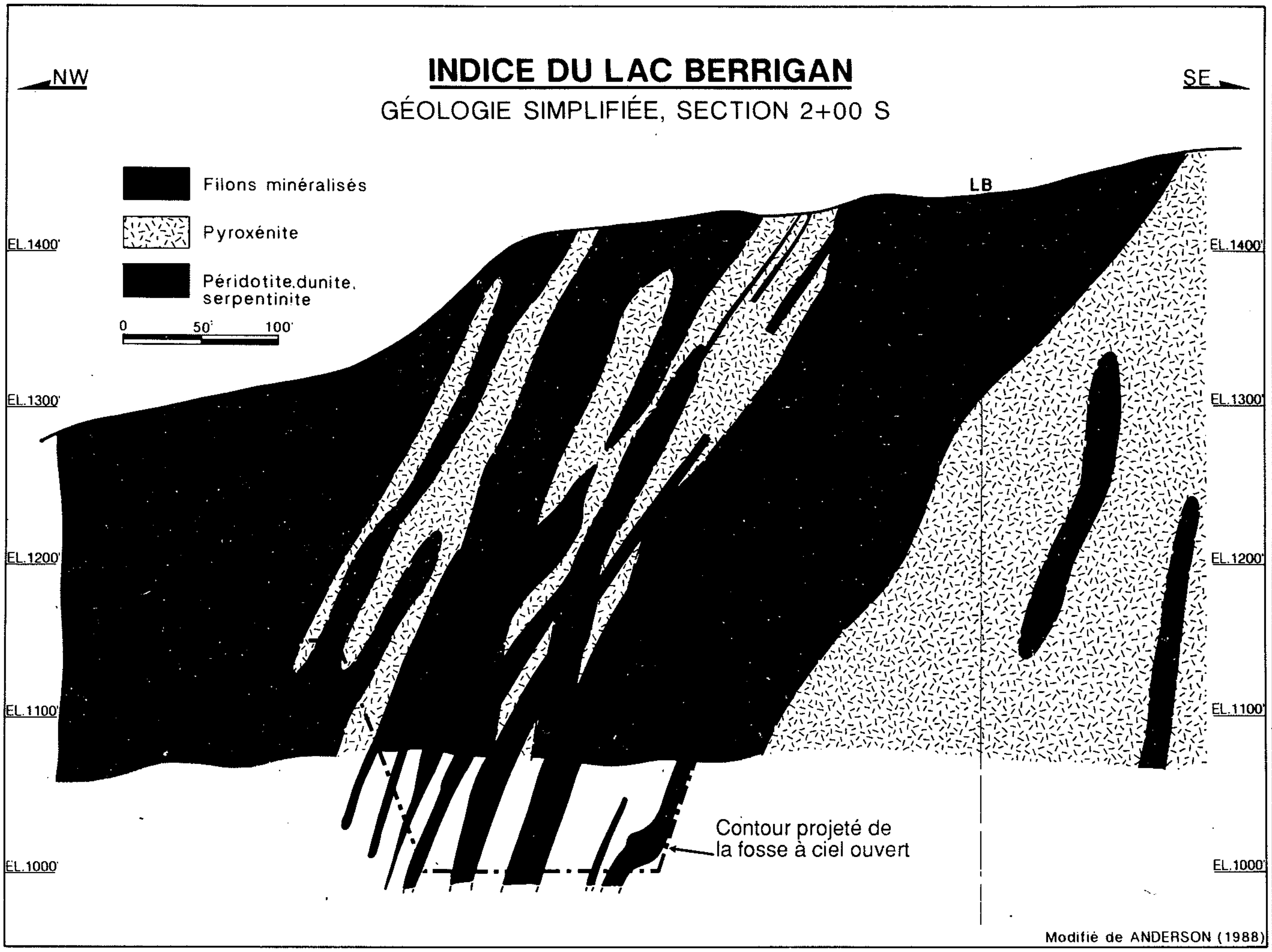

Figure 2B.2. Section simplifiée montrant les filons minéralisés et le contour projeté de la fosse à ciel ouvert. Les polarités dans les roches ignées font face vers le nord-ouest.

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THE NORBEAU GOLD MINE, ABITIBI GREENSTONE BELT, QUEBEC, CANADA*.

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INTRODUCTION

The Norbeau mine is a typical lode gold deposit hosted in Archean rocks of the Superior Province. It is a classical example of gold mineralization in a layered mafic sill, and a particular "sub-class" vein type of lode gold deposit in which the mineralization is almost entirely confined within the sill and where the fractures and vein patterns are complex (Dubé et al., 1989). These are common worldwide and some of the best examples are the gold deposits enclosed in the Golden Mile Dolerite sill at Kalgoorlie in Australia (Finucane, 1965; Boulter et al., 1987) and the San Antonio deposit in Canada (Gibson and Stockwell, 1948; Poulsen et al., 1986).

LOCATION

The mine property is located in McKenzie Township, approximately 5 km northeast of Chibougamau (Fig. 2C.1). The property is reached by a gravel road at the end of Chemin Merrill north of town. At km 5.1, turn left onto the "chemin de la mine Norbeau" and continue for 1.2 km to the mine gate. The New Vein outcrops are located 700 m past the gate. To reach the Sharpe Vein, continue on the gravel road for 1.9 km, turn left on the gravel road and continue 700 m past another gate.

GEOLOGY OF THE NORBEAU MINE

The Norbeau Mine operated from 1964 to 1969 and produced 419 029 tonnes of ore at a grade of 13.16 g/t Au and 1.64 g/t Ag. The mine is located on the south limb of the E-W ending Chibougamau Syncline. The local stratigraphic section, from base to top, comprises an ultramafic sill (Roberge sill), overlain by a mafic to intermediate sill (Ventures sill), followed by the volcaniclastic and pyroclastic rocks of the Blondeau Formation (Fig. 2C.2). The Bourbeau sill which hosts the gold mineralization at the Norbeau Mine was emplaced into the Blondeau Formation prior to deformation. It is a layered mafic intrusion composed of five different units; a basal peridotite-pyroxenite, leucogabbro, quartz ferrogabbro, quartz- granophyre ferrodiorite and a local upper ferrogabbro (Dubé and Guha, 1989). The contacts between these units as well as phase layering within the leucogabbro of the Bourbeau sill, and bedding within the Blondeau Formation typically strike E-W and dip north (260°/70°) (Fig. 2C.2). The rocks have undergone greenschist facies metamorphism.

The Norbeau deposit consists of gold-quartz veins in numerous ductile- brittle shear zones which cut the Bourbeau sill (Dubé and Guha, 1989; Dubé et al., 1989). Between these shears, strain is low and a relatively weak East-West regional schistosity (272°/74°) occurs sub-parallel to layering. Brittle faults with minor displacement and locally developed cleavage represent late-stage structures which strike both Northeast and Northwest.

The hydrothermal alteration associated with the gold mineralization is well developed and is characterized by a strong chloritization, carbonatization

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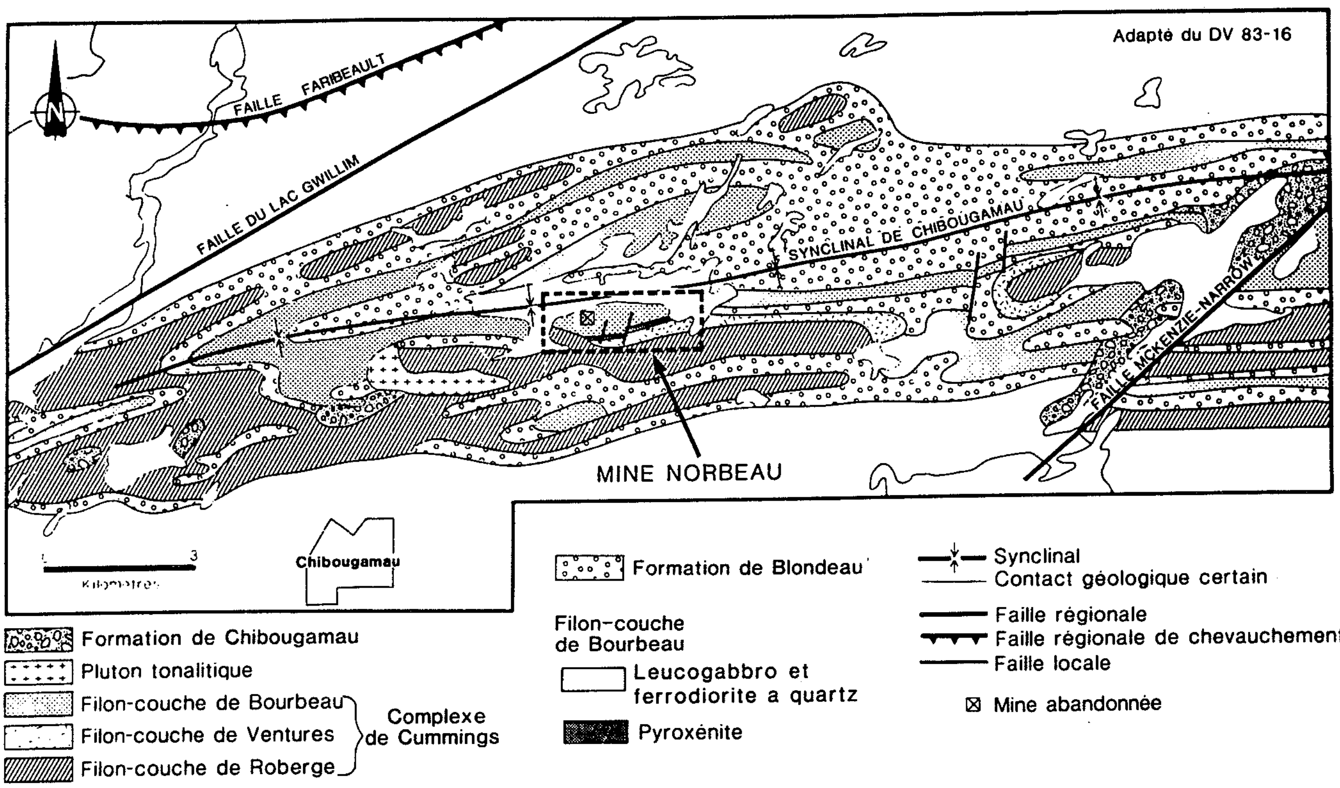

Figure 2C.1: Geological and geographical location of the Norbeau Mine (From Dubé and Guha, 1989).

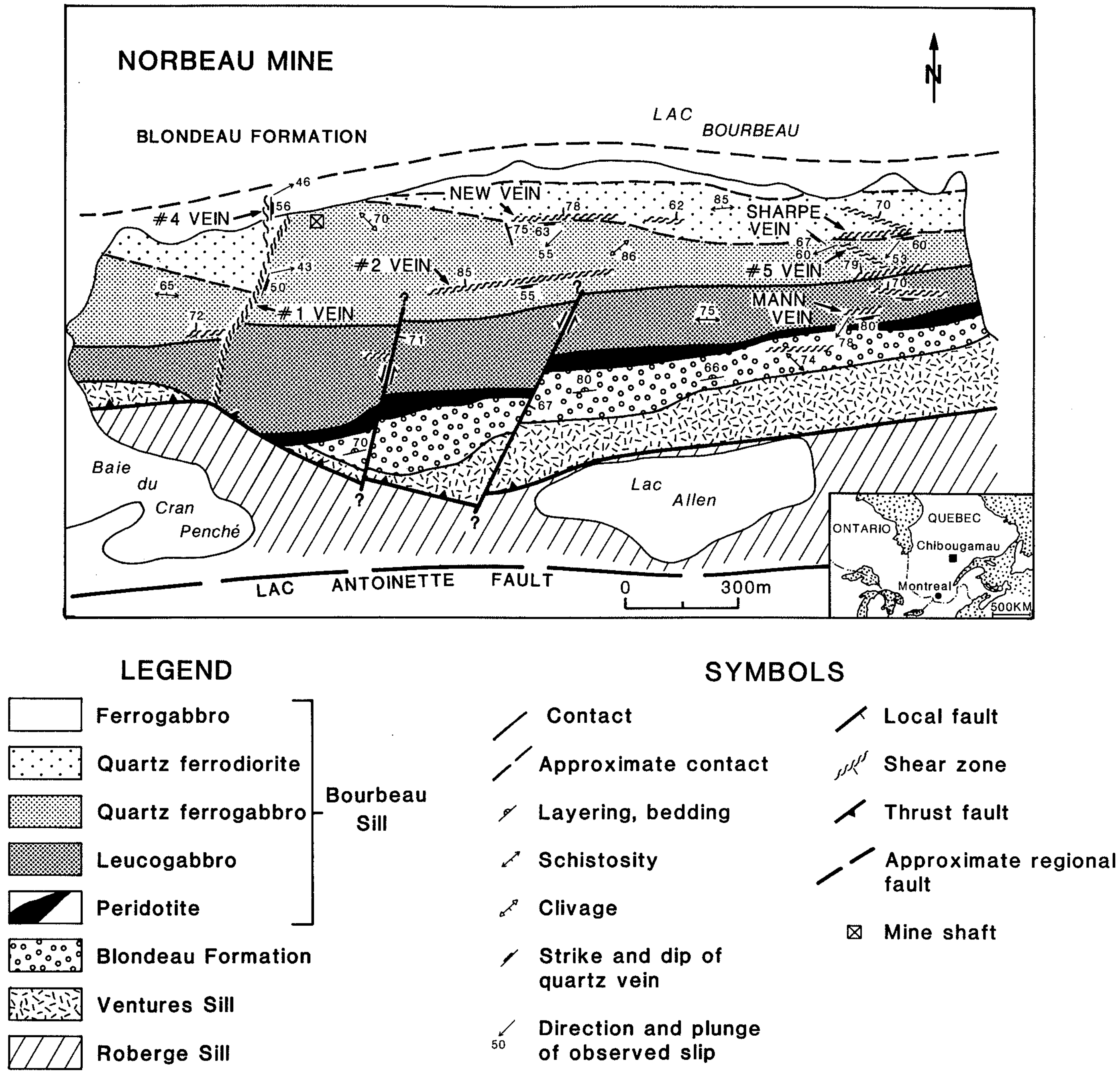

Figure 2C.2: Location map and geological setting of the Norbeau mine showing the stratigraphic units and the main structures (from Dubé et al.1989; modified from Dubé and Guha, 1989).

and sericitization of the wall rock. This hydrothermal alteration is not symmetrically distributed. The sericite and iron carbonate are mainly located in the footwall, where sub-horizontal quartz veins typically occur (Dubé and Guha, 1989). Gold mineralization consists of "steeply" dipping quartz veins containing a small amount (<5%) of sulphides, mainly pyrite and arsenopyrite with lesser pyrrhotite, chalcopyrite, sphalerite and free gold. Visible gold is sparsely distributed except in the Mann Vein where it is locally abundant. Microscopic studies suggest that gold is present as inclusions in pyrite and arsenopyrite, at grain boundary contacts between these sulphides and quartz, and filling fractures in sulphides and quartz (Dubé and Guha, 1989).

Shear Zones

Four different sets of auriferous shear zones occur in the Bourbeau sill (Dubé et al. 1989); the principal ones are shown in figure 2C.2. The most important mineralized shear zones, which furnished all the gold produced during the initial mining operations, are the northeast-trending (035°/50°) No 1 Vein, and the north-south-trending (000°/56°) No 4 Vein (Fig. 2C.2). Representatives of an E-W striking set are the New Vein (092°/63°), and the Sharpe Vein (090°/60°). Two lesser structures, the Mann Vein (085°/80°), and the No 2 Vein (082°/55°) as well as minor E-W quartz-bearing shear zones are also present. The No 5 Vein (120°/67°) is the only significant member of a northwest-striking set.

All of the shear zones record evidence of both ductile and brittle behaviour (Dubé et al. 1989). They are defined principally as zones of chlorite schist, within which schistosity consistently has a steeper dip than, and most commonly in the opposite direction from, the shear zones themselves. The schistosity has approximately the same strike as the shear zones. The oblique nature of the foliation to the shear zone boundaries, as viewed in cross-section, is taken to indicate ductile displacements of reverse sense. This is further supported by the presence of down-dip metamorphic mineral lineations which are of two types: one in the plane of the oblique schistosity and the other one in spaced fractures sub-parallel to the margins of the shear zones. Both types of lineation project onto the plane of each shear zone in its down-dip direction, though local deviations result in a minor sinistral horizontal displacement component on the East-West shears and a dextral component on the Northeast and North-South shears (Fig. 2C.2). While intense metamorphic foliation and lineation within the shear zones suggest a predominantly ductile deformation, the shear zones also display brittle characteristics. All shear zones contain a central quartz vein parallel to their boundaries and many extensional sub-horizontal quartz veins. Slickensides commonly occur along the walls of, and occasionally within the central veins. They contain slickenlines which are sub-parallel to mineral lineations in adjacent schists suggesting that they represent the last stages of the same progressive deformation event.

Despite their different orientations (Fig. 2C.2), all of the auriferous shear zones are interpreted to be contemporaneous (Dubé et al., 1989). They are all characterized by an oblique- reverse movement and they contain a central quartz vein as well as many extensional subhorizontal quartz veins. A detailed study shows that they share a common mineralogical and chemical pattern of hydrothermal alteration (Dubé, 1990). Sericite-ankerite-calcite-pyrite/ sericite-ankerite-calcite-chlorite, and chlorite-calcite form three distinct alteration facies which extend outward from each of the mineralized zones. A chemical mass balance study, based on Gresens (1967) equations, demonstrates a common leaching of Si and Na, and an increase of Ca, K, C, S, As and Au (Dubé, 1990).

Dubé et al. (1989) demonstrated that the four sets of shear zones at the Norbeau Mine were produced by a sub-vertical extension trending 346° and subparallel to the dip of the Bourbeau sill. The subhorizontal veins associated with the different shear zones are compatible with this. The other axes of deformation are located on a great circle oriented at 078°/36° and cannot be clearly distinguished from one another. However, there is an indication of maximum shortening parallel to the strike of the sill. According to Dubé et al. (1989), the sill behaved as a competent sheet in a matrix of "softer" sedimentary rocks and the Roberge and Ventures sills of ultramafic to mafic composition. Furthermore, layering within the sill resulting from magmatic differentiation may account for additional anisotropic strength. The development of shear zones at the Norbeau Mine can be logically attributed to the synchronous layer-parallel extension and shortening of the Bourbeau sill during regional deformation. The anisotropic strength of a stiff layer and its orientation may induce an internal strain different from the regional one and favour the development of local shear zones (Dubé et al., 1989).

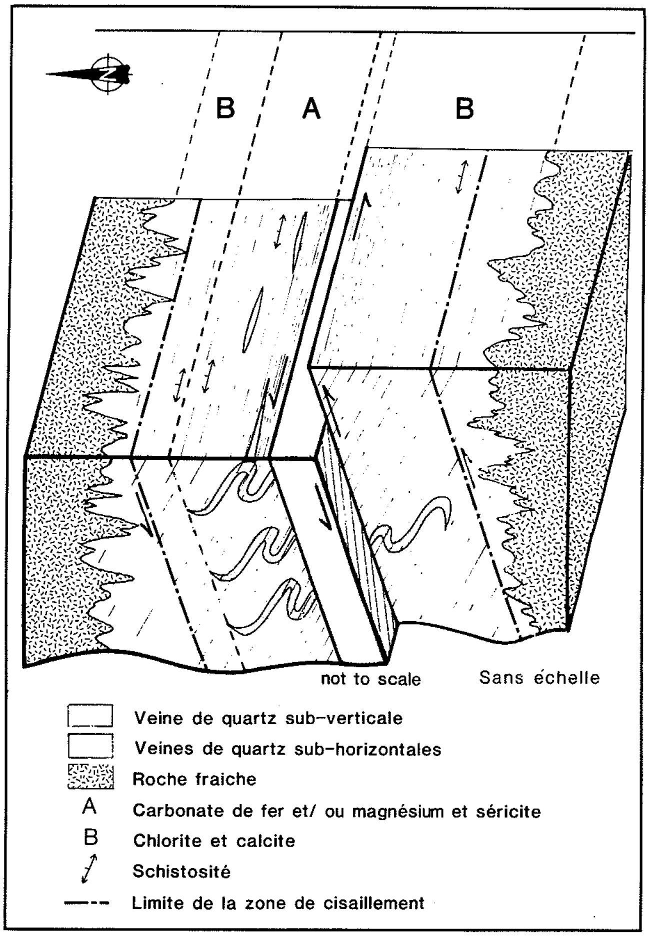

Figure 2c.3: Schematic block diagram showing structural relationships within the shear zones at Norbeau (From Dubé, 1990).

STOP DESCRIPTIONS

A brief description of the main structural features of the New Vein outcrops is presented. It should be noted that this description applies equally well to the Sharpe Vein and the No 1 Vein.

The mineralized steeply-dipping central quartz vein is enclosed in a 12 to 15 m wide ductile-brittle shear zone showing a strong E-W schistosity dipping to the north (270°/78°) whereas the quartz vein strikes sub-parallel to it but dips in the opposite direction. Dubé and Guha (1989) demonstrated that the shear envelope is sub-parallel to the steeply- dipping vein. A cleavage plane transecting the main schistosity and oriented sub-parallel to the main vein represents a possible shear plane. These features, combined with steeply dipping stretching lineations on the foliation plane and striations on the walls of the quartz vein, demonstrate that the shear zones moved in an oblique-reverse sense. The sub-horizontal veins are entirely compatible with this movement.

The alteration is characterized by an ankerite-sericite-pyrite zone close to the quartz veins, followed by an ankerite- sericite-calcite-chlorite zone mainly developed in the footwall and associated with the subhorizontal veins and a chlorite-calcite zone located in both the footwall and the hangingwall (Dubé and Guha, 1989).

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JOUR / DAY 3

GEOLOGY OF THE CAOPATINA SEGMENT

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Day 3 is devoted to typical lithologies, structures and mineralization in the southern Caopatina domain. A felsic volcanic centre (Lac des Vents Complex) built on and intercalated with tholeiitic basalts, and its associated mineralization, will be examined. The overlying sedimentary sequence (Caopatina Formation), affords a look at early folding in the belt defined by opposing structural facing. The dominant structural features of the southern terrane will be demonstrated and one of the major East-West faults (Doda) will be examined. Two shear-hosted gold prospects will also be visited. The different stops are located on the location map.

STOP 3.1: The Lac des Vents Complex.

Location

Turn west from the logging road L209S at kilometre 62 on the 2000 road. Proceed 10.2 km then turn left (northeast). Keep right at the 4.1 km intersection then turn left (west) 4.9 km further. Stop 3.1A is 500 m farther along the road. Continue until the next intersection 1.75 km, turn right (west), go 500 m further and turn left. Stop 3.1B is 1.45 km further, 100 m northwest of the road. Go back to stop 3.1A. Go east 500 m to the next intersection, turn right (south). Stop 3.1C is 600 M further, northwest of the road.

Geological Setting

The Lac des Vents Complex is a 2-to-2.5 km thick mafic-felsic centre composed of five distinct felsic units (FV-1 to 5) (Fig. 3.1). The entire sequence strikes northeast and dips 75-85° to the north. The sequence is overturned and youngs to the south. The first four units represent the construction of the edifice, wheras the last (FV-5) its destruction. Only units FV1- to -3 will be visited in this excursion. Units FV-1 to 3 are composed of the following components: (1) thick, massive to brecciated, dacitic lava flows, (2) felsic feeder dykes, (3) pyroclastic flows related to either magmatic or phreatomagmatic eruptions, (4) reworked pyroclastic material transported down-slope via high-density turbidity currents (Lowe, 1982) (5) pelagic sediments (shale), and (6) volcaniclastic (epiclastic) sediments deposited by low- and high-density turbidity currents (Lowe, 1982). Massive, brecciated, and pillowed basalt flows, as well as comagmatic sills constitute the interstratified mafic part of the edifice. A marine environment in moderate water depths of >200 m is implied by the background sediments, intercalated with the lower units. The uppermost unit, however, shows sedimentary structures related to wave-induced currents and/or storm influence, features indicating emergence of the volcanic edifice. A high-energy volcanic apron is suggested (Mueller and Chown, 1989).

Towards the top of the sequence some massive sulphide lenses, principally composed of pyrite, pyrrhotite with some chalcopyrite, were indentified and hydrothermal effects are recorded throughout the sequence. Reserves of up to 100 000 tonnes with 2% Cu were reported in the late fifties, however later work was not encouraging.

The emphasis of this excursion will be on felsic units FV-1 to -3 (Fig. 3.1) examining the various lava and pyroclastic flows, and related hydrothermal alteration, as well as mineralized showings in the black shale.

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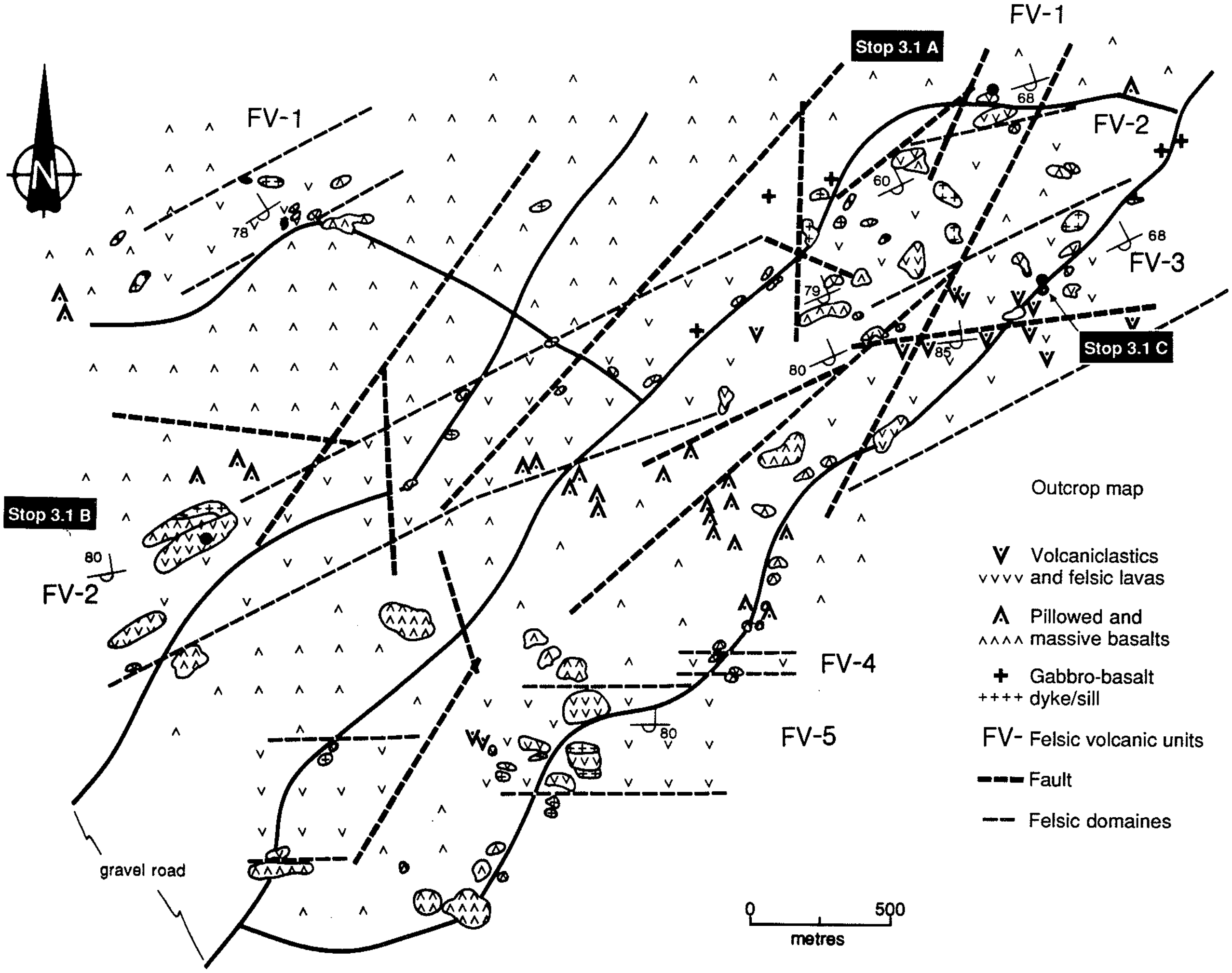


Figure 3.1: Lac des Vents felsic Complex, with location of stops.

Stop 3.1A

This outcrop zone (FV-1) represents the known base of the Lac des Vents Complex. A jig-saw puzzle felsic breccia occurs from the base nearly to the top of this outcrop zone (Fig. 3.2). At the base, a transition from massive to brecciated lava is observed. A pervasive hydrothermal alteration (epidotization/ chloritization) replacing the hyaloclastic (vitrophyric) matrix is related to this brecciation. Hydrothermal alteration resulted in the precipitation of secondary magnetite at the contact between breccia and pyroclastic turbidites. This is probably a form of seafloor alteration close to hydrothermal vents enhanced by later diagenetic processes. A thin dyke, interpreted as a degassing pipe, cuts across the breccia sequence, supporting the hypothesis of a proximal venting system. Furthermore, at the breccia/ pyroclastic turbidite interface, breccia displays a crude form of inverse grading. Grading and chilled pyroclasts are attributed to a hydroclastic or phreatomagmatic explosion at the magma/water interface.

The overlying pyroclastic turbidites feature a crude form of grading in dominantly parallel laminated beds. It is difficult to separate beds where masked by hydrothermal alteration. The overlying inversely graded breccia bed has the same characteristics as the lower one and is therefore a product of the same fragmentation process.

Stop 3.1B

Outcrop zone FV-2 is characterized by reworked pyroclastic breccias and subordinate tuff-lapilli tuff deposits (Fig. 3.3). These pyroclastic deposits are sandwiched between massive, in situ brecciated and pillowed flows at the base and a 5 m thick, brecciated felsic flow at the top. The 0.20-to-3m thick flows are composed of juvenile lithic fragments (felsic pyroclasts), cognate fragments (basalt clasts), and abundant accidental clasts (rip-up clasts of chert and background sediments), commonly set in a shaly matrix. The contacts between flows vary from erosive to diffuse. The heterolithic composition of components and abundance of boulder-sized rip-up clasts which are incorporated into the flows suggest an initial phreatomagmatic eruption that was re-transported down-slope via high-density turbidity currents causing the incorporation of accidental fragments into the flow.

The interstratified, laminated background sediments commonly contain pyrite nodules probably reflecting an early stage of diagenesis. The lower contact between basalts and laminated background sediments (argillite deposits) display a pervasive silicification due to contemporaneous hydrothermal activity in some places. These sediments have been eroded in turn by the subsequent reworked pyroclastic flows.

Stop 3.1C

The third felsic unit (FV-3) features the major, felsic constructional episode of the Lac des Vents complex. Primary pyroclastic flows (Fig. 3.4) of dominantly breccia-size fragments (> 6.4 cm, terminology of Fisher, 1966) are well exposed in this section. Flows are composed of angular to subangular, vesiculated to dense juvenile fragments, well-rounded pumice balls, and amoeboidal pyroclasts. The recrystallized matrix was originally volcanic glass or vapourized volcanic dust. The abundance of chilled margins around pyroclasts and amoeboidal-shaped fragments are indicative of a hot magma source. The initial eruptive mechanism could be either magmatic or phreatomagmatic in which water at the water/magma interface vapourized and acted as a catalyser for the ensuing eruption. Normal and inverse to normal grading is well developed in several flows. Contacts between flows may change laterally from erosive to sharp non-erosive to diffuse suggesting a change in transport process from turbulent to laminar.

An ideal sequence representing one eruptive event would be composed of the following: (1) massive or graded breccia (normal or inverse to normal grading), (2) massive to normal graded lapilli tuff, (3) laminated lapilli tuff, (4) laminated, in part graded, coarse-grained tuff, and (5) fine-grained tuff. Sequence 1 to 4 implies deposition from a high-density turbidity current with successive energy (velocity) dissipation and 5 is interpreted as settling or subaqueous fall out of fines from the subaqueous suspension cloud related to this eruptive event. The prevalence of coarse pyroclastic material and a subaqueous environment support a proximal vent interpretation.

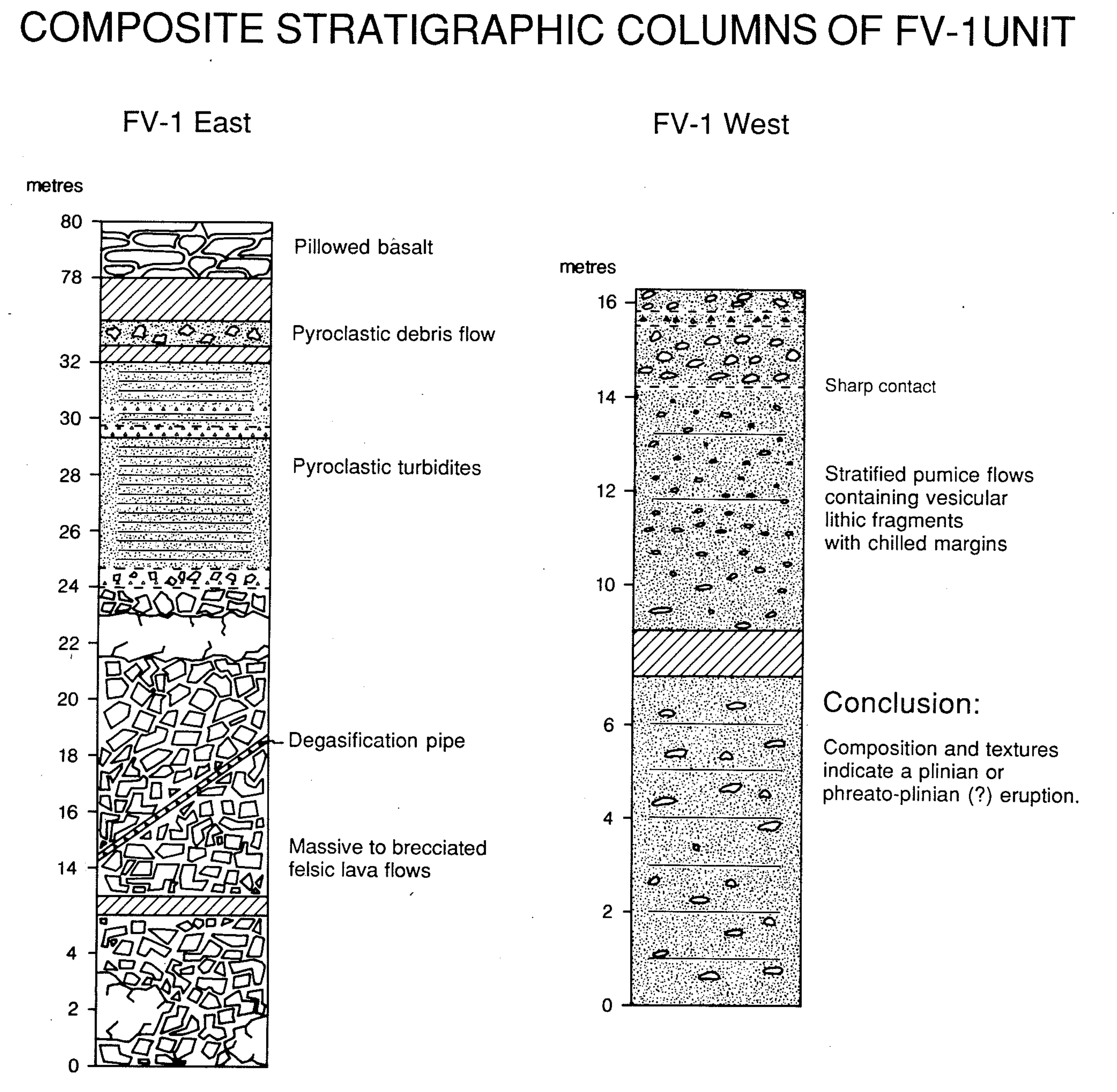
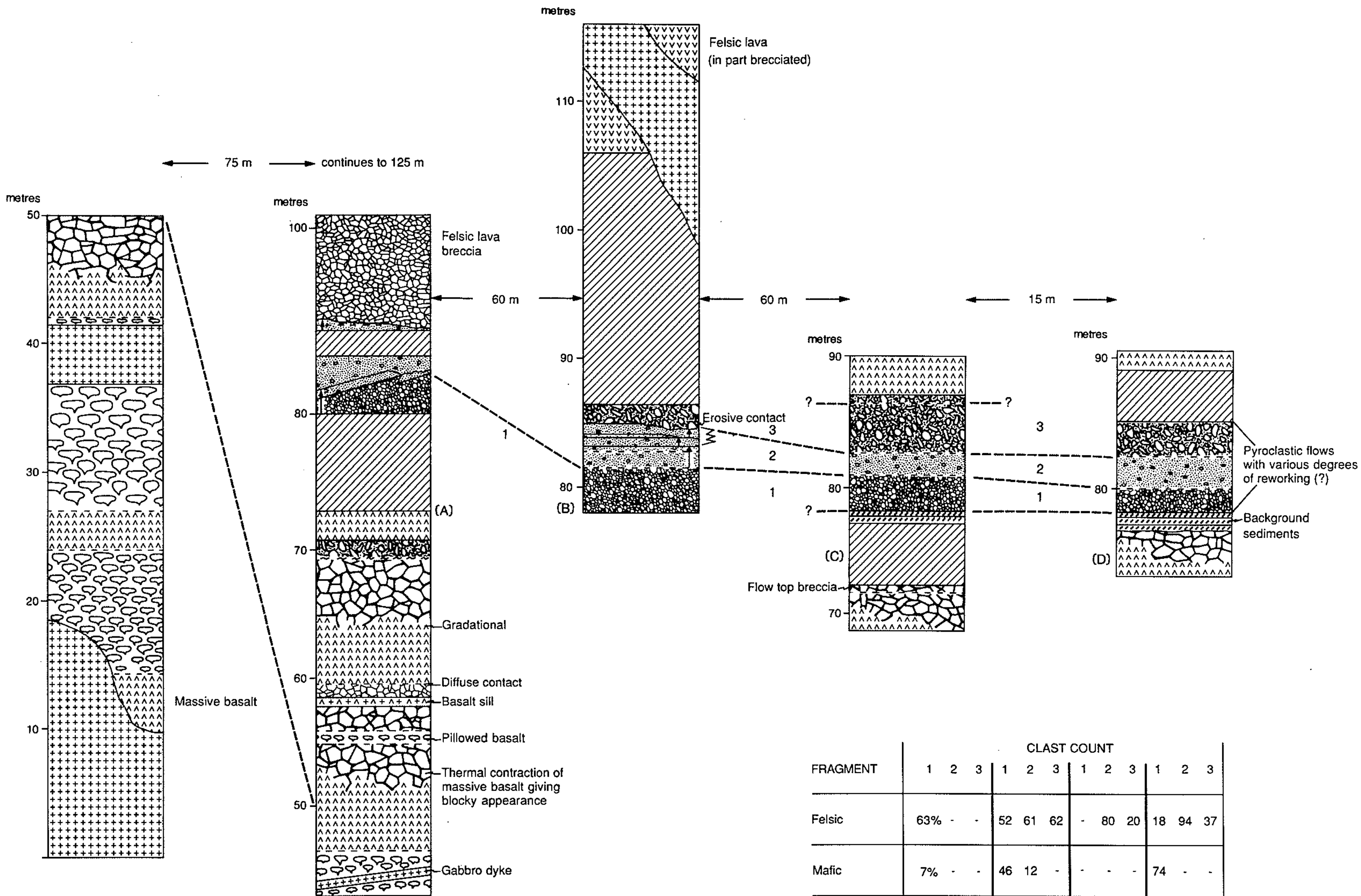


Figure 3.2: Composite stratigraphic columns of FV-1 Unit.

COMPOSITE STRATIGRAPHIC SECTION OF FV-2 UNIT



FRAGMENT	CLAST COUNT A 1	A 2	A 3	B 1	B 2	B 3	C 1	C 2	C 3	D 1	D 2	D 3
Felsic	63%	-	-	52	61	62	-	80	20	18	94	37
Mafic	7%	-	-	46	12	-	-	-	-	74	-	-
Sedimentary	30%	-	-	2	27	38	-	20	80	8	6	63
Column	FLOW A			B			C			D		

Figure 3.3: Composite stratigraphic columns of FV-2 Unit.

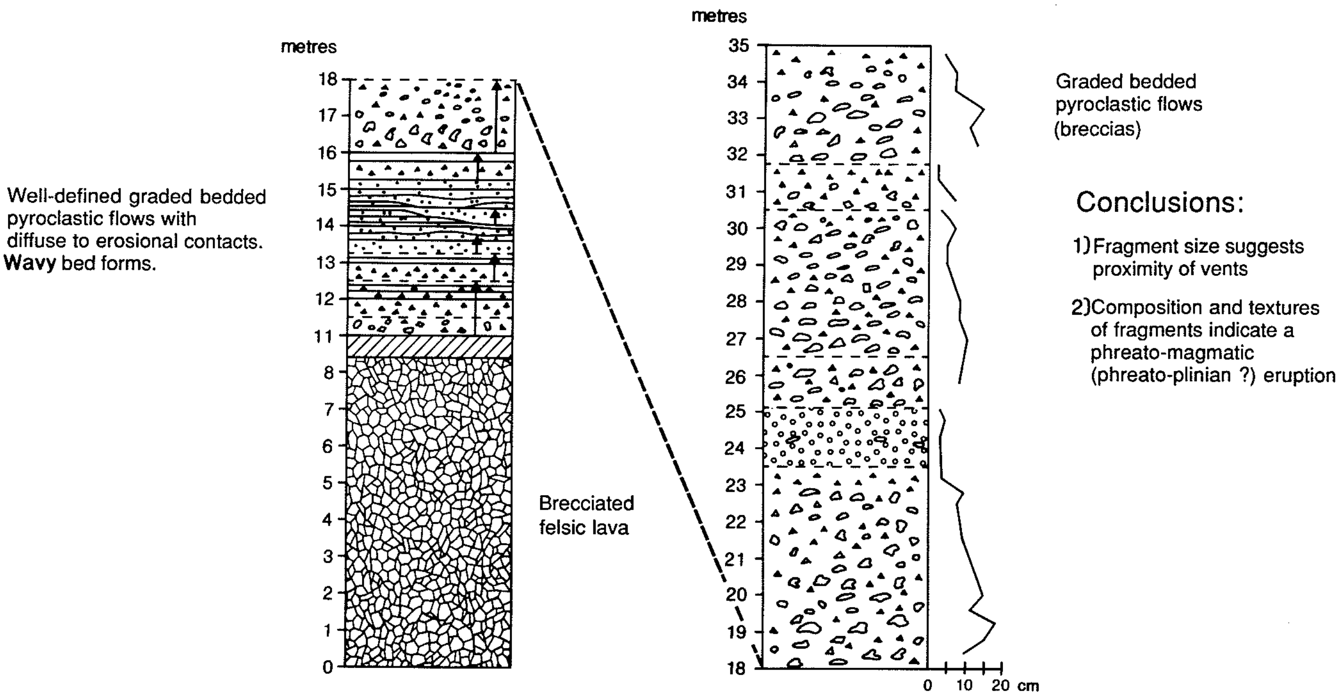


Figure 3.4: Detailed stratigraphic columns of pyroclastic flows (FV-3 Unit)

STOP 3.2: The Caopatina Formation

Location

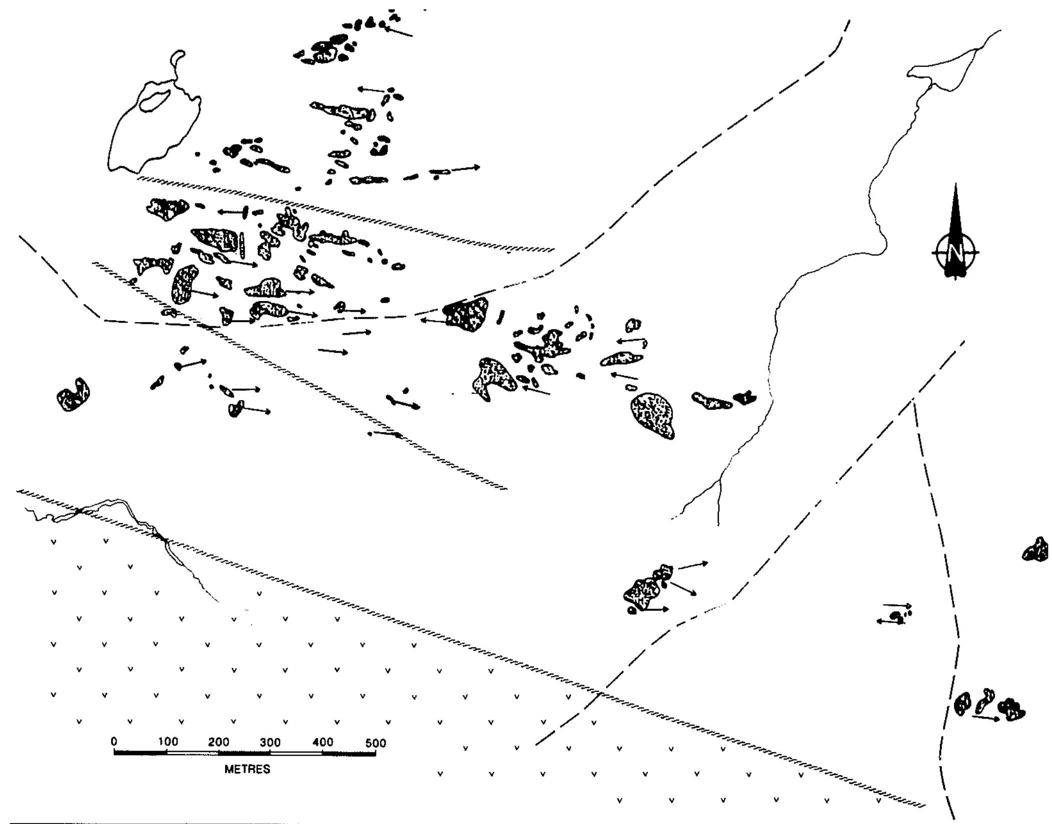
The best exposures of sedimentary rocks of the Caopatina Formation are in Druillettes Township, SW of Des Vents lake. Two outcrop zones will be visited in the area. Turn right on road 2000 at kilometre 62 of the logging road L209 south. Proceed 7.2 km, then turn left on road 7410. This road leads to the outcrop zone, at 4.8 km the first outcrop zone is 10 m south of the road and 1.4 km farther, the second outcrop zone is 25 m south of the road.

Geological setting

The outcrops are located on the southern limb of the Druillettes syncline. The rocks are relatively undeformed and consequently primary structures are easily recognizable. An angular relationship between the regional schistosity and bedding is common in this area, making the determination of the structural facing possible on several outcrops (Fig. 3.5, 3.6, 3.7).

Stop 3.2A:

The first outcrop zone is composed of interlayered wackes, siltstones and mudstones.



Structural facing in metasedimentary rocks of the Caopatina Formation

Figure 3.5: Attitude of stratification and younging directions, STOP 3.2.

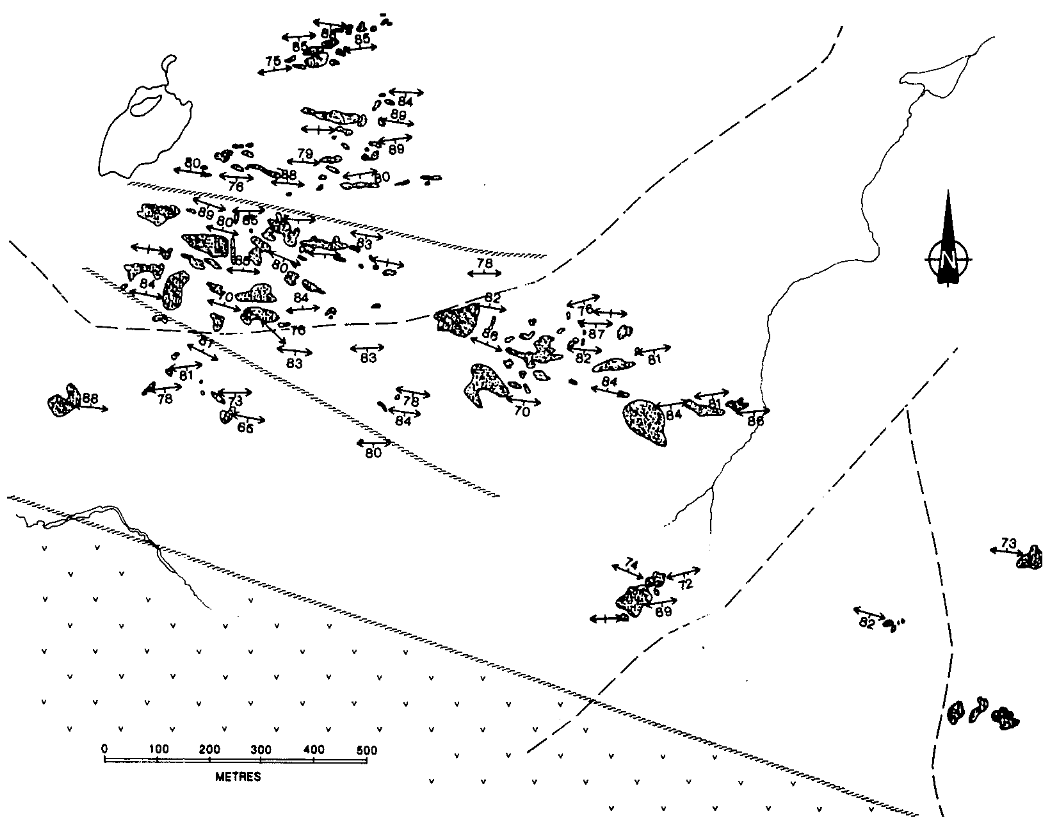
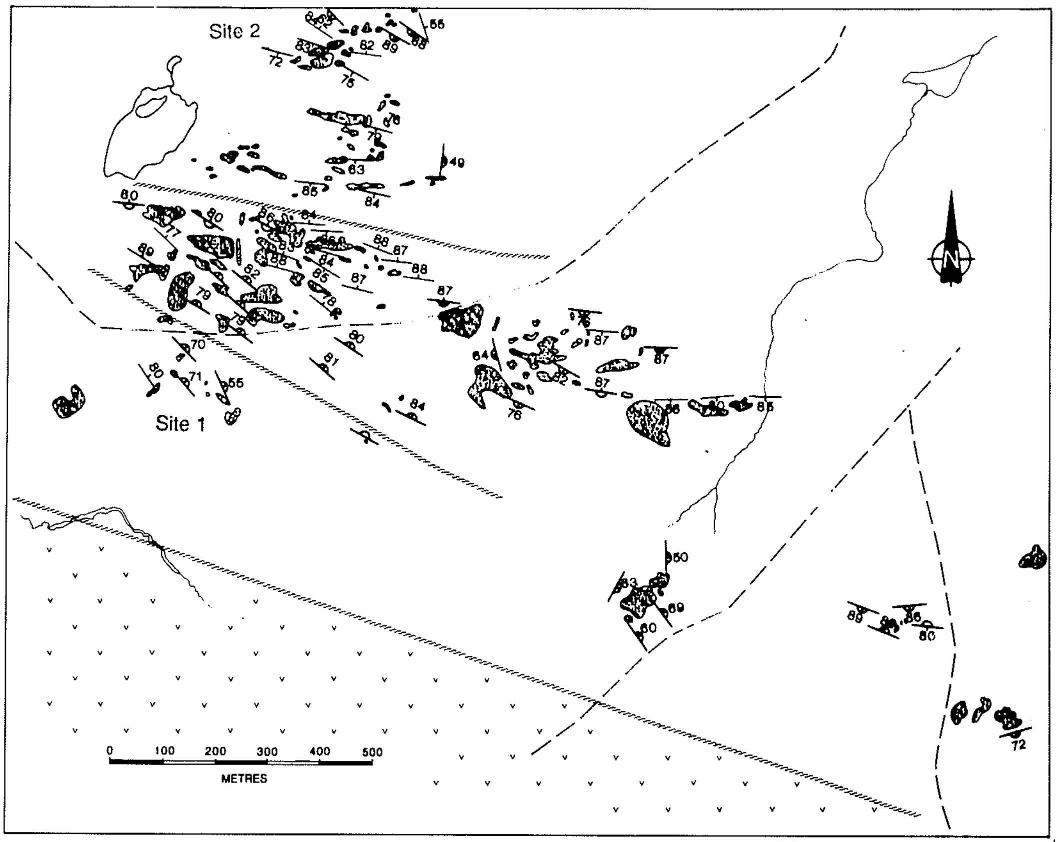


Figure 3.6: Attitude of regional schistosity, STOP 3.2.



Stratifications and stratigraphic facings in metasedimentary rocks of the Caopatina Formation

Figure 3.7: Structural facing, STOP 3.2.

Deformation is highly heterogenous, and intensely deformed domains, characterized by a strong schistosity separate relatively undeformed domains where the regional schistosity is represented by a spaced fracture cleavage. Attitude of the regional schistosity is relatively constant, and is approximately E-W with a sub-vertical dip, generally to the north. In intensely deformed domains, an oblique stretching lineation plunging 20 to 45 towards the east is measured and transposition of beds along the main schistosity can be seen. In relatively undeformed domains, stratification varies between N 310/70 to N 350/65. The intersection lineation between bedding and schistosity, and axes of minor sinistral fold hinges, plunge 60 east. Based on erosional surfaces, graded bedding, ball and pillow structures and other structures. The stratigraphic facing is E to NE indicating that the structural facing is towards the east.

Stop 3.2B

The second outcrop zone is mainly composed of feldspathic wackes, with thin interstratifications of siltstones defining a meter scale fold, steeply plunging west. An E-W axial plane schistosity is moderately to strongly developed. The stretching lineation is still oblique and moderately plunging towards the east. Sedimentary structures on the northern limb of the fold indicate a westerly structural facing.

These opposing structural facings could result from two different processes. Sheath folds formed during a shearing process may produce opposing structural facings. The attitude of the stretching lineation would then be expected to be intermediate to the fold axes of opposing limbs, a situation not observed in this area. The favoured interpretation is that opposing structural facing results from refolding of an pre-folded sequence (Lauzière et al., 1989). The absence of a cleavage associated with the earlier folding episode probably indicates that the sediments were not yet completely consolidated at that time.

STOP 3.3: The Doda fault

Location

Four outcrop zones will be visited in the vicinity of the fault. The outcrops are located on secondary logging roads off the L 209 south. Turn left at kilometre 65 on the L209 south, proceed for 5.85 km then turn right (south), the first stop is 2.3 km along this road. The first outcrops are located to the north, a few metres away from the road. The second outcrop zone is 200 m further east, on the main road. The third outcrop is located 2.75 km farther east, past an intersection at 2.1 km where you turn left. To go to the last outcrop, go east 100 m to the next intersection, and turn right (east). Follow the road for 8.1 km then turn right. The outcrop is located 1 km farther along this road.

Geological setting

The Doda fault is traced for over 120 km in the Druillettes syncline in the Caopatina segment. The E-W steeply-dipping fault forms a narrow zone in the west where it is constrained between the Lichen and Father plutons, but widens into a corridor up to 1 km in width within the supracrustal rocks to the east. South of Lac Caopatina the corridor is relatively wide and several excellent exposures have been stripped. Massive and pillowed basalts of the Obatogamau Formation, some of them charged with plagioclase phenocrysts, constitute the principal lithology of the fault zone. The stretching lineations of this part of the Doda fault plunge gently to the east, suggesting an important horizontal component of movement. This is in contrast to the down-dip plunge of lineations on most E-W faults in the Chibougamau region including the Doda fault farther west.

The outcrops of the fault zone were stripped and sampled for shear-hosted gold mineralization, but although some alteration and quartz veining is present, no significant values have been reported.

Stop 3.3A

On the first outcrop, the porphyritic basalts are deformed along an E-W plane dipping north 88°. The stretching lineation plunges 35° towards N 085°. Several kinematic indicators can be seen, not all of them showing the same sense of shearing. Asymmetric pressure shadows developed around the more resistant plagioclase phenocrysts suggest a sinistral component of movement. The displacement of parts of broken crystals, as in a deck of cards, also indicates sinistral movement.

A metamorphic overprint of mostly unoriented hornblende crystals can also be seen on this outcrop. This is typical of Grenville metamorphism.

Stop 3.3B

The second outcrop is composed of brecciated flows and possibly some highly deformed sedimentary horizons. As in the last outcrop, the lineation is gently plunging and asymmetric pressure shadows are observed. However a dextral component of movement is indicated by synthetic shear bands oriented N120/72°. Several examples of boudinage of more competent layers can be observed. Their presence is coherent with a sub-horizontal extension.

A regular tension-gash vein system is oriented N042°/82°. Combined with the compression that generated the shearing, it indicates a sinistral component of movement. However the diminishing angle observed between the veins and the main anisotropy could be produced by a transposition resulting from subsequent dextral shearing.

Stop 3.3C

The third stop shows two shear systems in interaction. One is oriented N075°, the other N110°. In the first system (N075°), the angle developed between the schistosity and the shearing direction is enhanced by an epidote alteration and clearly indicates a sinistral component of movement. Dextral sliding on a book shelf structure is also compatible with an overall sinistral movement on the main shear zone.

Stop 3.3D

At the fourth stop, slightly asymmetric ductile boudins suggest a dextral shearing component, whereas more fragile-type boudins suggest a later sinistral movement.

On some outcrops of the Doda fault, sub-vertical stretching lineations along discrete corridors, indicate that a complex deformation process, or distinct successive movements took place along the fault

The partipant will see that the kinematic history of the Doda fault is not fully understood and that the influence of the later Grenville Orogeny has yet to be evaluated.

STOP 3.4: Eratix

Location

The outcrops of the Eratix prospect are located on Soquem claims. Secondary roads off the L209 south logging road give access to this outcrop zone. Turn right (north) 200 m north of kilometre 48, follow the road for 4.25 km then take another road leading in a NW direction, for 1.1 km. The large stripped outcrops can be seen to the north. A ten to fifteen minute walk along an occasionally damp road leads to the spectacular outcrops.

Geological setting

The basalts composing these outcrops lie stratigraphically within the Obatogamau Formation. Porphyritic pillowed basalts are common, pillows up to 3 m in diameter are observed. Several feldspar and quartz porphyritic dykes of tonalitic composition, probably related to the Eau Jaune Complex, are also present. The structural trend follows a N-S to NNE (N020°) direction. The basalts are well preserved, hyaloclastic material is commonly observed between pillows. The presence of molar tooth shaped pillows suggest proximity to a fold hinge. South of the showings the Obatogamau Formation is consistently overturned with tops to the south. These showings were first located by a spectacular mineralized boulder with visible gold (hence the name Eratix). Stripping and drilling of the claims followed the discovery of the boulder. Gold values are associated with smokey quartz veins in a double alteration halo of quartz muscovite and carbonate alteration. Disseminated sulphides are less common here than in similar gold showings in the area, and exploration results so far have not shown consistent values.

Stop 3.4A

The map in figure 3.8 shows the approximate outline of the outcrops as well as the location of some points of interest. The Eratix shear is an E-W structure oriented N085° and diping south 75°. Gold bearing quartz veins are approximately parallel to this direction. The stretching lineation within is steep. It plunges 53° towards N190° with an 80° west pitch suggesting that the movement was mainly in a sub-vertical direction. At point 1, the shear direction is slightly deflected towards N120°. Small, local asymmetric Z folds are associated with a crenulation cleavage to the left of the main anisotropy. Its orientation is N 075°/65.

At point 2, a younger shear system, oriented at N045°/78 shows a well developed CS fabric. C is parallel to the major structure while S is oriented N250°/85 implying a sinistral movement compatible with the deflection of the earlier schistosity. A gently plunging stretching lineation (15° towards NE) associated with this younger system is also coherent with the proposed horizontal movement.

On limited vertical sections within the main shear, a plane crosscutting the schistosity is observed. It is possible to interpret this plane (N087°/58°) as a C-plane, which was combined with the schistosity oriented N090/80°, would indicate thrusting from south to north. This plane seems parallel to the quartz vein located within the shear zone. At this point, symmetrical alteration fronts on either side of NNE (N020°) fractures can be seen. Small quartz veins commonly occupy these fractures.

At point 3, the NE shearing is locally intense and the C and S planes are sub-parallel (N225/85).

At point 4, the CS fabric, as well as the reorientation of older structures along the shear zone, indicate a sinistral movement.

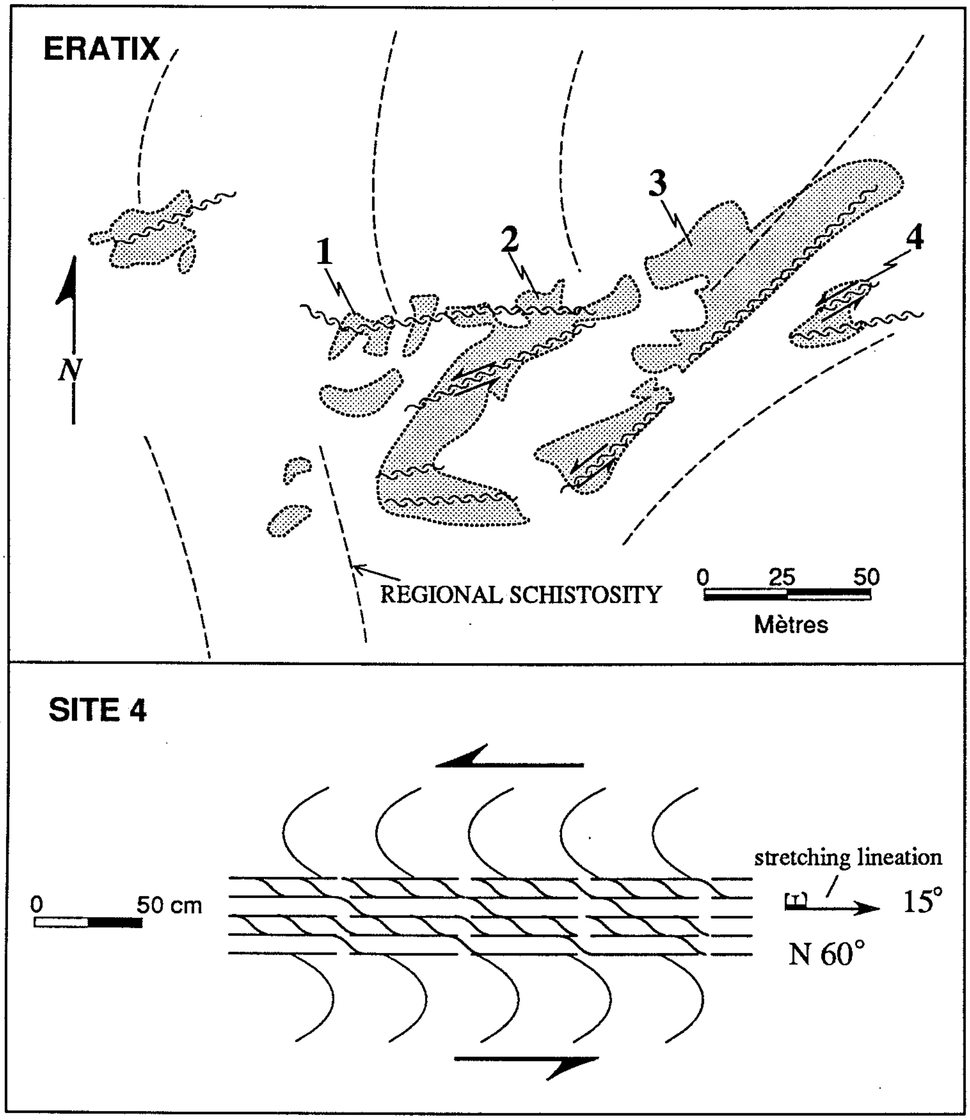


Figure 3.8: Geology of the Eratix showing

STOP 3.5: Mondor Showing

Location

Turn west from the logging road L209S at kilometre 38.8 on a secondary road. Keep to the left at intersections at 2 and 3.9 km, and turn right at 4.4 km continue north, stop 3.5A is at 6 km and stop 3.5B at 9.1 km.

Geological Setting

The host rocks are locally porphyritic pillowed flows, pillow breccias and thin gabbro sills of the Obatogamau Formation. The formations strike N-S and dip west with tops to the east in this section. Numerous porphyritic tonalite dykes believed to be related to the Eau Jaune Complex immediately to the southwest cut across the basalts.

The showing, discovered in 1957, was stripped and drilled in 1984-86. Mineralization is contained in quartz veins located in shear zones oriented N120-130°. The quartz veins contain disseminated sulphides and have anomalous gold values. Results did not justify further exploration.

Stop 3.5A (optional)

Outcrop beside the road shows a particularly good example of a lava tube. The tube, 3 m by 2 m, contains numerous shelves representing different lava levels formed within the active feeder tube.

Stop 3.5B

The main shear zones (Fig. 3.9), striking N120-130°, are complemented by a second set of shear zones striking N-S. The two result in an anastomosing pattern (Fig. 3.10), the whole zone striking about N140°, nearly parallel to the regional schistosity in this sector. The principal shear zones indicate a sinistral sense of movement and the N-S shears dextral, resulting in a conjugate system superimposed on the regional schistosity. The steep stretching lineation is related to the development of regional schistosity, and the superimposed conjugate shears do not appear to have developed a pronounced stretching lineation.

The shear zones are enclosed in a 5-metre-thick aureole of carbonate alteration where the shearing is most intense (Fig. 3.9). Quartz veining is most evident within the zone of intense alteration. The veins are small, discontinuous bodies generally less than 1 metre in length and averaging 30 cm in thickness.

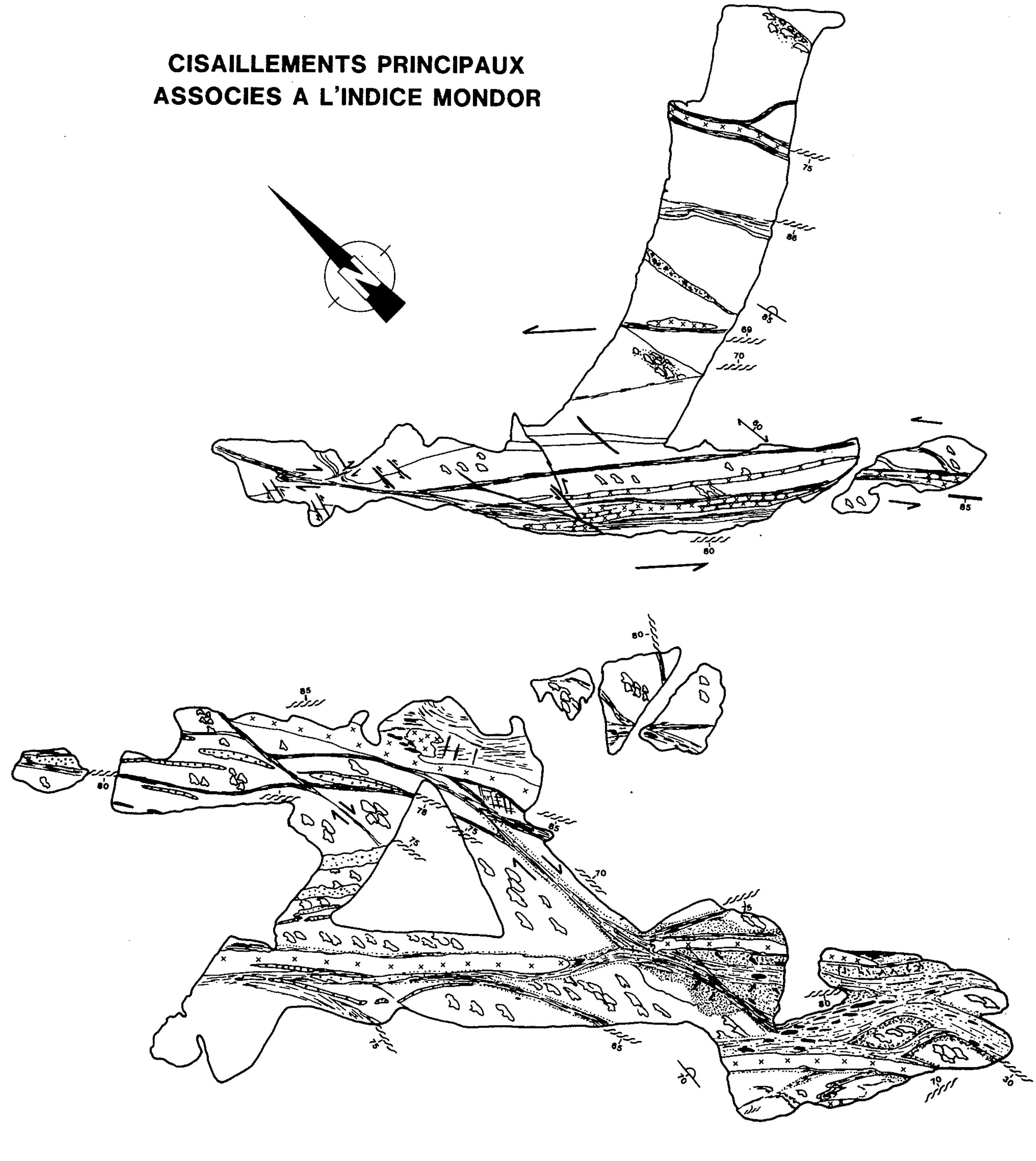


Figure 3.9: Geology of the Mondor showing with principal shear zones.

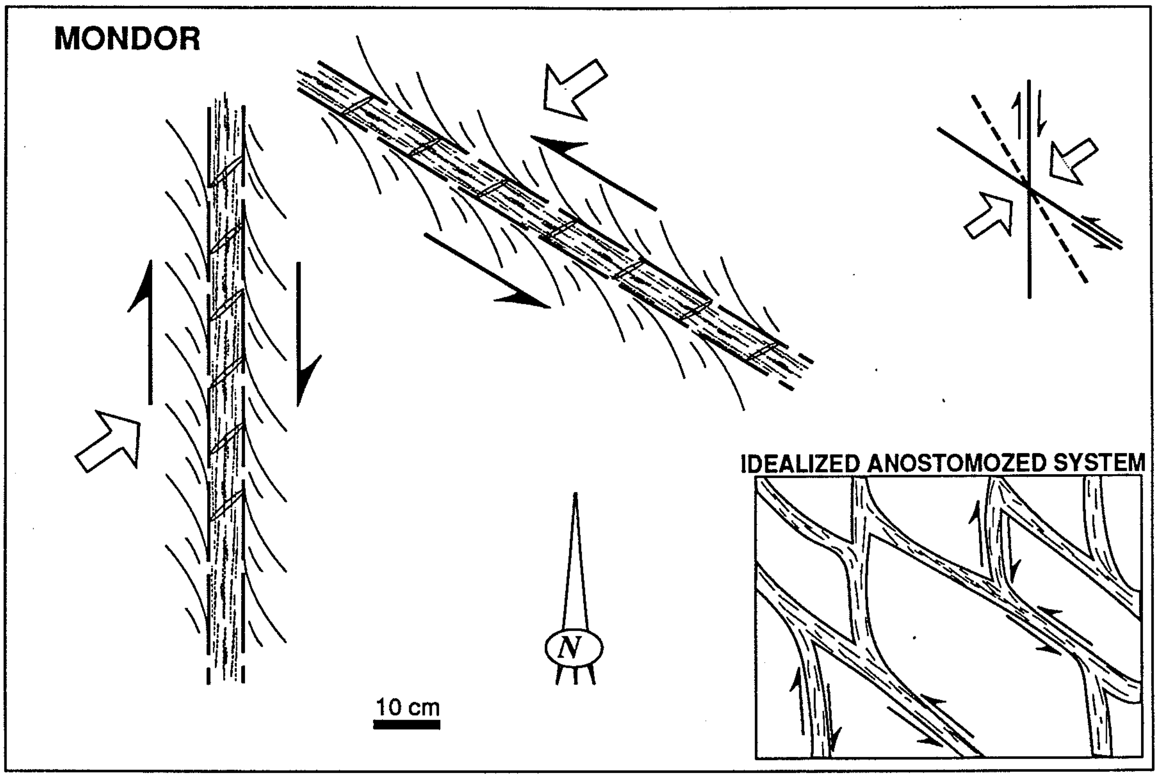


Figure 3.10: Conjugate shears producing an anastomozed shear system.

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JOUR / DAY 4A

GÉOLOGIE DE LA MINE JOE MANN

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Localisation

La mine Joe Mann (anciennement Chibex) est une mine d'or et de cuivre située à la limite des cantons de Rohault et La Dauversière (feuillet SNRC 32G/8) à environ 64 km au sud-ouest de la ville de Chibougamau. On s'y rend en empruntant le chemin non pavé en direction ouest à l'intersection du kilomètre 191,5 de la route Chibougamau - St-Félicien (route 167). On roule ensuite pendant 19 km pour atteindre la mine.

Historique de la mine

L'histoire de cette mine est longue et pleine de rebondissements. La première découverte d'or sur la propriété a été faite par Rider Storn durant la campagne de terrain 1950. Les travaux qui ont suivi conduisirent à la mise à jour de la zone Principale durant l'hiver 1951. La compagnie Chibougamau Explorers Ltd, devenue à partir de 1956 Anacon Mines Ltd, mit la propriété en valeur en creusant un puits de 450 pieds en 1952 et en l'approfondissant successivement à 1250 pieds en 1954 puis à 1850 pieds (profondeur actuelle) en 1959. Entre 1956 et 1960, l'exploitant avait extrait de la mine 685 868 tonnes de minerai à 6,8 g/t (0,22 oz/tonne) d'or et 0,50% de cuivre. A la fin de 1960, la mine était fermée pour refinancement, profondissement du puits et d'autres développements souterrains. En 1961, le feu détruisait le concentrateur et les installations étaient démantelées.

En 1970, la propriété est transférée à une nouvelle compagnie, la Chibex Mining Corp. (rebaptisée Chibex Ltée en 1972). Des travaux préliminaires délimitèrent des

réserves de 1 207 619 tonnes à 0,51% de cuivre et 8,4 g/t (0,23 oz/tonne) d'or pour les zones Principale et Nord au dessus de 1350 pieds. En 1973-74, Chibex fonça une rampe d'exploration 1,5 km à l'ouest du puits et on entreprit de dénoyer la mine. La mine et le concentrateur débutèrent leurs opérations au début de 1975 avec une production de 750 tonnes par jour. En 1976, des difficultés financières et une mauvaise récupération des métaux forcèrent Chibex à cesser ses activités et à vendre ses intérêts à Meston Lake Resources Inc. en 1979. La production au cours de la période 1974-1975 s'est élevée à 169 000 tonnes de minerai à une teneur de 0,38% de cuivre et 4,23 g/t (0,136 oz/tonne) d'or.

En 1981, suite à un accord entre MLR et la Société de Développement de la Baie James, la mine est de nouveau dénoyée et réhabilitée. On la rebaptise mine Joe Mann, du nom du fameux prospecteur de la région de Chibougamau. Cependant, à la fin de la même année, un conflit entre les deux compagnies entraîne la fermeture de la mine. En 1984, les deux parties parviennent à une entente qui permet à Meston Lake Resources, devenue depuis une filiale à part entière de Campbell Resources (décembre 1987), de devenir propriétaire à 100% de la propriété en échange d'une redevance retenue par SDBJ sur la production potentielle.

La mine Joe Mann est entrée officiellement en production en avril 1987 avec une production journalière de 750 tonnes et des réserves de 910 000 tonnes à 7,54 g/t (0,22 Au oz/tonne) d'or. Cette même année, la découverte d'une nouvelle zone minéralisée dans le prolongement est de la zone Principale porta les réserves de la mine à 3,2 million de tonnes à une teneur de 7,85 g/t (0,229 oz/tonne). Cet événement amena la compagnie à entreprendre un programme d'expansion important qui devrait permettre de passer d'une production d'environ 45 000 tonnes de minerai en 1989

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à 100 000 tonnes en 1992. Ce programme comprend, entre autre, le creusage d'un nouveau puit à 4 compartiments jusqu'à une profondeur de 680 mètres (2050 pieds).

Géologie de la mine

Cette description de la mine Joe Mann est tirée d'un rapport (Dion et Guha, 1988) constituant la première étape d'une étude de la métallogénie de l'or dans la partie orientale du segment Caopatina- Desmaraisville réalisée pour le compte du Ministère de l'Énergie et des Ressources du Québec. La mine a aussi été l'objet d'un mémoire de maîtrise par Wagner (1979).

La stratigraphie de la mine (Figures 4A.1 et 4A.2) est composée, du nord vers le sud et de la base vers le sommet, d'un filon-couche de gabbro (± 300 m), de basaltes déformés et altérés (3 à 50 m), d'un mince niveau (60 m) de «rhyolite» ou de tufs felsiques surmonté de nouveau par des basaltes. Cette séquence est typique de la partie supérieure de la Formation d'Obatogamau. Les couches ont une orientation est-ouest avec un pendage subvertical et une polarité vers le sud.

La minéralisation aurifère se présente dans des veines de quartz- carbonate décimétriques logées dans trois zones de cisaillement E-W (275/85N) de type ductile-cassant, subparallèles entre elles et à la stratigraphie et distantes l'une de l'autre d'une centaine de mètres (Fig. 4A.2). Un filon-couche de gabbro est l'hôte des zones Nord et Principale, alors que la zone Sud se trouve dans la rhyolite. La nouvelle zone minéralisée découverte en 1987 dans le prolongement E de la zone Principale possède les mêmes caractéristiques que celle-ci. Sur une section longitudinale (Fig. 4A.3), on observe que ces structures minéralisées forment deux lentilles montrant un fort plongement vers l'est. Certaines veines peuvent être suivies latéralement et verticalement sur plus de 150 m. Le minerai extrait de la mine Joe Mann provient en grande partie (plus de 70%) de la zone Principale et, accessoirement, de la zone Sud. Cette dernière n'est exploitée qu'à partir du niveau 1050 pieds. La zone Nord n'a été exploitée jusqu'ici que localement en raison de son manque de continuité et des teneurs en or erratiques qui la caractérisent.

L'épaisseur des cisaillements varie de 8 à 20 m dans le gabbro et à moins de 5 à 6 m dans la rhyolite. Les veines de quartz-carbonate aurifères occupent le centre des zones de cisaillement et sont généralement subparallèles à l'orientation de celles-ci. La direction de la schistosité principale est parallèle à la surface enveloppante des zones déformées soit environ 275°. La schistosité montre cependant un fort pendage vers le sud (80 à 85°S) définissant ainsi une relation angulaire avec les murs de la zone de cisaillement de pendage nord. Cette relation indique un mouvement vertical inverse, i.e. un chevauchement du nord vers le sud. La présence d'une linéation d'étirement minérale à fort plongement (80-85°), parallèle à l'orientation des lentilles minéralisées, constitue un autre argument important en faveur d'un mouvement vertical. Cette linéation d'étirement est particulièrement bien définie dans la zone Sud.

Une autre évidence de mouvement inverse est donné par les plis d'entraînement en Z à charnière subhorizontale dans la zone de cisaillement Principale. Ces plis d'amplitude centimétrique à métrique déforment à la fois la schistosité, les veines de quartz et les dykes felsiques. D'autres indicateurs cinématiques à l'échelle microscopique confirment ces observations (fabrique C-S, porphyroclastes asymétriques ou «shear bands»).

Les zones minéralisées sont recoupées et déplacées par des failles tardives NE à pendage est ou ouest. Ces failles ont joué dans plusieurs sens mais le mouvement apparent principal semble senestre. Des dykes de diabase protérozoïques de 2 à 3 m d'épaisseur ont été injectés préférentiellement dans les failles NE à pendage est. On observe aussi dans la partie ouest de la zone Nord un système minéralisé de direction NNW-SSE.

Le gabbro «sain» (non cisaillé ni altéré) est une roche homogène de couleur vert foncé à vert moyen, à grains moyens à grossiers, avec un indice de coloration variant entre 70 et 90%. Le filon-couche montre peu d'indices de différenciation mis à part la présence de lentilles de gabbro plus leucocrate (I.C. 50-70). Le litage magmatique est indistinct. Cette roche présente un assemblage albite- hornblende (ferro-tschermakite)-quartz- épidote-magnétite typique d'un métamorphisme au faciès supérieur des schistes verts.

Les zones de cisaillement qui recoupent ce gabbro sont constituées en périphérie de roches plus ou moins schisteuses montrant une altération rétrograde en chlorite - carbonate. En s'approchant des zones minéralisées, la déformation et l'altération deviennent plus intenses et le gabbro est transformé en schiste à biotite

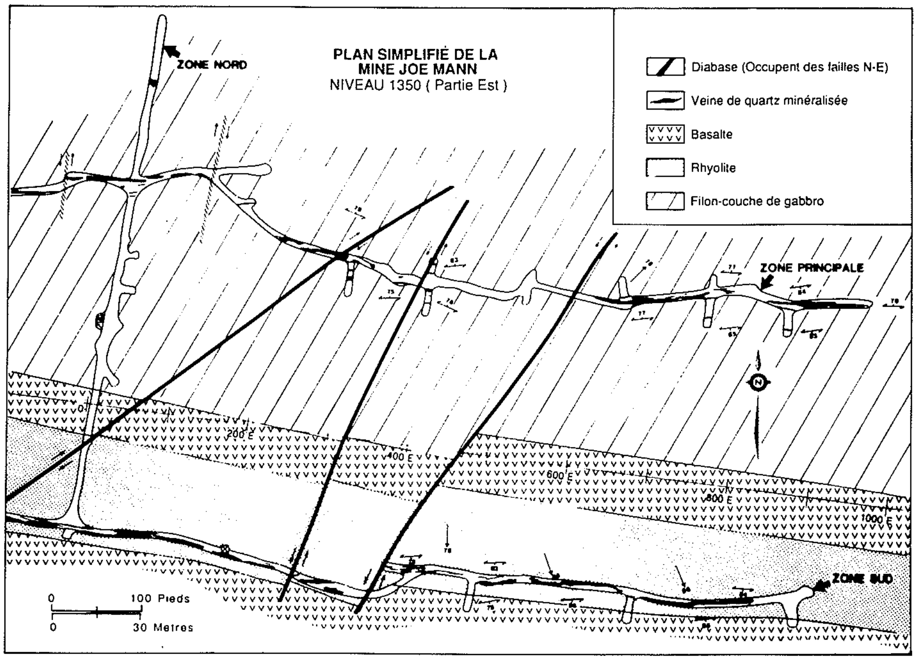

Figure 4A.1. Plan simplifié de la mine Joe Mann, niveau 1350. (Tiré de Dion et Guha, 1988).

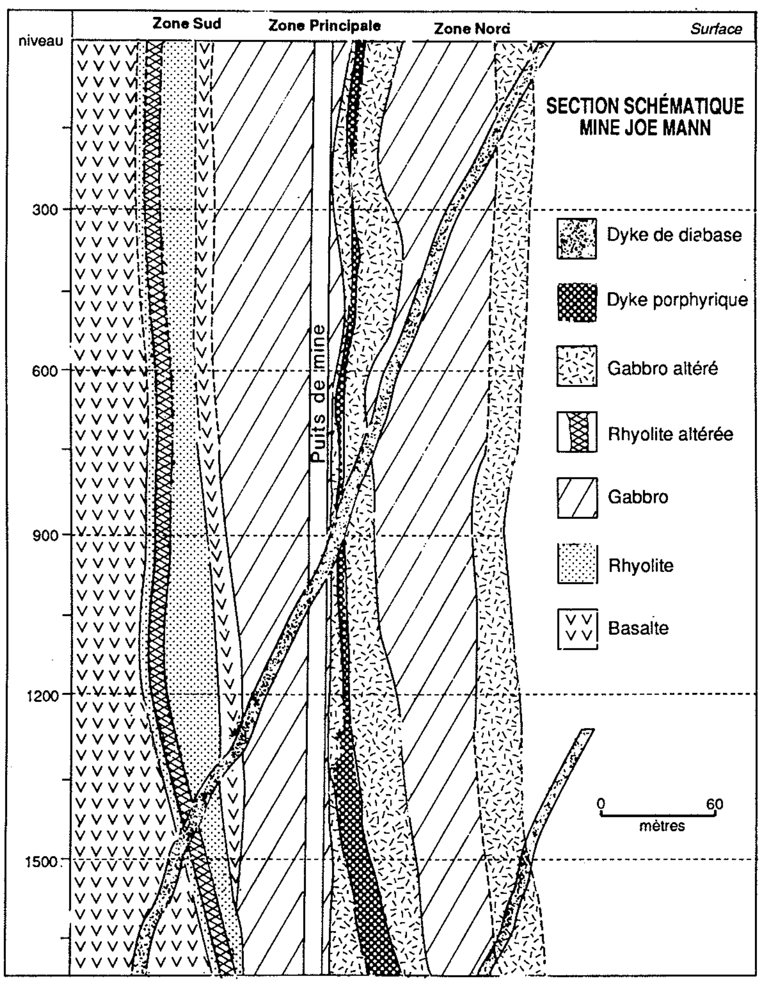

Figure 4A.2. Section transversale schématique de la mine Joe Mann. (Tiré de Dion et Guha, 1988).

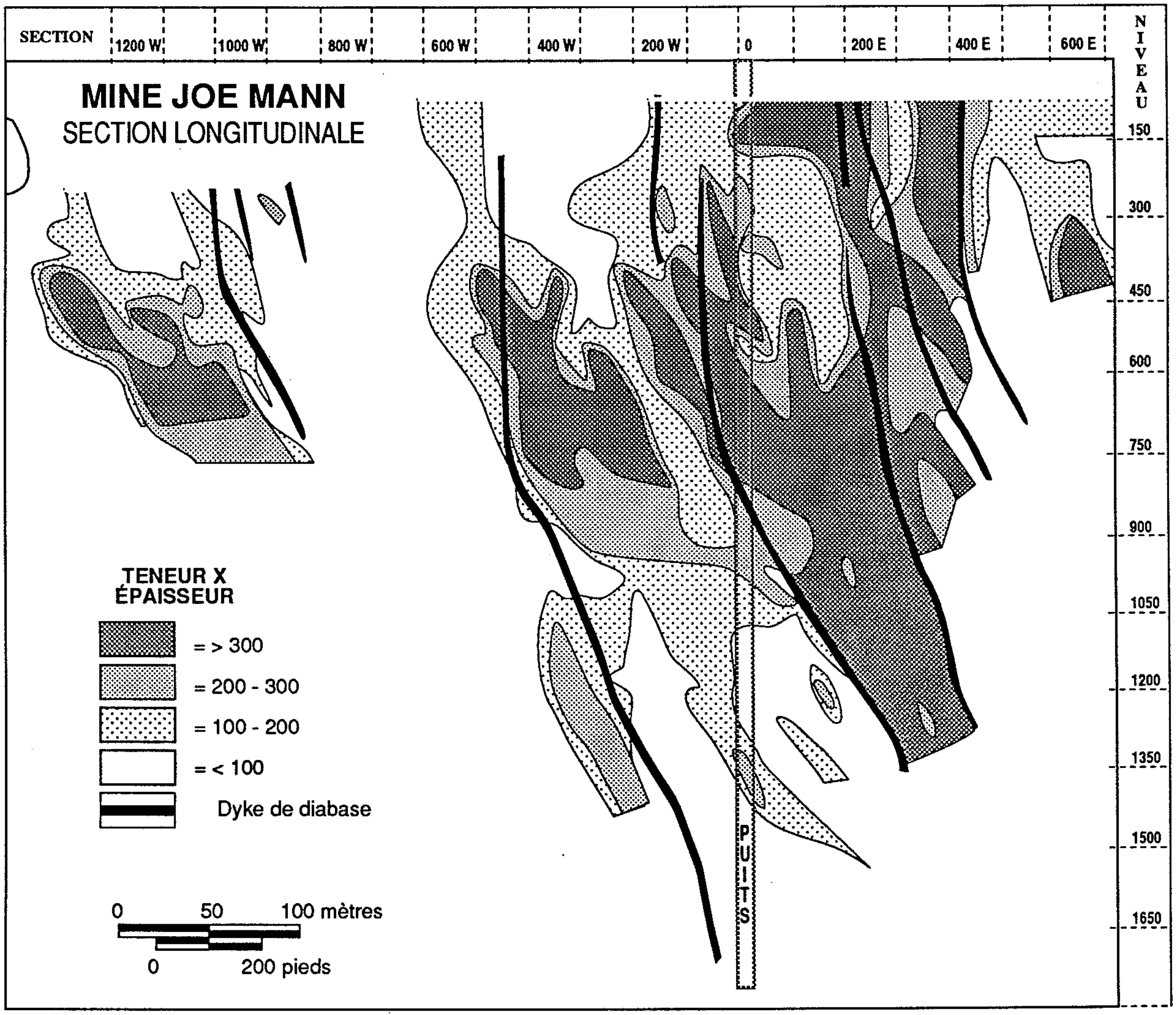

Figure 4A.3. Section longitudinale de la zone Principale, mine Joe Mann, avec les valeurs contourées de l'épaisseur de la zone minéralisée multipliée par les teneurs. (D'après les plans de la mine).

- sulfures ± carbonates ferrugineux ± albite ou en schiste à chlorite - carbonate ferrugineux - séricite - albite - sulfures. Les veines de quartz aurifères sont logées à l'intérieur de ce gabbro altéré, dans la partie centrale de la zone de cisaillement. Dans la zone Principale, les veines sont associées à deux variétés de dykes felsiques. Les dykes porphyriques, de couleur gris à rose, sont composés de 20 à 50% de phénocristaux de plagioclase idiomorphes baignant dans une matrice quartzo-feldspathique à grains fins. La déformation est typiquement faible et se limite généralement aux bordures des dykes. L'autre variété est typiquement aphyrique à microporphyrique, d'une couleur beige à rosée et se caractérise par une déformation moyenne à forte. Les deux types de dykes sont souvent directement en contact avec la veine et sub-parallèles à celle-ci. L'altération hydrothermale dans la zone fortement cisaillée se manifeste par un enrichissement en potassium, rubidium, barium, soufre, cuivre et or et un appauvrissement en calcium et magnésium.

La zone Sud occupe un cisaillement qui recoupe l'unité "rhyolitique" (en fait de composition dacitique). Il s'agit d'une roche massive, aphanitique, de couleur beige pâle à rosée, contenant moins de 10% de séricite, de biotite, de chlorite et de magnétite disséminées dans une matrice quartzo- feldspathique granoblastique. La zone la plus déformée montre une altération hydrothermale intense en séricite et sulfures. Cette altération correspond à un enrich- issement en silice, potassium, soufre, cuivre, arsenic et or et un appauvrissement en magnésium, calcium, sodium et carbone total (i.e. CO_2). Les veines ont une forme irrégulière et montrent un fort boudinage. La zone Sud est aussi la seule où l'on observe de l'arsénopyrite en quantité importante.

Les veines sont encaissées dans une roche fortement cisaillée, altérée et minéralisée en pyrite, pyrrhotine et chalcopyrite. Ces sulfures sont disposés en lentilles et veinules parallèles à la schistosité. Les veines sont composées essentiellement de quartz blanc vitreux avec un peu d'albite verte, de carbonate ferrugineux et d'une quantité variable mais généralement inférieure à 15% de sulfures. Ces veines se caractérisent par leur structure rubanée ou laminée définie par l'alternance de rubans de quartz et de roche encaissante. Elles montrent souvent une structure bréchiques et sont parfois plissées et boudinées. La majeure partie de la minéralisation en sulfures des veines proprement dites est contenu dans ces fragments d'éponte. La minéralisation aurifère, quant à elle, est concentrée essentiellement dans les veines de quartz. Celles-ci contiennent couramment de l'or visible en veinules ou en petites plages alignées dans les fractures recoupant les cristaux de quartz ou d'albite. Dans les épontes, l'or est aussi relié aux veinules de quartz. Les teneurs ne semblent pas être directement proportionnelles au pourcentage de sulfures.

Description des arrêts
Visite de la mine

Les contraintes de temps ne nous permettront qu'une visite de la zone Principale. De plus, en raison des travaux d'exploitation, il nous est très difficile de planifier un itinéraire à l'avance.
Points d'intérêts de la zone Principale:

- La zone de cisaillement E-W avec une schistosité de même direction. Notez la présence d'une linéation minérale subverticale.

- Les veines de quartz laminées à bréchiques, boudinées et parfois plissées associées à deux types de dykes felsiques.

- Le gabbro cisaillé, chloritisé et carbonaté, magnétique par endroit, transformé en schiste à biotite-sulfure ou en schiste à chlorite-carbonate-séricite à proximité des veines de quartz.

- La minéralisation en chalcopyrite- pyrrhotine-pyrite associée aux veines de quartz. Il est courant d'observer de l'or visible libre dans le quartz cataclasé.

- Les failles NE tardives matérialisées par les dykes de diabase protérozoïques. Ces failles recoupent la minéralisation avec un déplacement apparent senestre (vers le nord).

Visite de la tranchée à l'ouest de l'ancien chevalement Cette tranchée (Fig. 4A.4) d'orientation N-S met à jour la zone Principale et la zone Nord de la mine Joe Mann. La zone Sud se situerait de l'autre côté du chemin, au flanc de la petite colline. En cheminant vers le nord de la tranchée, on remarque d'abord la présence de basaltes coussinés qui passent bientôt à un gabbro "sain" lité à grains moyens à grossiers. La zone de cisaillement E-W située un peu au nord de la ligne de canton coïncide avec la structure contrôlant la zone Principale. La lentille minéralisée se situe cependant plus à l'est, près du chevalement. La texture et la minéralogie originale de la roche cisaillée sont complètement détruites. On passe progressivement d'un schiste à chlorite-carbonate à un schiste à biotite plus ou moins riche en sulfures. Les veines de quartz sont relativement rares mais on note encore la présence des deux variétés de dykes felsiques observées sous terre.

On revient ensuite au gabbro peu déformé du départ. La zone Nord est située à l'extrémité de la tranchée, à proximité du petit étang. Elle se caractérise surtout par son orientation qui varie de NW à E-W. On note aussi la présence de dykes de diabase de direction NE. Une structure NW très étroite subsidiaire au cisaillement principal E-W loge une veine de quartz montrant parfois de bonnes valeurs aurifères (on rapporte des échantillons titrant plus de 130 g/t).

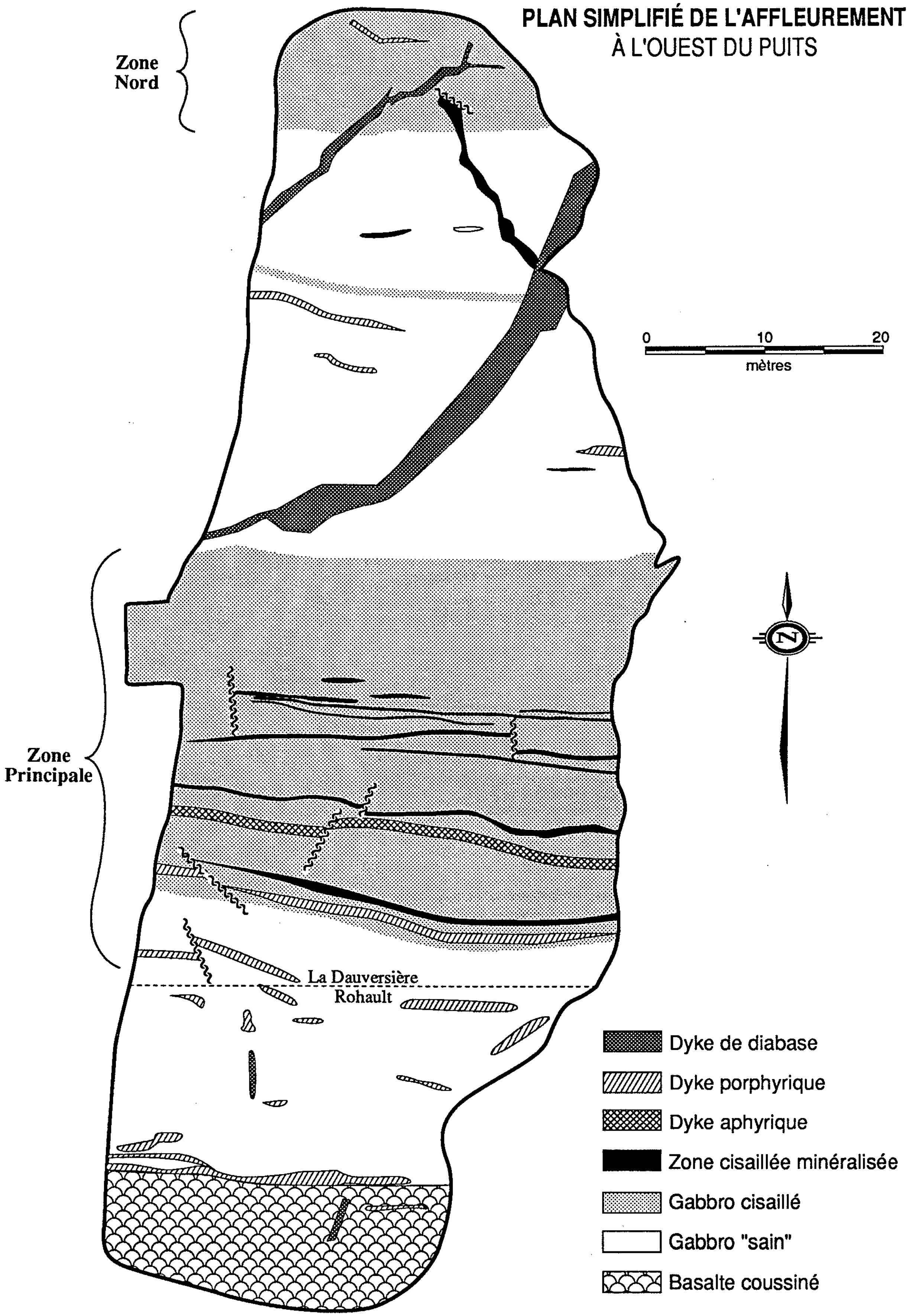

Figure 4A.4. Plan simplifié de la tranchée située à environ 120 m à l'Ouest de l'ancien puits. (Modifié de Hébert, 1980)

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JOUR / DAY 4B

MÉTALLOGÉNIE DE LA RÉGION DE CHAPAIS

Rémi Morin[1], Marc Boisvert[2]

Introduction

La division Opemiska de Minnova à Chapais est située dans le quart SW du canton de Lévy à 35 km au SW de Chibougamau (Fig. 4B.1). Les premiers indices minéralisés furent découverts en 1929 par Léo Springer, mais ce n'est qu'en 1954 que la production débuta avec l'ouverture du puits Springer suivi de Perry en 1958, Robitaille en 1969 et Cooke en 1977. La division Opemiska a produit, à ce jour, 23 534 942 t à 2,24% Cu et 1,17 g/t Au. Le contrôle structural (principalement les failles) à l'intérieur du Complexe de Cummings caractérise tous les dépôts de la division Opemiska à l'exception de la zone 8-5 de la mine Cooke.

Environnement géologique

Dans la région de Chapais, l'assemblage volcano-sédimentaire, d'âge archéen est divisé en deux groupes: le Groupe de Roy, à la base et le Groupe d'Opémisca, au sommet. Les roches volcano-sédimentaires sont envahies par plusieurs intrusions de composition variée. La plus importante est le Complexe de Cummings (Fig. 4B.2). Le Complexe de Cummings (Duquette 1976) comporte 3 filons-couches séparés mais génétiquement reliés: le filon-couche de Roberge à la base, surmonté par le filon- couche de Ventures puis le filon-couche de Bourbeau. Chapais, ce complexe comporte deux des trois filons-couches: le filon- couche de Ventures atteint une épaisseur de 940 m et le filon-couche de Bourbeau d'une épaisseur de 700 m. Dans le secteur sud de la mine Cooke, un important dyke de leucogabbro, d'une puissance de 200 m environ, recoupe les roches ultramafiques du Complexe de Cummings. Ce leucogabbro est constitué de phénocristaux idiomorphes de pyroxène (30-35%) altérés en amphibole et chlorite, dont la taille varie de 1 à 2 cm, qui baignent dans une matrice gloméroporphyrique de plagioclase saussuritisé. Les roches de la région de Chapais ont été déformées lors de l'orogenèse kénoréenne. Des plis isoclinaux orientés EW auxquels sont généralement associés une schistosité bien développée, donnent le grain tectonique régional. La région est également traversée par de nombreuses failles dont les principales sont: la faille Kapunapotagen d'orientation ESE, la faille "Chiboug Copper" d'orientation ENE et la faille Gwillim orientée NE.

Description de la mine Springer.

La mine Springer est situé à la limite NNW de la localité de Chapais. Le gabbro ophitique du filon-couche de Ventures constitue la roche encaissante de cette mine. La complexité du gisement réside dans le patron structural. Ainsi les zones minéralisées occupent des cassures EW à pendage vers le nord qui sont parallèles au plan axial d'un anticlinal synforme. Nous croyons que ce réseau de fractures a pu se développer lors de la formation des plis régionaux (déformation de charnière). Ceci pourrait être dû à la fragilité des phases grossièrement cristallisées et à la texture ophitique du filon-couche de Ventures (Watkins et Riverin 1982).

La veine principale (veine #3) était la plus étendue avec une longueur de 900 m, une largeur de 6 m et une extension en profondeur supérieure à 1 000 m. On y a extrait 6 491 793 t à 2,61% Cu et 0,69 g/t Au. La veine #7 avec une dimension de 606 m de long, 2,4 m de largeur et une extension en profondeur de 1 000 m est la seconde en importance; on y a extrait 616 320 t à 1,88% Cu et 2,37 g/t Au. Outre ces deux veines, une quinzaine d'autre sont été exploitées à ce jour, pour un total pour la mine Springer de 12 468 000 t à 2,56% Cu et 1,23 g/t Au.

La composition minéralogique des veines EW est caractérisée par la présence de chalcopyrite et dans une moindre part de pyrite et de pyrrhotite avec des

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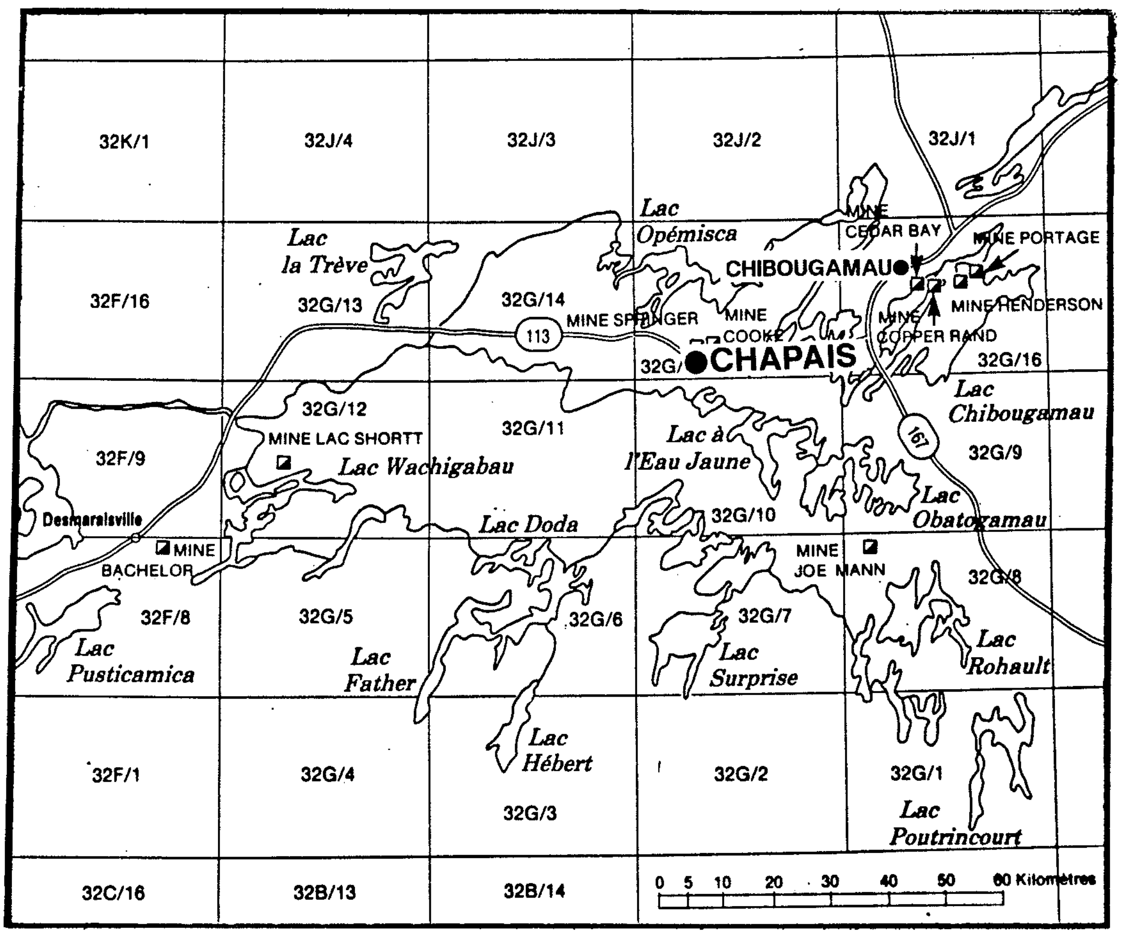

Figure 4B.1 Localisation de la région de Chapais

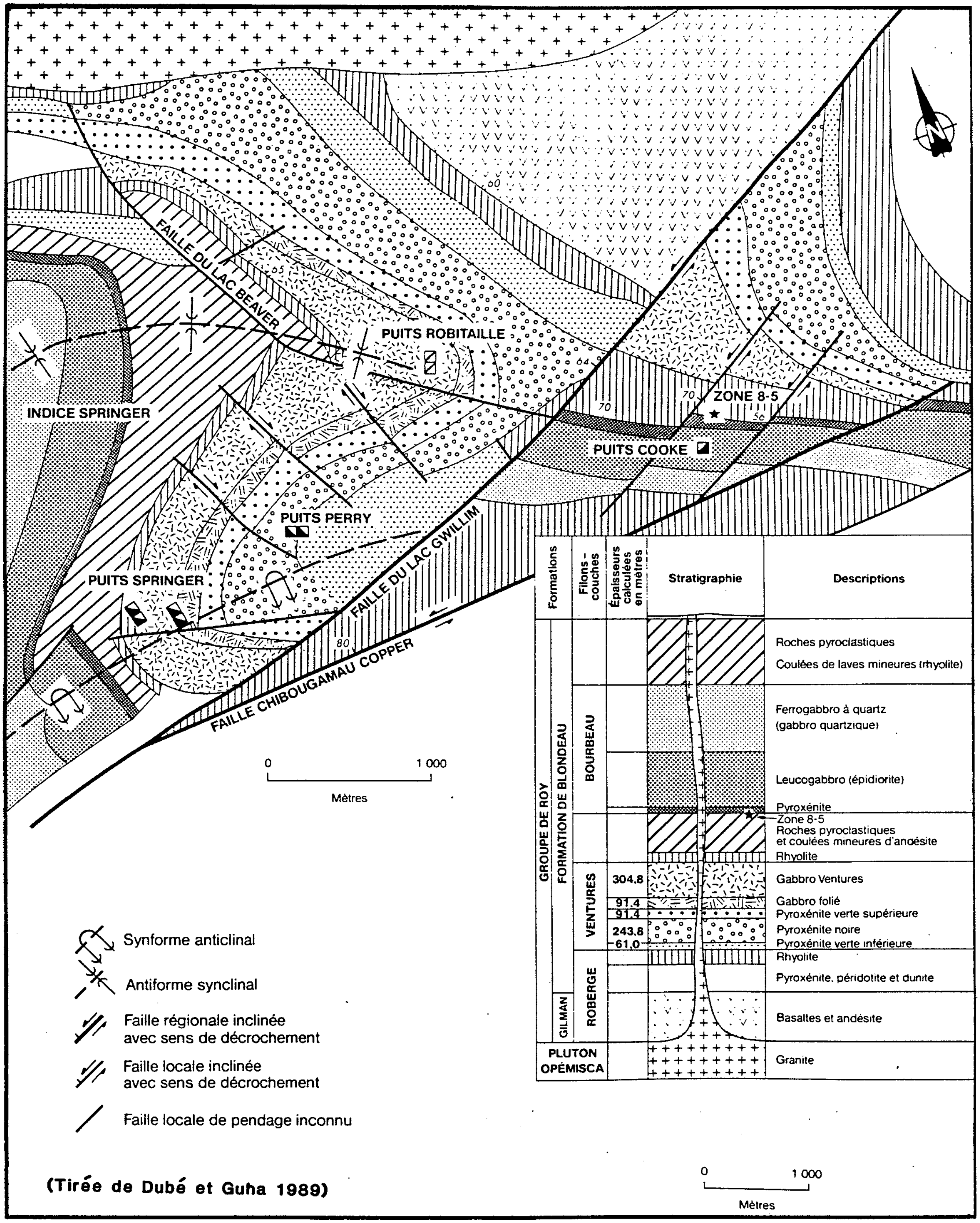

Figure 4B.2 Géologie de la région de Chapais quantités mineures de sphalérite, galène, molybdénite et arsénopyrite. Le quartz est le minéral principal de la gangue. La veine contient également de la calcite et de faibles quantités de biotite, actinote, chlorite et axinite.

Mine Perry.

La mine Perry est situé à 400 m à l'est de la mine Springer. L'environnement géologique de la mine Perry est constitué principalement par le gabbro ophitique du filon-couche de Ventures et, dans une moindre part, par le gabbro folié du même filon-couche. Les zones minéralisées occupent une série de fractures parallèles orientées NNW à pendage vers le NE. Elles sont localisées sur le flanc nord d'un anticlinal synforme où elles sont suborthogonal à la trace axiale du pli. Nous croyons que ces fractures sont associées à la formation des plis régionaux avec déformation de flancs et à la fragilité des phases grossièrement cristallisées du filon-couche de Ventures.

Les zones B et D exploitées étaient les plus étendues avec une longueur de 455 m, une largeur de 12 m et une extension de 600 m de profondeur pour la zone B et une dimension de 330 m de long, 7,6 m de largeur et une extension de 750 m de profondeur pour la zone D. On y a extrait respectivement 3 309 103 t à 2,45% Cu et 0,34 g/t et 1 870 742 t à 2,05% Cu et 0,28 g/t Au. La production totale de la mine Perry à ce jour, s'élève à 8 890 720 t à 2,16% Cu et 0,24 g/t Au. La composition minéralogique des veines est similaire à celle de la mine Springer.

Mine Robitaille.

La mine Robitaille est située à 2,4 km au NE de la localité de Chapais. L'environnement géologique et le type de minéralisation sont similaires à ceux des mines Springer et Perry. On a extrait (dans la zone de "Beaver Lake") 203 037 t à 1,87% Cu et 0,21 g/t Au.

Mine Cooke.

La mine Cooke est située à 2,8 km à l'est de Chapais. L'environnement géologique de la mine Cooke est occupé principalement par le leucogabbro du filon-couche de Bourbeau et, dans une moindre mesure, par le ferrogabbro à quartz du même filon-couche. Deux zones minéralisées principales (veines 7 et 9) ont été exploitées à la mine Cooke. Ces deux veines (Dubé et Guha, 1986) correspondent à un ensemble de plusieurs zones de cisaillement minéralisées d'orientation EW à NW. La largeur du cisaillement varie généralement de 2 à 3 m.

Les veines renferment de la chalcopyrite avec des proportions beaucoup moindres de pyrrhotine, d'arsénopyrite et de pyrite (Dubé et Guha, 1986). La gangue est constituée de quartz et de carbonate. La mine Cooke diffère des mines à l'ouest de la faille Gwillim (Springer, Perry, Robitaille) par une teneur plus faible en chalcopyrite et des teneurs plus élevées en or et arsénopyrite. On y a extrait depuis ses débuts en 1976, 1973 188 t à 0,66% Cu et 5,04 g/t en Au. La zone 8-5, pour sa part, est une petite masse de sulfures massifs localisée dans les volcanoclastites de la Formation de Blondeau à la base du filon-couche de Bourbeau. Elle contient de la pyrrhotine, de la chalcopyrite, de la sphalérite et, dans une moindre part, de la pyrite, de la têtra édrite et des traces de galène et de linnadite (Bélanger et al., 1984).

Arrêt No 1.

Veines #1 et #2 de la mine Springer à l'intérieur du filon-couche de Ventures (Complexe de Cummings).
Localisation: 200 m au sud-ouest du puits Springer

Cette série d'affleurements (Fig. 4B.3) nouspermet d'observer deux veines minéralisées (veine #1 et veine #2) exploitées à la mine Springer. La roche encaissante est le gabbro ophitique du filon-couche de Ventures. Ce gabbro est constitué de plagioclase, de clinopyroxène et de magnétite titanifère et se distingue facilement par sa texture ophitique à grains grossiers de plagioclase.

Les veines #1 et #2 orientées EW et à pendage abrupt vers le nord sont parallèles à la veine 3 sise plus au sud. Nous pouvons observer à l'intérieur de ces veines des minéralisations de chalcopyrite, de pyrite et de pyrrhotine. Le quartz et la calcite constituent les principaux minéraux de la gangue.

Arrêt NO. 2

Zone de cisaillement typique de la mine Cooke
Localisation: 1 km à l'est de la mine Cooke. De Chapais, suivre la route 113 vers l'est sur 5,5 km, tourner à gauche (nord) sur la route secondaire et parcourir une distance de 2 km. Affleurement à l'ouest de la route.

Cette série d'affleurements (Fig. 4B.4) nous permet d'observer le passage entre le leucogabbro et le ferrogabbro à quartz du filon-couche de Bourbeau présent à la mine Cooke

Le leucogabbro est massif, de couleur blanchâtre, tacheté de vert et possédant une texture subophitique. Il renferme principalement en proportion équivalente du pyroxène et du plagioclase et, dans une moindre part, de la chlorite et du leucoxène. Le ferrogabbro à quartz est de couleur noire tacheté de blanc et sa texture est subophitique à panidiomorphique. Il est composé de pyroxène et de plagioclase, de 5 à 10% de grains de quartz arrondis et de leucoxène.

Outres ces deux lithologies, des dykes de leucogabbro à phénocristaux idiomorphes de pyroxène recoupent les deux gabbros du filon-couche de Bourbeau

Côté structural, ces affleurements permettent d'observer des cisaillements EW à NW typiques de la mine Cooke. Les zones de cisaillement ont une puissance qui varie de 2 à 3 mètres.

Remerciements

Nous tenons à remercier la division Opemiska de Minnova à Chapais pour leur excellente collaboration.

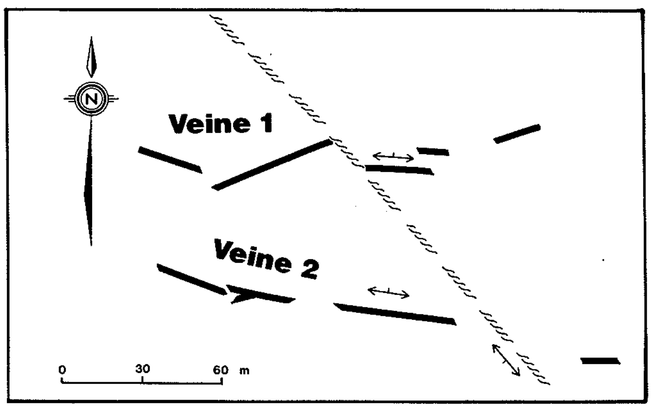

Figure 4B.3 Arrêt #1, veines 1 et 2 du puits Springer

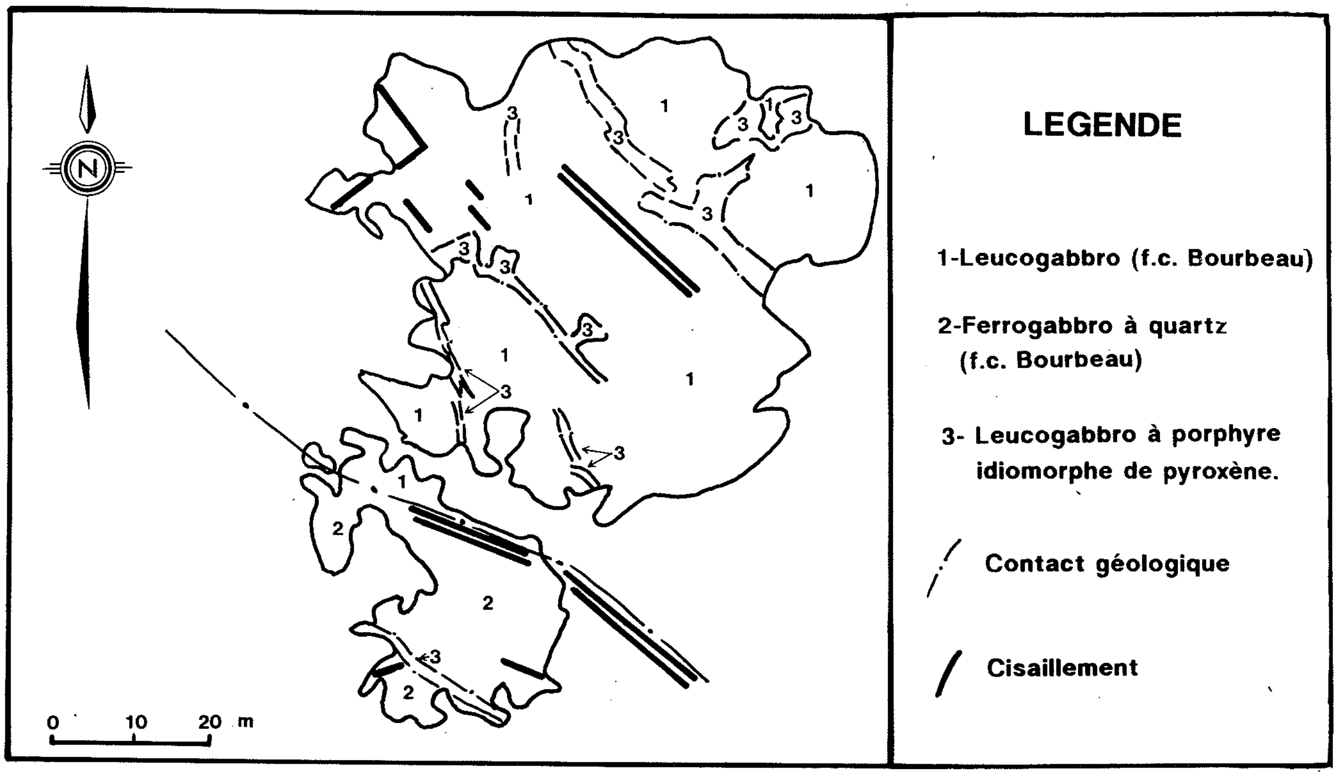

Figure 4B.4 Cisaillement de la Mine Cooke.

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INDICES AURIFERES PRES DE LA MINE DU LAC SHORTT, ABITIBI, QUÉBEC

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INTRODUCTION

L'excursion que nous proposons permettra de voir quelques-unes des unités lithologiques de la région ainsi que quelques indices aurifères dont le contexte de minéralisation est différent de celui de la mine du lac Shortt.

LOCALISATION

La région du Lac Shortt est située à quelques 120 km au sud-ouest de Chibougamau. Cette région fait partie de la Province de Supérieur du Bouclier Canadien et elle est située dans la portion nord de la Ceinture de roches vertes de l'Abitibi. Plus spécifiquement, la région du lac Shortt constitue la partie occidental du segment de Caopatina- Desmaraisville (Fig. 1).

GÉOLOGIE RÉGIONALE

La géologie de la région que nous visitons est connue par les travaux de Shaw (1940), Lamothe (1981 et 1983), Sharma et Lacoste (1981) et Giovenazzo (1983 et 1986). Récemment, Sharma et Gobeil (1987) ont révisé la géologie de l'ensemble de la région (Fig. 5A.1) et ont proposé un nouveau schéma stratigraphique global.

Lithostratigraphie

Le schéma stratigraphique de Sharma et Gobeil (1987) divise les roches archéennes de la région en deux grandes unités lithostratigraphiques: la Formation d'Obatogamau (Cimon, 1976), surmontée par la Formation du ruisseau Dalime. La Formation d'Obatogamau consiste en une succession de roches de plusieurs kilomètres d'épaisseur formée de basaltes massifs, coussinés et bréchiques communément porphyriques à phénocristaux de plagioclase. Cette formation inclut une unité de roches pyroclastiques et de rhyolites de moins de 1 km d'épaisseur appelée le Membre de Wachigabau. Une unité de rhyodacite située sur l'île Opawica est aussi stratigraphiquement en dessous de la Formation du ruisseau Dalime. La Formation du ruisseau Dalime, au sommet de la séquence stratigraphique, est formée de roches pyroclastiques et de roches sédimentaires volcanogènes.

Intrusions

Les intrusions archéennes majeures sont le complexe anorthosique de la rivière Opawica, le complexe mafique et ultramafique des Chutes de l'Esturgeon situé au nord du lac Shortt (Fig. 5A.1) ainsi que des plutons granitoïdes. Le complexe des Chutes de l'Esturgeon (Lamothe, 1983) a une épaisseur d'au moins 1350 m et est formé de plusieurs intrusions stratiformes différenciées comprenant une pyroxénite ou une péridotite à la base qui passe graduellement vers le sommet à un gabbro quartzifère.

Métamorphisme et structure

Les roches archéennes de la région sont plissées et généralement métamorphisées au faciès des schistes verts. Régionalement on rencontre plusieurs plis dont la trace de la surface axiale est orientée entre NE et E. Près du lac Shortt, les plis sont pour la plupart isoclinaux à trace axiale ENE avec des intervalles interaxiaux de 2 à 4 km.

Dans l'ensemble de la région on renconnaît une famille de failles majeures de direction ENE à E. Les failles Opawica et du lac Shortt en sont des exemples qui se manifestent par de la roche cisaillée et localement carbonatée. Une seconde famille de failles, plus tardives, de direction NNE à NE est aussi rencontrée régionalement. De plus, on trouve localement quelques failles de directions NW.

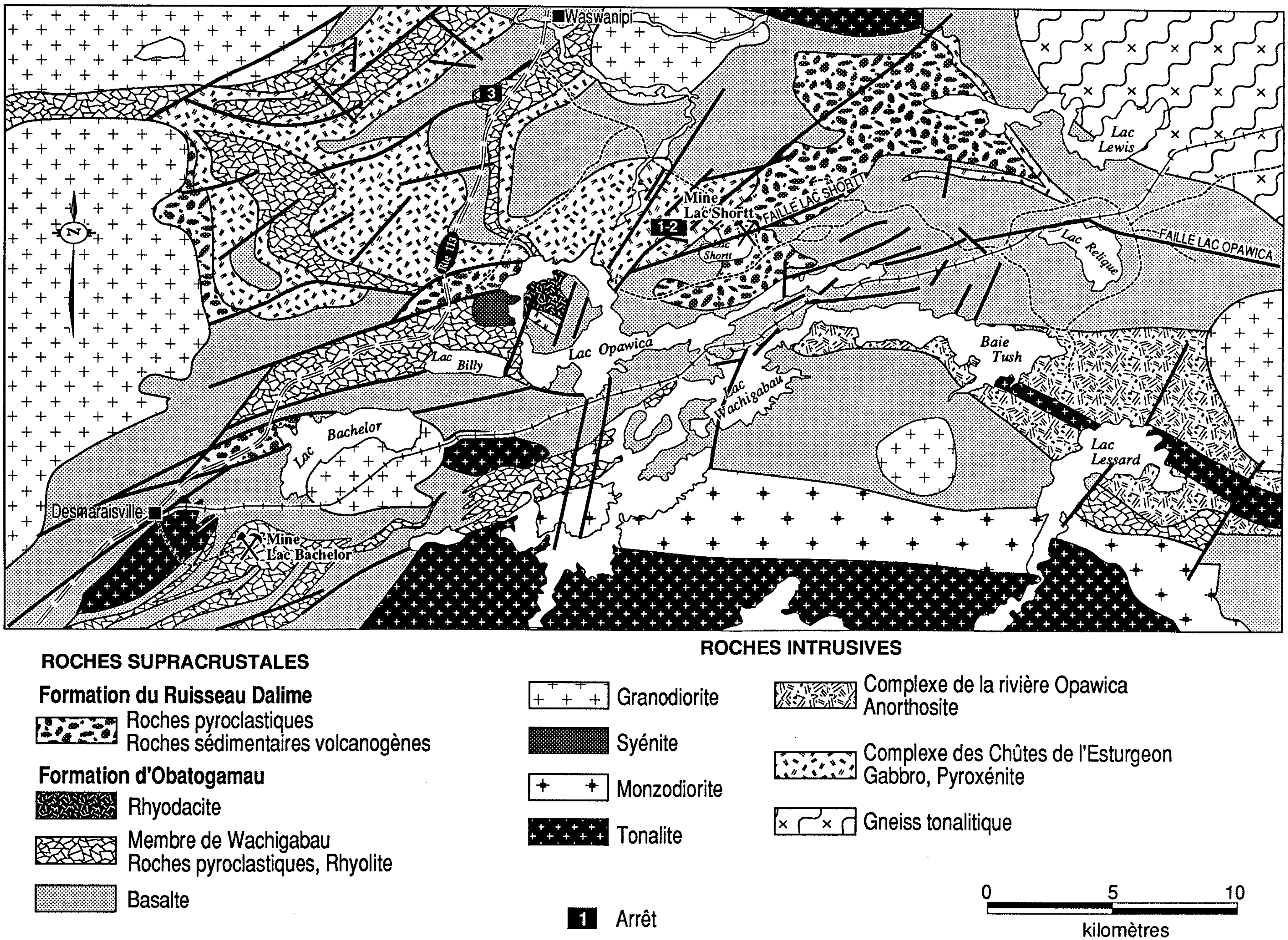

Figure 5A.1. Géologie de la région du lac Shortt et localisation des arrêts.

Une schistosité subverticale dont l'attitude varie entre N70° et N90° est omniprésente dans la région et correspond assez bien au plan axial des plis ENE à E. On rencontre parfois un clivage de crénulation N20° et on rapporte la présence locale d'un clivage antérieur à la schistosité dominante.

Gitôlogie

La région renferme le gîte d'or du lac Shortt, le gîte d'or du lac Bachelor ainsi qu'une multitude d'indices aurifères. Des travaux de métallogénie (Brisson et Guha, 1988 et sous presse) ont permis de proposer une classification qui divise en trois types les minéralisations aurifères de la région du lac Shortt (Brisson et Guha 1989): (1) «filon avec quartz et dissémination en association avec des failles fragiles et fragiles- ductiles», (2) «filon avec quartz en voûte anticlinale et dissémination associée», et (3) «zone d'altération et de mylonitisation». Les roches hôtes des minéralisations sont des roches volcaniques, volcanoclastiques, sédimentaires et intrusives. Les gîtes du lac Shortt et du lac Bachelor font partie du type «zone d'altération et de mylonitisation» et montrent tous les deux: (1) un contrôle structural par des couloirs de déformation de direction ENE, (2) des assemblages d'altération aurifère semblables (carbonates - hématite - feldspath potassique - pyrite), et (3) une association avec des intrusions (Morasse, 1988; Lauzière, 1989; Quirion, ce volume).

VISITES

L'excursion propose de visiter trois indices aurifères situés près du lac Shortt (Fig. 5A.1). Les deux premiers arrêts sont accessibles en empruntant vers l'ouest un chemin qui croise la route de la mine du lac Shortt à environ 2 km au nord de la mine. L'arrêt 1 est situé juste à coté de ce chemin à environ 1 km de l'intersection avec la route de la mine. Sur le même chemin, à environ 500 m plus à l'ouest de l'arrêt 1, on rencontre un embranchement de direction nord qui mène à son bout à l'arrêt 2 (environ 500 m). L'arrêt Non 3 est accessible en marchant environ 500 m dans un chemin situé du côté ouest de la route 113 à quelques 4 km au sud de Waswanipi.

Arrêt 1

Cet indice constitue un exemple de minéralisation aurifère de type «filon avec quartz et dissémination en association avec des failles fragiles et fragiles-ductiles» (Brisson et Guha, 1989).

L'affleurement nous met en présence d'un gabbro du complexe des Chutes de l'Esturgeon qui est injecté par des dykes de composition intermédiaire à phénocristaux d'amphibole. Les roches sont recoupées par un réseau de veinules et veines de quartz. Le quartz se retrouve dans des fractures subverticales, plutôt régulières, de directions E et NE et dans les fractures irrégulières d'un couloir bréchique de direction approximative E. Une altération plutôt restreinte accompagne les veinules et veines de quartz. Cette altération est révélée par la présence de carbonates de fer et localement de pyrite. L'or se retrouve dans les veines de quartz et dans ses épontes contenant des carbonates et de la pyrite.

Arrêt 2

Cet indice est un autre exemple de minéralisation aurifère de type «filon avec quartz et dissémination en association avec des failles fragiles et fragiles-ductiles» (Brisson et Guha, 1989). L'affleurement (Fig. 5A.2) montre essentiellement la lave massive, coussinée et bréchique de la Formation d'Obatogamau. La lave est intercalée d'une unité de roches sédimentaires siliceuses et ferrugineuses perturbées par des plis syn-sédimentaires (slumping). Toutes ces roches sont injectées par quelques dykes de composition intermédiaire à phénocristaux d'amphibole. Cet assemblage de roches est recoupé par des veines subverticales de quartz de directions N90°, N60°, N45° et N10°. Le quartz occupe des fractures le long desquelles la roche n'est pas foliée. Une figure depincement et ouverture suggère un sens de mouvement senestre le long d'une veine de direction N60°. Les roches sont altérées de façon restreinte le long des veines et cette altération se manifeste surtout par la présence de carbonate de fer avec localement de la pyrite. L'or est parfois visible dans le quartz. Des valeurs anomales en or sont aussi obtenues dans la partie pyritisée de l'éponte des veines.

Arrêt 3

Cet indice représente la minéralisation aurifère de type «filon avec quartz en voûte anticlinale et dissémination associée» de Brisson et Guha (1989). L'affleurement expose une interstratification de roches volcanoclastiques de composition felsique et de laves de

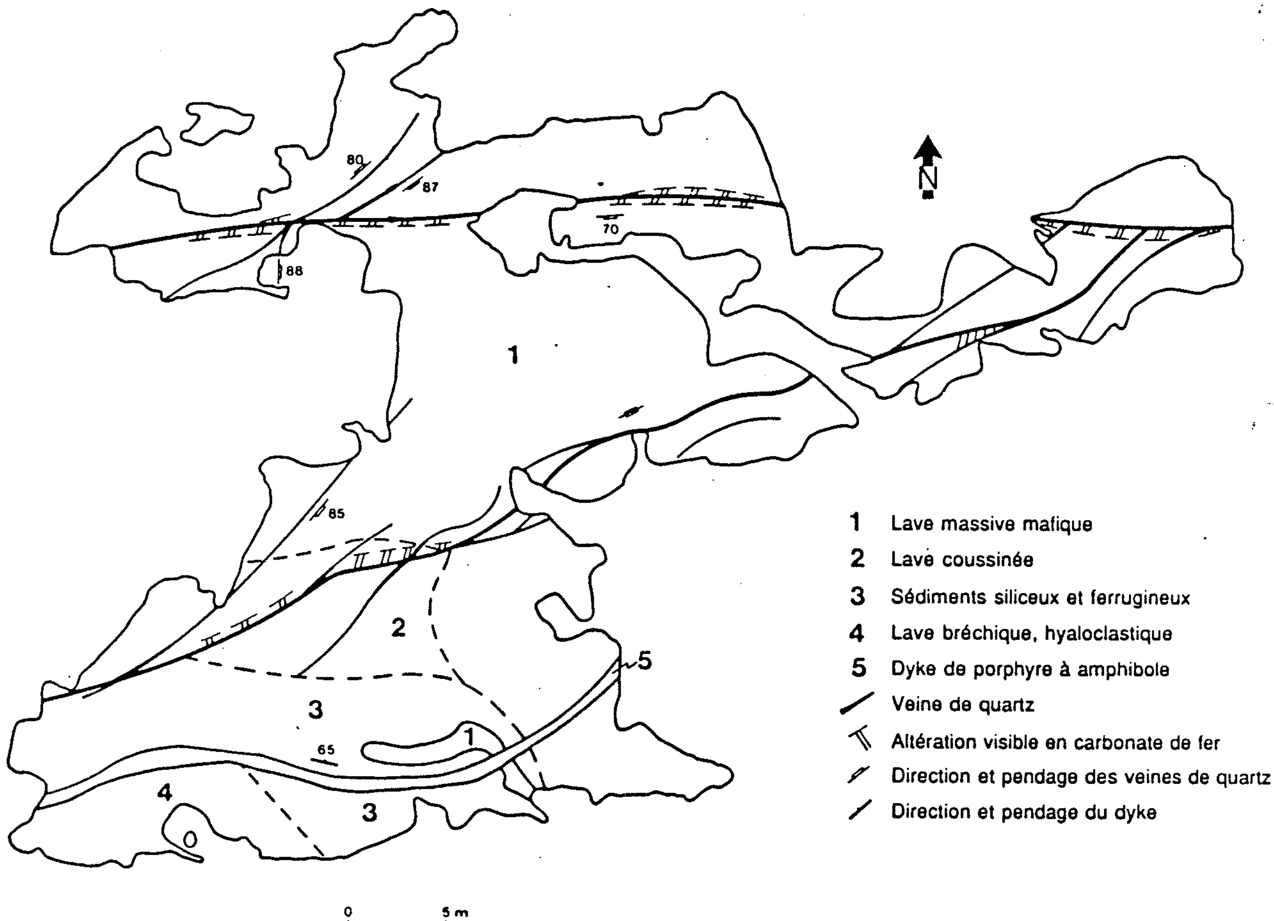

Figure 5A.2. Arrêt 2: géologie

composition mafique qui font partie du Membre de Wachigabau. L'affleurement se situe dans la charnière d'un anticlinal antiforme de direction E et présente des couches à faibles pendages qui sont marquées par la schistosité régionale subverticale de direction E.

L'indice présente plusieurs veinules et veines de quartz qui recoupent la roche et une veine principale de quartz située entre les couches formant une voûte. Les veines sont accompagnées d'une altération qui s'exprime par la présence de carbonates de fer, de séricite, de chlorite et localement de pyrite. Les valeurs anomales en or sont retrouvées dans les veines et dans la roche pyriteuse dans les épontes des veines.

REMERCIEMENTS

Nous remercions les dirigeants et les géologuesdes compagnies Falconbridge Ltée et Minnova Inc. grâce à qui ces visites sont rendues possibles.

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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

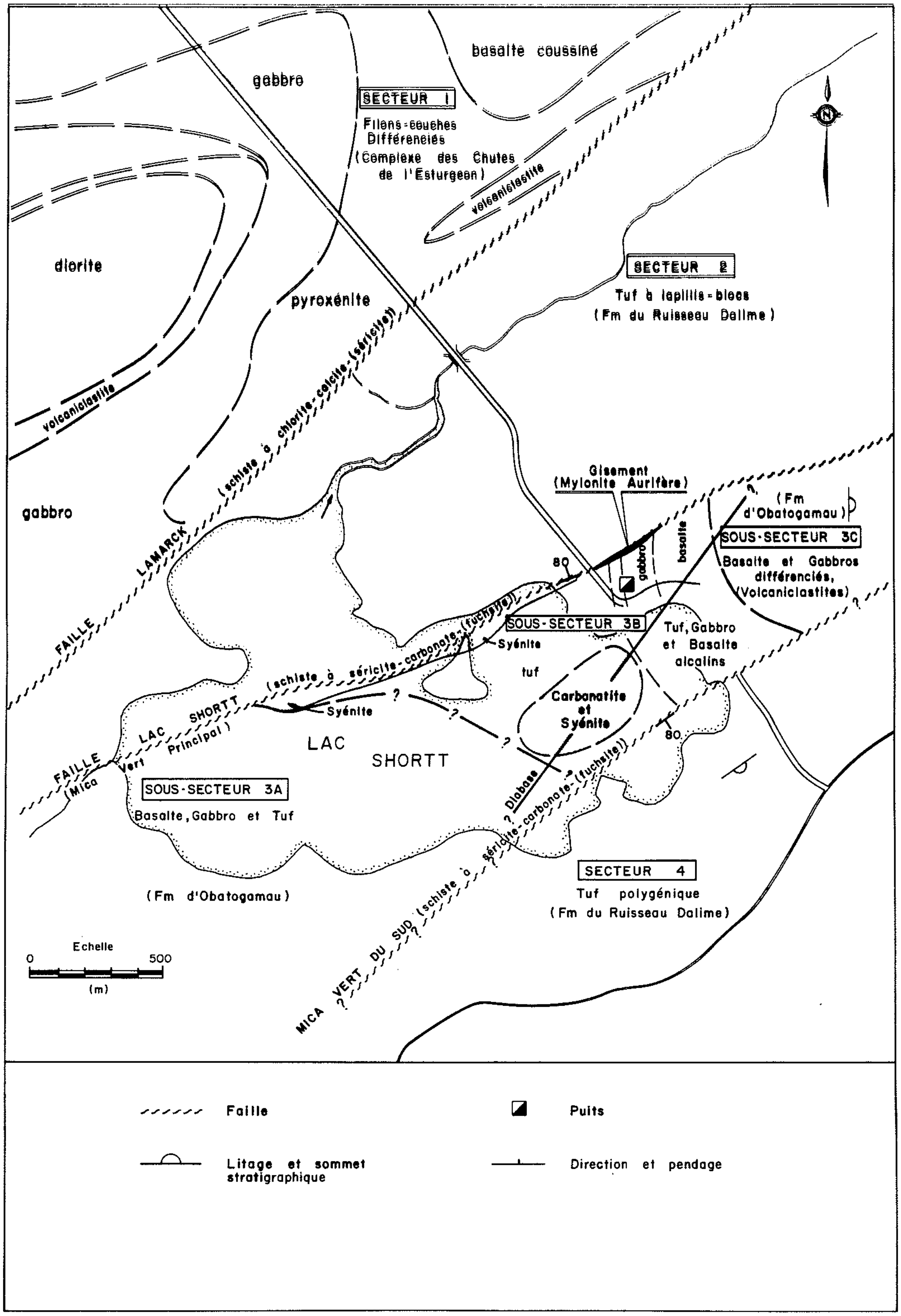

Figure 5B.1. Géologie de la propriété Lac Shortt. Les failles Lamarck, Lac Shortt et le Mica Vert du Sud divisent le territoire en 4 secteurs distincts.

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 glomérocristaux 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
	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
1	106-2	47.90	0.63	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
COMPLEXE DES CHUTES	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
DE L'ESTURGEON	S-21	50.30	0.61	3.58	11.40	0.21	15.10	17.40	0.72	0.07	0.06	0.18	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	0.16	0.08	49
	S-24	46.30	0.56	13.60	10.00	0.18	10.30	11.70	1.20	0.23	0.17	0.39	43
2	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	0.13	5.76	4.80	4.22	1.85	0.33	0.13	116
TUF (FM DU	2166	51.00	1.02	15.60	11.90	0.14	3.07	4.28	2.79	1.53	0.39	2.89	121
RUISSEAU DALIME)	3141	55.30	0.73	15.10	6.93	0.12	2.64	6.08	2.65	1.99	0.35	4.20	147
3	103-1	51.90	1.90	13.50	12.20	0.24	4.01	7.64	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	3.92	0.60	0.41	1.23	172
3A OUEST DU GISEMENT	111-1	54.50	0.80	12.60	8.65	0.19	3.95	6.04	3.98	1.06	0.21	4.02	98
	111-2	46.20	2.68	12.20	16.30	0.29	4.51	5.46	3.52	0.48	0.10	3.75	86
	05	55.60	0.48	14.10	5.08	0.13	1.86	7.63	6.17	3.26	0.17	2.95	< 1
	11-AE	48.85	0.98	15.80	10.39	0.17	3.61	5.36	5.80	2.62	0.15	3.12	25
	11-01	45.80	1.65	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	4.36	0.18	0.12	6.88	< 1
3B EPONTE INFERIEURE	29	46.10	1.52	12.30	16.40	0.25	3.52	5.51	6.11	1.49	0.39	4.08	97
DU GISEMENT	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	6.68	4.36	1.66	0.43	1.28	68
	3235-8	50.40	0.68	14.60	7.14	0.25	3.96	7.30	6.70	3.65	0.44	4.79	75
	3149-2	46.30	0.56	14.90	11.40	0.19	9.06	11.10	1.48	0.42	0.12	2.51	28
	3149-3	47.30	0.64	14.30	11.90	0.21	7.60	11.30	2.65	0.16	0.17	0.93	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
3C EST DU GISEMENT	3149-5	50.60	0.88	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	8.11	50
	3149-7	58.20	0.60	16.90	6.47	0.05	3.23	3.71	4.66	0.46	0.29	0.44	120
	3149-8	50.20	1.04	13.70	14.90	0.24	6.07	7.86	2.80	0.12	0.29	0.06	57
4	3147-4	57.30	0.48	13.30	6.98	0.21	2.22	7.28	3.53	1.43	0.30	5.02	116
TUF POLYGENIQUE (FM	3153-3	62.10	0.28	15.60	4.80	0.09	1.23	2.80	3.27	2.56	0.21	2.71	122
DU RUISSEAU DALIME)													
	03	51.90	0.65	13.70	9.30	0.21	1.62	3.63	0.20	11.80	0.12	5.52	1470
	07	50.00	0.16	17.50	7.37	0.18	0.20	6.96	0.94	10.90	< 0.01	5.43	3903
SYENITES INTRUSIVES	15	47.70	0.93	11.10	10.30	0.32	2.08	6.64	0.53	8.76	0.55	9.92	395
(SOUS-SECTEUR 3B)	09	53.80	0.12	21.60	2.87	0.07	0.40	2.85	0.65	12.70	0.01	2.15	2689
	3142-1	48.30	0.11	15.40	5.06	0.24	0.38	9.89	0.72	11.20	0.28	8.29	1872
	3235-3	51.50	0.20	17.30	3.20	0.16	0.87	7.24	1.86	9.19	0.18	6.31	1459
	11-BD	53.30	0.48	15.70	5.31	0.17	1.30	5.19	5.80	4.99	0.08	6.24	34
	11-C	55.50	0.13	19.60	2.38	0.05	0.29	2.58	2.10	12.00	0.10	2.04	1695
FENITES	3175-AG	44.35	0.48	12.70	5.68	15.68	3.46	11.95	2.05	7.80	0.37	8.65	134
(SOUS-SECTEUR 3B)	3175-BF	48.25	0.52	15.25	5.70	0.25	2.54	8.44	1.96	9.12	0.47	6.05	517
	3175-CE	52.40	0.24	20.40	4.18	0.12	1.63	5.05	1.70	11.15	0.09	3.41	1370
	3175-D	53.40	0.08	20.30	1.20	0.10	0.29	5.00	0.69	12.10	< 0.01	3.09	367
MYLONITE AURIFERE	20	37.00	1.40	10.60	13.80	0.34	2.57	7.85	0.25	8.94	0.17	11.20	51
(SOUS-SECTEUR 3B)	3166-2	37.90	1.43	10.40	15.50	0.32	3.08	8.14	1.58	6.48	0.19	15.48	84

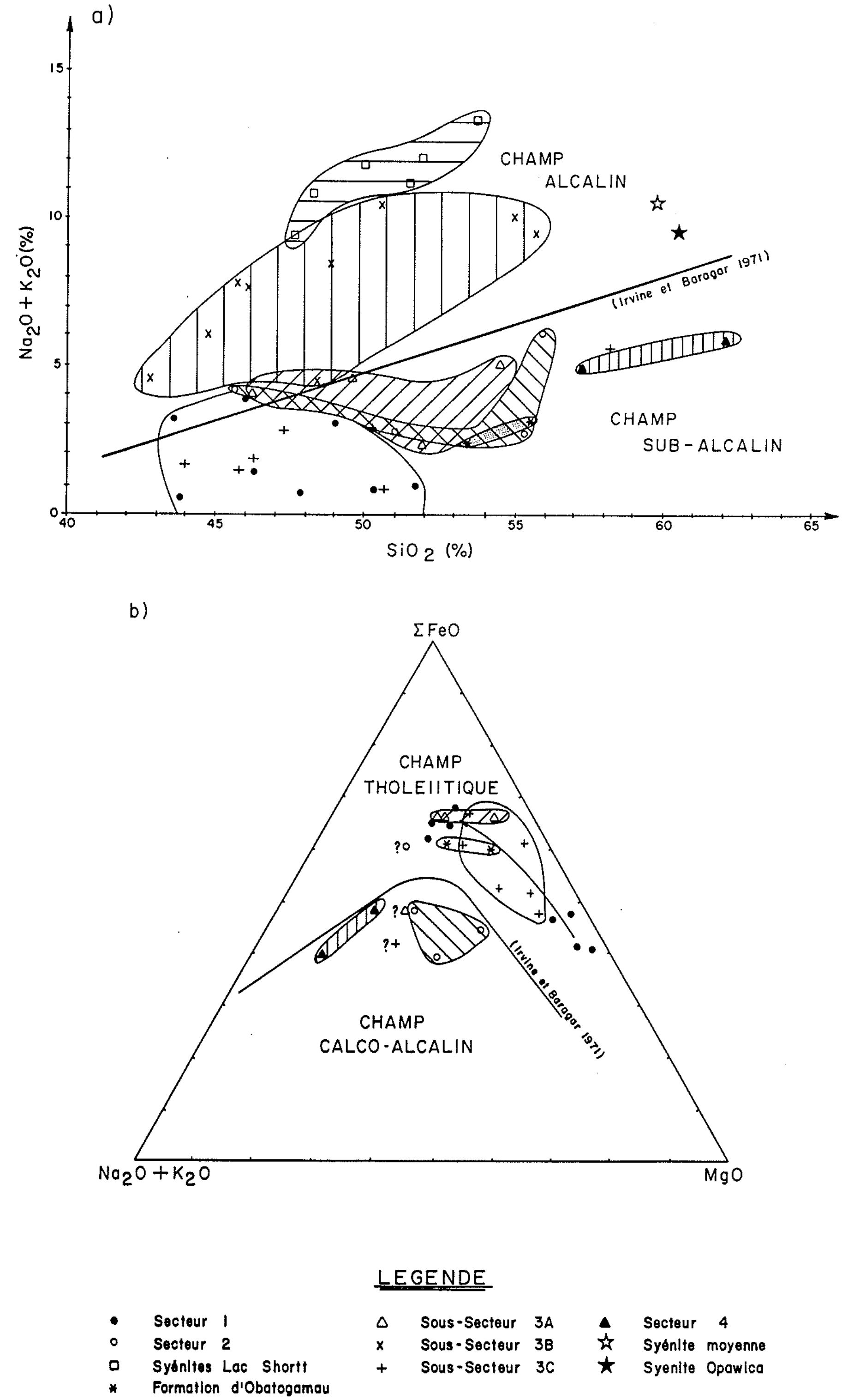

Figure 5B.2. Géochimie des roches de la mine Lac Shortt a) Diagramme alcalis versus Si02 et b) Diagramme AFM. Echantillons de la Formation d'Obatogamau tirés de Ludden et al., 1984; Syénite moyenne d'après Le Maître, 1976; Syénite Opawica d'après une moyenne de 21 syénites tiré de Averill et al., 1989.

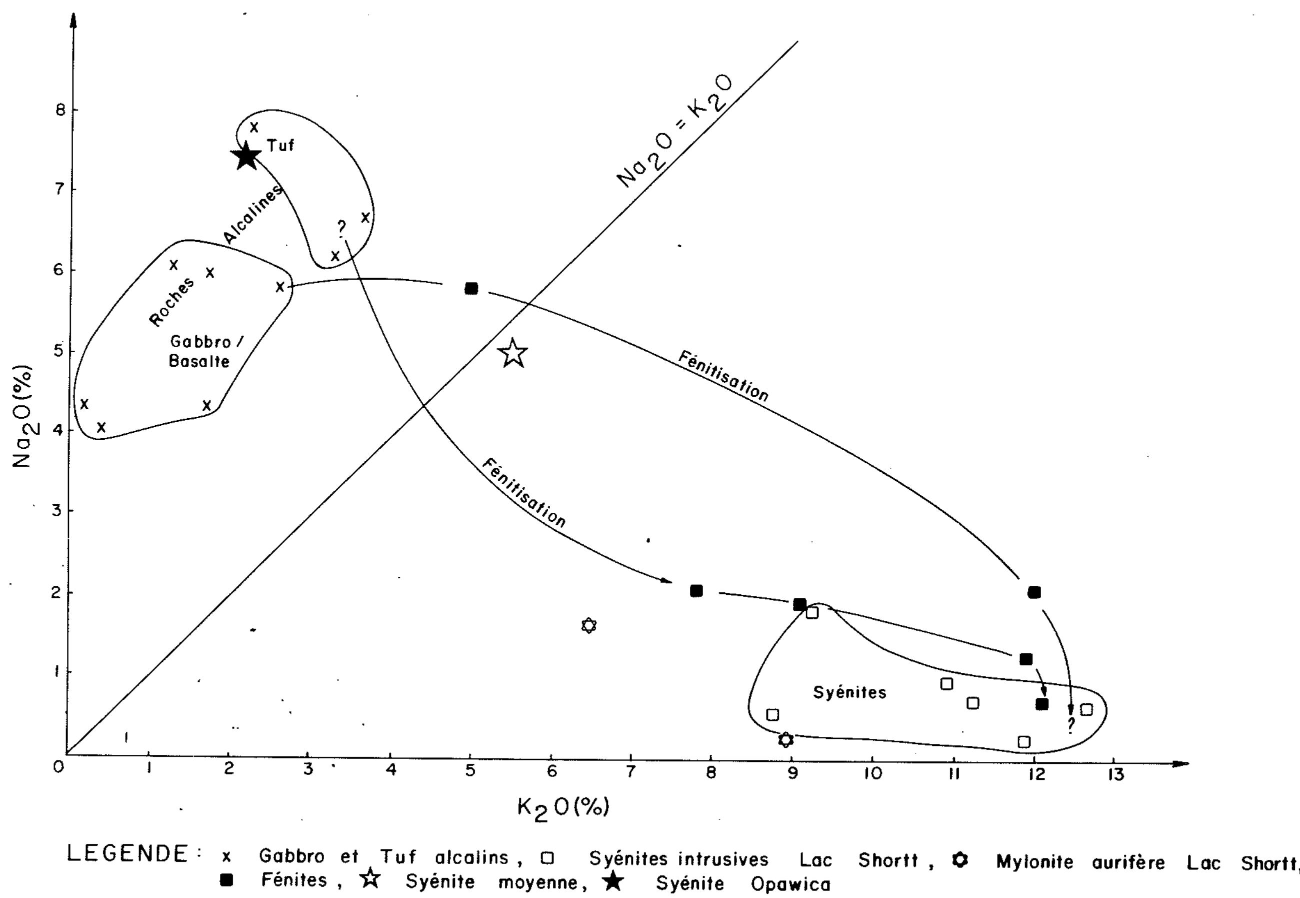

Figure 5B.3. Diagramme Na20 versus K20: caractérisation de la fénitisation à la mine Lac Shortt. Syénite moyenne selon Le Maître, 1976; Syénite Opawica d'après une moyenne de 21 syénites tiré de Averill et al., 1989.

la faille Lac Shortt sont géochimiquement anormales par rapport à la moyenne des syénites qui selon Le Maître (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 FeO 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

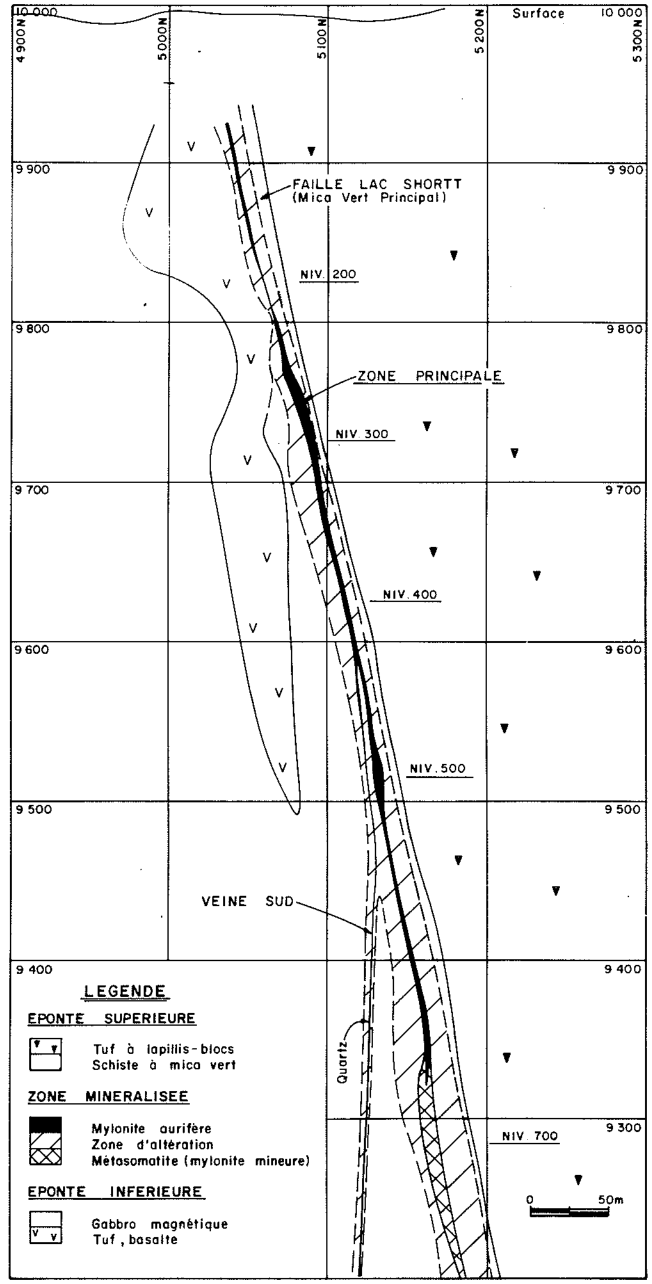

Figure 5B.4 Section transverse dans le gisement Lac Shortt. Zone Principale: section 5325E; Veine Sud: section composite 5350 à 5400 E.

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

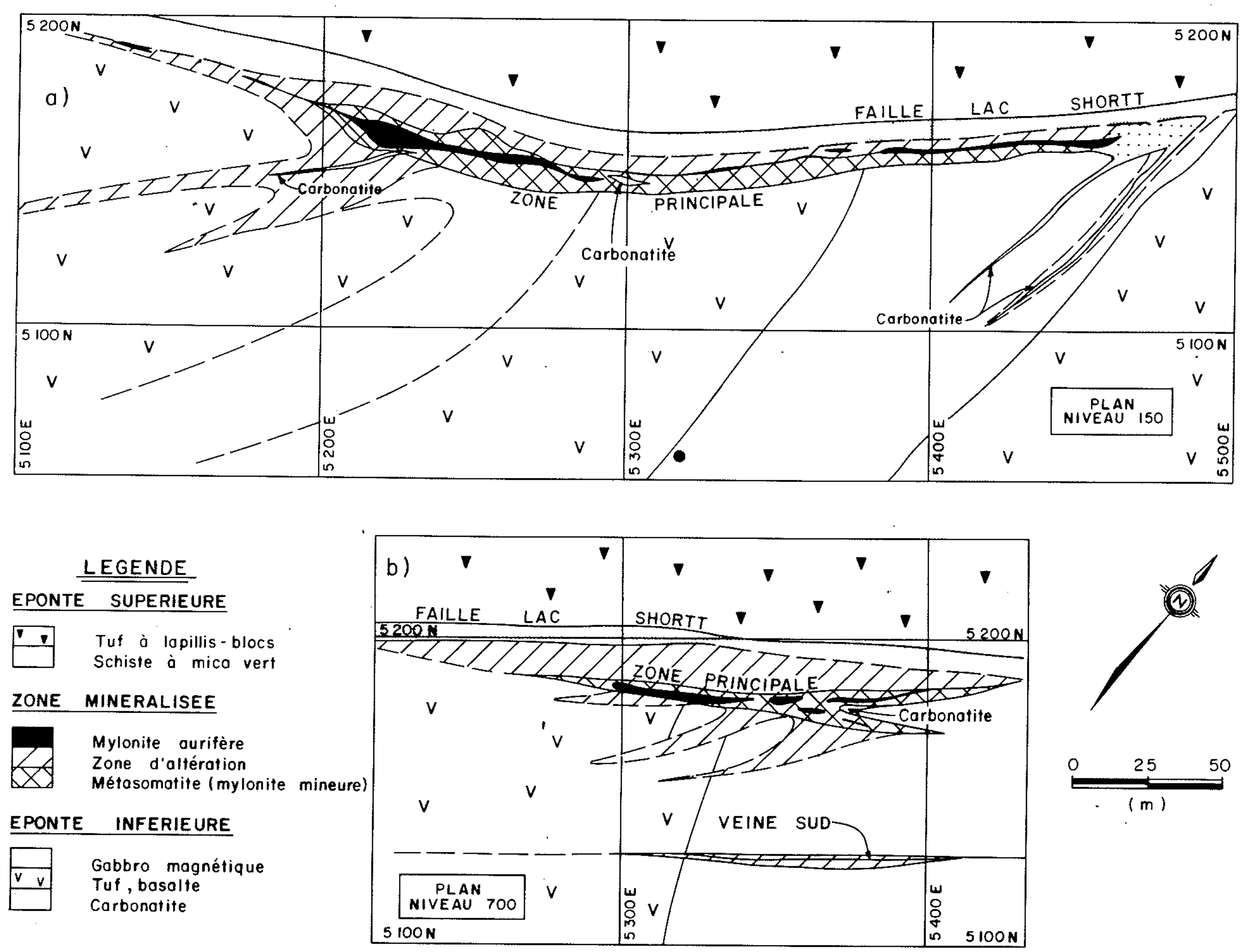

Figure 5B.5. Géologie des niveaux a) 150 et b) 700 à la mine Lac Shortt.

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).

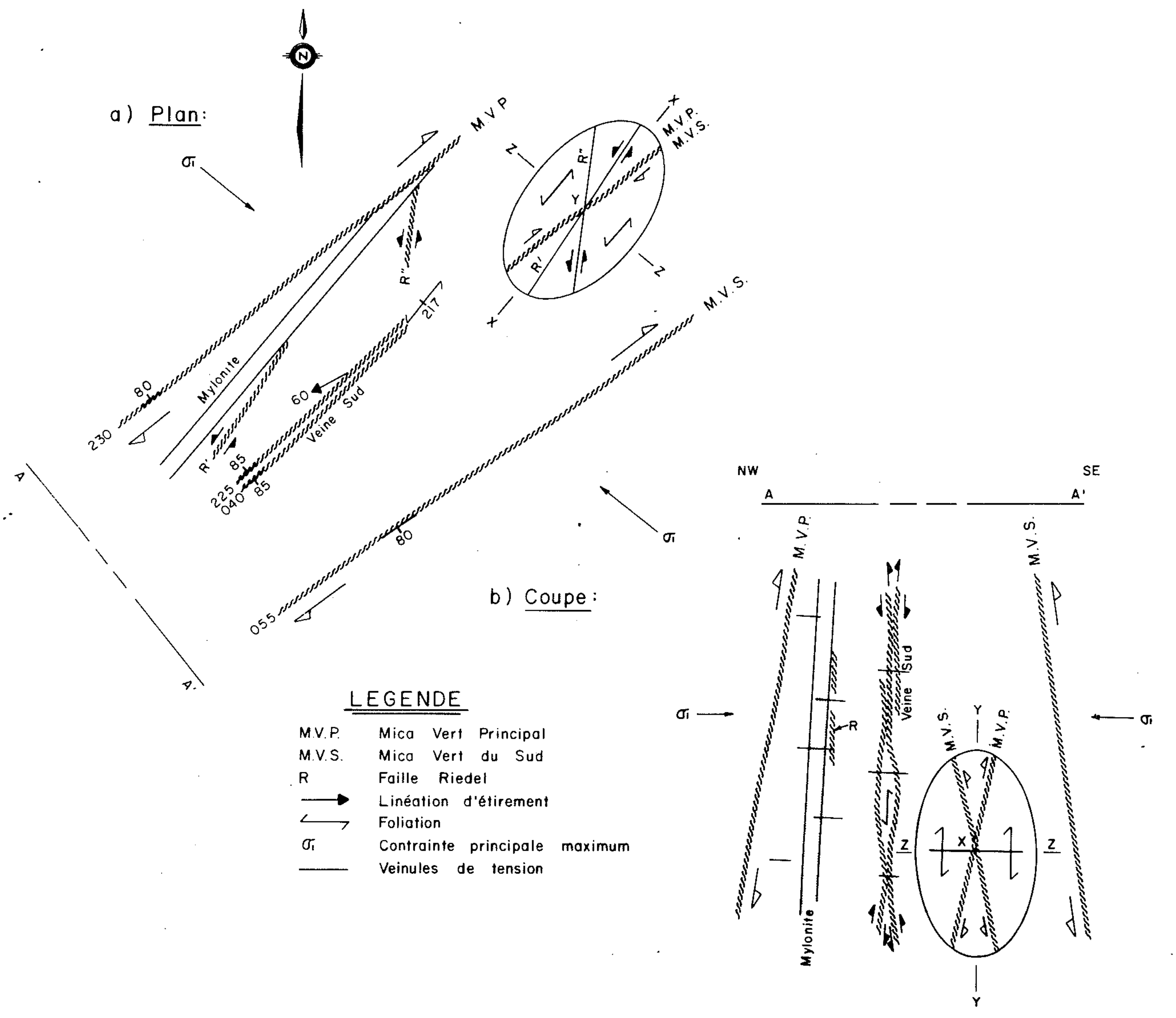

Figure 5B.6. Analyse des éléments structuraux à la mine Lac Shortt.

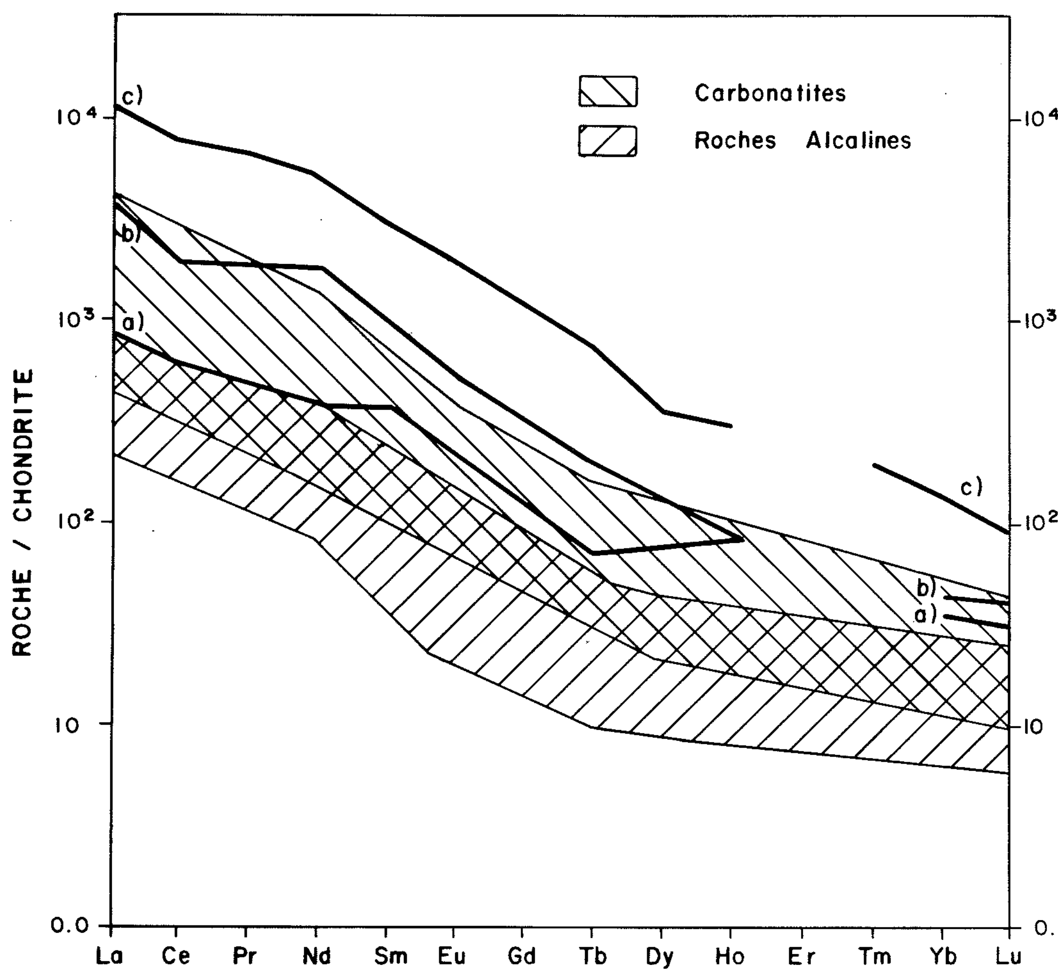

Figure 5B.7 Patrons des terres rares moyens pour la syénite (a), la carbonatite (b) et sol résiduel (c) du Lac Shortt. Normalisations d'après un composé de 9 chondrites (Haskin et Frey, 1968). Champs des carbonatites et des roches alcalines d'après Moller et al., 1980.

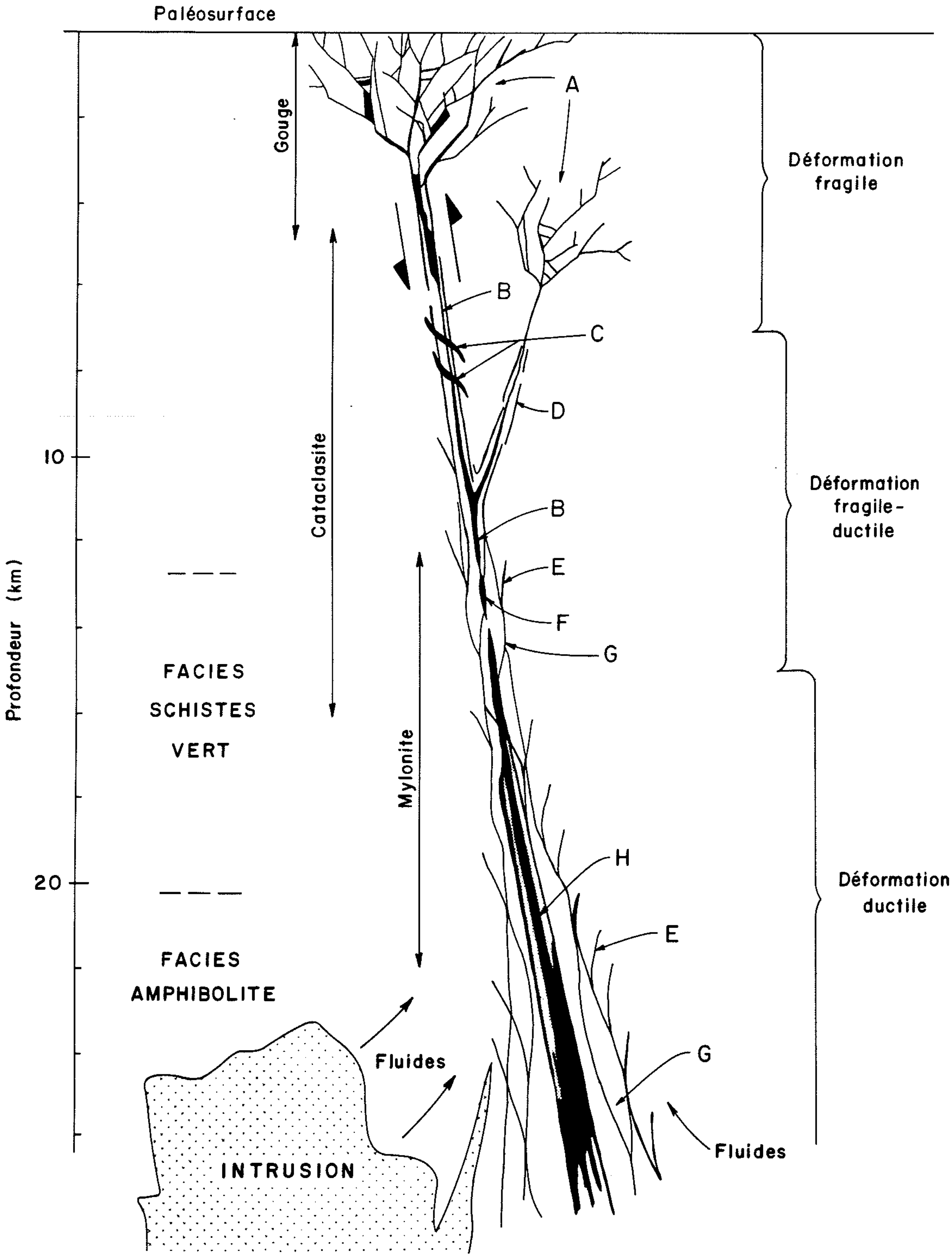

Figure 5B.8. Représentation schématique idéalisée d'une zone de cisaillement aurifère. Veines bréchiques (A), veine de cisaillement (B), veines de tension (C), faille Riedel (D), foliation (E), boudinage (F), anastomose (G), veine de remplacement (H), (modifié de Sibson et al., 1988).

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	274.00	175.00
Sc	0.58	0.80	0.43	0.72	0.79	10.00	1.90
La	286.78	887.02	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	-	-	-	-	-	7.50	6.10
Yb	8.40	9.10	9.90	6.80	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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