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FIFTH PROGRESS REPORT ON VOLCANOLOGICAL AND SEDIMENTOLOGICAL WORK IN ROUYN-NORANDA AREA

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Erich Dimroth
Sciences de la Terre
Université du Québec à Chicoutimi
Chicoutimi, Québec

Claude Larouche
Department of Geology
Carleton University
Ottawa, Ontario

and

Pierre Trudel
Département de Génie Minéral
Ecole Polytechnique
Montréal, Québec

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P R E F A C E

This report outlines the results of the field season 1976 and, thereby, marks the conclusion of the first phase of volcanological and sedimentological work done by the group of geologists led by the senior author. Work done to date falls in two categories, namely: (1) Regional studies of the stratigraphy, volcanic evolution, structure and metamorphism of selected areas and (2) studies of the volcanic and sedimentary facies and their interpretation. During the first phase of the present research programme, regional studies were completed in the area surrounding the central mining camp of Noranda in the north, south and east. These regional investigations were carried out by M. Larouche (1974), P. Boivin (1974), R. Côté (unpub., Ecole Polytechnique) and P. Trudel (this report), and have been reported previously. Studies of volcanic and sedimentary facies were carried out by N. Tassé (1976, facies of pyroclastic flows), G. Provost (unpub., Ecole Polytechnique, facies of rhyolite flows), M. Rocheleau (unpub., facies of sedimentary rocks, Université de Montréal), and the senior author (Dimroth 1977, facies of autoclastic volcanic rocks, this report: facies of mafic flows). Results of these studies also have been reported in the earlier reports of this group.

This report consists of three parts: (1) An account of facies relations of mafic flows based on previous mapping, and on the work by R.C. Bald in 1976, written by E. Dimroth, (2) a report on the stratigraphy and structure of Cléricy area by P. Trudel and (3) a report on the intrusive history in the area south of Rouyn by C. Larouche.

FACIES OF MAFIC - INTERMEDIATE FLOWS

Erich Dimroth

PREFACE

Only in one circumstance has the formation of pillows and of pillow breccias actually been observed (Moore et al., 1973). Where sub-aerial basalts enter the sea, lava-filled sacs grow from fissures at the toe of pahoehoe lobes. These sacs detach, disintegrate in part, and avalanche downslope. The resulting flows consist of two divisions: (1) A division of pillow breccia with avalanche cross-bedding occupies the space between the ancient sea-flow and sea level; (2) a division of massive lava or pahoehoe forms the top of the flow. Downward pointing pahoehoe lobes fed the flow-foot breccia. The typical sequence of these shore-line facies of basalt flows is shown in Fig. 1A.

Flows that formed entirelly below the surface of the sea have a radically different organization (Fig. 1B). They consist of a massive or pillowed division at the base, overlain by pillow breccia. Such flows have never been observed in the process of formation. Their origin must be inferred from the properties of solidified flows now exposed on land.

Here I will propose a hypothesis on the mode of formation of sub-aqueous basalt and andesite flows based on three sets of observation: (1) Information on the sequence of facies within individual flows, collected during previous summers but not described in previous reports;

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(2) Information on the relation between the facies sequence within single flows and the petrography of aquagene tuff and (3) Information on the shape of pillows and their relations, and on the relations between the massive facies and its flow-top breccia, based on mapping in 1976 by R.C. Bald.

Previous Work

The observation of advancing lava flows at the shore-line of Hawaii confirms previous ideas on the origin of flow-foot breccias by Fuller (1931). Noe-Nygaard (1940) described and interpreted pillow breccias that formed sub-glacially, below glaciers. Rittmann (1958), Silvestri (1963) and Re (1963) assume that submarine eruptions will form a thick deposit of brecciated glass beneath which the lava will spread. Continuous granulation will occur at the top of the underlying lava; on the other hand, lava will be injected into the overlying granulated material where it will form pillows. Carlisle (1963) relates the form of lava globulation to the rate and regularity of the lava advance. He suggested that regular lava flow produced pillows at the early stage of an eruption. Increasingly explosive eruption would lead to granulation and, finally, to the production of tephra, so that pillow breccias and, finally, stratified hyalotuff would overlie the pillowed facies. Reid and Dewey (1908), and Henderson (1953) suggested that pillows were bodily emplaced buoyed by steam. Lewis (1914) interpreted pillows as forming by budding of lava filled sacs from the surface of a flow.

DESCRIPTION OF FACIES

Massive Facies

The basal contact of the massive facies against the underlying flow generally is relatively flat: Depressions between underlying pillows are filled up and some relief may be present where the flow overlies hyaloclastic breccias; lobe-shaped projections into the underlying breccia are less common. The upper surface is flat and regular in sequence 1a, whereas numerous irregular lobes project upward into the overlying breccia in sequences 2a and 3a(Fig. 2). A chilled glassy crust, a few cm thick, is present at the lower and upper contacts.

The thickness of the massive facies varies from about 2 m to over 30 meters. The size and habit of phenocrysts does not vary within thin flows, except for the presence of skelletal overgrowth that become larger toward the interior of the flow. Thin flows are composed of microlitic basalt, thick flows are microlitic at the margin whereas their center has a gabbroic texture.

Generally, the massive facies shows same flow layering, defined by alternating lenses, several cm thick, of material slightly enriched in feldspar and in chlorite. Variolitic flows display spectacular flow-layering, because varioles commonly coalesce to form feldspar-rich layers 50 cm or more thick, alternating with chloritic bands that contain isolated varioles.

Columnar joints have been observed here and there in massive

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flows. More common is an irregular polygonal jointing. Joint surfaces are striated and joint polygons locally have been deformed, suggesting that the joints are a primary feature due to thermal contractions during cooling.

Pillowed Facies

The second facies consists of closely packed pillows. Here and there, pillows have subequal size and shape. More commonly, pillows have very inequal sizes and shapes (Fig. 3A, B). Very large pillows commonly are elongated and have quite irregular contours. Smaller pillows are ellipsoidal to bun-shaped and the smallest pillows, interstitial between large pillows, are sub-spherical. The long axes of pillows do not necessarily parallel the flow contact; pillows systematically inclined to the flow contact have been observed (Fig. 3C, Côté and Dimroth, 1976). Many pillows are interconnected and incompletely separated pillows are extremely common (Fig. 3D, E).

Pillows are covered by a crust, 1 to several cm thick, of shattered devitrified sideromelane. The interior of the pillows is composed of microlitic basalt. Morphology and spacing of microlites change systematically away from the pillow surface: Few, widely spaced small quench crystals of plagioclase and pyroxene are present in the outermost zone. Toward the interior of the pillow these become larger, more numerous and eventually lose their quench crystal morphology. Telluric phenocrysts are present in the pillow center and in the glassy crust, but those in the microlitic zones show an overgrowth with quench morphology. Fluidal texture parallel to the pillow surface is not uncommon and is shown by (a) enrichment of quench crystals and phenocrysts in bands parallel to the pillow surface and (b) by orientation of quench microlites and phenocrysts parallel to the pillow surface. Surface-parallel flow ribboning should not be conformed with sub-concentric contraction cracks.

Vesicularity of pillows varies from less than 1 to over 50 per cent, and vesicle size from less than 0.5 mm. to several cm. Generally, vesicles are enriched in the periphery of pillows, in particular in the upper periphery. Pipe vesicles a few mm across and several cm long, with radial orientation have been observed. Some pillows have tabular voids , oriented parallel to stratification, in their upper center.

Two systems of contraction cracks have been observed. A system of sub-concentric contraction cracks is always present in the glassy crust and is very common in the outer part of the pillow. A system of radial cracks overprints the concentric system in the outer and median parts of the pillow *. The inter-pillow spaces are filled either with void-filling chert or carbonate or , more commonly, are filled with a hyaloclastite composed of globules and granules of sideromelane.

Transitions from the massive to the pillowed facies take two forms: Incompletely formed pillows at the top of the massive facies are an intermediate stage. On the other hand, the extremely large pillows shown in Fig. 3B may be interpreted as patches of massive lava within what is essentially a pillowed flow. In the first case, the transition

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* Fig. 3F does not exist

is in a vertical direction, in the second a stage of the lateral transition from the massive to the pillowed phase probably is exemplified.

Pillow Breccia

Pillow breccias are composed of isolated pillows and/or pillow fragments set in a matrix of shattered glass fragments. Isolated pillows generally have very irregular, commonly ameboid shapes and are fairly small. Very small isolated pillows are sub-spherical. Otherwise, the isolated pillows are identical to the pillows of the pillowed facies.

Pillow fragments are angular or edge-rounded. They are easily recognized where part of the chilled margin is preserved. However, in many cases, the chilled crust has been completely separated from the fragments derived from the microlitic interior of the pillows. In-situ brecciation of pillows is very uncommon: In a few outcrops we have observed pillows in the process of fragmentation (Fig. 3G). In general, however, pillow fragments have been separated so that the original pillow cannot be reconstructed on the outcrop surface.

The isolated and broken pillows are set in a matrix of hyaloclastite. The petrography of the hyaloclastite will be described below; it is very largely a function of the vesicularity of the pillows. Carlisle (1963) described the typical petrography of the hyaloclastic matrix of poorly vesiculated pillow breccias. The matrix consists of two elements (1) Sub-pherical globules - in fact very small pillows - and (2) granulesthat formed by the thermal shattering of globules and of pillow rinds.

The contact of the massive facies against the overlying pillow breccia (Facies sequence 2a, 3a, Fig. 2) is very irregular. Irregular lobes of massive basalt protrude into the pillow breccia; pillows bud off the tips of these lobes. All stages of incomplete pillow formation have been observed (Fig. 3H).

The transition from the pillowed facies to the pillow breccia is gradational. In flow sequences 2b, 2c, 3b and 3c, the volume of interstitial hyaloclastite between pillows increases upward until an open fabric of isolated pillows set in a hyaloclastic matrix is obtained. Broken pillows generally are present at that stage, and the writer has not in general,been able to make the sharp distribution between broken - and isolated - pillow breccia proposed by Carlisle (1963).

Hyalotuff

Stratified hyalotuffs are at the top of the sequence in facies sequence 3a - 3e. They are graded tuff breccias. Graded beds of stratified hyalotuff have a thickness varying from a few meters to less than a cm. The thicker beds consist of two divisions: (1) A lower division of tuff breccia, massive or with graded bedding but without internal sedimentary structures and (2) An upper division of stratified tuff, generally with parallel stratification. Oblique stratification has been observed.

Hyalotuffs consist of fragments of microlitic basalt set in a matrix of hyaloclastic shards. They are monolithologic, that is all fragments have the same type of phenocrysts and are identical petrographically except in their quench crystals and the related devitrification features. Phenocrysts within the hyalotuff are identical in petrography to the phenocrysts in underlying massive of pillowed facies of the flows.

The lowermost contact of stratified hyalotuff against underlying broken pillow breccia may not everywhere be recognizable unambiguously: Thick beds of hyalotuff generally have a lower division that shows poor graded bedding, which is the only distinctive characteristic setting it apart from a broken-pillow breccia. Hyalotuffs may contain complete or, more commonly, broken pillows.

PETROGRAPHY OF HYALOCLASTITES

Three petrographic types of hyaloclastite are easily identified, namely: (1) microlitic hyaloclastite, (2) sideromelane shard hyaloclastite and (3) pumiceous hyaloclastite.

Microlitic Hyaloclastite

Microlitic hyaloclastite is essentially composed of fragments of microlitic basalt, although a few sideromelane shards may be present. The fragments of microlitic basalt generally are fairly large - one to several cm across - and they may be angular or edge-rounded. Fragments generally show effects of unequal chilling, and chilled crusts may be present within fragments, indicating that cooled crust of the flow has here and there been engulfed by lava. Here and there, quenching at the edge of the fragments is visible.

Microlitic hyaloclastites show a fairly high degree of tectonic brecciation, and are easily confounded with tectonic breccias. However, the inequal chilling of fragments at the margin shown by inequal size of quench crystals, and the presence of internal chilled crusts is characteristic. Where some sideromelane shards are present, the origin as a hyaloclastic breccia is unambiguously determined.

Sideromelane Shard Hyaloclastite

Sideromelane shard hyaloclastites are defined by the predominance of fragments of poorly vesiculated devitrified sideromelane. Some microlitic fragments, derived from the interior of pillows, are generally present. These hyaloclastites have been well described by Carlisle (1963) to whom the reader is referred. Characteristics are: (1) Presence of globules, that is sub-spherical bodies composed of sideromelane, (2) Presence of elongated pieces of pillow rims; (3) In-situ brecciation of globules and pillow rims and (4) Sharp-edged, convex concave configuration of the granules produced by the thermal shattering of (1), (2) and (3). Of course, hyalotuffs do not show in-situ brecciation.

Pumiceous hyaloclastites

Pumiceous hyaloclastites form from highly vesiculated basalts

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and andesites. They consist of strongly vesiculated pillows and pillow fragments and fragments of pillow rims, odd-shaped, highly vesiculated, glass shards and bubble wall shards. Vesiculation of fragments may attain up to 50 per cent. The convex-concave shards of sideromelane shard hyaloclastites are absent.

Generally, the fragment shapes are difficult to recognize in pumiceous hyaloclastites because the original porosity has been reduced by compaction and pressure solution rather than by cementation of the pore-space. Furthermore, much of the original fragmental material was fairly fine-grained, with grain sizes below 1mm, and the outlines of these fine-grained fragments are poorly visible. However, coarser grained fragments are present.

Effects of sea-flow metamorphism

Hyaloclastic breccias appear, at first sight, to be polymict due to differential metasomations during sea-floor weathering and sea-floor metamorphism. However, in feldspar-porphyric flow breccias, all fragments contain petrographically identical phenocrysts in about the same proportion.

The rapidly quenched basalt or andesite glass has been chloritized and/or partly silicified, thus, is now composed of chlorite and/or in part of silica. Relict textures suggest that the chloritized and silicified glass formed from palagonite. Traces of the textures suggest the former presence of fibro-palagonite and of gel-palagonite.

Non-porphyritic pillows and pillows containing few phenocrysts of feldspar consist of three zones. The outer zone (1) derived from quenched glass, is now composed of chlorite, that locally has been partly silicified. This zone is strongly fractured due to thermal contraction. Zone 2, up to 15 cm thick, consisted originally of glass containing quench crystals of plagioclase. During devitrification, albite spherulites formed around each of the plagioclase quench crystals. Spherulites are isolated in the outer part of zone (2) but coalesce inward. Thus, most of zone 2 consists of a white weathering rock very rich in feldspar containing very little chlorite. Toward the center of the pillow (zone 3) the size of quench microlites of plagioclase increases, and the proportion of albite overgrowth on plagioclase decreases. Consequently the center of the pillow has a more normal basaltic composition. In many cases, fractures separate the three zones so that chloritized fragments, fragments composed mostly of albite spherulites and fragments of microlitic basalt are present; all are derived from the same lava.

The spherulitic zone is absent in strongly feldspar-phyric andesite. Flow breccias of this rock contain fragments of two kinds: The thermally shattered quenched glass, generally highly vesiculated, has been chloritized. The microlitic part of the pillow has a normal andesitic composition, thus is relatively rich in albite microlites. Quench crystals of olivine and pyrozene may be present in both types but their presence does not otherwise change the devitrification textures of the compositional differentiation during devitrification.

VERTICAL SEQUENCE OF FACIES WITHIN FLOWS.

Fig. 2 represents the spectrum of facies sequences observed in Rouyn-Noranda area. The classical facies sequence of sub-aqueous basalts, described by Carlisle, is sequence 3c. Not all sequences are equally common. The most common sequences are 1, 2a, 3a and 1c, 2c and 3c; other sequences are relatively uncommon.

There is a definite relationship between the facies sequence and the petrography of the accompanying hyaloclastite. In facies sequence 1a, microlitic hyaloclastite is present. All facies sequences 1 contain sideromelane shard hyaloclastites. Facies sequences 2 contain sideromelane shard hyaloclastites (in poorly vesiculated basalts) or pumiceous hyaloclastites (in well vesiculated basalts and andesites). Pumiceous hyaloclastites are predominant in facies sequences 3.

Repetitions of sequences within a single flow were observed. Particularly, the facies sequence 2a, has been observed to be repeated, due to readvance of the flow over an already formed flow-top breccia.
Another type of repetition has been observed south of Lake Dufault where rather thick flows consist of tabular pillows, exceeding the outcrop size in length. These can be interpreted as a repetition of massive facies 30-50 cm thick.

MODEL OF FACIES RELATIONS.

INFERRED MODEL OF FORMATION OF FACIES.

Formation of pillows.

As noted pillows contain textures proving that they were rapidly cooled from the outside: Their outer crust consists of devitrified glass devoid of microlites but, eventually, containing telluric phenocrysts. Below the glass crusts quench microlites are present; the number and size of the microlites increases away from the surface of the pillow. Quench microlites very commonly show fluidal texture and flow banding also is present. Flow banding commonly is folded into recumbent folds parallel to the pillow surface close to the outer skin of the pillow, proving that viscosity increased toward the skin of the pillow. These features prove that pillows formed by interaction of lava and water at temperatures slightly above the liquidus temperature for crystallization of plagioclase.

Imbrication of pillows and budding of pillows from imbricated lava lobes proves that some pillows formed by dynamic processes during lava flowage analogous to pillows budding from pahoehoe lobes observed by Moore et al. (1973). However, I have also observed pillowed basalt dykes cutting tuffs. In this case pillows must have formed by a static process, after the flow across the dyke cross-section ceased. However, the very common occurence of incompletely formed pillows suggests that pillow formation is, in general, a dynamic process and that pillows form or multiply by budding.

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Some pillowed flows have very great lateral extent but, nonetheless , show great regularity of their top surface and little variability of thickness. This excludes the common concept that pillows are emplaced by avalanching of lava-filled sacs downslope, since such deposits would have a small length/thickness ratio, great thickness variation, and a humocky surface. Emplacement of pillows by a steam-buoyed density current (Baragar, oral common., 1966) also is excluded since such mass flows cannot float large pillows unless they contain a high proportion of fine-grained material, which is absent. My tentative conclusion is that pillows of flows having large length/thickness ratio, small thickness variation, and a regular top surface are tube-fed and that such flows are the submarine equivalent of tube-fed pahoehoe lava. However, the presence of lava-tubes in pillowed flows remains to be proved.

Formation of pillow breccias.

The shape of the fracture surfaces of glass shards and pillow fragments in pillow breccias are evidence for fragmentation by thermal strain: Glass shards show perlitic cracks, and have typical convex-concave boundaries where strain systems are not influenced by abundant vesicles. In porphyritic flows, perlitic fractures not uncommonly surround feldspar phenocrysts, due to differential contraction of glass and phenocrysts during cooling. Pillow rims preferentially show cracks subparallel to the pillow surface whereas the crystallized zones of a pillow show an older, concentric set of contraction cracks, overprinted by a younger set of radial cracks.

Glass shards in pillow breccias are sharply angular; pillow fragments show the same zonation of inward increasing crystallinity typical of whole

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pillows and completely independant of the fragment boundaries. Thus, fragmentation of pillows and of the glass crust took place at temperatures below the softening point of basalt glass, and at temperatures where crystallization of quench microlites in the pillow interior had ceased.

In addition to these fragments, some pillow breccias contain globules, rounded glass bodies that generally are shattered by thermal contraction cracks. There is a complete transitions from globules, round glass bodies 1 or a few cm across, to very small pillows, perhaps 5 cm across that already contain a few quench crystals of plagioclase in the center, to small pillows, a few cm across, and finally to normal and large pillows. Thus, I believe that the globules form by the same process as pillows, by extrusion of liquid lava from a pahoehoe lobe, lava tube or from a freshly formed larger pillow.

Absence of any kind of sedimentary structure in pillow breccias and, in particular the characteristic presence of in-situ breccias proves that pillow breccias formed in place.

Evidence outlined above suggests that pillow breccias formed by interaction between sea water and basalt lava at intermediate temperatures (below the softening point of basalt glass). Such interaction was localized in the interspaces between pillows at the time the pillows cooled below the liquidus temperature of plagioclase and affected pillows only at a relatively late stage of their cooling. The sequence from closely packed pillows to hyaloclastites, with isolated pillow breccias and broken-pillow breccias as intermediate stages, is a sequence of increasing intensity of lava-water interaction. Pillow breccias overlying massive flows have to be interpreted in the same way, fragmental material forming from lava lobes and pillows, and new lava lobes continuously injecting the fragmental material from below. In that case, pillow breccias may already form during flow age and then may show imbrication (Côté and Dimroth, 1976).

Formation of hyalotuffs.

The structural sequence of the graded hyalotuffs described above is identical to the graded stratified facies of mass-flow deposits of Walker and Mutti (1973). Thus, I interprete these tuffs as emplaced by turbidity currents. Very likely, all gradations from hyalotuffs emplaced by turbidity currents to hyalotuffs emplaced as high-density grain flows are present; however the whole spectrum of the sedimentary facies of these rocks has not yet been studied. The following conclusion follow from the interpretation of the sedimentary structures of hyalotuffs: (1) Hyalotuffs, although composed of material derived from the underlying flow,are allochthonous and have been emplaced as mass-flows. (2) Dispersal of hyalotuffs occured at a location higher topographically than their locality of deposition (that is closer to the source of the underlying flow). (3) Dispersal likely was caused by minor explosions, although gravitational instability on steeper slopes is a possible alternative.

Hyalotuffs consist to a very large part of glass shards, with a minor component of pillow fragments and microlitic fragments. Thus, it is difficult to state at which temperature the lava-water interaction responsible for fragmentation took place. Furthermore, most of the hyalotuffs of the area are strongly vesiculated and occur in facies 3a to 3b, although poorly vesiculated hyalotuffs, in facies sequence 1a' to 1c' have been observed locally. The writer believes that hyalotuffs are the product of local explosive interaction between lava and water, generally starting at high temperatures, and generally related to eruption of highly vesiculated lavas.

FACIES SEQUENCE AND INTENSITY OF LAVA-WATER INTERACTION

In the inferred mechanisms of formation of pillows, pillow breccias and hyalotuffs, facies have been related to the nature and intensity of water-lava interaction. It follows that the facies series 1a to 1c is a serie of increasing intensity of high-temperature interaction between lava and water in the absence of significant low temperature interaction. Since basalt glass contracts and fractures during chilling, this is possible only if the flow has been shielded from further lava water interaction after the initial pillow formation by a layer of water vapour.

Facies sequence 2a - 2d is a sequence of increasing intensity of high-temperature interaction between lava and water followed by lava-water interaction at low temperatures. Since pillow breccias commonly overlies the massive and not only the pillowed facies, high temperature and low-temperature interaction are independent. Such flows have not been shielded against the influx of water from above by a layer of vapour after the initial pillow formation.

Facies sequence 3a - 3e is identical to sequence 2a - 2d, except that explosive interaction between lava and water took place . Similar relations exist between facies sequence 1a to 1c and the uncommon facies sequence 1a' to 1c'.

Lateral transitions between facies sequences

At the present time, individual flows have not been followed for distances sufficiently long to actually walk out the transition from one facies sequence into the other. Thus, my conclusions are presently based on measured vertical sections. In the interpretation of these sections, it is assumed that all flows at one and the same locality have essentially the same character.

In vertical section, facies sequences 1a, b and c as well as 2a, b, c, d are commonly associated. However, very common is also the association of sequences 2a, 2b or 1b and 1c. The presence or absence of hyalotuff cannot yet be related to any regularity.

Undoubtedly, facies sequences are related to proximity to the source. I assume that the facies sequences containing a massive phase are more proximal than the facies sequences containing pillows or pillow breccia. As above, presence or absence of hyalotuff cannot be used at present to estimate proximity, except that hyalotuffs are always more distal than the site of the causative explosion (which is not necessarily at the eruptive vent).

INTRUSIVE ROCKS OF THE BELLECOMBE - REMIGNY AREA

Claude Larouche, Department of Geology, Carleton University.

The field-work essential to the study of part of a large Archean batholith was continued last summer. The batholith intrudes the Pontiac metasediments south of Rouyn-Noranda.

Two north-south cross-sections have been made from Cloutier to the south of Remigny based mainly on shorelines and road cuts. This work allowed correlation of the four areas visited in 1975. Samples have been collected for petrographic and chemical studies. The mineralogy and the sequence of the various intrusions were determined. The information thus obtained will be used in a Master's thesis at Carleton University.

The reader is referred to previous reports (Dimroth et al., 1974, 1975) for the presentation of the regional geology and the description of the different generations of intrusive rocks present in this area.

The homogeneous member (Dimroth et al., 1974) predominates throughout the area and varies from a quartz-poor monzodiorite to a quartz-rich granodiorite composition. The metaluminous character of the homogeneous member leads principally to the crystallization of biotite and hornblende as the main mafic constituents. The heterogeneous member is present in variable amount and fills a complicated system of dykes.

The rock names of Chagnon (1968) have been changed to conform to conventional petrographic classification. In order to avoid confusion Chagnon's equivalents will be given in parenthesis.

HOMOGENEOUS MEMBER

Monzodiorite (hornblende-pyroxene syenite) forms a distinct massif east of Lac Frechette to the north of the cross-sections. It is separated from the granodiorite (hornblende granite, porphyritic hornblende granite) by a syncline of Pontiac metasediments. This granodiorite underlies a large area south of the monzodiorite, extending as far as Rémigny.

Variations in composition, texture and grain size are locally present in the monzodiorite. North of Lac Barrière, this rock is finer grained and is in contact with a meta-peridotite. The contact is sharp and the ultramafic rock is composed of talc, chlorite and pyrite. Metasomatism of the peridotite contributes to the development of biotite. The contact of the monzodiorite with the Pontiac Group is not very well exposed, but where exposed it is sharp and intrusive, some dykes of monzodiorite cutting the Pontiac. Lineations, well developed in some localities by the uniform vertical orientation of hornblende, become sparse and invisible over a large area. Small xenoliths are numerous at the margin. The country rock is believed to have led to hybridation, with an increase in the percentage of mafic minerals. Megacrysts of potassic feldspar are not uniformaly present.

The porphyritic hornblende granodiorite (porphyritic hornblende granite) predominates south of the cross-sections and encloses large blocks of monzodioritic composition. The contact between the porphyritic granodiorite and the large blocks shows gneissification developed over a few inches in the monzodiorite at some places, a sharp demarcation characterised by the concentration of mafics at the contact zone at other places, and lastly a gradual decrease in the amount of megacrysts of potassic feldspar in the granodiorite approaching the contact.

Inclusions of Pontiac Group, mafic and ultramafic rocks are also present in the porphyritic granodiorite. The mafic inclusions are small (5 cm) and elongated, defining and orientation. The ultramafic inclusions are blocks of all scale, with biotite rims.

A non-porphyritic granodiorite (hornblende granite) rich in biotite is also present. The relation between the porphyritic hornblende granodiorite and the biotite hornblende granodiorite is believed to be one of progressive transition. This gradation is supported by correlation between the decrease in size of potassic feldspar, enrichment in biotite at the expense of hornblende, and increase in the amount of quartz present progressing from porphyritic to non-porphyritic unit.

HETEROGENEOUS MEMBER

The later heterogeneous member (oligoclase, microcline granite) consists of a complicated system of dykes including muscovite pegmatites granites and aplites. A generation of lamprophyre dykes cuts most of the acidic dykes. Two generations of diabase, in place porphyritic, represent the last intrusion.

Joints system spaced from a few inches to several feet are well developed in both units and contain epidote, quartz and biotite.

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A few thin sections have been studied to date to evaluate the possibility of metasomatic origin, as proposed by Chagnon, of potassic feldspars present as megacrysts in the porphyritic hornblende granodiorite.

Generally, there are two generations of potassic feldspar: a minor population, average size 1.5 mm, of anhedral interstitial grains in the matrix and a major population which dominates as large crystals with an average size of 2 cm. Poikiloblastic or sieve texture is present in the large crystals: hornblende, plagioclase, biotite with or without chlorite, apatite, sphene and opaque minerals are distributed as inclusions within the potassic feldspar. The distribution is not random but exhibits a zonal pattern. The long axes of the minerals occuring as inclusions are always oriented parallel to the margin of the potassic feldspar and mantle the core at different stages. Frequently, a reaction rim is formed around plagioclases, biotites and hornblendes present as inclusions.

It is also believed that the presence of worm-like grains of quartz in parts of some plagioclases producing a texture similar to myrmekite, may be related to the crystallization of the potassic feldspar.

These observations supported by the fact that most of the xenoliths present in the porphyritic hornblende granodiorite do not contain megacryst, suggest a magmatic origin for the potassic feldspar.

MINERALIZATION

A few minor evidences of mineralization have been observed in the area. Most of them were previously known and have already been investigated.

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Work was carried on, north of Lac Chabot, in search of radioactive minerals in pegmatites. Pegmatites found along the road west of Lac Pian and those south-east of Lac Beaumesnil gave high readings on a geiger counter. Smoky quartz and black mineral are present in this pegmatite, no analysis of this black mineral is yet available. South-east of Lac Caire a few crystals of beryl (average 2.5 inches) are found associated with quartz. Smoky quartz is present and a mineral of the colombite-tantalite series was analysed.

STRATIGRAPHIE ET TECTONIQUE DE LA REGION DE CLERICY

Pierre Trudel, Ecole Polytechnique

1. Introduction:

Cette étude constitue le sujet d'une thèse de doctorat à l'Ecole Polytechnique de Montréal. La région considérée consiste en une bande de roches volcaniques orientée NW-SE d'environ 19 km de longueur par 10 km de largeur, et couvre une partie des cantons Destor, Dufresnoy, Cléricy et Joannès. Elle est limitée au nord par le contact entre le Groupe de Kewagama et le Groupe de Blake River, au sud par une ligne qui joindrait le Lac Dalembert et le Lac Marillac, à l'est par la faille Davidson Creek et à l'ouest par la route 101. Toutes les roches étudiées appartiennent au Groupe de Blake River (d'âge Archéen); ce Groupe est constitué essentiellement de roches volcaniques métamorphisées. La région a été cartographiée en entier au cours des étés 1974, 1975 et 1976; quelques vérifications seront sans doute nécessaires pendant la saison 1977. Voici les résultats obtenus à date:

2. Stratigraphie:

2.1. Relation avec le Groupe Kewagama:

Le Groupe de Kewagama (également d'âge Archéen) est composé de roches métasédimentaires (graywackes et schistes argileux). Son contact avec le Groupe de Blake River n'est pas exposé dans la région étudiée, mais aux endroits où les affleurements de ces deux Groupes sont les plus rapprochés (dans un cas ils sont séparés d'à peine quelques mètres), l'attitude de la stratification est la même pour les deux Groupes et les

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polarités sont dans les deux cas vers le sud. Il semble donc raisonnable de supposer que le Groupe de Blake River repose en concordance sur le Groupe de Kewagama, bien que des changements de faciès latéraux dans le Groupe de Blake River près de son contact avec le Groupe de Kewagama nous laissent supposer que ce contact peut être localement faillé.

2.2. Stratigraphie à l'intérieur du Groupe de Blake River:

Sur des bases lithologiques et géochimiques, on peut subdiviser les roches volcaniques du Groupe de Blake River en plusieurs séries. Entre la route 101 et le Lac Dufresnoy, nous avons établi la stratigraphie suivante, de la base vers le sommet:

La série tholéiitique Duparquet-Destor est peu épaisse (environ 600 m) et consiste en des coulées basaltiques massives et à coussins; elle semble reposer en concordance sur le Groupe de Kewagama et est caractérisée par la présence d'un horizon-repère très bien marqué de lave variolaire à environ 300 m de la base. Cet horizon a pu être tracé sur une distance d'au moins une quinzaine de Kilomètres à l'est de la route 101. Les basaltes sont très faiblement vésiculés (en général moins de 5% et souvent moins de 2%) et les brèches de coulée très rares, ce qui indiquerait une déposition en eau très profonde (probablement plus de 2,000 mètres).

La série calco-alcaline Reneault mesure environ 2,500 m d'épaisseur et semble reposer en concordance sur la précédente. Elle est composée d'un ensemble de roches volcaniques différenciées, de composition basaltique à rhyolitique. Elle débute à la base par une séquence de brèches de coulée, de coussins et de coulées pyroclastiques de composition andésitique très grossièrement porphyrique en feldspath (phénocristaux de 3 à 5 mm) et se termine au sommet par un horizon très bien marqué de coulées pyroclastiques grossières de composition andésitique à dacitique (les pyroclastiques de Reneault).

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Ces coulées pyroclastiques sont porphyriques en feldspath (phénocristaux de 3 à 5 mm); elles sont composées surtout de fragments d'andésite porphyrique en feldspath et faiblement vésiculée, de quelques fragments de ponce et de cristaux et fragments de cristaux de feldspath dans une matrice formée de matériel volcanique fin. Elles sont très grossières à l'ouest de la route 101 et s'acheminent vers un faciès de plus en plus distal vers l'est, où la taille des fragments et l'épaisseur des lits diminue constamment. Dans la région étudiée, juste à l'est de la route 101, la taille maximum des fragments est de 15 cm et les lits mesurent environ 1 m d'épaisseur; cet horizon a été suivi sur une dizaine de km où il disparaît à l'est du Lac Dufresnoy, dans le coeur du synclinal de Cléricy: à cet endroit, le tuff est très fin et mélangé avec du matériel sédimentaire (tuffite); les lits sont millimétriques à centimétriques.

Les basaltes et andésites de cette série sont bien vésiculés (en général plus de 10%, jusqu'à un maximum de 30%) et les brèches de coulée sont très abondantes, ce qui indiquerait une mise en place dans un milieu d'eau relatiement peu profonde. Malheureusement, peu de travaux ont été faits sur la vésicularité des laves mafiques d'affinité calco-alcaline et une comparaison directe ne peut être effectuée avec les laves tholéiitiques étant donné les propriétés différentes des magmas (contenu en volatils, température, viscosité, etc...).

La série tholéiitique Dufresnoy mesure au moins 2,000 m d'épaisseur dans la région étudiée; elle semble reposer en concordance sur la série précédente. Au SW du Lac Dufresnoy, elle forme une séquence homoclinale (sommets vers le SW) de coulées monotones de basaltes et d'andésites aphanitiques; on ne peut y suivre aucun horizon-repère. Les brèches de coulée sont assez abondantes et les laves possèdent une vésicularité moyenne d'environ 10%, ce qui implique qu'elles se seraient déposées dans un milieu d'eau

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beaucoup moins profonde que la série Duparquet-Destor-Manneville.

Une autre série tholéiitique semble être présente au SE du village de Cléricy: elle consiste en un empilement monotone de basaltes massifs et à coussins. Il y a très peu de brèches de coulée et les basaltes sont très faiblement vésiculés, ce qui indique un épanchement en eau profonde. La position stratigraphique de cette série par rapport aux autres n'est pas encore connue et il est probable qu'en cet endroit la situation soit compliquée par des failles. Cette série a été le site de minéralisations métalliques intéressantes (chalcopyrite et pyrrhotine) le long de zones cisaillées et silicifiées; de nombreux travaux miniers y ont été effectués et plusieurs se poursuivent encore.

3. Tectonique:

3.1. Plis à grande échelle:

L'ensemble de la structure de la région est contrôlée par de grands plis isoclinaux qu'on peut tracer sur plusieurs kilomètres. Ce sont des plis de phase P_1^0 ; la trace de leur plan axial est déterminée par cartographie sur le terrain et a une direction générale NW-SE et un pendage sub-vertical. A l'échelle régionale, la stratification S_0 possède également cette même attitude. Malheureusement, il nous est très difficile de connaître la plongée de ces plis, car:

1°) On n'observe en aucun endroit une charnière de ces plis,

2°) Il n'y a souvent aucune schistosité associée à cette phase de plissement; en effet, la schistosité S_1 est très souvent absente et lorsque présente, elle est très faiblement développée.

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Cependant, la très grande continuité des horizons-repères tracés favorise une faible plongée pour ces plis, probablement vers l'est, tel que suggéré par le fait que l'on a retrouvé l'horizon-repère des pyroclastiques de Renault dans un trou de forage environ 1 km à l'est de l'endroit où il est aperçu pour la dernière fois sur le terrain.

Les principaux grands plis isoclinaux que nous avons tracés à l'échelle régionale sont les suivants:

- Le synclinal de Cléricy a été tracé sur une distance de plus de 10 km, de la partie nord du Lac Dufresnoy jusqu'au nord du village de Cléricy. Il s'agit d'un pli isoclinal légèrement déversé vers le NE; son plan axial a une direction NW-SE et un pendage sub-vertical.
- Un anticlinal a été tracé sur 3 km de long de la rive SW du Lac Dufresnoy; il est probablement relié au synclinal de Cléricy et comme ce dernier, son plan axial a une direction NW-SE et un pendage sub-vertical. Il est légèrement déversé vers le NE.
- Le synclinal Harvie se situe dans la série tholéiitique localisée au SE du village de Cléricy; il s'agit d'un petit synclinal serré dont le plan axial a une direction NNW et un pendage sub-vertical. La trace de ce plan axial est déplacée par une série de petites failles de décrochement de direction E-W: ces failles sont toutes sénestres et les déplacements varient de 150 à 500 m.
- L'anticlinal du Lac Imau a été cartographié au NW du Lac Imau (canton Cléricy, rang V) par l'inversion de polarité dans les

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laves à coussins. Son plan axial a une direction NW-SE et l'un de ses flancs contient une séquence de roches sédimentaires (graywackes interlités avec des schistes argileux) de position stratigraphique actuellement inconnue.

Cette tectonique simple de grands plis isoclinaux caractérise le canton Destor et la partie nord du canton Dufresnoy, mais dans la partie sud du canton Cléricy, ces plis P_1^0 sont repris par une deuxième phase de déformation et on observe l'apparition de plis P_2^0 très bien développés.

3.2. Plis de type P_2^0:

Ces plis ont une amplitude beaucoup plus faible que ceux précédemment décrits. Nous avons observé des charnières de ces plis en quelques endroits, notamment:

1°) Canton Cléricy, rang V, lot 7.

2°) Canton Cléricy, rang IV, lots 22 et 23.

Dans ces zones de charnière, la schistosité de S_2 est très forte, orientée E-W avec un pendage abrupt, et les coussins montrent un écrasement extrême, avec leur grand axe maintenant parallèle à S_2. Cependant, des chambres remplies de quartz dans la partie supérieure des coussins nous indiquent que la stratification à ces endroits possède effectivement une direction N-S avec un faible pendage (environ 30°) vers l'ouest. La schistosité S_1 n'est pas présente dans ces zones de charnière.

3.3. Microtectonique:

Quatre phases de déformation ont été identifiées dans la région:

- La schistosité S_1 est plan axial des grands plis isoclinaux que l'on observe à l'échelle régionale. C'est une schistosité de flux; elle est rarement développée et lorsque présente, généralement de faible intensité. Lorsqu'on l'observe en présence de S_2, elle est crénulée par S_2 et fait un angle d'environ 30° avec cette dernière, dans le sens des aiguilles d'une montre. Les linéations L_2^1 ($S_1 \wedge S_2$) plongent vers l'ouest abruptement (50 à 60°). Partout où nous avons observé S_1, son attitude était sub-parallèle à celle de la stratification.

- La schistosité S_2 est une schistosité de crénulation. Elle est développée à travers toute la région étudiée et mesurable sur presque chaque affleurement. Son intensité est variable, mais son attitude est très constante: direction ENE à E-W et pendage très abrupt. Elle correspond à une phase compressive et est grandement responsable des déformations internes dans les roches: aplatissement des coussins, des varioles, des amygdules, etc... Elle est très bien développée dans la partie sud du canton Cléricy où elle reprend les plis isoclinaux de première phase pour former des plis de type P_2^0.

- Deux systèmes de kink, l'un dextre orienté NNE et l'autre sénestre orienté NNW sont aussi présents. Ils sont probablement contemporains et représentent une phase de déformation tardive: on ne les observe que dans les zones de forte déformation où ils reprennent S_2. Ils n'influencent pas la configuration des plis importants dans la région. Ces deux plans ont des pendages sub-verticaux.

Nous avons noté quelques zones où la déformation est très forte et le développement de la schistosité extrême:

- Sur la rive nord de la Plage Destor,
- Dans une zone de direction E-W entre la route 101 et la pointe extrême ouest du Lac Dufresnoy,
- Dans une zone de direction NW-SE au SE de Lac Dalembert.

3.4. Failles:

Plusieurs failles sont présentes dans la région étudiée; ce sont le plus souvent des failles transversales de décrochement dont le sens et l'amplitude du décrochement sont très variables. Nous avons déjà mentionné celles qui déplacent la trace du synclinal Harvie au SE de Cléricy, mais il y en a plusieurs autres.

Nous avons également des évidences pour une faille longitudinale importante qui oblitère la séquence des pyroclastiques de Reneault sur le flanc nord du synclinal de Cléricy.

Les microfailles de décrochement sont ectrêmement abondantes et on les observe qui déplacent (de quelques centimètres) les minces lits de sédiments et de pyroclastiques, les bordures de coussins, les petits dykes, etc... Elles ont été mesurées systématiquement (direction, pendage, sens et amplitude du décrochement), mais leur compilation n'est pas encore complétée.

4. Métamorphisme:

L'étude du métamorphisme de la région n'est pas complète et il

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convient de distinguer plusieurs types de métamorphisme (Dimroth et al., 1974); cependant, l'intensité du métamorphisme régional semble augmenter d'une façon générale du nord vers le sud (c.a.d. de la base vers le sommet de la séquence) dans la région étudiée. En effet, nous observons successivement du nord vers le sud:

- Une zone de la chlorite où l'augite est remplacée par l'actinote. Il faut noter que la chlorite est présente partout dans la région et que cette zone est plutôt une zone négative caractérisée par l'absence des minéraux prehnite, pumpellyite, augite fraîche et biotite.

- Une zone de la biotite dans la partie à l'extrême sud de la région, où l'on observe pour la première fois l'apparition de la biotite dans les basaltes.

Nous avons également observé des basaltes à coussins métamorphisés au faciès amphibolite dans le canton Clericy, rang VIII, lots 4 et 5, mais dans ce cas, il s'agit d'un métamorphisme de contact, ces affleurements étant juste en bordure d'une intrusion syénitique importante.

5. Découverte de croûtes oxidées sur des coussinets

Nous avons découvert des croûtes composées de magnétite et autres oxides de fer sur la bordure de coussinets et de fragments de coussinets à un endroit de la région. Des croûtes semblables ont déjà été découvertes par Spence (1976) à l'est de Rouyn, et par Dimroth (communication orale) dans le canton de Dufresnoy. Les croûtes semblent analogues aux croûtes oxidées des basaltes sous-marins récents, ce qui leur confère une importance considérable.

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