ARCHAEAN SEDIMENTATION:

ALLUVIAL FAN AND TURBIDITE DEPOSITS,

LITTLE VERMILION LAKE, NORTHWESTERN ONTARIO

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By

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SCOPE AND CONTENTS:

A sequence of sedimentary rocks within a greenstone belt at Little Vermilion Lake, near Sioux Lookout, in northwestern Ontario, has been mapped in detail, and 5 facies have been established.

The lowest 1,700 m of the sedimentary succession rests unconformably on a thick sequence of volcanics, and is composed of a conglomerate facies and an arkose facies. The conglomerates occur in beds up to 20 m in thickness, are laterally impersistent, and comprise pebbles, cobbles and boulders (up to 1.5 m in diameter) of basic to acidic volcanic rocks and granodiorite, with an interstitial matrix of arkose. The conglomerates are commonly massive, and exhibit no cross-stratification or graded beds, but some are imbricated.

The arkoses commonly occur as interbeds less than 1 m thick between massive conglomerates, or occur as thick sequences which are commonly massive, but which exhibit some trough cross-

bedding, parallel lamination and graded bedding.

The arkoses and conglomerates are succeeded by 100 m of felsic tuff, 60 m of greywacke and argillite, 270 m of basic tuff, and a 600 m thick section of graded greywackes and argillites, with individual beds commonly less than 30 cm thick.

The arkoses and conglomerates are interpreted as alluvial fan deposits, whilst the greywackes and argillites are interpreted as turbidites. The tuffs mark explosive volcanic activity which accompanied the deepening of the basin of deposition.

The petrographic characteristics of the sediments indicates that the provenance area was composed of basic to acidic volcanic rocks, together with a large extent of granodiorite.

The granodiorite may have intruded the volcanics, prior to the deposition of the sediments.

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CHAPTER 1

ARCHAEAN SEDIMENTATION - A REVIEW OF SOME OF THE PROBLEMS, AND THE POTENTIAL CONTRIBUTIONS OF SEDIMENTARY STUDIES

1. Introduction

Archaean greenstone belts of the Canadian Shield not only provide us with the earliest available records of the formation and development of the shield, but also assume economic importance by virtue of their valuable mineral deposits.

During the last decade, attention has been mainly focused on the volcanic rocks of greenstone belts, whilst there have been fewer publications dealing in detail with the sedimentary rocks. Several reconstructions of lithofacies distributions within Archaean basins have been published (Goodwin and Shklanka, 1967; Goodwin, 1969; Goodwin and Ridler, 1970; Hutchinson et al., 1971; Ridler, 1971) but their general applicability to all greenstone belts has not been established.

In this present study, the sedimentary rocks in part of a greenstone belt near Sioux Lookout have been mapped in detail, and the sedimentological interpretations are discussed in the context of basin development.

However, a short review of the problems involved in Archaean sedimentation will first be presented, and the potential contributions of sedimentary studies to the interpretation of greenstone belts will be outlined.

2. Archaean sedimentary problems

The common sedimentary suites within greenstone belts have been known for many years (Pettijohn, 1943), but interpretations of these rocks in terms of depositional processes, depositional environments, basin geometry, tectonic setting and provenance, have rarely been attempted in the context of modern sedimentological thought.

Detailed reviews of previous work and of the ideas relating to Archaean sedimentation have been given by Pettijohn (1943, 1970), but the current problems of Archaean sedimentation can be divided into 4 categories, which are outlined below.

(a) Stratigraphy and correlation

An accurate stratigraphy must be established for the sedimentary and volcanic rocks of a greenstone, by detailed mapping. Neither the sediments nor the basin evolution can be fully understood without such a stratigraphy.

Unfortunately, the maps produced prior to the 1930's commonly did not record in detail, nor interpret the stratigraphy of the greenstone belts (Pettijohn, 1937, p.155), and in many later reports of map areas, detailed stratigraphic studies are lacking.

Comparisons of the stratigraphic sections of various greenstone belts enable the basin evolutionary histories to be compared, and thus models for the basin developments to be established.

(b) Sedimentary interpretations

The major Archaean sedimentary lithologies can be listed in decreasing order of abundance as sandstones (commonly greywackes), conglomerates, argillites, and iron-bearing formations. There are specific problems associated with the deposition of each lithologic type, but some lithologies are better understood than others.

Prior to about 1940, much of the coarse clastic detritus of greenstone belts was thought to have been deposited under continental conditions of sedimentation (Pettijohn, 1943), but these early interpretations can now be revised due to the many advances in sedimentology which have occurred during the last 10 to 15 years, particularly in the areas of sediment transport processes, sedimentary structures and recent sediments.

In recent papers which emphasise the processes and environments of Archaean sediment deposition rather than the petrography, turbidites and related deposits are shown to be prevalent. Donaldson and Jackson (1965) suggested that thick sequences of greywacke-argillite couplets in the North Spirit Lake area of northwestern Ontario were turbidites, and that turbidity currents deposited most of the sediments in this area. McGlynn and

Henderson (1970) and Henderson (1972) demonstrated that most rocks in the Yellowknife supergroup of the Slave structural province are greywackes which were deposited by turbidity currents. Walker and Pettijohn (1971) examined the sedimentary rocks at East Bay in Minnitaki Lake and concluded that greywackes and also graded conglomerates were turbidites; and Campbell (1971) interpreted most sediments of the Rice Lake Group in Manitoba as being turbidites and related mass flow deposits.

No convincing descriptions of shallow agitated-water or terrestrial Archaean clastic sediments have been forthcoming in recent years, although Donaldson and Jackson (1965) do mention the possibility of shallow-water sedimentation accounting for the channels and cross-bedding found in ungraded sandstones of the North Spirit Lake area. Also, Holubec (1972, photo. 1) very briefly suggested that cross-stratified greywacke of the Cadillac Group in the Rouyn-Noranda area of Quebec, was deposited in shallow water; and Glikson (1971a) postulated a shallow-water origin for the Kurrawang Beds in Western Australia (but had reservations regarding this suggestion, as the sediments are immature).

(c) Basin geometry and tectonic influence

A number of difficulties are involved in determining the original size and shape of the Archaean depositional basins. Perhaps the greatest of these problems is the complex structure of most greenstone belts. The common vertical attitude of the sediments provides only a cross-section of the sedimentary pile and thus the locations of the original basin margins are in doubt.

The structural complexities also hinder paleocurrent determinations, as do the highly indurated, and commonly horizontal outcrops. The time-consuming process of block-sampling and sectioning of lithologies containing cross-stratification is necessary in order to determine paleocurrent directions in most cases.

Also, it has been suggested by Pettijohn (1943, p.941) that the conglomerates of greenstone belts represent wedge-shaped basin-margin accumulations, but the recent recognition of conglomerate turbidites in the Archaean (Walker and Pettijohn, 1971) casts doubt on this interpretation for all Archaean conglomerates.

Careful analyses of the sediments of greenstone belts may also shed light on tectonic processes such as faulting which may have operated during sedimentation.

(d) Provenance and nature of Archaean crust

Two of the most intriguing and difficult questions regarding Archaean geology are whether there was a sialic crust before deposition occurred in greenstone belts, and also whether intrusion and erosion of the volcanic piles occurred prior to the deposition of sediments. Archaean sedimentary rocks are potentially very important in this respect as they contain clasts within the conglomerates which may have been derived from sources external to the depositional basin.

However, Bass (1961) concluded that the "granitic" fragments in conglomerates in western Quebec and eastern Ontario were derived from flows or shallow intrusives emplaced contemporaneously with, and genetically related to the associated Archaean volcanics. Also, Ayres (1969) concluded that the Archaean sediments of Lake Superior Park were volcaniclastic, and were derived largely from felsic pyroclastic eruptions.

However, the situation may be rather different in other areas. Donaldson and Jackson (1965) considered the quartz budget in the sediments of the North Spirit Lake area, and concluded that a significant sedimentary and (or) granitoid provenance was reflected. Also, Walker and Pettijohn (1971) established the volume percentages of "granitic" clasts in some conglomerates at Minnitaki and Abram Lakes, and concluded that up to 50 percent by volume of the clasts ranged from leucocratic quartz diorites to granites, and thus there must have been plutonic rocks in the provenance area. McGlynn and Henderson (1971) and Henderson (1972) also considered a sialic provenance likely for the Archaean sediments in the Slave structural provenance.

When discussing the provenance of Archaean sediments the most important factors to consider are the clast types in the conglomerates and the quartz budget and rock fragments in the sandstones. Some studies in the past have quoted number percentages of clasts in the conglomerates, but it is the <u>volume</u> percentages which must be determined if provenance predictions are to be of value (Donaldson and Jackson, 1965).

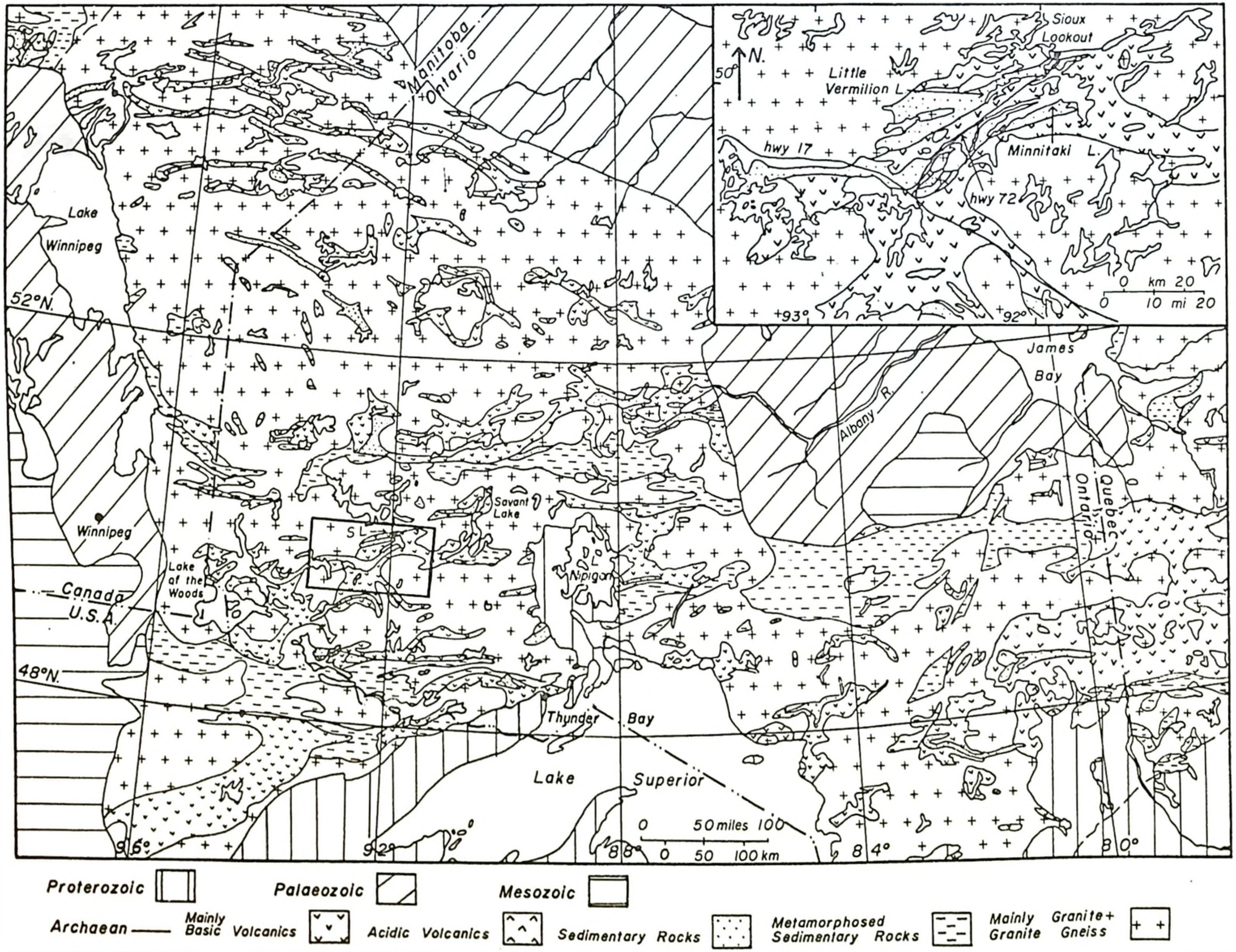
3. Sioux Lookout area: the thesis problem

The western part of the Superior Province of the Canadian Shield consists of a series of elongate, easterly trending greenstone belts, surrounded and intruded by younger Archaean granitic rocks (fig. 1). The greenstone belt in which Sioux Lookout is located, extends in an east-northeasterly direction from Lake of the Woods to Savant Lake.

To the south and southwest of Sioux Lookout the greenstone belt is about 25 km. wide, and has a symmetrical arrangement of 2 belts of sedimentary rocks, each up to 6 km. in width, which are separated from each other, and from the surrounding granites by 3 volcanic belts (fig. 1). The granitic rocks which border the greenstone belt have intruded the greenstones (Hurst, 1933).

Previous sedimentary studies within this greenstone belt near Sioux Lookout had shown that additional work in the area could prove profitable. Pettijohn (1934, 1935) showed that the area immediately north of Little Vermilion Lake in the northern sedimentary belt (fig. 1) was underlain by a thick sequence of arkoses and conglomerates, which was lithologically and stratigraphically distinct from a sequence of well-bedded greywacke-argillite couplets. Also, Walker and Pettijohn (1971) described another sedimentary sequence at Minnitaki Lake, within

Figure 1. Distribution of greenstone belts within the western part of the Superior Province of the Canadian Shield. Geology is from the Geological Survey of Canada, map no. 1250A - "Geological Map of Canada", and from the Ontario Department of Mines, map no. 1958B - "Geological map of the Province of Ontario".


Manitoba
Ontario
Sioux Lookout
N.
50°
Little Vermilion L.
hwy 17
Minnitaki L.
hwy 72
0 km 20
0 10 mi 20
93°
92°
Lake Winnipeg
52°N.
James Bay
Albany R.
Savant Lake
Ontario
Quebec
Winnipeg
SL
L. Nipigon
Lake of the Woods
Canada
U.S.A.
48°N.
Thunder Bay
Lake Superior
0 50 miles 100
0 50 100 km
96°
92°
88°
84°
80°
Proterozoic
Palaeozoic
Mesozoic
Archaean
Mainly Basic Volcanics
Acidic Volcanics
Sedimentary Rocks
Metamorphosed Sedimentary Rocks
Mainly Granite
Granite + Gneiss

the southern sedimentary belt. The sedimentary rocks at Minnitaki Lake were concluded to be turbidites and related deposits, which differed from Pettijohn's earlier suggestions of the origin of the sediments in the northern sedimentary belt (Pettijohn, 1934).

Thus, a detailed examination of the sedimentary rocks in the northern sedimentary belt (Abram Group - see Chapter 2) around Little Vermilion and Abram Lakes (fig. 2) was undertaken, with the following objectives in mind:

(i) To map the area, elucidate the detailed stratigraphy, and describe the sedimentary lithologies and structures;

(ii) To suggest the varying processes and conditions of sedimentation, in the light of recent advances in sedimentology;

(iii) To outline the sequence of basin filling from the stratigraphy;

(iv) To attempt to draw conclusions regarding the nature and tectonic setting of the provenance area, and its relationship to the basin of deposition.

During 3 months of the summer of 1971, field mapping, stratigraphic measurement, pebble analysis of the conglomerates, and sampling of the lithologic units were carried out. Petrographic analyses of these samples and 8 whole rock chemical analyses have since been completed.

CHAPTER 2

GEOLOGY OF THE SIOUX LOOKOUT AREA

1. Previous work

Hurst (1933) gave a summary of the earliest geological investigators who passed through the Sioux Lookout area in the early 1900's, but reports from these investigations are only brief descriptions of the notable rock types encountered.

Hurst (1933) was the first to produce a map of the area, and he classified the volcanic rocks as "Keewatin", and the sedimentary rocks as "Timiskaming".

In the 1930's, a series of papers by Pettijohn discussed the conglomerates at Abram Lake (1934), the stratigraphy and structure of Vermilion Township (1935), the geology of East Bay in Minnitaki Lake (1936), and Archaean geology and correlational problems of northwestern Ontario and Minnesota (1937).

Johnston (1966a,b, 1967a,b, 1968, 1969) mapped the Sioux Lookout area at a scale of 4 inches to 1 mile, and the most recent work is a sedimentary study by Walker and Pettijohn (1971) in the East Bay of Minnitaki Lake.

2. Structure and stratigraphy of the greenstone belt

Both the northern and the southern boundaries of this greenstone belt are marked by intrusive granites and granite gneisses (Hurst, 1933; and fig. 2). In the Sioux Lookout area, the northern boundary is exposed at Pelican Lake, where recrystallisation, injection and assimilation of greenstones has taken place. According to Hurst (1933, p.15) "the greenstones have been largely converted to amphibolite, and in many places they have been penetrated by lit par lit injection of granite. The granite gneiss contains many streaks and patches of partially assimilated greenstone ".

The southern margin of the belt is generally obscured by drift, but both granite and gneiss crop out near the boundary. Numerous pegmatite dikes cut the gneiss, whilst aplite dikes and quartz veins intrude the greenstones near the granite contact (Hurst, 1933).

Hurst considered all the granites to be of Algoman (Kenoran) age.

The greenstone belt at Sioux Lookout consists of 5 zones, which have east-northeasterly trends (figs. 1 and 2):

(a) Northern Volcanic Belt;

(b) Northern Sedimentary Belt (Abram Group);

(c) Central Volcanic Belt;

(d) Southern Sedimentary Belt (Minnitaki Group);

(e) Southern Volcanic Belt.

(a) <u>Northern Volcanic Belt</u>

The northern volcanic belt was described by Pettijohn (1935).

He classified the rocks as Keewatin in age, and divided the belt into 3 units:

(iii) Younger volcanics - 610+ m. thick

(ii) Patara sediments - 380-460 m. thick

(i) Older volcanics - 2,7000+ m. thick

The Older volcanics occur in the north, adjacent to the granites, and comprise massive basic lavas together with pillow lavas (fig. 3) indicating a southerly top direction (fig. 2). The Patara sediments are volcanogenic and consist of greywackes, argillites, tuffs, and breccias and conglomerates of greenstone and felsite fragments (commonly less than 10 cm. in diameter) (fig. 4). No granitic detritus is present within the Patara sediments.

The Younger volcanics, at the south of the northern volcanic belt, comprise massive and pillowed lavas (again indicating a southerly top direction in most places - fig. 2), volcanogenic breccias and tuffs, and quartz and feldspar porphyries.

(b) Northern Sedimentary Belt (Abram Group)

Pettijohn (1935) used the term "Abram Series" for the rocks of the northern sedimentary belt, but "Abram Group" is used here as a lithostratigraphic rather than a time-stratigraphic term. Pettijohn (1935) also divided the Abram Group into 2 formations:

(ii) Daredevil Formation - 650+ m. thick

(i) Ament Bay Formation -2,150 m. thick

However, for descriptive purposes it is convenient here to establish 3 formations of the Abram Group (fig. 6):

(iii) Little Vermilion Formation (dominantly greywackes) - 570+ m. thick

(ii) Daredevil Formation (dominantly tuffs) - 430 m. thick

(i) Ament Bay Formation (dominantly arkoses and conglomerates) - 1,700 m. thick

Formations (ii) and (iii) are equivalent to Pettijohn's "Daredevil Formation". Pettijohn's value of 2,150 m. for the thickness of the Ament Bay Formation is thought to be excessive, whilst the thickness of the remainder of the Abram Group was underestimated by Pettijohn.

The Abram Group is in part unconformable on the northern volcanic belt, but the contact is faulted for most of its length. Most of the Abram Group also youngs in a south to southeasterly direction (fig. 2).

The stratigraphy and structure of the Abram Group at Little Vermilion and Abram Lakes are discussed in detail in sections 4 and 5 of this chapter.

(c) <u>Central Volcanic Belt</u>

The central volcanic belt is separated from the Abram

Figure 2. Geology of the greenstone belt at Sioux Lookout, northwestern Ontario. Sources - Pettijohn (1936), Johnston (1966a,b, 1967a,b, 1968, 1969), Walker and Pettijohn (1971), and author's mapping.

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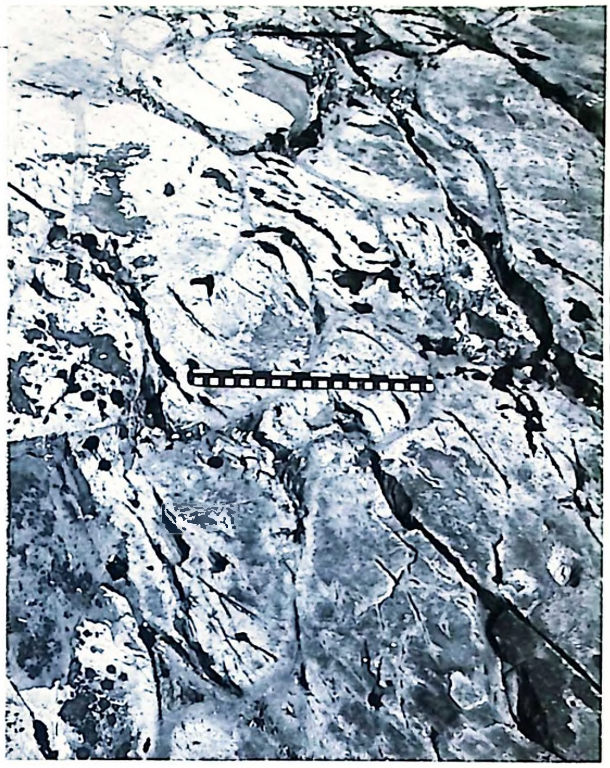


Figure 3. Typical pillow lava of the Northern Volcanic Belt. Arrow indicates top direction. Scale in inches and centimetres. Vermilion Lake.

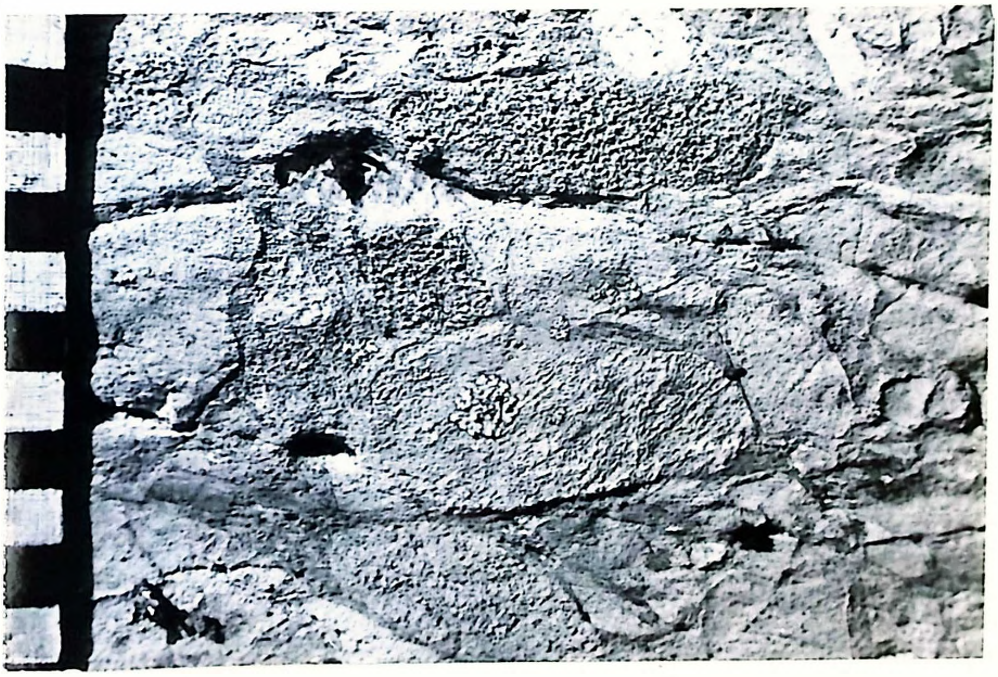


Figure 4. Volcanogenic conglomerate of the Patara sediments, consisting of greenstone clasts in a chloritic matrix. Scale in centimetres. Vermilion Lake.

Group of sediments by the Little Vermilion fault which extends through the length of Little Vermilion Lake, and can be traced eastward through Abram Lake (fig. 2).

Way-up indicators within the central volcanic belt are more scattered than those in the northern volcanic belt, but most top determinations from pillow lavas indicate a younging of the volcanics in a general southerly direction (fig. 2 and Johnston, 1966a,b, 1967a,b, 1968). Only one determination out of 15 is at variance with this statement in that part of the central volcanic belt to the west of Minnitaki Lake. The generally southwesterly orientations of the pillow lava top directions in Northeast Bay of Minnitaki Lake probably result from the intrusion of the nearby diorite and granite bodies (fig. 2).

The faulted relationship of the central volcanic belt with the Abram Group of sediments makes their time-relationship uncertain, but the southerly top directions in the central volcanic belt suggest that the central volcanics are younger than the Abram Group.

(d) Southern Sedimentary Belt (Minnitaki Group)

The Minnitaki Group (defined by Walker and Pettijohn, 1971) is more tightly folded and more extensively faulted than the Abram Group, but Walker and Pettijohn (1971) mapped 5 sedimentary formations in East Bay of Minnitaki Lake, which although folded, are predominantly older in the north of East

Bay than in the south.

The relationship of the Minnitaki Group to the central volcanic belt is uncertain in East Bay because of the faulted nature of the area, and the probable faulted contacts (fig. 2). Pettijohn (1936, p.349) suggested that the central volcanic belt was younger than the Minnitaki Group of sediments, although he did point out that this view had not been proved. However, interpretations of the same contact by Hurst (1933, p.12) and Johnston (1969, p.17) along Pickerel Arm of Minnitaki Lake (fig.2), are at variance with the view expressed by Pettijohn. They suggested that a transition occurred from the volcanics of the central volcanic belt, through pyroclastic material, to the sediments of the Minnitaki Group (see Johnston, 1969, for a map of this area).

This latter view was also supported by Armstrong (1950, p.18), from his mapping in Echo Township, to the west of Pickerel Arm, and about 5 km. due south of Maskinonge Lake (fig. 2).

(e) Southern Volcanic Belt

The relationship of the Minnitaki Group to the southern volcanic belt is also uncertain, but the contact is probably faulted. Pettijohn (1936) found that both sediments and volcanics younged to the north in Twinflower Bay (fig. 2), and thus suggested that the southern volcanic belt was older than the Minnitaki Group. Farther west, south of Pickerel Arm (fig. 2), an ambiguous relationship exists between the southern volcanic belt and the Minnitaki Group. Pettijohn (1937, p.174) shows the boundary as a fault which truncates a synclinal axis in the southern volcanics, but Johnston's map (1969) shows no such syncline and very few top determinations in the volcanics. Johnston also states (1969, p.17) that "the contact shows a gradation from sedimentation to volcanism, or vice versa, with age relationships between the metasediments and metavolcanics being unknown.".

Thus it is possible that the southern volcanic belt is older than the Minnitaki Group, but the published data are conflicting.

(f) Summary of age relationships within the greenstone belt

The northern volcanic belt is older than the Abram Group of sediments which, in turn, is probably older than the central volcanic belt. The Minnitaki Group of sediments is probably younger than the central volcanic belt, and the southern volcanic belt is possibly older than the Minnitaki Group. Thus it is probable that the Abram and Minnitaki Groups of sediments are not of the same age.

3. Metamorphism

The metamorphic grade of most rocks throughout the greenstone belt is low, and within the Abram Group the metamorphic minerals present in the rock matrices are commonly quartz, colourless mica, chlorite and albite, with a small amount of epidote. Secondary biotite is found in some places, but is not common. Most larger detrital grains in the sediments have not been completely recrystallised, and many (especially the quartz grains) appear to have suffered no alteration (see Chapter 5).

This mineral assemblage is characteristic of the lower to middle greenschist facies of metamorphism, and represents a low temperature thermal event (Winkler, 1967). Higher metamorphic grades occur near the granitic intrusions, but Little Vermilion Lake is at least 7 km. from the nearest granites, and thus contact metamorphic effects are negligible.

At Little Vermilion Lake many sedimentary structures are visible within the rocks (Chapter 3), and few outcrops have schistose textures. Thus, penetrative movements did not accompany the recrystallisation.

4. Structure and stratigraphy of the Abram Group in the Little Vermilion Lake area

(a) Introduction

The major structural and stratigraphic features of the Abram Group of sediments between Vermilion and Little Vermilion Lakes (fig. 2 and map 1) are summarized as follows.

(i) An unconformity exists between the quartz porphyry body at Vermilion Lake, and the overlying basal conglomerate of the Abram Group. This quartz porphyry has intruded greenstones of the northern volcanic belt, and hence the Abram Group is younger than the northern volcanic belt.

(ii) Most of the sedimentary rocks indicate a top direction to the south-south-east. However, an anticline exists near the northern boundary of the sedimentary belt, and a north-east trending syncline occurs near the southern boundary of the belt, between Little Vermilion and Maskinonge Lakes (fig. 2).

(iii) Both the northern and southern boundaries of the sedimentary belt are marked by faults for most of their lengths, which cut out both the northerly limb of the anticline at Vermilion Lake, and much of the southerly limb of the Little Vermilion Lake syncline (figs. 2 and 5B).

(b) <u>Stratigraphic relationships between the major sedimentary units</u>

The detailed stratigraphic sequence of the sedimentary facies within the Abram Group at Little Vermilion Lake is shown in fig. 6, but the 5 stratigraphic units discussed below are tabulated as follows.

Formation	Stratigraphic unit	Thickness	Stratigraphic position
Little Vermilion	(v) Greywacke	570+ m	2,130-2.700 m
Daredevil	(iv) Basic tuff	270 m	1,860-2,130 m
	(iii) Greywacke	60 m	1,800-1,860 m
	(ii) Felsic tuff	100 m	1,700-1,800 m
Ament Bay	(i) Arkosic sandstone and granite-bearing conglomerate	1,700 m	0-1,700 m

The base of the sedimentary sequence is only exposed at one locality throughout the whole area, where a basal conglomerate containing greenstone and quartz porphyry clasts rests unconformably on a quartz porphyry body (for lithological descriptions see pp.35 and 36). Pettijohn (1935) noted this unconformable relationship, and also found that the quartz porphyry body is interlayered with, and apparently intrusive into agglomerates of the underlying volcanic succession. This interpretation was also supported by mapping during the present study (fig. 2 and map 1).

The basal conglomerate of the Abram Group passes southward (up section) into typical arkosic sandstone, which together with the granite-bearing conglomerate facies form the Ament Bay

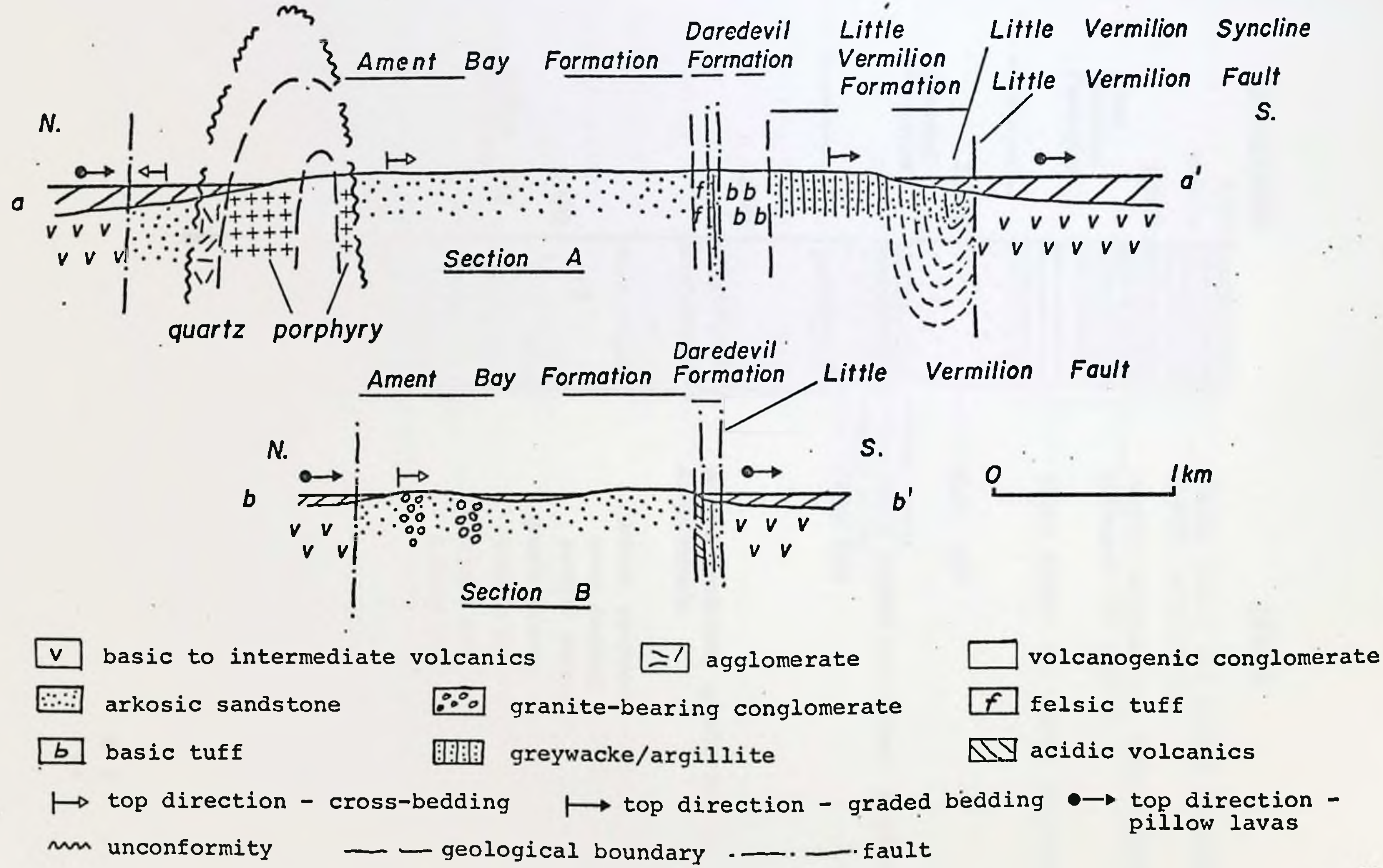


Figure 5. Stratigraphic sections through the Abram Group at Little Vermilion Lake. Lines of sections shown on figure 2.

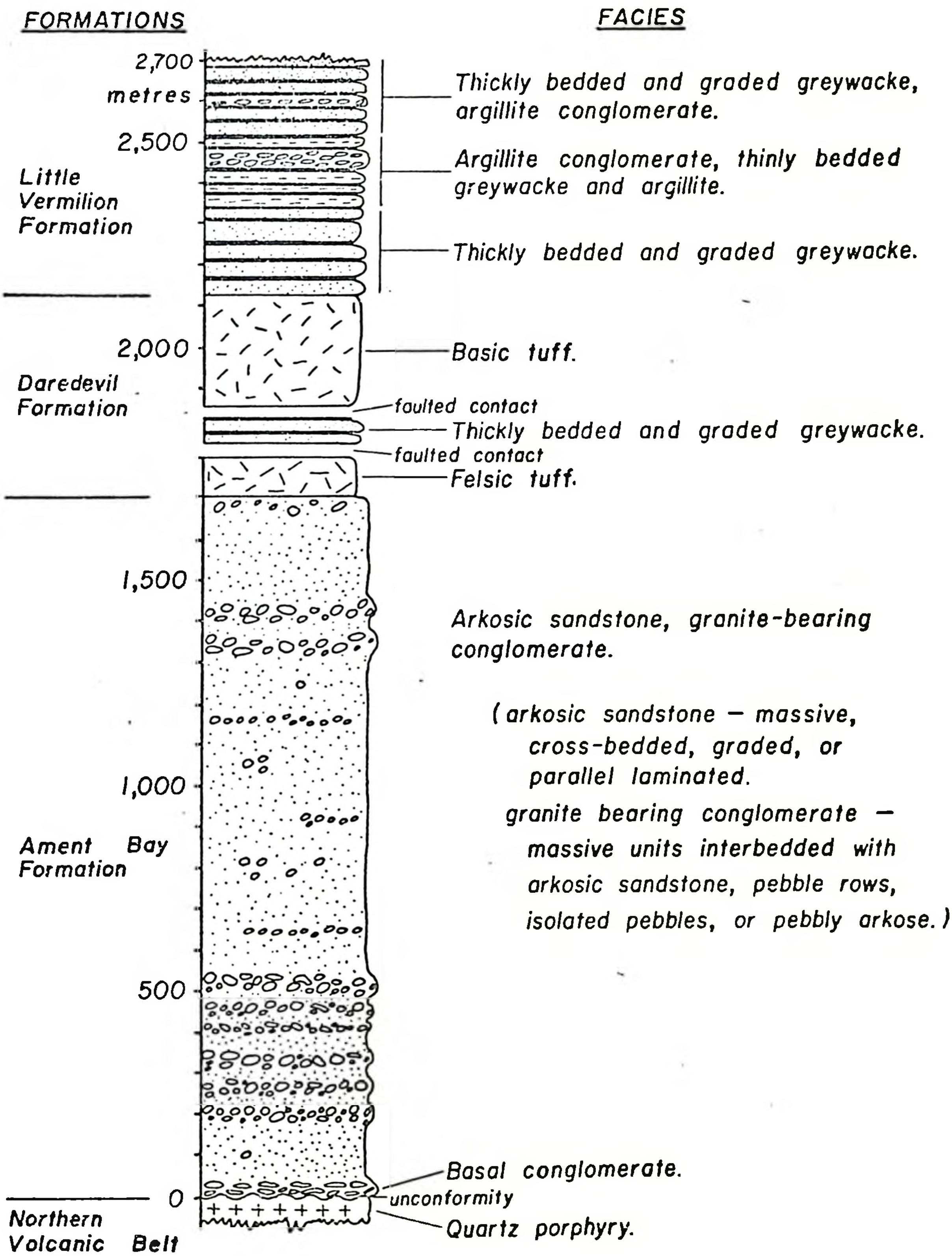


Fig. 6 Stratigraphy of the Abram Group at Little Vermilion Lake

Formation, and persist for approximately 1,700 m. of the stratigraphic section (fig. 6).

The contact of the Ament Bay Formation with the overlying felsic tuff facies of the Daredevil Formation is apparently conformable, but the actual contact is not exposed (see Facies 3, p. 87). The felsic tuff facies occupies 100 m. of section and has a faulted contact with the overlying unit of the greywacke facies. This contact is marked by a low swampy area 10 to 20 m. wide, and the strike of the felsic tuff is about 80°, whilst that of the greywacke is about 40° (map 1).

This greywacke unit is 60 m. thick, and probably also has a faulted contact with the overlying basic tuffs (fig. 6), the boundary at Little Vermilion Lake again being marked by a swampy area.

The top of the basic tuff facies is apparently conformable with the greywackes of the Little Vermilion Formation (see Facies 4, p. 97), but the top of the greywacke facies is not observable, due to the presence of the Little Vermilion Syncline.

(c) Structure

All the bedding surfaces throughout this area either dip at very high angles, or are vertical, and in most places the bedding surfaces continue straight across the width of outcrop.

Shoreline outcrops within the Abram Group were found to be by far the most useful for a sedimentary study. Outcrops away from the shore are commonly lichen-covered, and are generally only of use in determining the lithology and the most pronounced of sedimentary structures. Also, most shoreline outcrops have been smoothed as a result of Pleistocene erosion, and hence, almost all outcrops described in this study are essentially cross-sections through the sedimentary beds.

Way-up determinations within the sedimentary rocks were possible using graded bedding within the greywackes, and using cross-bedding within the arkosic sandstones (see fig. 2 for locations of way-up determinations).

(i) Folding

Two major fold axes occur within the sedimentary belt. The northern anticlinal axis probably passes through the quartz porphyry body at Vermilion Lake (figs. 2 and 5A). At the southern boundary of the quartz porphyry body, the succession youngs to the south. However, the quartz porphyry terminates to the east, and at locality A at Vermilion Lake (fig. 2) the basal conglomerate of the sedimentary succession is exposed to the north of the quartz porphyry, suggesting the top direction is to the north at this point. Also, at the nearby island B (fig. 2), an outcrop of typical arkosic sandstone occurs, which contains excellent cross-bedding (fig.28) also indicating a top direction to the north.

The anticlinal axis does not extend eastward from the quartz porphyry body into Ament Bay, and so must either have a rather steep easterly plunge with the northern limb essentially dying out eastward, or the axis must be displaced northward by faulting. All the way-up determinations around Ament Bay indicate a top direction to the south-east (figs. 2 and 5B). Also, all the top determinations at Little Vermilion Lake to the east of Ament Bay, indicate a southerly top direction, as do many greywacke beds in the south-west part of the lake.

However, a few northerly way-up determinations in the greywacke beds at the south-west end of Little Vermilion Lake indicate the presence of the Little Vermilion syncline (figs. 2 and 5A), which trends south-westward to Maskinonge Lake where graded beds again indicate a northerly top direction.

(ii) Faulting

Apart from the one outcrop of basal conglomerate mentioned above, neither the northern nor the southern boundaries of the sedimentary belt are exposed, but it is likely that faults mark these contacts with the adjacent volcanic belts.

The Little Vermilion Fault forms the southern boundary of the sedimentary belt, and in the south-western part of Little Vermilion Lake it truncates the Little Vermilion Syncline, as well as the Little Vermilion Formation and much of the Daredevil Formation (fig. 2). In the eastern part of Little Vermilion Lake, the fault occurs between volcanics and arkosic sandstones or conglomerates, and is marked in places by very small, schistose, carbonatized outcrops.

The northern boundary of the sedimentary belt is not exposed for most of its length, but the presence of a few schistose arkosic sandstone outcrops along the south shore of Vermilion Lake (map 1) may indicate the presence of a fault. Also, the northern limb of the anticline is truncated, which further implies faulting.

Several smaller north-easterly trending faults occur, which separate the felsic tuff facies and the basic tuff facies from the 60 m. section of greywacke in the south-west of Little Vermilion Lake (fig. 2). The greywacke section to the south-east of Ament Bay (map 1), is also fault bounded, and the section of black slate at island A in the east of Little Vermilion Lake may have faulted contacts.

Apart from these faults which trend to the north-east and east-north-east, there are a few which have a northerly trend. The two best examples of the north-trending faults are first, a fault which is inferred to trend through the narrows at the entrance to Ament Bay (fig. 2) because a north-easterly striking acidic volcanic unit (probably part of the felsic tuff facies of the Daredevil Formation - see p. 87) which crops out on the east side of the narrows, does not occur on the west side. Thus, some sinistral offset has taken place, which could amount to about 200 m. laterally if the acidic volcanic unit is in fact the stratigraphic equivalent of the felsic tuffs further west. The second example is in Vermilion Lake, to the north of

Ament Bay (fig. 2), where a northward trending fault occurs with a dextral displacement of about 400 m.

The time relationship between the northerly and north-easterly trending faults is uncertain, because neither set of faults definitely displaces the other.

(iii) Cleavage

In most outcrops of arkosic sandstone or granite-bearing conglomerate, cleavage is not visible, and thin sections from these outcrops show no preferred orientation of grains. However, cleavage is not uncommon in greywacke and does occur in a few places within the more massive arkoses and conglomerates. Displacements of a few cm occur along the cleavage plane in some places.

The strikes of most cleavages at Little Vermilion Lake lie between 70 and 110°, and either lie in the plane of the bedding, or are oriented at angles up to 40° in a more southerly direction than the bedding, but with the cleavage plane remaining essentially vertical. Thus, cleavage and bedding intersections are also approximately vertical.

In a few beds of greywacke, the graded nature of the beds is emphasised by refracted cleavage (fig.53).

5. Structure and stratigraphy in the Abram Lake area

(a) Stratigraphy

There are no unambiguous way-up structures within the sediments exposed around the central part of Abram Lake (between Frog Rapids and Prospector Island - fig. 2), but it is likely that here the sedimentary succession youngs to the north-north-west. This is suggested because of the similarity of the stratigraphy at Little Vermilion Lake (fig. 6), and that at the east end of Abram Lake (near the Sturgeon River), where the top direction is indicated by graded bedding in greywackes (fig. 2). At these 2 localities the sequences young to the south, and pass from arkosic sandstones and granite-bearing conglomerates, to felsic tuffs, and then to greywackes. In the central part of Abram Lake a thick sequence of felsic tuffs lies to the north of the arkosic sandstones and granite-bearing conglomerates. Therefore the sedimentary sequence in this central part of the lake probably youngs to the north.

(b) Structure

A synclinal axis must exist between the central part of Abram Lake, and the eastern end of the same lake, to account for the difference in top directions, but its location is uncertain (fig. 2).

The southern boundary of the sedimentary succession at Abram Lake is marked by a fault, which passes through island A and Long Island (fig. 2) in the central part of the lake, and continues through the islands in the eastern arm of Abram Lake. On these islands the rocks are highly sheared, and in places carbonatization has occurred (Johnston, 1967b). This fault may be a continuation of the Little Vermilion Fault.

The northern boundary of the sedimentary belt in the Abram Lake area is nowhere exposed, but is probably in fault contact with the northern volcanic belt. This fault probably passes through the southern part of Pelican Lake, near Frog Rapids, and its westerly extension may pass through Patara Lake to Vermilion Lake (fig. 2). The volcanogenic Patara sediments of the northern volcanic belt crop out at Vermilion Lake, along the north shore of Patara Lake (Pettijohn, 1935), and along Highway 116 (fig. 2). The meta-sedimentary rocks exposed at Pelican Lake are also probably extensions of the Patara sediments, but they are highly contorted and "have probably been influenced by 'shallow-depth' granitic intrusions similar to the small dome-like bodies in the metavolcanics to the north" (Johnston, 1967b). The Patara sediments continue eastward from Pelican Lake until they are truncated, probably by faulting, near the east end of Abram Lake (fig. 2).

Thus, the northern boundary of the northern sedimentary belt is probably marked by a fault, which truncates an anticlinal axis at Vermilion Lake, the Younger volcanics of the northern volcanic belt between Vermilion Lake and Patara Lake, and the Patara sediments near the east end of Abram Lake (fig. 2). This fault is apparently displaced northward by a south-east trending fault near the east end of Abram Lake (fig. 2).

The relationship between the felsic tuffs in the central part of Abram Lake and the Ament Bay Formation at the east end of Little Vermilion Lake is not certain, but there may be a synclinal axis between the two areas, as shown in figure 2.

CHAPTER 3

FACIES FIELD DESCRIPTIONS

1. Introduction

During field mapping of the rocks around Little Vermilion Lake, it became apparent that the Abram Group could be divided into 5 facies in the stratigraphically unrestricted sense. Lithology and sedimentary structures were the criteria used to distinguish these facies in the field.

The following facies are present within the Abram Group -

(1) Granite-bearing conglomerate

(2) Arkosic sandstone

(3) Felsic tuff

(4) Basic tuff

(5) Greywacke, with subfacies -

(a) thickly-bedded and graded greywacke;

(b) thinly-bedded greywacke and argillite;

(c) argillite conglomerate.

The greywacke facies has been divided into 3 subfacies due to the wide range of both the sedimentary structures and lithologies of the greywackes.

The stratigraphic distribution, and the relationships between the facies types are shown in figure 6. The following upward sequence is apparent - granite-bearing conglomerate and arkosic sandstone (0-1700 m) - felsic tuff (1700-1800 m) - thickly bedded and graded greywacke (1800-1860 m) - basic tuff (1860-2130 m) - thickly bedded and graded greywacke (2130-2310 m) - thinly bedded greywacke and argillite, argillite conglomerate, and thickly bedded and graded greywacke (2310-2700 m).

The field characteristics of the 5 facies are described in detail below.

2. Granite-bearing conglomerate facies

(a) Introduction

Granite-bearing conglomerates are characterized by the presence of pebbles, cobbles and boulders of various lithologies, together with a matrix (commonly about 30 percent by volume) of arkosic sand. The facies derives its name from the conspicuous leucocratic medium grained "granitic" clasts which constitute nearly 40 percent of some conglomerate outcrops (fig. 7). These clasts have granodiorite compositions and granitic textures (see Chapter 5). The conglomerates are poorly sorted, and most clasts are well rounded.

Within the granite-bearing conglomerate facies, four varieties were distinguished in the field,

(i) Basal conglomerate; exposed at only one outcrop, where it rests unconformably on a quartz porphyry body.

(ii) Massive conglomerate with arkosic sandstone interbeds; individual conglomerate units are up to 20 m in thickness, and the arkosic sandstone interbeds are commonly less than 1 m thick.

(iii) Pebble bands within arkosic sandstone; the thickness of the bands are commonly less than the diameters of the largest included clasts.

(iv) Pebbly arkosic sandstone; the clasts are dispersed and matrix supported.

Types (ii) and (iii) are the most common, and the massive conglomerate of type (ii) volumetrically comprises most of the granite-bearing conglomerate facies. All types (except the basal conglomerate) occur throughout the stratigraphic range of the Ament Bay Formation, and are never found in the Daredevil or Little Vermilion Formations (fig. 6). The thickest units of massive conglomerate are found in the lower part of the Ament Bay Formation (on the north side of Ament Bay), and also towards the top of the formation (at Kenneally Lodge, and in the east of Little Vermilion Lake). The geographic distribution of this facies is shown on map 1.

Due to the patchy nature of the exposure around Little Vermilion Lake it is difficult to trace conglomerate units along strike. In some places, where a thick unit of predominantly conglomerate is mapped as being replaced laterally by arkosic sandstone, it may be the result of faulting and not an original depositional characteristic. For example, at the small bay to the west of the entrance to Ament Bay (map 1), a unit of predominantly granite-bearing conglomerate (at least 120 m thick) is replaced within about 100 m along strike by a thick sequence of arkosic sandstone. In this case it is likely that a fault separates the conglomerates and the arkoses (map 1).

However, the localized distribution of massive conglomerate throughout the Ament Bay Formation (fig. 2) makes it unlikely that thick conglomerate units persist far along the strike. At Little Vermilion Lake no unit of predominantly massive conglomerate extends for more than 1 km, and no such unit has a thickness greater than about 300 m. However, the conglomerate unit which crops out on the islands in Abram Lake (fig. 2) is over 3 km in length, and has a maximum thickness of about 350 m.

The relationships between the various conglomerate types and the arkosic sandstone facies became apparent by measuring field sections containing granite-bearing conglomerates and arkosic sandstones (Appendix 1). Four major relationships are - (1) Massive conglomerates up to 20 m in thickness are commonly interbedded with thinner bands of massive or cross-bedded arkosic sandstone. (2) In only a few places are pebble bands found within these interbedded arkoses. (3) Pebbly arkosic sandstone is not commonly associated with massive conglomerates, but occurs within thick sequences of predominantly arkosic sandstone. (4) Pebble bands and isolated pebbles are common in thick arkosic sandstone sequences.

In the Ament Bay Formation the granite-bearing conglomerate is never interbedded with rock types other than the arkosic sandstone facies. At Minnitaki Lake (fig. 2) conglomerates were described by Walker and Pettijohn (1971) as being graded and having thin slate partings at their tops. No such slate partings, nor any other argillite bands are present within the granite-bearing conglomerates at Little Vermilion or Abram Lakes.

However, one problematical slate occurrence was found at Island A in the eastern part of Little Vermilion Lake (fig. 2). Here, an 80 m section of well-cleaved black slate occurs between highly sheared granite-bearing conglomerate adjacent to the Little Vermilion Fault, and typical conglomerate and arkose at the north end of the island. This slate may represent part of the greywacke facies of the Daredevil Formation (fig. 6), which has been faulted into its present position.

(b) Conglomerate types

The 4 types of granite-bearing conglomerate outlined above, are described in detail below.

(i) Basal conglomerate

Only one locality was found which shows the base of the sedimentary succession (fig.2). This is located near the south shore of Vermilion Lake.

Here, the conglomerate consists of poorly sorted, rather angular fragments of quartz porphyry, felsic volcanics, and greenstone, together with a few quartz "eyes" up to 5 mm in diameter, set in a fine grained chloritic matrix. The largest clast (composed of quartz porphyry) measures 40 cm by 15 cm, but most fragments are less than 10 cm in length. This conglomerate has a very sharp contact with a massive, slightly schistose quartz porphyry body, which is identical in lithology to the quartz porphyry fragments in the conglomerate (fig. 10). The quartz porphyry weathers to a pale pink or cream colour, is predominantly felsic, and contains scattered euhedral quartz phenocrysts, up to 1 cm in size.

The conglomerate clasts have their long axes aligned parallel to the contact with the quartz porphyry body, but no sedimentary structures such as grading or cross-bedding were observed in the outcrop. The alignment parallel to the contact is probably tectonic in origin, as most clasts are stretched, and in places it is difficult to distinguish the matrix from the clasts.

The contact of the basal conglomerate with the quartz porphyry body is sharp and slightly curved (concave towards the conglomerate) along the 2 m length of exposure (fig. 10), but when traced laterally this appears to be a minor irregularity in an otherwise straight contact.

The sharp contact of the two rock types, plus the presence of quartz "eyes" and quartz porphyry fragments (identical to the underlying quartz porphyry) within the basal conglomerate, are good evidence that this surface is an erosional unconformity, immediately overlain by angular debris.

The quartz porphyry shows no obvious signs of weathering prior to the deposition of the conglomerate, and has a constant texture throughout the width of the outcrop, and also within the conglomerate clasts. Other outcrops near the base of the sedimentary succession are also rather sheared, but fragments of quartz porphyry and greenstone can be distinguished. However, these outcrops are generally small and poorly exposed.

(ii) Massive conglomerate with arkose interbeds

These massive conglomerates are clast supported, and the interstices are filled with arkosic sand, which is indistinguishable in the field, both in composition and grain size, from adjacent arkose beds. In most massive granite-bearing conglomerates this arkosic matrix comprises about 30 percent by volume of the rock (fig. 7), and is of coarse sand grade.

Most pebbles, cobbles and boulders in undeformed massive conglomerates are rounded to well rounded, but leucocratic clasts are commonly more spherical than the basic greenstone clasts. The massive conglomerates are poorly sorted, and maximum clast diameters at various outcrops range from about 10 cm to 1.5 m (the largest boulder in the whole area). Clast size and shape are discussed in more detail under separate headings below.

Massive conglomerate beds can reach 20 m in thickness, but arkosic sandstone beds commonly divide the conglomeratic sequences, so that most conglomerate beds are between 1 m and

10 m thick. Some conglomerate outcrops show a crude bedding, due to clast size changes (fig. 11), or due to a rough alignment of the long axes of clasts parallel to the contacts with arkose interbeds. This alignment is particularly noticeable when large numbers of greenstone pebbles are present, but when outcrops include many of the more spherical granitic clasts, no bedding is discernible.

At several outcrops around Little Vermilion Lake, pebbles within conglomerate units are inclined at angles to the known strike directions (fig. 12). In these cases (where no tectonic foliation is obvious) the structure is thought to be of primary sedimentary origin, akin to imbrication. Because this type of structure is also found within pebbly arkosic sandstone and pebble bands within arkosic sandstone, it will be described in more detail under a separate heading below.

No cross-stratification on any scale was observed in the conglomerates, and graded bedding is rare. At a few localities a decrease in the maximum clast sizes does occur toward the top of a conglomerate bed (fig. 13), but this is atypical of most conglomerate occurrences. Regular changes of the maximum clast sizes throughout the thickness of a bed are not present in any of the conglomerate units.

Arkosic sandstone beds are commonly interbedded with the massive conglomerate units, and range in thickness from a few cm (fig. 13) to over 4 m. These arkose interbeds are normally massive, but in a few cases they are cross-bedded (see Arkosic sandstone facies description).

The bases of most massive conglomerate beds are sharp and commonly straight (fig. 12). Individual clasts may protrude into the underlying arkose bed, but this is probably a loading phenomenon. The tops of granite-bearing conglomerate beds are usually more irregular than the bases, due to some pebbles being completely enclosed within the overlying arkose bed (fig. 11).

Only one example of a massive conglomerate filling a small scour in arkosic sandstone was found throughout the Ament Bay Formation. At an island in Closs Lake (map 1), a scour at least 20 cm deep and 1 m wide truncates faint parallel laminations in the underlying arkose, and is filled with arkose and granite-bearing conglomerate (fig. 14). The basal 10 cm or so of the scour-fill consists of very coarse sand with small pebbles, whilst above this the typical massive conglomerate occurs, with pebbles and cobbles up to 10 cm in diameter.

<u>Clast types of massive conglomerates</u> - The percentages of clast types in various outcrops of massive conglomerates were measured in the field by a point counting method, using a constructed 50 cm square grid with wires at 5 cm intervals. At some small or dirty outcrops only 121 points were counted, but at cleaner, larger outcrops, up to 484 points were counted. The results of these measurements are plotted as histograms in

figure 7.

Greenstone is commonly the most abundant clast type (both numerically and volumetrically), but conspicuous leucocratic granitic clasts comprise nearly 40 percent of some conglomerate outcrops. In the field, the term "greenstone" was used for clasts which are dark green in colour and which are very fine grained. These clasts are probably metabasalt fragments, and are similar in appearance to the typical Archaean greenstone of the volcanic sequences in greenstone belts. A few clasts which were termed greenstones in the field are of coarser grain size, and probably represent metagabbros or metadiorites.

Granitic clasts are typically equigranular, and consist of quartz and feldspar crystals up to 2 mm in diameter. Ten thin sections from granitic clasts were point counted, and the rocks classified as granodiorite (Chapter 5).Rare granitic clasts (6 were noted throughout the whole study area) contain xenoliths of greenstone up to about 3 cm in size (fig. 15). These xenoliths are angular to rounded in outline, and are apparently identical in composition to the greenstone clasts in the conglomerates.

Other clast types present within the granite-bearing conglomerates include felsic fragments (probably originally acidic volcanics), "chert", quartz porphyry, felsic tuff, vein quartz, and a few foliated granitic (or gneissic) clasts. Petrographic

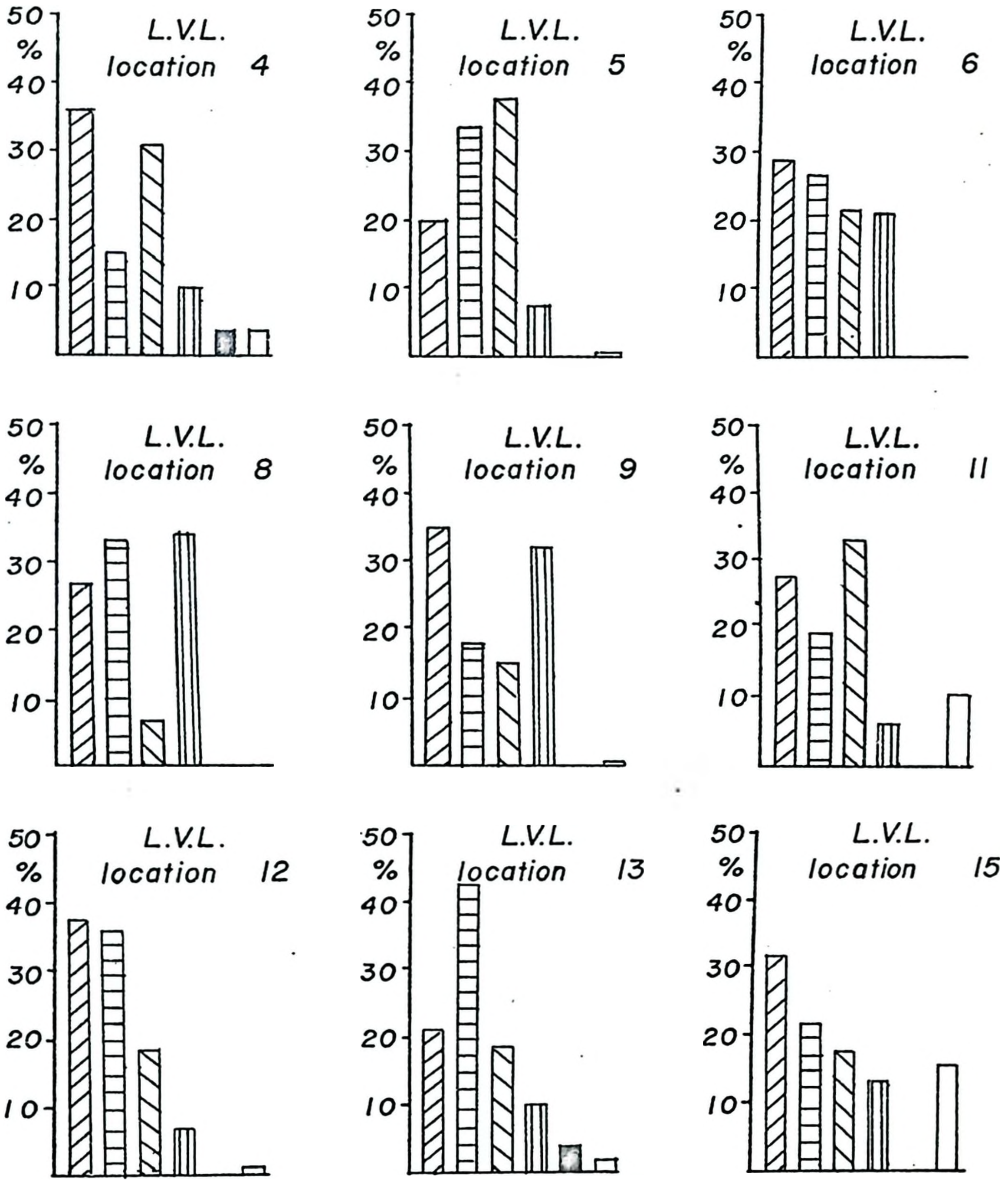


Figure 7a. Percentages of clast types within outcrops of granite-bearing conglomerate at Little Vermilion Lake (L.V.L.). Locations are shown on map 1.
▨ matrix, ▤ greenstone, ▧ granitic, ▥ felsic (including "chert"), ■ foliated granitic, □ remainder (quartz and feldspar porphyry, felsic tuff and "vein quartz").

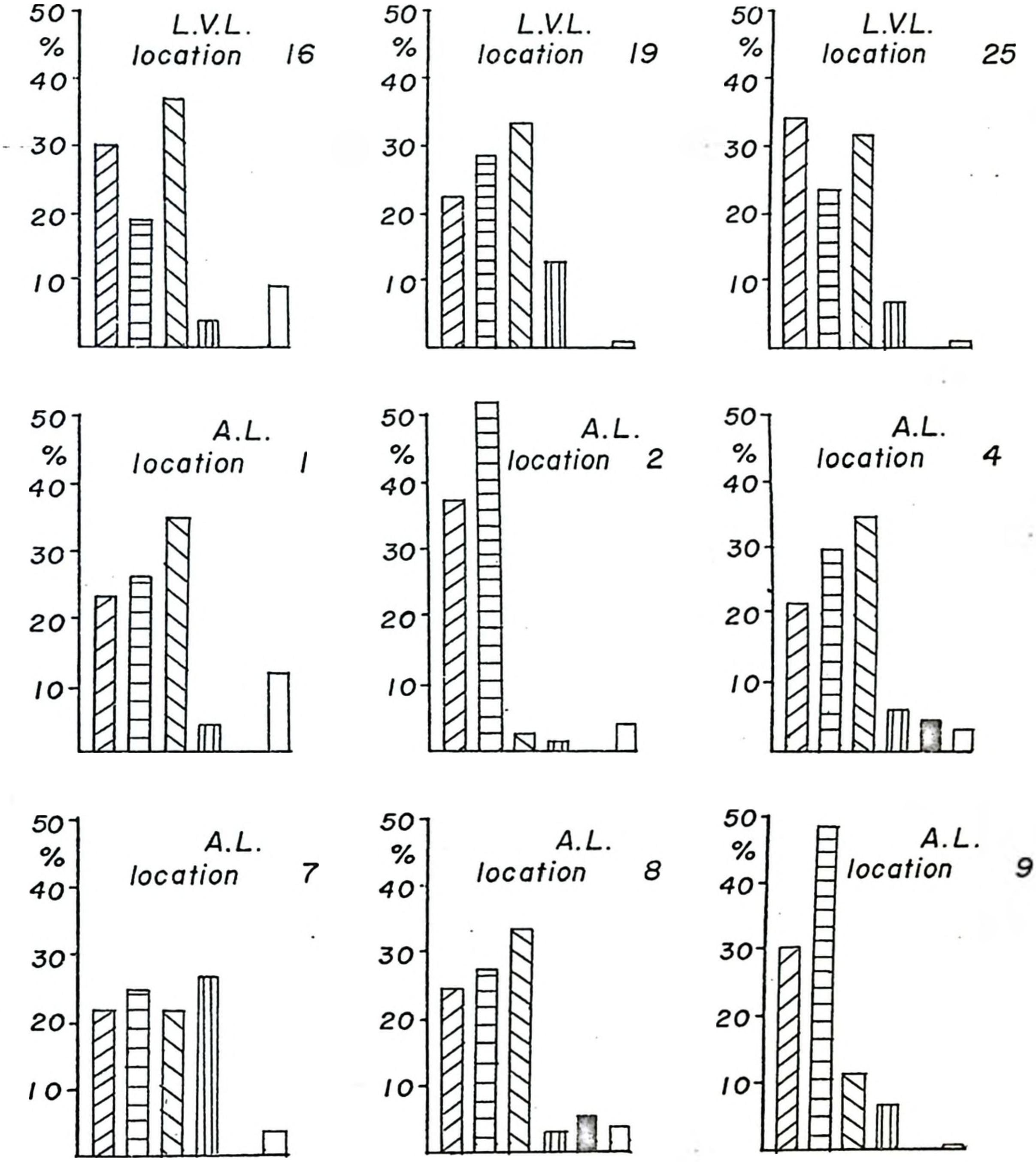


Figure 7b. Percentages of clast types within outcrops of granite-bearing conglomerate at Little Vermilion Lake (L.V.L.) and Abram Lake (A.L.). Locations are shown on maps 1 and 3.

descriptions of these clasts are given in Chapter 5.

Considerable variability of clast composition distributions exists between outcrops of granite-bearing conglomerates (fig. 7), but the two most important results from these measurements are - (1) The percentage of matrix varies from 20 to about 38 percent. (2) Greenstone clasts are commonly the most abundant volumetrically, followed by granitic and felsic clasts.

Clast sizes of massive conglomerates - The nature of exposures of the Ament Bay Formation is such that only 2-dimensional examination of the conglomerates is possible. Pleistocene glacial scour has smoothed the outcrops, producing random cross-sections through the clasts. Variations in the coarseness of the massive conglomerates were estimated in the field by determining the mean of the longest visible axes of the 10 largest clasts, at individual outcrops. This measure is hereafter referred to as D/10. Only those outcrops which showed little or no obvious tectonic deformation were used for this purpose. The sizes of the outcrops affect these mean values to a small extent, but most determinations were carried out on outcrops with areas of about 10 to 20 square metres.

There is no consistent change in the D/10 values along the length of the sedimentary belt at Little Vermilion Lake (fig. 8). However, the outcrops of massive conglomerate in the lower part of the sequence (along the northern side of Ament Bay and north-west Closs Lake) have smaller values of D/10 (from

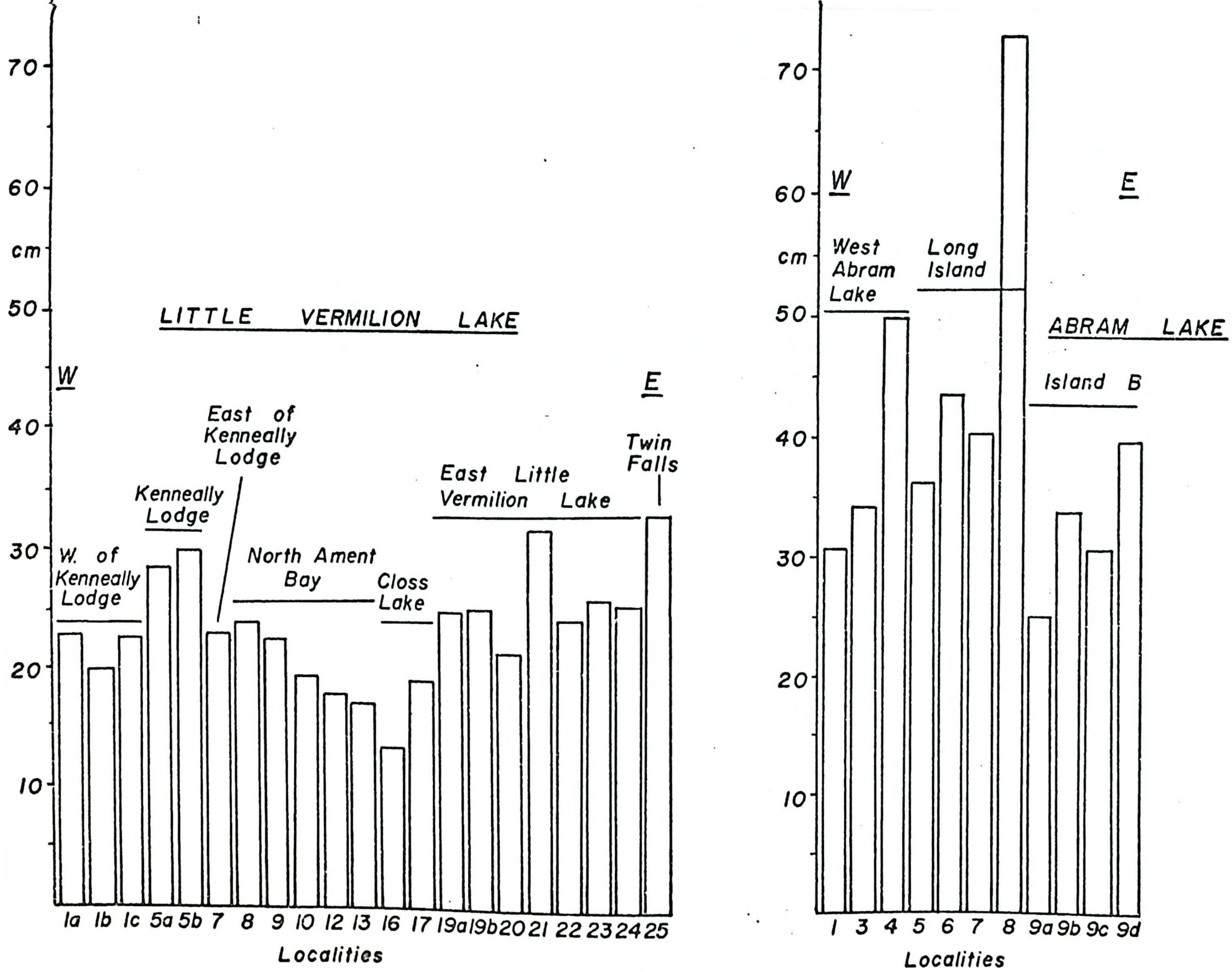


Figure 8. Plots of the mean length of the longest axes of the 10 largest clasts (D/10) at outcrops of granite-bearing conglomerate. The locations are arranged from west to east (left to right) (see maps 1 and 3). The letters represent measurements on separate beds at individual locations.

about 14 cm to 24 cm) than do the conglomerates higher in the sequence (at Kenneally Lodge and near Twin Falls, fig. 2), which range from about 20 cm to 34 cm. This variation may not strictly reflect the stratigraphic position of the conglomerates, but may be due to the geographically localized accumulations of slightly coarser granite-bearing conglomerates.

However, there are differences of maximum boulder sizes between Little Vermilion Lake and Abram Lake, farther east (fig. 8). D/10 values at Abram Lake range from 25 cm to 70 cm, whilst those at Little Vermilion Lake range from 14 cm to 33 cm. The largest boulder present in the whole area occurs at Long Island in Abram Lake (fig. 2), and has a maximum diameter of 1.5 m (fig. 16). As at Little Vermilion Lake, there is no consistent change of maximum clast sizes along the conglomeratic belt at Abram Lake. The suggestion of Walker and Pettijohn (1971, fig. 5) that boulder sizes decrease southwestward is only correct in as much as the pebbles on Little Vermilion Lake are smaller than those on Abram Lake. There is no gradual size decrease of the type they suggested.

Estimates of the size distributions of clasts within individual massive conglomerates were obtained, in order to further compare granite-bearing conglomerate units along the length of the sedimentary belt. This was carried out by measuring all clasts greater than 1 cm in length within a 60 cm square. These data are presented in figure 9.

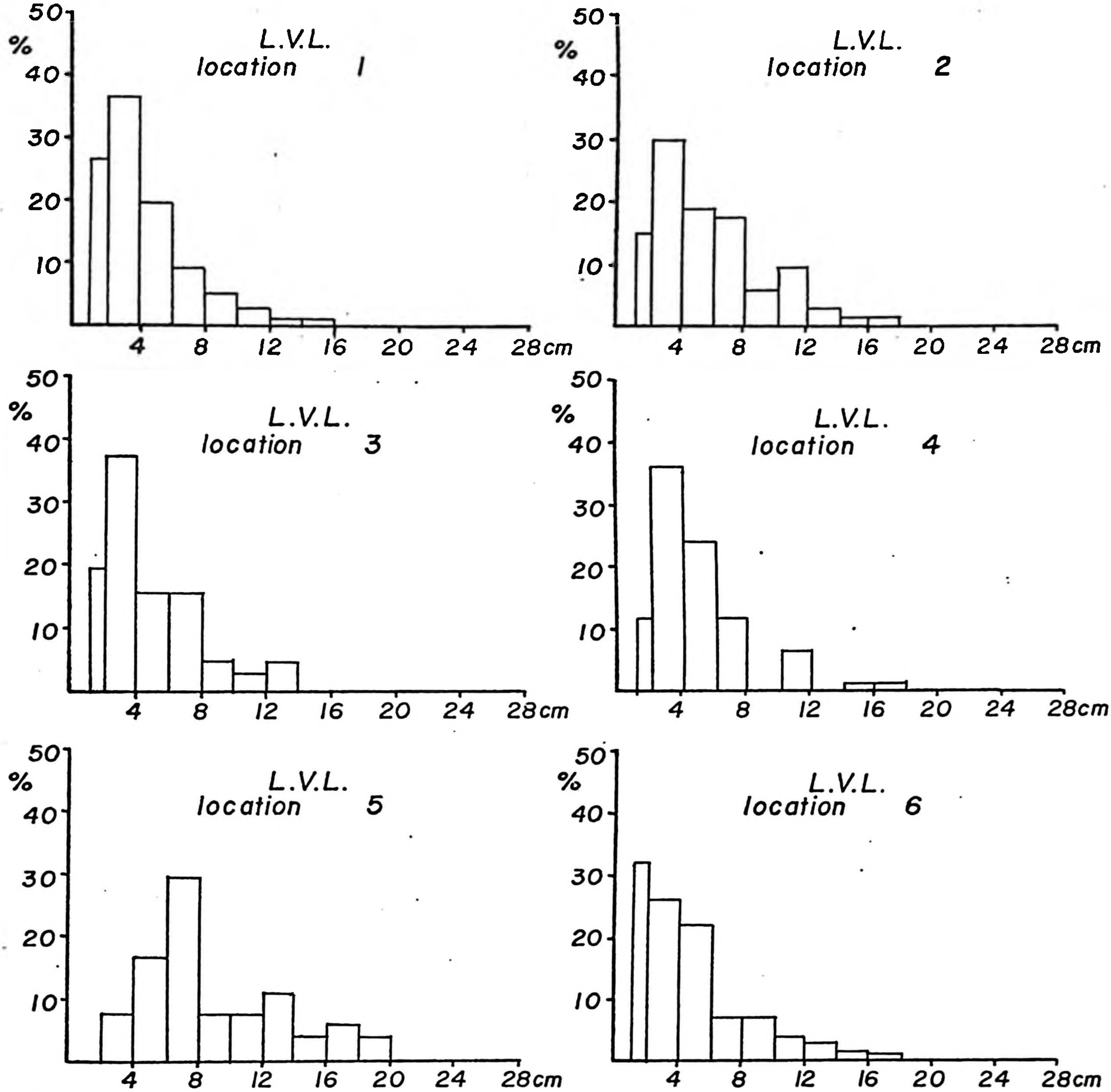


Figure 9a. Size percentage distributions of clasts within granite-bearing conglomerates at Little Vermilion Lake (L.V.L.) and Abram Lake (A.L.). The abscissa indicates the length of the apparent long clast axes. The locations are shown on maps 1 and 3.

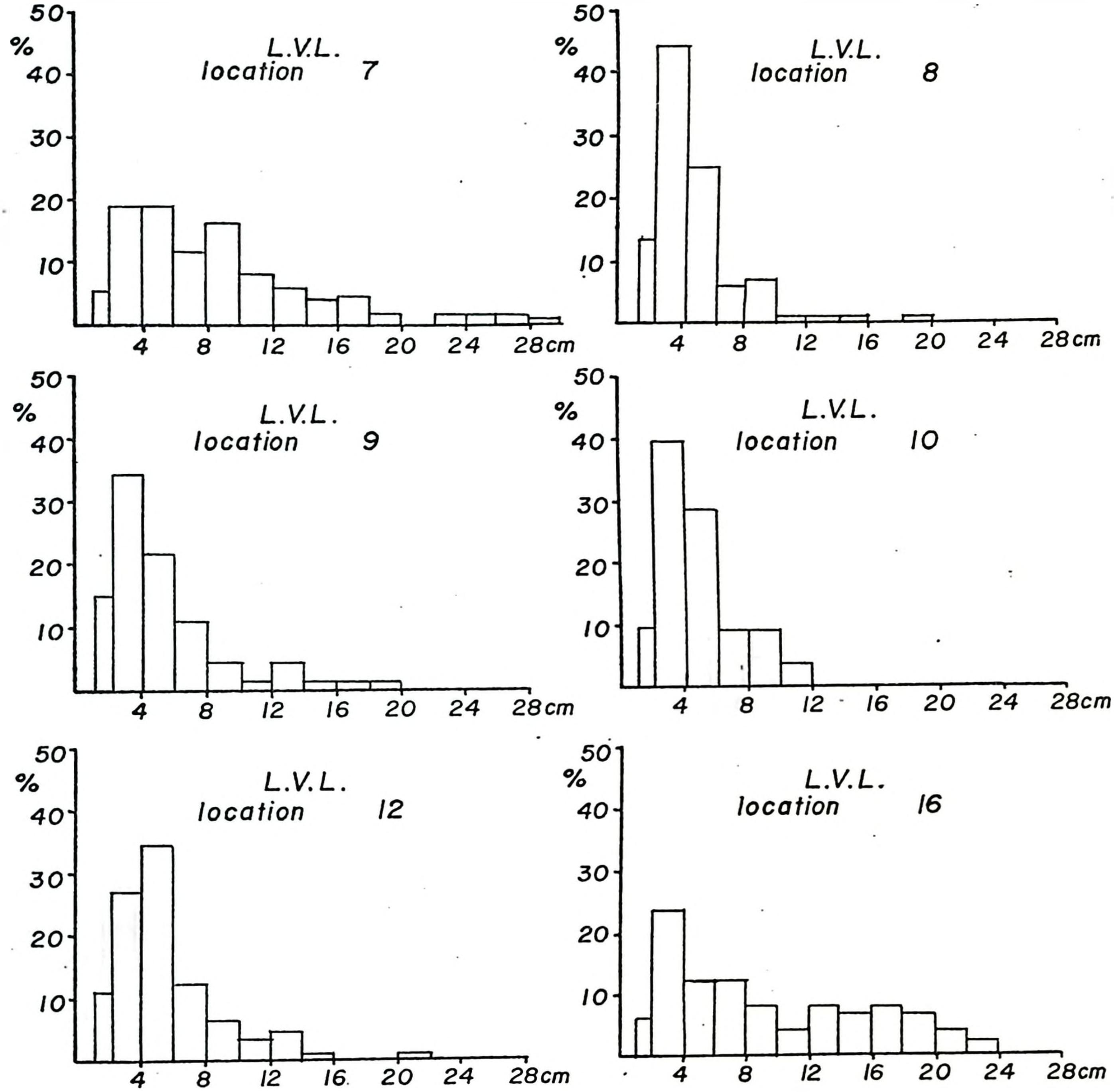


Figure 9b. Size percentage distributions of clasts within granite-bearing conglomerates

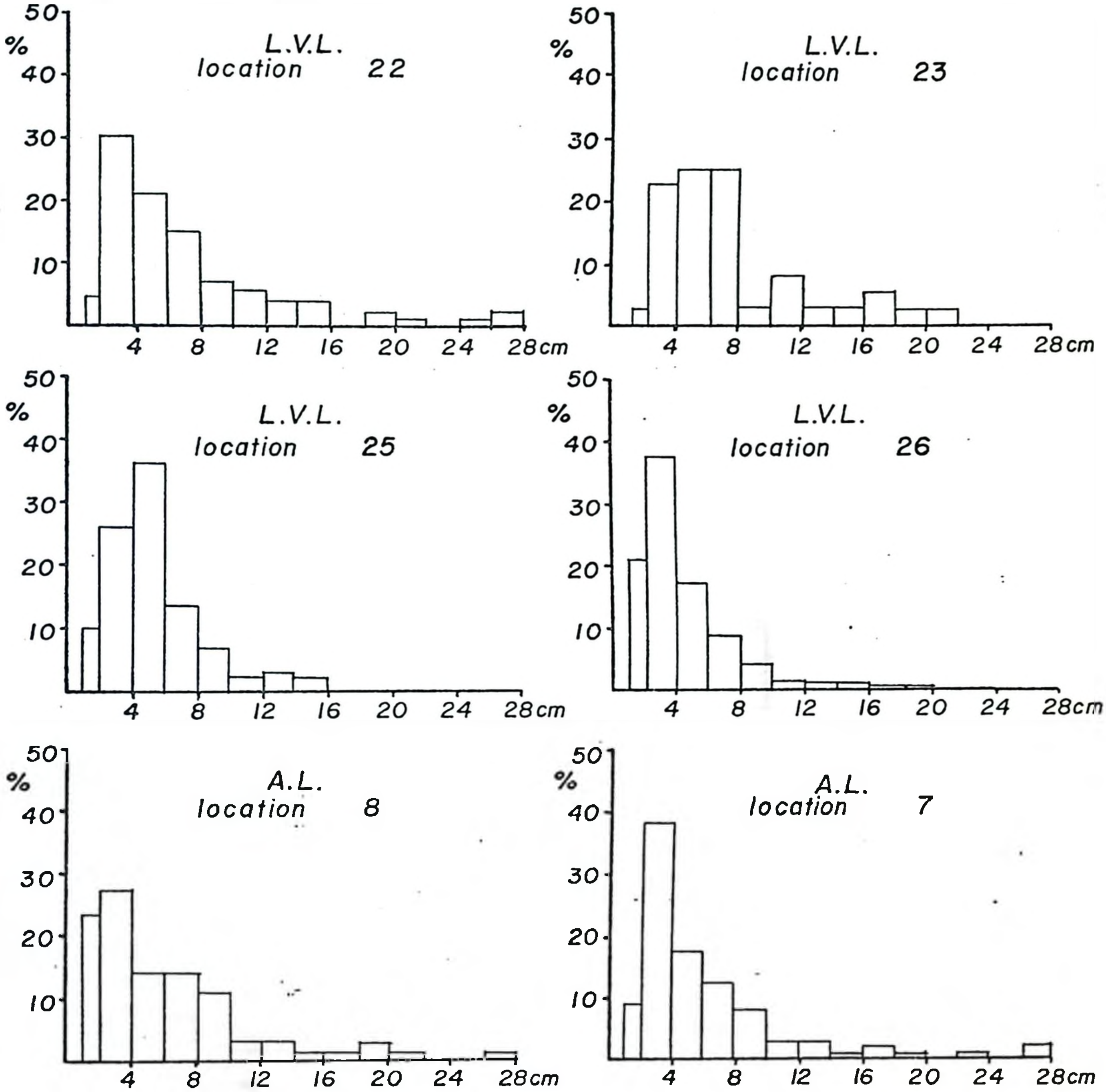


Figure 9c. Size percentage distributions of clasts within granite-bearing conglomerates

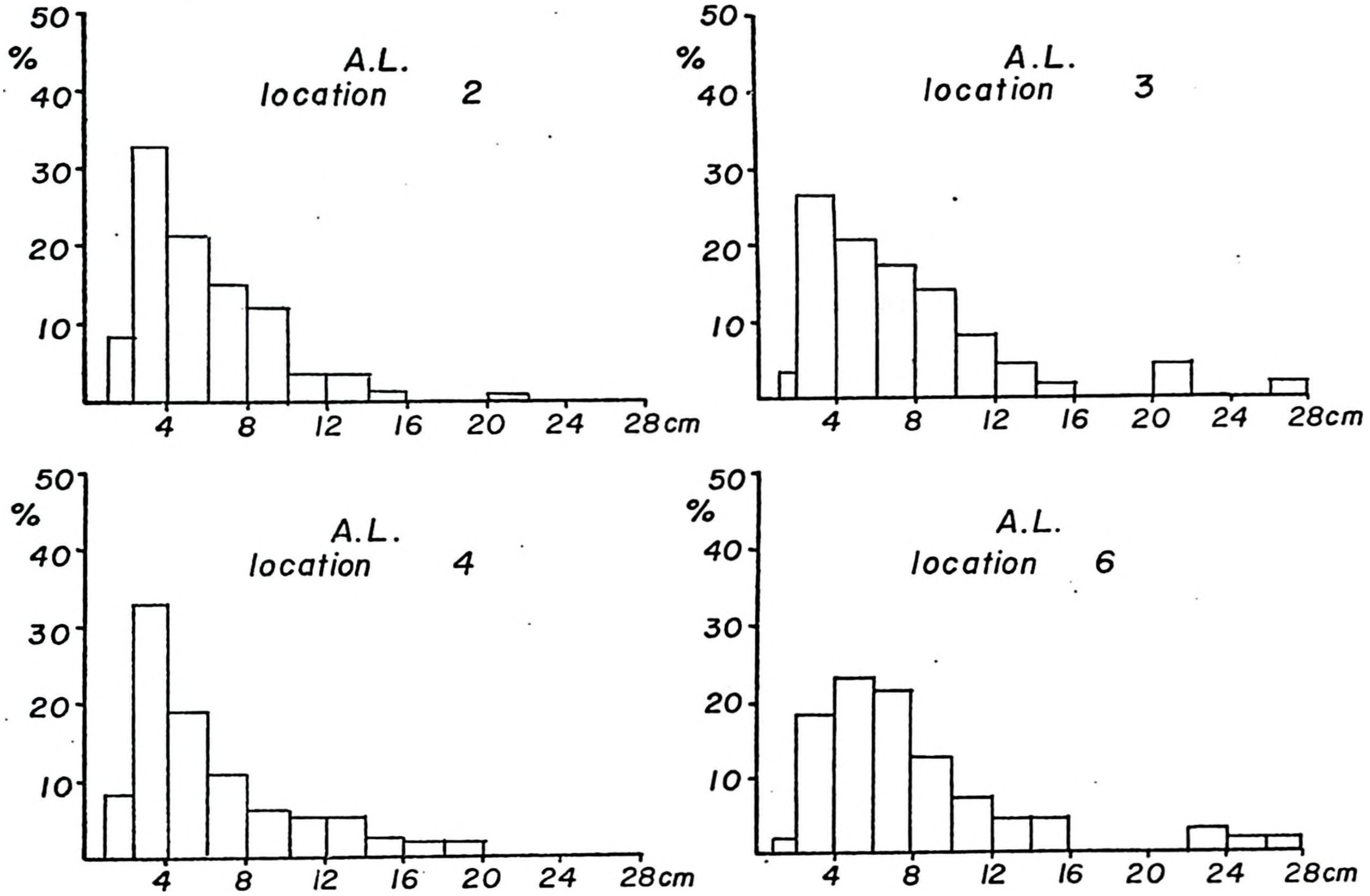


Figure 9d. Size percentage distributions of clasts within granite-bearing conglomerates

Some variations from the general unimodal pattern of these distributions are caused by clast composition differences between the massive conglomerate outcrops. Distributions at localities 5, 7, and 16 (fig. 9) at Little Vermilion Lake are less peaked than other distributions, due to greater concentrations of granitic clasts (38, 32, and 37 percent by volume) than in most outcrops (fig. 7). Because most granitic clasts are larger than greenstone clasts in all of the outcrops of massive conglomerate, a higher concentration of granitic clasts increases the percentage of large clasts, and thus the distribution becomes less peaked. The similarity of the other sorting distributions makes it probable that the various conglomerates were deposited under similar conditions.

Clast shapes of massive conglomerates - When conglomerates containing greenstone and granitic clasts are deformed, the greenstones are squashed and greatly elongated, whilst the granitic clasts show much less elongation. This was demonstrated by Hsu (1971, p.137) when he compared the average shortening strains of major pebble lithologies in the Seine conglomerate, and found that the ductility ratio between volcanic and granitic pebbles was 8.4:1.0. A high ductility ratio is also demonstrated by figure 17, where a highly deformed conglomerate near Sioux Lookout shows a granitic clast with a length:width ratio of 2:1, whereas greenstone and other volcanic clasts are greatly elongated - some having length:width ratios greater than 30:1.

Thus, in those instances where greenstone clasts are not obviously compressed or bent around the more resistant clasts, it is likely that the granitic clasts have undergone negligable deformation. This is the case at many of the outcrops of granite-bearing conglomerate, and here, estimates of the original sphericity of the granitic clasts were made by measuring the apparent width (W) and length (L) of the clasts (mutually perpendicular axes), and dividing one by the other (W/L). This gives the 2-dimensional elongation of the clasts.

There is a wide spread of elongation values for granitic clasts within massive conglomerates, but about 40 percent of the clasts measured had W/L values greater than 0.70 (table 1). Thus many of the granitic clasts were probably originally equant in shape.

Only visual estimates of the roundness of conglomerate clasts were carried out in the field, and the majority of granitic clasts (together with many other clasts of different lithologies) were designated as rounded or well-rounded (e.g. fig. 12).

(iii) Pebble bands within arkosic sandstone

Pebbles and cobbles commonly occur in thin rows or bands within thick sequences of arkosic sandstone (fig. 18), but are seldom found in sequences which predominantly consist of massive conglomerate (see Appendix 1). These rows are usually thinner than the diameter of the largest clast present (commonly less

Location	$(W/L)_{10}$ for granitic clasts	$(W/L)_{10}$ for greenstone clasts
2	0.68	0.64
4	0.72	0.50
7	0.64	0.51
8	0.67	0.58
9	0.72	0.52
12	0.65	0.60
15	0.68	0.40
16	0.66	0.56
22	0.70	0.48
26	0.73	0.60

Table 1. Two dimensional elongation values for clasts within massive granite-bearing conglomerates at various locations along the length of Little Vermilion Lake. The $(W/L)_{10}$ value is the mean of 10 W/L values from clasts at each location. The locations are shown on map 1.

than 15 cm in diameter), and in some places persist for only a few metres along strike (fig. 19).

The clasts within these pebble bands are similar in composition and roundness to those in the massive conglomerates, but the maximum clast sizes in these bands are smaller than in most massive conglomerates. The granitic clasts are commonly larger than the greenstone clasts, are usually well-rounded, and many are almost circular in outline. Greenstone clasts are well-rounded to subrounded, and are more elongate than the granitic pebbles.

In some cases, pebble bands occur between distinct beds of arkosic sandstone, and a change in the sand grain size is evident between the arkose beds. In these instances the coarser arkose is found at the base of the upper bed, in which grading is occasionally present. However, pebble bands are also common within massive, structureless arkose, where no change of grain size occurs on either side of the band (fig. 19).

In most places, the clasts in pebble bands are in contact with each other and the band is several pebbles thick. Some bands, however, consist of isolated pebbles scattered along flat bedding surfaces. At one locality at Closs Lake (fig. 2) a row of isolated pebbles was found separating 2 beds of cross-bedded arkose (fig. 20). Here, the long axes of the pebbles are oriented at varying angles to the bedding.

Clusters of pebbles also occur at a few localities, and usually consist of one large, well-rounded clast (up to about 15 cm in diameter), with many smaller pebbles grouped on either side. In most instances this arrangement of pebbles is found in massive structureless arkose, and in a few cases there are more pebbles on one side of the large clast than on the other (fig. 19). Thus, the overall configuration of a pebble cluster is that of a small lens or wedge of pebbles within arkose, which is aligned parallel to the bedding.

Imbricate arrangements of pebbles exist in some pebble bands, but this alignment is parallel to the tectonic foliation in some places (fig. 18), making interpretation of the structures difficult (see section below on Imbrication).

(iv) Pebbly arkosic sandstone

Accumulations of pebbles and cobbles comprising up to 45 percent of the outcrop are found at a few localities within thick sequences of arkosic sandstone (map 1). Also, single isolated pebbles are common within massive arkosic sandstone units (fig. 21). Most clasts in these pebbly arkoses are isolated from each other by the massive arkosic sandstone, but a few clasts touch each other (fig. 22). The clasts are commonly less than 10 cm in length, and rarely exceed 15 cm. The compositions of the clasts are similar to those in the massive conglomerates, and most clasts are well-rounded.

Pebbly arkosic sandstone is only a locally developed conglomerate type, and is confined to beds 1 or 2 m thick, within otherwise massive arkosic sandstone. At one exposure at Closs Lake a band of pebbly arkose about 70 cm thick (fig. 22) dies out laterally (within about 1 m) and is replaced by massive arkosic sandstone. Here, the pebbly arkose has a slightly curved, scoured contact with the underlying massive arkosic sandstone bed.

In a few examples of pebbly arkosic sandstone the clasts are aligned at angles of about 35° to the bedding, which represents a type of imbrication (fig. 23; see also the section below on Imbrication).

No pebbly arkosic sandstone was observed to be associated with massive granite-bearing conglomerate.

(c) Imbrication

At a few localities of the granite-bearing conglomerate facies a preferred orientation of many pebbles is apparent. In some cases this is obviously a tectonic fabric, as the pebbles have been stretched and aligned within outcrops containing pronounced cleavage planes. However, at a few other localities, clast alignment occurs at angles to the bedding within granite-bearing conglomerates which exhibit no obvious tectonic structures (such as cleavage), and where greenstone clasts are not flattened (some being circular in outline). In these cases the structure is thought to be true imbrication, produced by primary sedimentary processes.

Eliptical pebbles of greenstone exhibit the best alignment, and some are stacked such that the inclined lower surface of one clast rests on the inclined upper surface of an adjacent clast (fig. 12). In these cases, virtually all clasts with length:width ratios greater than about 2:1, have a common direction of dip relative to the bedding planes. The angle of inclination of these imbricate pebbles is commonly 30° to 40°, but values range from about 10° to 55° (table 2).

Convincing imbrication was found in massive conglomerates at only 2 localities in the Little Vermilion Lake area, the best being on Island B in Ament Bay (see fig. 12). An imbricate stacking of pebbles also occurs within some pebble rows (table 2, and fig. 18), and a similar type of clast alignment occurs within a few pebbly arkoses (fig. 23). In this latter case, pebbles do not touch each other, but are inclined at angles of about 30° to the bedding. It is unlikely that pebbles (particularly greenstones) set in a sandstone matrix could be rotated by tectonic forces after the sediment had become lithified, and still remain virtually undeformed (P.M. Clifford, personal communication). Thus, it is probable that the common pebble alignment in some pebbly arkoses, also resulted from primary sedimentary process.

Seven localities were found in the Little Vermilion Lake area which exhibited reasonably convincing imbrication (table 2). At 6 of these localities the pebbles dipped in a roughly westerly direction relative to the present bedding orientation. This implies that in most cases the depositional currents had an easterly component to their flow directions (see Chapter 4).

Location	Conglomerate type	Angle of clast imbrication with respect to the bedding	Strike direction = rough current direction
Near 3	pebbly arkosic sandstone	31° (from 44 measurements)	towards 60°
Near 5	pebble row	21° (from 13 measurements)	towards 70°
6	massive conglomerate	up to 40°	towards 70°
14	pebble row	up to 50°	towards 45°
15	massive conglomerate	32° (from 46 measurements)	towards 45°
15	pebble row	35° (estimated average)	towards 225°
22	pebble row	30° (estimated average)	towards 55°

Table 2. Angles of imbrication and rough current directions from imbricated granite-bearing conglomerates at various locations along the length of Little Vermilion Lake. The locations are shown on map 1.

Figure 10. Basal conglomerate of the Ament Bay Formation overlying quartz porphyry at Vermilion Lake. The conglomerate comprises quartz porphyry and greenstone clasts. Arrow indicates top direction. Scale in centimetres.

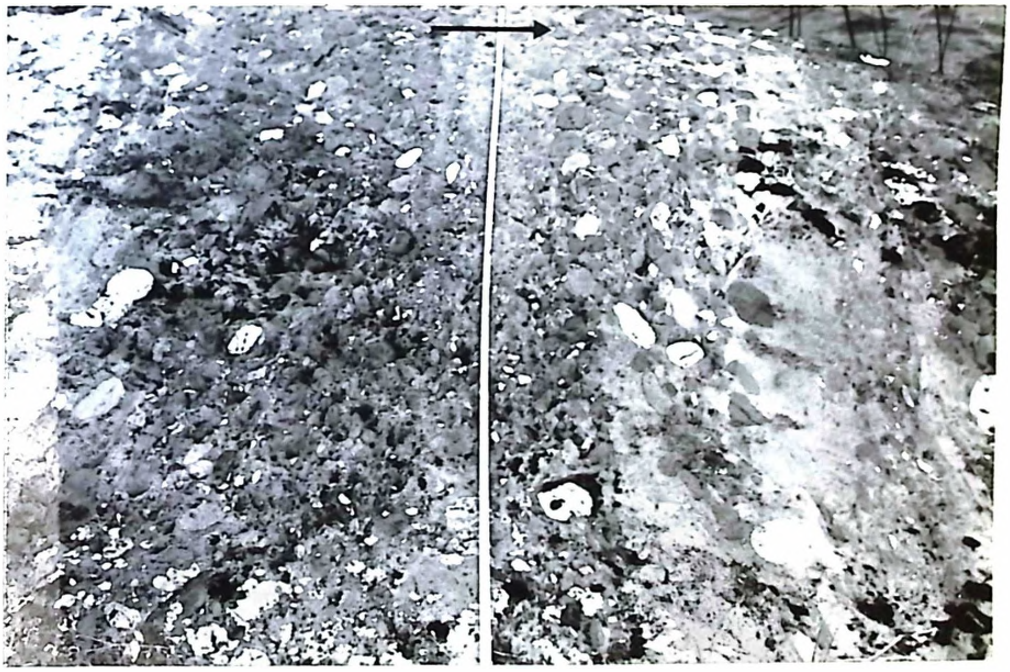


Figure 11. Crudely bedded granite-bearing conglomerate, with arkosic sandstone interbeds. Arrow indicates probable top direction. Scale in centimetres. West of Island A, Abram Lake.

Figure 12. Imbricated granite-bearing conglomerate resting sharply on massive arkosic sandstone. Average orientation of clasts is approximately parallel to the centimetre scale. Arrow indicates top direction to the southeast. Island B, Ament Bay.

Figure 13. Granite-bearing conglomerates, one of which has a graded top, together with thin interbeds of arkosic sandstone. Arrow indicates top direction. Scale in inches and centimetres. Kenneally Lodge, Little Vermilion Lake.

Figure 14. Granite-bearing conglomerate filling erosional channel in parallel-laminated arkosic sandstone. Arrow indicates top direction to the southeast. Scale in inches and centimetres. Closs Lake.

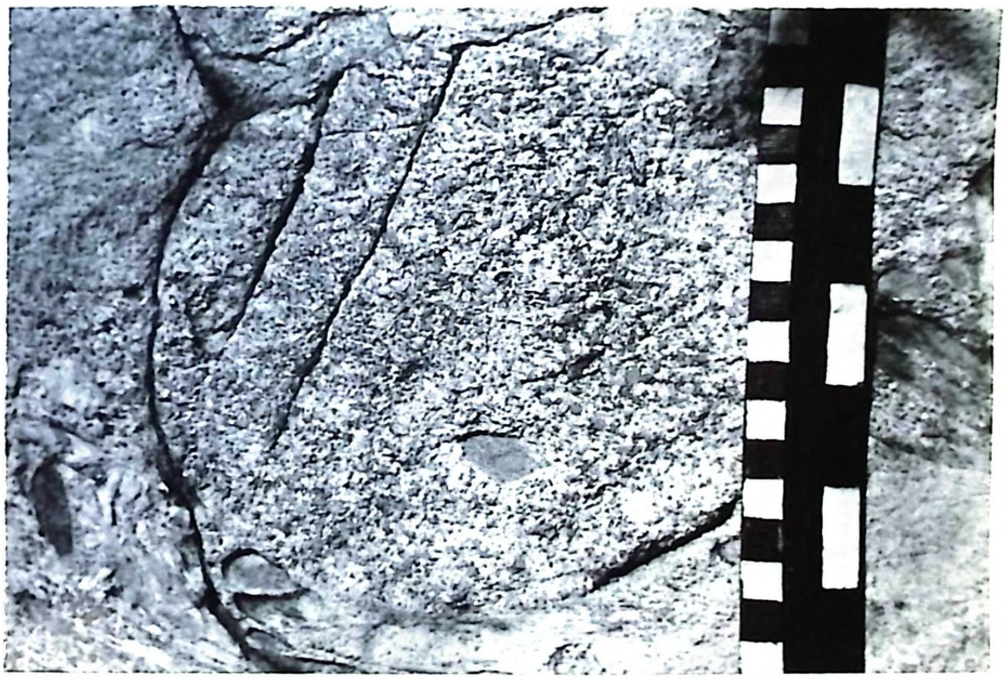


Figure 15. Granodiorite clast containing xenolith of greenstone, within arkosic sandstone. Scale in inches and centimetres. West of Kenneally Lodge.

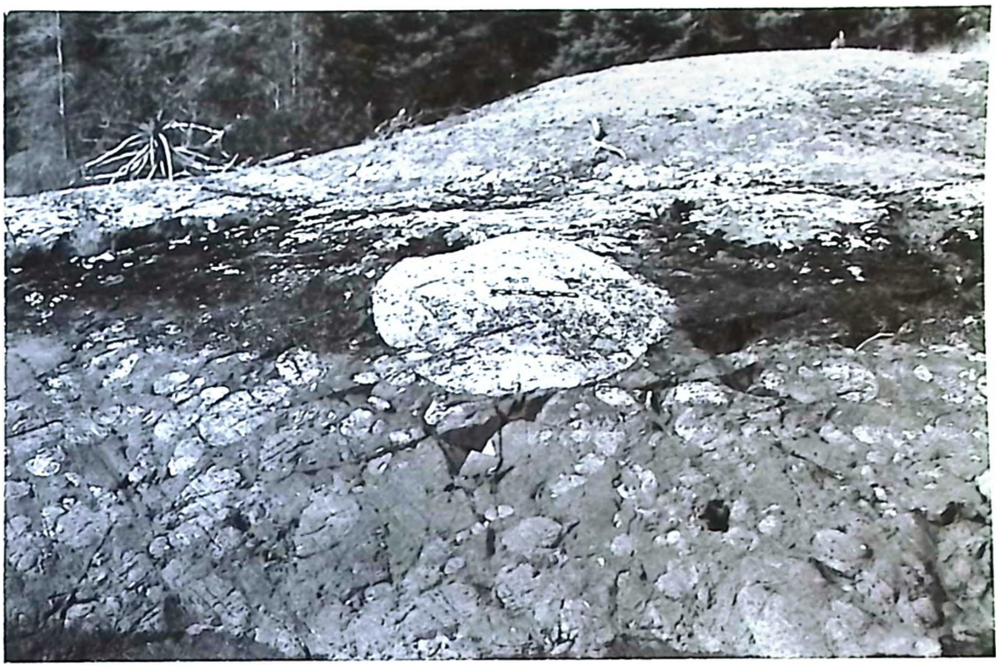


Figure 16. Massive granite-bearing conglomerate at Long Island in Abram Lake. Scale on large granodiorite boulder is 30 cm long.

Figure 17. Highly deformed granite-bearing conglomerate. Note great elongation of greenstone clasts relative to large granodiorite clast. Along highway, east of Sioux Lookout.

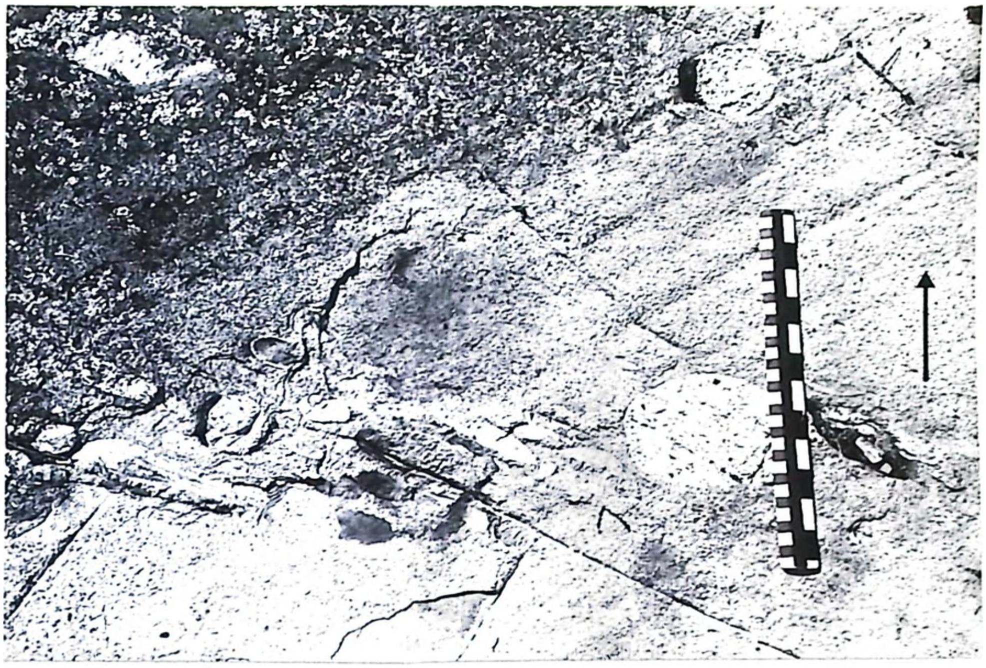


Figure 18. Pebble band within massive arkosic sandstone. "Imbrication" may be tectonically influenced as it is parallel to cleavage. Diagonal markings are due to glacial scour. Arrow indicates top direction. Scale in inches and centimetres. Near Kenneally Lodge.

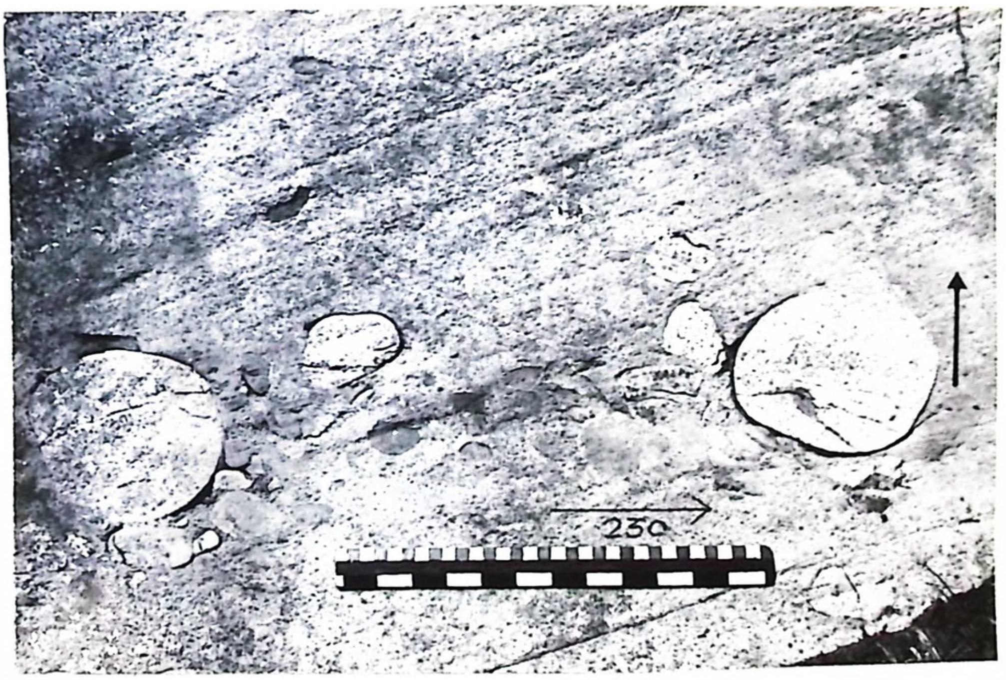


Figure 19. Pebble band within massive arkosic sandstone. Note tendency of clasts to form "pebble clusters". Diagonal markings are due to glacial scour. Arrow indicates top direction. Scale in inches and centimetres. South Ament Bay.

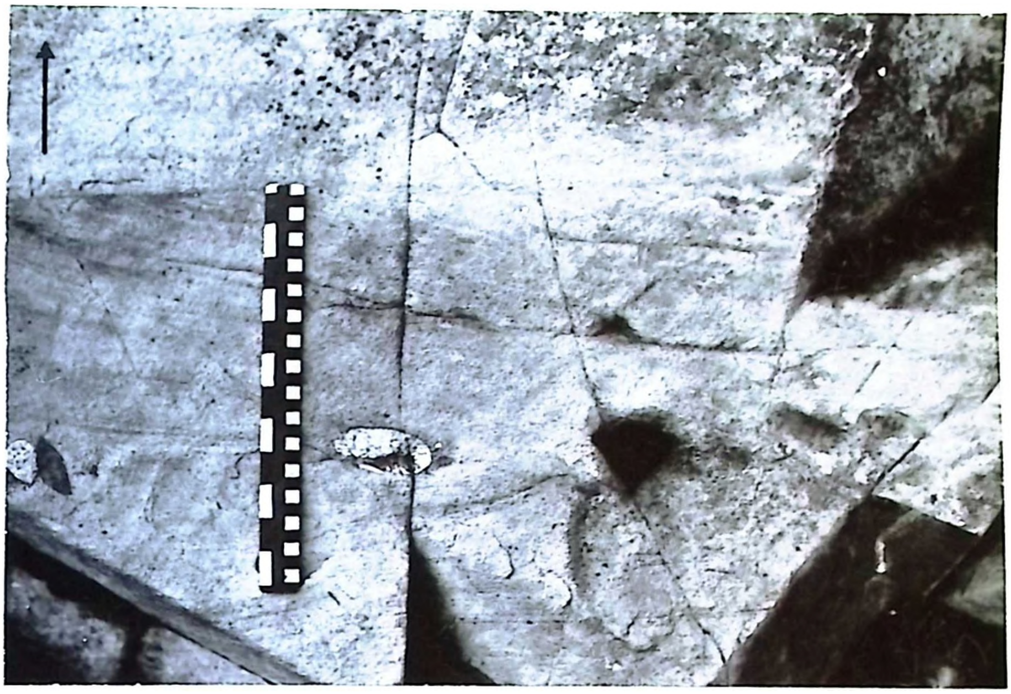


Figure 20. Trough cross-bedding within arkosic sandstone. Discontinuous pebble row separates sets of cross-strata. Arrow indicates top direction to the southeast. Scale in inches and centimetres. South Closs Lake.

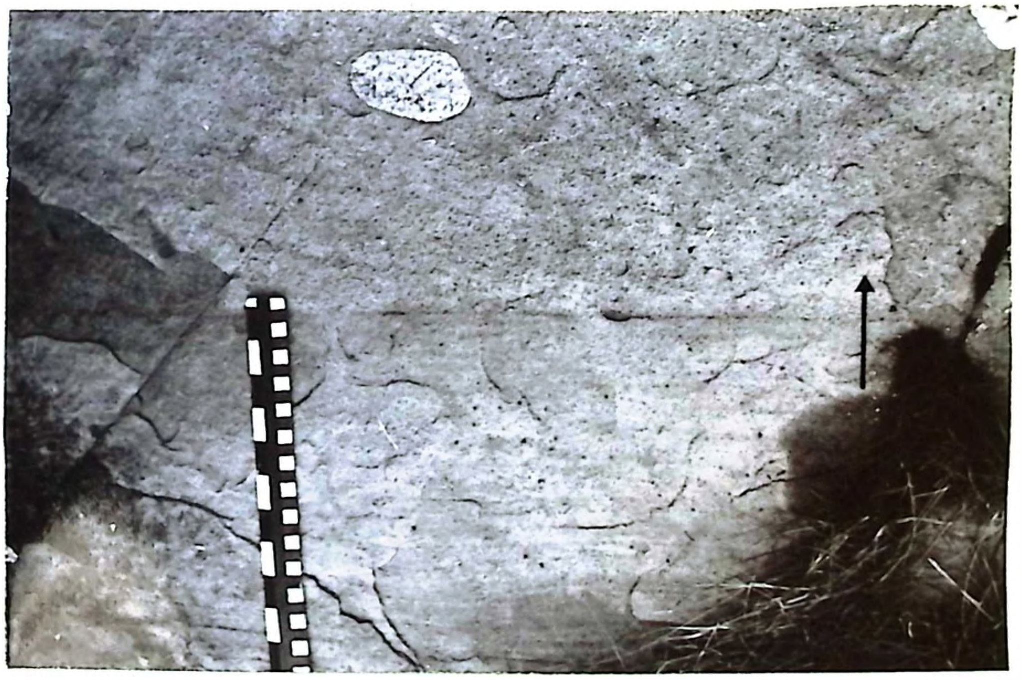


Figure 21. Massive arkosic sandstone enclosing isolated clast and overlying parallel-laminated arkose. Note straight sharp bedding surface. Arrow indicates top direction. Scale in inches and centimetres. West of Kenneally Lodge.

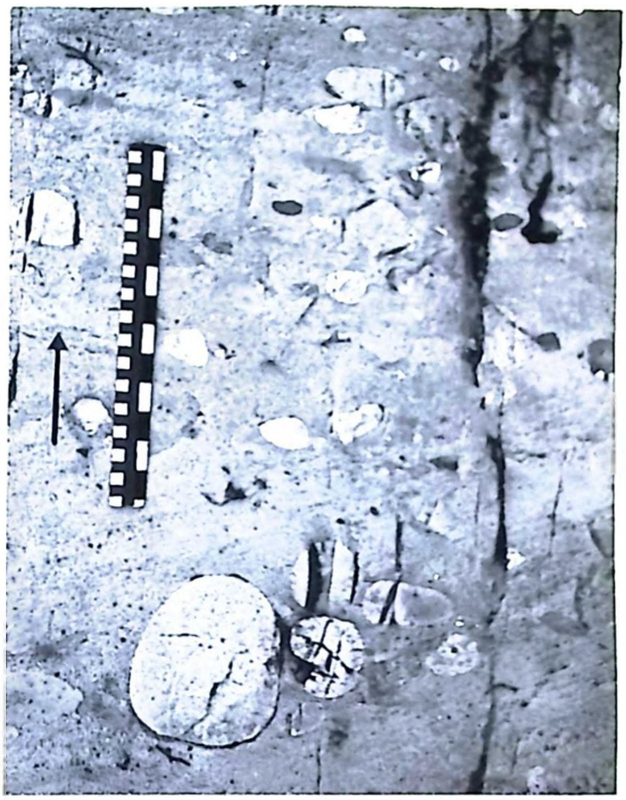


Figure 22. Pebbly arkosic sandstone with a pebble cluster at the base of the bed. Note dispersed nature of the clasts within coarse arkose. Diagonal markings are due to glacial scour. Arrow indicates top direction. Scale in inches and centimetres. South Closs Lake.

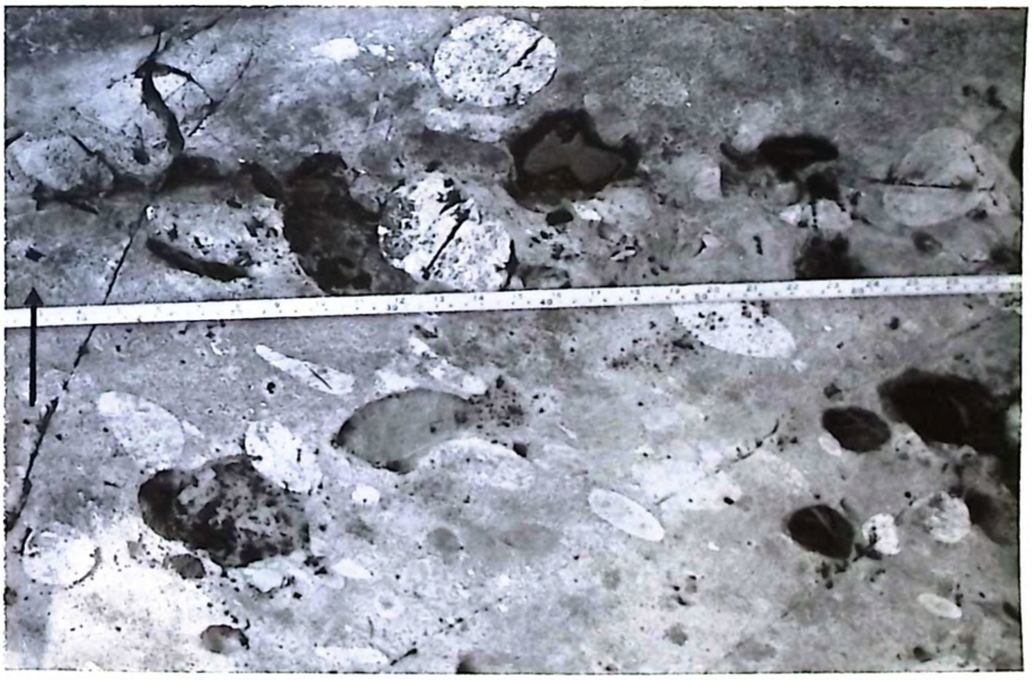


Figure 23. Imbricated pebbly arkosic sandstone. Note massive nature of arkose, and undeformed character of greenstone clasts. Arrow indicates top direction to the southeast. Scale in inches and centimetres. West of Kenneally Lodge.

3. Arkosic sandstone facies

(a) Introduction

The arkosic sandstone facies comprises the bulk of the Ament Bay Formation at Little Vermilion Lake, and is intimately associated with the granite-bearing conglomerate facies. The best exposures of this facies occur around the shoreline of the north-east half of Little Vermilion Lake, and around Closs Lake (fig. 2). Arkosic sandstones also crop out at Abram Lake, but they are not found in such thick sequences as at Little Vermilion Lake.

On weathered surfaces in the field, typical arkosic sandstones show distinct grains of subangular and subrounded white feldspar and grey quartz (up to 2 mm in diameter), which comprise most of the rock. This gives rise to pale grey, massive out-crops of arkose, which differ distinctly from the well-bedded, darker outcrops of the finer grained greywackes higher in the stratigraphic sequence.

The term "arkosic sandstone" is used for this facies despite the fact that the sandstones commonly prove to contain more than 15 percent matrix when studies in thin sections (Appendix 2).

Thus, classifications such as Pettijohn's (1957) would term these sandstones "feldspathic greywackes", and according to Okadu's classification (1971), they would be "feldspathic wackes". However, "arkosic sandstone" (or "arkose") is a useful field term, already established in the literature on Archaean rocks, and will be retained throughout this thesis.

(b) Arkosic sandstone occurrences

Arkosic sandstone occurrences can be divided into 2 types, depending on their bedding features and on their relationship with the granite-bearing conglomerate facies:

(i) Thick arkosic sandstone sequences; several hundred metres of arkose occur with no major granite-bearing conglomerate units.

(ii) Interbeds of arkosic sandstone between massive conglomerate units; the arkose beds are commonly less than 1 m thick.

(i) Thick arkosic sandstone sequences

A thick section of predominantly arkosic sandstone occurs between about 550 m and 1,300 m of the stratigraphic section of the Ament Bay Formation in the central part of Little Vermilion Lake. Two such sequences were measured, along the west shore of Closs Lake (fig. 2), and along the peninsula to the west of Kenneally Lodge (see Appendix 1). These measured sections contain no conglomerate bodies thicker than about 1 m.

At many small outcrops of arkosic sandstone, bedding surfaces are not present, and some arkose sequences can be traced laterally for over 10 m without any bedding being apparent.

However, where bedding surfaces are visible, they either take the form of grain size changes, or are marked by pebble bands. In the former case, boundaries between arkosic sandstones of different grain size are commonly sharp and straight, and usually have the coarser arkose in the upper bed (fig. 21). Individual arkose beds are rarely thinner than 50 cm, and at some localities only one bedding surface is present within the 2 or 3 m outcrop section. Due to the restricted width of most outcrops, it is rarely possible to trace bedding for more than about 5 m along the strike, but most bedding surfaces do extend across the full width of the outcrop. At no locality does the arkosic sandstone facies consist of regularly spaced beds of arkose.

Bedding surfaces within arkosic sandstone sequences are also commonly marked by thin bands of pebbles extending in straight lines across the outcrops. These pebble bands are commonly less than 15 cm thick, and consist of either scattered pebbles along the bedding (fig. 20), or of clasts which touch each other (fig. 18). Isolated pebbles (usually less than 10 cm in diameter) are also fairly common throughout arkosic sandstone sequences (fig. 21).

The distribution of pebble bands and isolated pebbles throughout the arkosic sandstone facies does not show any obvious depositional pattern. For example, pebble bands are not regularly spaced, nor are they preferentially associated with any sedimentary structure in adjacent arkose beds. Isolated pebbles occur sporadically throughout the arkosic sandstone facies (see Appendix 1).

Sedimentary structures such as graded bedding, cross-bedding, and parallel laminations occur within some beds of arkosic sandstone. At single outcrops, only one type of sedimentary structure is generally found, and vertical sequences of sedimentary structures are rare. These structures are described in detail under a separate heading below.

Thick sequences of arkosic sandstone, such as those at Little Vermilion Lake, do not occur at Abram Lake. Here the granite-bearing conglomerate facies comprises a greater proportion of the sedimentary sequence than it does at Little Vermilion Lake, and most arkose beds are found between conglomerate units.

(ii) Interbeds of arkosic sandstone between massive conglomerate units

Within those sections of the sedimentary succession where the granite-bearing conglomerate facies predominates (north side of Ament Bay, Kenneally Lodge, Abram Lake), arkosic sandstone units occur as interbeds between the conglomerates.

These arkosic sandstone interbeds are commonly less than 1 m in thickness, and can be as thin as 3 cm (fig. 13). They are either parallel-sided across the width of outcrop, or they wedge out in one or both directions along strike (fig. 24). The tops of these arkose units are commonly straight and sharp (fig. 25), but in some places clasts from the overlying conglomerate unit protrude into the arkose bed. The bases of the arkoses are usually more irregular, and pebbles and cobbles are commonly enclosed within the arkosic sandstones near to their base. The boundaries of the arkose interbeds with the underlying conglomerates are sometimes gently curved, and are concave upward (see section A in Appendix 1).

The laterally impersistent lenses of arkose are not common, and the smallest found was about 3 m in length. However, it is suggested here that some of the more extensive, apparently parallel-sided arkose interbeds may in fact be wide lenses of arkose within predominantly massive conglomerate units, their lateral truncations being obscured due to lack of exposure. This is suggested due to the fact that at Island B in Ament Bay 3 beds of conglomerate (each about 20 cm thick), together with 2 parallel-sided arkosic sandstone interbeds (both about 15 cm thick), merge westward, so that a single massive conglomerate unit (about 1 m thick) is formed, within a distance of about 10 m. Thus, other apparently parallel-sided arkosic sandstone interbeds may represent arkose lenses within granite-bearing conglomerates.

Most interbeds of arkosic sandstone consist of massive, medium to very coarse grained arkose (using the Wentworth scale). In the field these arkose units appear to be identical to the arkosic matrix within the adjacent conglomerates.

No bedding planes parallel to the conglomerate beds are found within these single arkose units, but cross-bedding is present in some places, and grading is occasionally found.

(c) Sedimentary structures

(i) Cross-bedding

Within some outcrops of the arkosic sandstone facies are found structures which resemble cross-bedding. In a few cases it is difficult to determine whether these "laminations" were produced tectonically, or whether they are original depositional structures. In most cases however, there can be no doubt that the structures are of primary sedimentary origin, as grain size changes occur between adjacent cross-strata, and these cross strata are commonly truncated at their upper boundaries by further textural changes (fig. 26).

Occurrences of cross-bedding are few, and only 15 good examples were found throughout the Ament Bay Formation at Little Vermilion Lake (see Appendix 1). Most occurrences of cross-bedding were found within the thick sequences of arkosic sandstone, but a few examples are present within arkoses interbedded with granite-bearing conglomerates (fig. 27).

Cross-bedding was only found in the Little Vermilion Lake area, and appeared to be absent at Abram Lake.

Most of the cross-stratification occurrences are examples of low-angle trough cross-bedding (e.g. figs. 28, 29, 30). Cosets of cross-bedding seldom exceed 2 m in thickness, and individual sets are commonly between 5 and 40 cm thick. The maximum number of cross-bedding sets within any coset is about seven (fig. 28).

In all of the cross-bedding occurrences the cross-strata have low-angle contacts with the lower bounding surfaces (figs. 28 and 30). The tops of the cross-strata are also commonly truncated at low angles (15° to 20°), but an angle of 45° was recorded at one locality (fig. 31). The lower bounding surfaces of the sets of cross-strata are generally smooth, and gently curved or planar. The lower bounding surfaces of cosets are occasionally marked by isolated pebbles or pebble rows (fig. 20). Only one example of a symmetrical cross-bedded trough was found within arkosic sandstone. This occurs at a locality in Closs Lake, where the trough is 6 cm deep and about 40 cm wide, and is situated at the top of an otherwise massive arkose bed (fig. 32). A pebble row followed by pebbly arkose overlies this cross-bedded unit.

The individual cross-strata are usually marked by very thin (1 mm or so) dark laminae, which are visible either due to slightly greater concentrations of chloritic material in the laminae than in the surrounding arkose (fig. 30), or are narrow physical partings within the arkosic sandstone (fig. 31). Slight grain size changes between laminae are common, and in a few cases the grain size changes by 3 or 4 Wentworth grades (from very coarse to fine arkose) between adjacent cross-strata (fig. 26). Occasionally, small pebbles (up to about 2 cm in diameter) are found on the toe, or part-way down the length of the cross-strata.

These pebbles commonly have their long axes parallel to the cross-bedding, but at one locality, pebbles at the foot of cross-strata are inclined at higher angles to the true bedding than are the cross-strata (fig. 33), but dip in the same direction.

In some places, a gradual lateral transition can be seen, from cross-bedded arkose to massive structureless arkose, but in most places the lack of outcrop prevents tracing cross-bedding along strike. Cross-bedded arkose is also commonly associated vertically with massive arkose. In these places boundaries between cross-bedded and massive arkose either consist of sharp bedding contacts, commonly marked by a pebble row (fig. 30), or are rather more gradational (fig. 31).

In a few places grading is associated with cross-bedding, and figure 29 shows a 60 cm bed of graded arkose beneath an occurrence of trough cross-bedding. Here, the lower part of the graded bed has an irregular contact with the underlying massive arkose, and contains a few cobbles (up to 12 cm in diameter). The bed grades from very coarse arkose at the base to medium grained, cross-bedded arkose at the top.

Repeated grading can also occur within a set of cross-bedded arkose (fig. 31). The graded sets are up to 30 cm thick, and grade from coarse to fine arkose.

Another type of structure associated with cross-bedding was found at Island B in Vermilion Lake (fig. 2). This consists of slight convolution of thin parallel laminations, which occur in fine to medium arkosic sandstone (fig. 34). This structure occurs approximately 2 metres east of the cross-bedding shown in figure 28. The sandstone laminae are from 1 mm to 2 cm in thickness, and are in the form of very low angle (5° to 10°) cross-beds (behind scale in figure 34). The laminations beneath the scale in figure 34 have a convolute form, with a wavelength of about 30 cm, and an amplitude of 2 cm. The asymmetry of this structure suggests that the depositional current had a component of flow towards the left of the figure, in a general westerly direction with respect to present bedding. This was the only occurrence of convolute lamination found throughout the arkosic sandstone facies.

Of the fifteen occurrences of cross-bedding in the Ament Bay Formation, 7 have trough forms with sets of cross-strata dipping in opposite directions. Thus it is not possible to deduce the direction of flow of the depositional current. In 7 other cases, where a single set of cross-strata exists, or the cross-strata of several sets have a common direction of dip, a westerly component of the current direction is suggested (figs. 26, 30, 31, 33 and 34). At one other locality, a single set of cross-bedding exists where the cross-strata dip in an easterly direction. A discussion of paleoflow is given in Chapter 4.

(ii) Parallel lamination

At a few localities around Little Vermilion and Abram Lakes, outcrops of arkosic sandstone contain parallel laminated units up to about 5 m thick. The lamination extends across the widths of the outcrops, and is parallel to the bedding. These laminations are commonly less than 1 cm in thickness (fig. 35), and are occasionally as thin as 1 mm.

In most places the laminations are demonstrably of primary sedimentary origin because textural changes occur between the laminations. The grain sizes within adjacent laminations commonly differ by 2 Wentworth grades (from coarse to fine sand). Rarely do sand grains exceed 1 mm in diameter (fig. 35). In some places, however, the structure is due to chloritic laminae extending across the outcrops, and in these instances it is difficult to determine whether the laminations are the result of primary sedimentary processes, or are due to shearing.

Parallel lamination occurs at outcrops within thick arkosic sandstone sequences, but rarely occurs in arkose units interbedded with granite-bearing conglomerates.

At some outcrops, parallel lamination is associated with some other sedimentary structures, but the sequences of structures differ in each case. At one locality at Closs Lake, parallel lamination occurs within arkosic sandstone immediately above a pebble band (about 10 cm thick). The parallel laminated unit is about 50 cm in thickness, and the laminations die out vertically, giving rise to massive arkose. However, at Kenneally Lodge (fig. 2), parallel lamination occurs for about 2 m within arkose, immediately beneath a massive granite-bearing conglomerate unit several metres thick (fig. 36). Also, at one locality near Kenneally Lodge, a 50 cm thick unit of massive arkosic sandstone occurs above a pebble band, and parallel laminations are present for a further 3 m in the arkose immediately overlying this massive unit. The boundary between the massive and parallel laminated units is gradational, but the top of the parallel laminated arkose is not exposed.

Thus, no regular repeated sequences of sedimentary structures exist within the arkosic sandstone facies.

(iii) Graded bedding

At a few localities around Little Vermilion Lake, single beds of arkose are found with upward decreases in the sand grain size (fig. 37). The way-up of the stratigraphic section is known in these instances by the presence of cross-bedding nearby in the sequence.

The beds are 30 to 40 cm thick, but in some instances the grading dies out when traced for only 2 or 3 m along strike. This is the case at Island C, near Twin Falls (fig. 2), where the graded arkose bed thickens laterally and gives way to pebbly arkosic sandstone.

The bases of these beds are sharp, but the upper boundaries are commonly indistinct, where the graded beds pass into massive arkose.

The bases of graded beds are commonly composed of very coarse sand, but can also contain isolated pebbles or cobbles (fig. 37). Medium to fine sand usually comprises the uppermost parts of these rare graded beds.

At Abram Lake, the stratigraphic top of the succession is thought to be to the north (see Chapter 2). If this is the case, then some reverse grading is present (fig. 38). This example is from an arkose unit interbedded with conglomerate, on Island B in Abram Lake (fig. 2), and grades from medium sand at the base of the bed, to coarse pebbly arkose at the top, which then passes into massive conglomerate.

Well developed graded or reversely graded beds are uncommon within the arkosic sandstone facies, and those beds which <u>are</u> graded pass upward into massive finer arkose, never into argillite.

(iv) <u>Other bedding features</u>

A channel-fill of very coarse massive arkose is found within a bed of fine to medium grained arkose, in the north of Ament Bay (fig. 39). The channel occurs within a 1 m thick arkose unit, which is interbedded with massive granite-bearing conglomerates. The scour is approximately 2 m wide and 50 cm deep. The base is rather diffuse in places, but faint parallel laminations in the arkose beneath the channel-fill, were observed to be truncated by coarser arkosic sandstone fill. This was the only example of a channel-fill found within the arkosic sandstone facies.

Figure 24. Interbed of massive arkose, wedging out within massive granite-bearing conglomerate. Note the straighter upper boundary of the interbed than the lower boundary. Arrow indicates top direction to the southeast. Scale in centimetres. Kenneally Lodge.

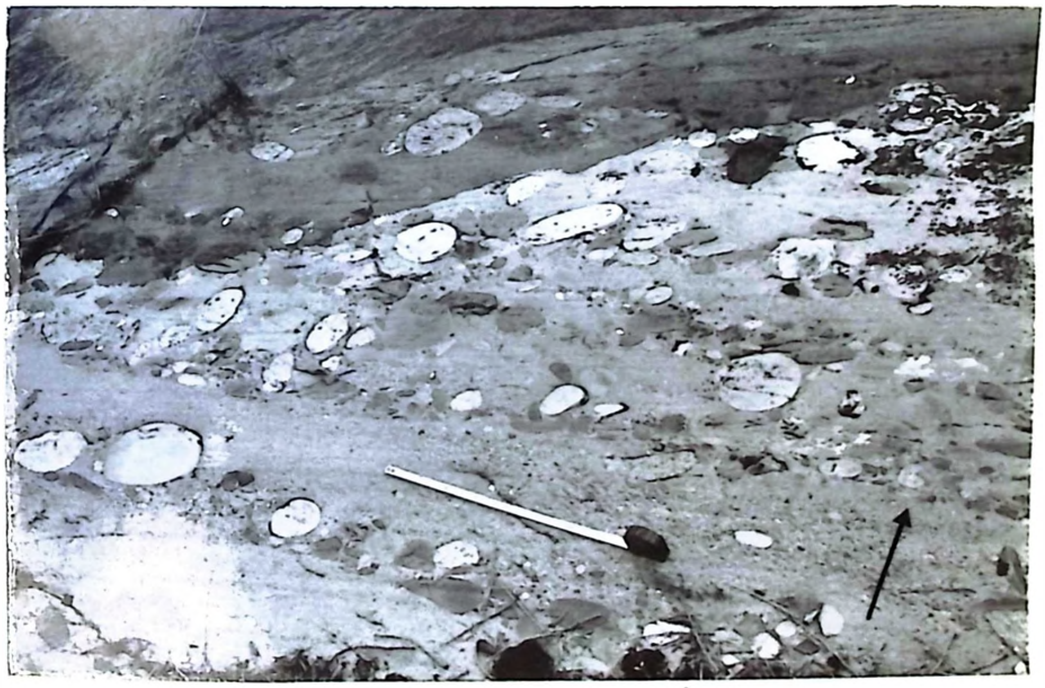


Figure 25. Interbedded pebble bands and arkosic sandstone. Note imbricate nature of clasts near the centre of the photograph indicating a westerly component of the depositional current, and also the undeformed nature of the greenstone clasts. Arrow indicates top direction to the southeast. Scale in centimetres. Island B, Ament Bay.

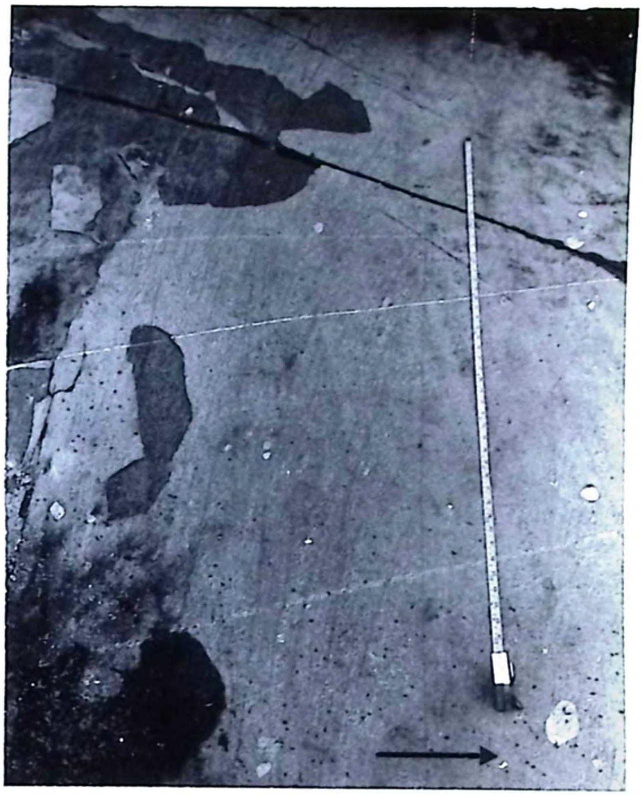


Figure 26. Low angle cross-bedding within coarse to very coarse arkosic sandstone. Note textural changes between individual cross-strata, and small pebbles part-way along the strata. A westerly component of the depositional current is indicated. Arrow indicates top direction to southeast. Scale in inches and centimetres. Island B, Ament Bay.

Figure 27. Low angle trough cross-bedding within an interbed of arkosic sandstone between massive granite-bearing conglomerate beds. Arrow indicates top direction to the southeast. Scale in inches and centimetres. North Ament Bay.

Figure 28. Low angle trough cross-bedding within fine to medium grained arkosic sandstone. Arrow indicates top direction to the north. Scale in inches and centimetres. Island B, Vermilion Lake.

Figure 29. Graded arkosic sandstone passing upward into trough cross-bedded arkose. Irregular base of the bed contains cobbles. Arrow indicates top direction to southeast. Scale in inches and centimetres. West of Kenneally Lodge.

Figure 30. Cross-bedded arkosic sandstone unit overlying massive arkose, with a pebble row marking the bedding surface. A westerly component of the depositional flow direction is indicated, as arrow indicates top direction to southeast. Scale in inches and centimetres. Closs Lake.

Figure 31. Trough cross-bedded arkosic sandstone exhibiting grading within individual sets (thin arrows). A westerly component of the depositional flow direction is indicated as thick arrow indicates top to the southeast. Scale in inches and centimetres. West of Kenneally Lodge.

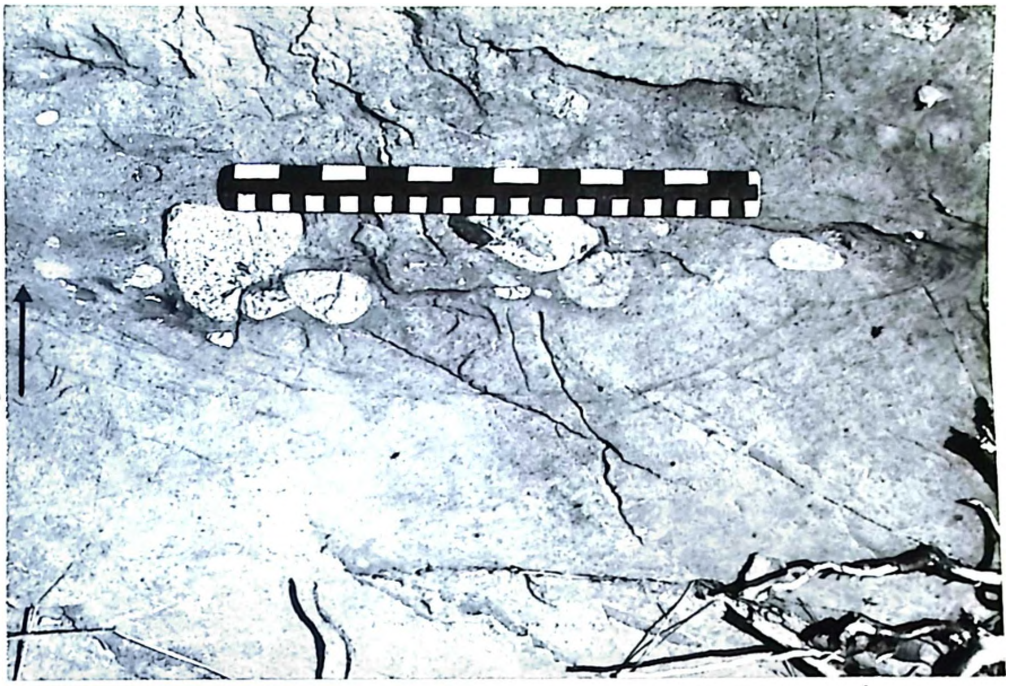


Figure 32. Symmetrical trough cross-bedding within arkosic sandstone, beneath massive arkose bed with a pebble row at the base. Arrow indicates top direction to the southeast. Scale in inches and centimetres. North Closs Lake.

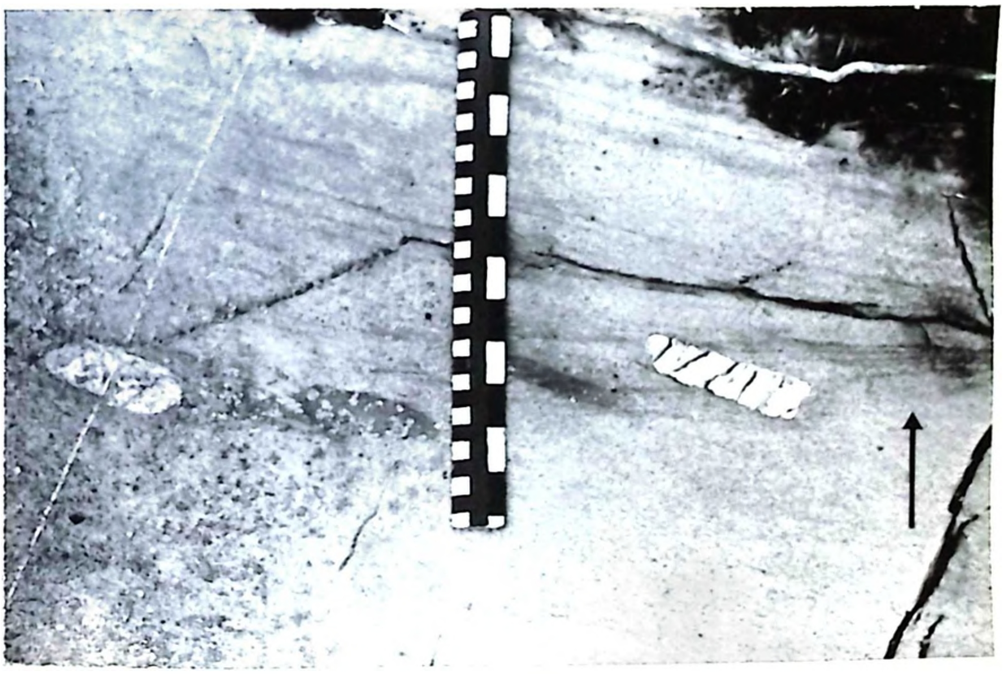


Figure 33. Cross-bedded arkosic sandstone with pebbles at the toes of the cross-strata, overlying massive arkose. A westerly component of the depositional current is indicated, as the arrow indicates top direction to the southeast. Scale in inches and centimetres. Island B, Ament Bay.

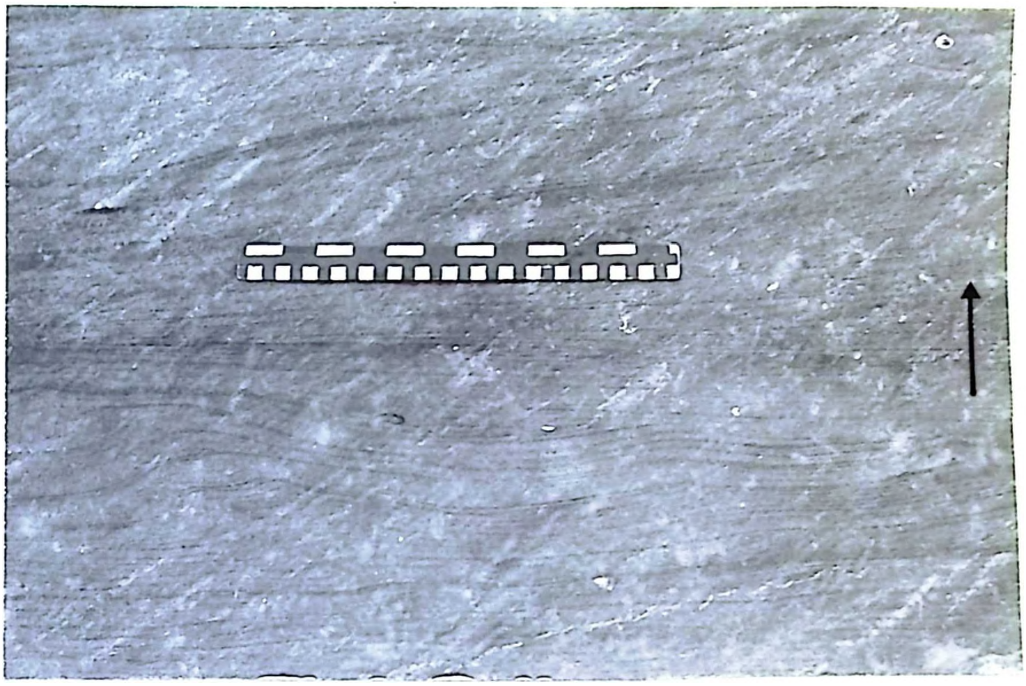


Figure 34. Convolute lamination within fine grained arkosic sandstone. Low angle cross-bedding is also evident. Westerly component of the depositional current is indicated, as the arrow indicates top direction to the north. Scale in inches and centimetres. Island B, Vermilion Lake.

Figure 35. Parallel lamination within fine to coarse arkosic sandstone. Note textural changes between laminations. Arrow indicates top direction. Scale in centimetres. West of Kenneally Lodge.

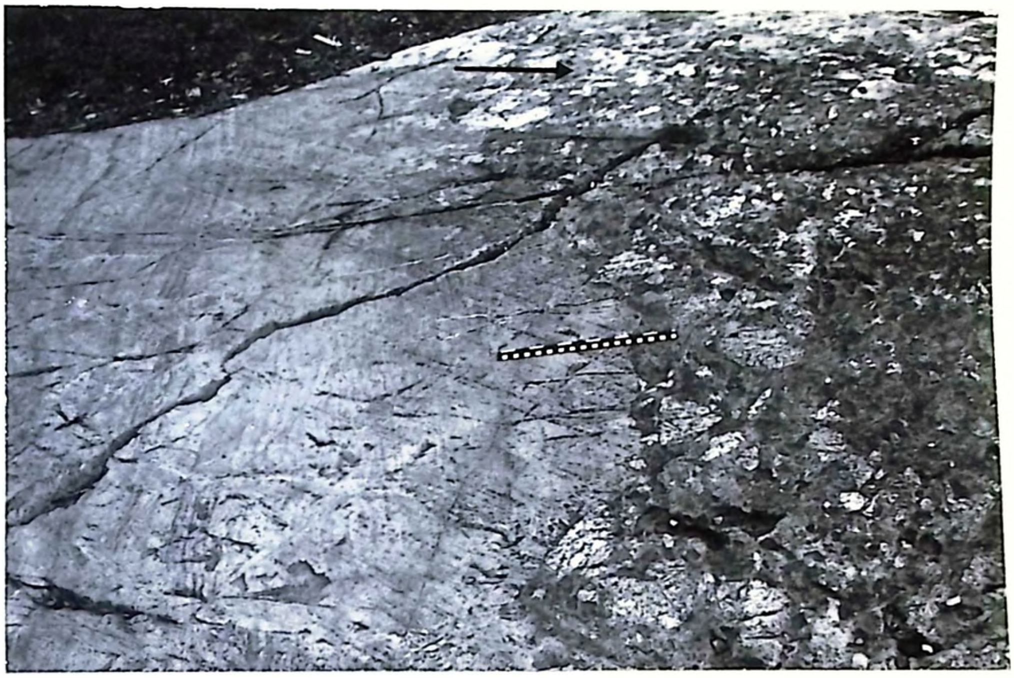


Figure 36. Massive granite-bearing conglomerate with an irregular base, overlying parallel-laminated arkosic sandstone. Arrow indicates top direction. Scale in inches and centimetres. Kenneally Lodge.

Figure 37. Graded arkosic sandstone bed including a felsic cobble at the base, passing upward into massive arkose. Arrow indicates top direction to the southeast. Scale in inches and centimetres. West of Kenneally Lodge.

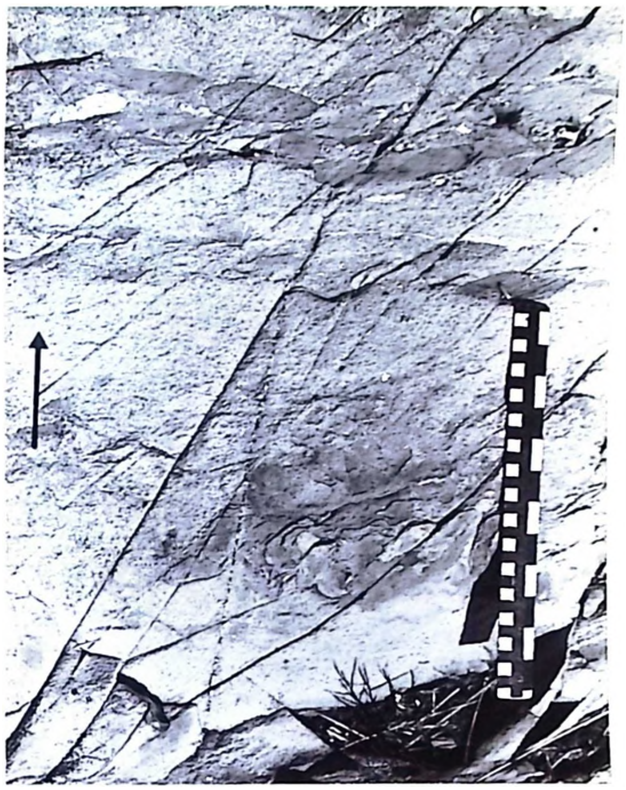


Figure 38. Reversely graded arkosic sandstone bed. Arrow indicates probable top direction to the north. Scale in inches and centimetres. Island B, Abram Lake.

Figure 39. Channel in-fill of massive coarse arkose within parallel-laminated interbed of arkose, between massive granite-bearing conglomerate beds. Arrow indicates top direction to the southeast. Scale in centimetres. North Ament Bay.

4. Felsic tuff facies

The felsic tuff facies crops out at both Little Vermilion Lake and Abram Lake. Because there are certain differences between the felsic tuffs of the two areas, the facies at each lake will be described separately.

(a) Felsic tuff at Little Vermilion Lake

The felsic tuff facies forms the lowest 100 m of the Daredevil Formation (1,700 to 1,800 m of fig. 6), and occurs directly above the arkosic sandstones and granite-bearing conglomerates of the Ament Bay Formation. The felsic tuff unit is overlain by a 60 m section of the greywacke facies, and is exposed along an approximately south-west trending belt, from the western part of Little Vermilion Lake to Daredevil Lake (fig. 2). It is uncertain what happens to the felsic tuff to the west of Daredevil Lake, but the unit may die out, because Johnston (1966a) mapped no tuff at Blackfox or at Hooch Lakes (fig. 2).

(i) Lithologies and structures

Most felsic tuff occurs as massive, granular exposures, which weather to a white colour due to the large amounts of plagioclase in the rocks (see Chapter 5, and fig. 40). A small amount of grey quartz grains are scattered throughout the felsic tuff, but these commonly comprise only 5-10 percent of the rock (fig. 65). Most grains of feldspar and quartz are angular or sub-angular, and are up to 5 mm in diameter (fig. 40).

The appearance of the felsic tuff in the field is somewhat similar to that of massive arkosic sandstone, but the great abundance of feldspar, the small amount of quartz, the angularity of the fragments, and the coarse grain size, distinguish the felsic tuff.

No bedding surfaces or sedimentary structures were observed within the felsic tuff, and very few clasts other than the feldspar and quartz grains are present. However, at a few localities, the felsic tuff outcrops contain 1 or 2 dark green beds of very fine grained chloritic material (fig. 41). These beds are commonly less than 30 cm thick, are approximately parallel-sided, and either continue across the width of the outcrop, or abruptly pinch out within the felsic tuff. In places, angular chloritic fragments less than 2 cm in diameter are present. The chloritic beds commonly have a schistose texture, and both sides of a bed have sharp straight contacts with the felsic tuff (fig. 41).

No grading (or any other sedimentary structure) was observed within the chloritic beds, and no clasts are present in the very fine grained material. However, a very few grains of quartz and feldspar (up to about 3 mm in diameter) are scattered within some chloritic bands.

On the shore of Little Vermilion Lake, only 6 chloritic bands were observed within 80 m or so of felsic tuff, and each band was less than 20 cm in thickness. However, along the north-west shore of Daredevil Lake (approximately 2.5 km south-west of the outcrops at Little Vermilion) several chloritic beds exceed 1 m in thickness, and one unit is at least 10 m thick (the contacts with felsic tuff not being exposed).

A possible lateral equivalent of the felsic tuff facies occurs along the shore of Little Vermilion Lake, to the east of the entrance to Ament Bay (fig. 2 and map 1). Here, a massive but well jointed very fine grained rock is exposed at a few poor outcrops. The rock has a hard cherty character in the field, but subhedral plagioclase crystals (less than 1 mm in length) can be seen in thin sections. Grains of polycrystalline quartz are also present, set in an extremely fine grained quartzose matrix.

This lithology may originally have been an acidic lava, but no flow structures were observed in the small outcrops.

(ii) Stratigraphic relations

At Little Vermilion Lake, the boundary between the felsic tuff facies and the underlying pebbly arkose of the granite-bearing conglomerate facies is poorly exposed, but originally may have been conformable. The contact is now marked by a band of shearing (about 5 m wide), but the strikes of the beds on either side of the contact are approximately parallel, and no major fault separates the two facies.

No felsic tuff units occur within the underlying arkosic sandstone or granite-bearing conglomerate facies.

The boundary between the felsic tuff facies and the overlying section of the greywacke facies is faulted (see Chapter 2), thus the original relationship between the rock types is obscured. However, this section of greywacke is found at both Little Vermilion Lake and Daredevil Lake (fig. 2), and it is probable that the greywacke was originally deposited on top of the felsic tuff, and that the greywacke section does not represent a localized "faulted-in" unit.

(b) Felsic tuff at Abram Lake

Thick successions of felsic tuffs occur around the central part of Abram Lake (between Frog Rapids and Island B), and also at the east end of Abram Lake near the Sturgeon River (fig. 2). The felsic tuff facies which crops out in the central part of Abram Lake will be discussed below, and the exposures in the eastern part of the lake will be only briefly mentioned, as little detailed study was undertaken in this more highly sheared eastern area. Some of the outcrops in the central part of the lake which were previously mapped as arkose (Johnston, 1967b), were re-mapped in this study as tuffs (map 3). Thin section analyses also supports this latter interpretation (see Chapter 5).

(i) Lithologies and structures

In the central part of Abram Lake, the stratigraphic succession youngs to the north-west (see section on stratigraphy below), and the contact of the granite-bearing conglomerate facies with the felsic tuff facies, is exposed at Island B (fig. 2, map 3, and fig. 42). The felsic tuff facies is here conformable on massive granite-bearing conglomerate, and consists of crudely bedded, white weathered, fine grained felsic rock, which commonly has a shallowly-pitted surface (fig. 42). In places, the tuff is difficult to distinguish from fine grained arkosic sandstone. However, no sand-size grains of quartz are visible in the tuff (either in the outcrops or in most thin sections), and this, together with the rougher weathered surfaces than most arkosic sandstones, facilitates distinction between them.

At Island B (fig. 2), the felsic tuff facies is irregularly bedded, but at other outcrops of similar tuff at Island A (fig. 2), bedding is uncommon. Beds on Island B seldom exceed 1 m in thickness, and are recognisable due to the development of agglomerate phases within some units. In these agglomerate units most fragments are angular, are less than 30 cm in length, are aligned parallel to the bedding, and are composed of felsic tuff (fig. 43). The bedding surfaces are rarely very sharp, and in a few beds, upward grading occurs from felsic fragments a few cm long, to fine grained felsic tuff. However, grading is rare and dies out along strike.

Some of these tuffaceous beds may have been reworked by water currents, but few fragments are rounded to any degree, and no sedimentary structures, such as cross-bedding, are present within the beds.

At Island A in Abram Lake (fig. 2), most outcrops of felsic tuff are rather similar in appearance to the tuff at Island B, but the exposures are more massive.

Other outcrops of the felsic tuff facies nearer to Frog Rapids are composed of rather different lithologies. One outcrop of fine grained felsic tuff occurs at locality C (fig. 2), and contains chloritic bands which commonly pinch out in one or more directions along the strike. These chloritic bands are similar to those found within the felsic tuff facies at Little Vermilion Lake. They have straight, sharp boundaries, are massive and very fine grained, and most are about 5 cm in thickness (fig. 44).

The other types of tuffaceous deposits around the central part of Abram Lake, include the following:

(i) Finely laminated (from 1 mm to about 1 cm) felsic tuff (fig. 45) with some thin argillite interbeds, which crop out on either side of Frog Rapids (map 3);

(ii) Contorted, alternating, extremely fine grained chloritic and silicic tuff layers, within 1 or 2 m of section. These bands range from less than 1 mm in thickness to about 5 cm (fig. 46), and crop out to the east of Frog Rapids;

(iii) A massive fine grained felsic tuff, which overlies a quartz and feldspar porphyry body (fig. 47), at locality D (fig. 2).

(ii) Stratigraphic relations

In the stratigraphic successions at both Little Vermilion

Lake and the east end of Abram Lake, the felsic tuff facies overlies sequences of arkosic sandstones and granite-bearing conglomerates. Thus, although there is no reliable way-up data for the central part of Abram Lake, it is probable that a similar situation exists there. This implies that the stratigraphic succession in the central part of Abram Lake youngs to the north-west.

The base of the felsic tuff facies is exposed at Island B, where it conformably overlies granite-bearing conglomerate, but the top of the stratigraphic interval of felsic tuff is not exposed at Abram Lake. The structural control in this central part of Abram Lake is rather poor, but if the whole section is homoclinal, with vertically dipping beds (as is the case at most outcrops), there could be over 1,000 m of tuff present at Abram Lake. If this is the case, then the thickness of tuff at Abram Lake is approximately three times that of the felsic plus basic tuffs at Little Vermilion Lake (fig. 6).

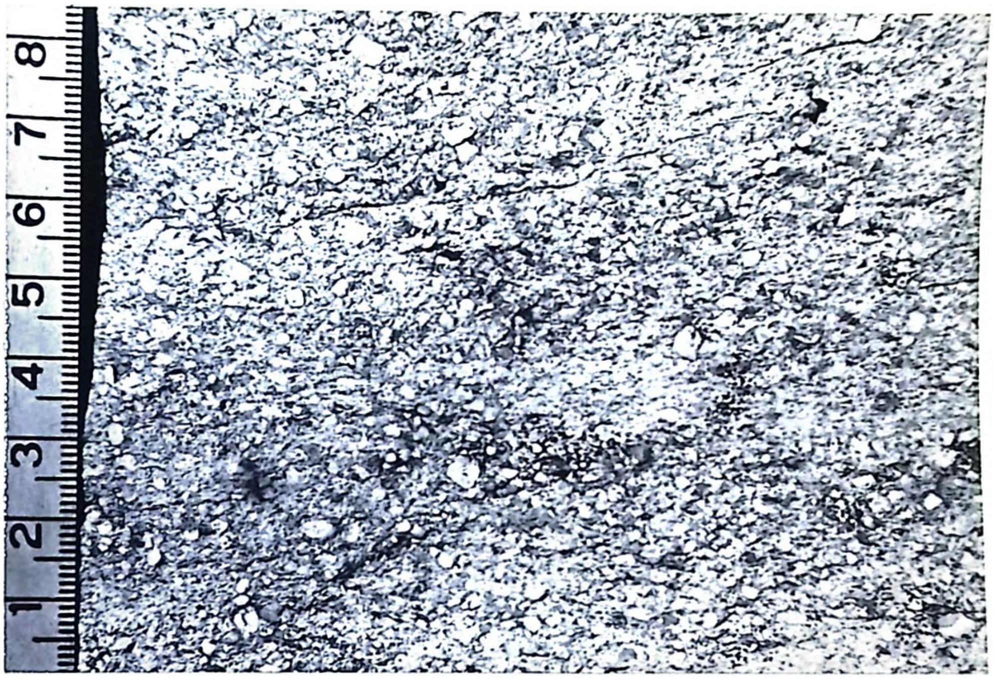


Figure 40. Massive felsic tuff outcrop. Note angular white feldspar grains and grey quartz grains. Scale in centimetres. West Little Vermilion Lake.

Figure 41. Schistose chloritic bed between massive felsic tuff units. Arrow indicates top direction. Scale in inches and centimetres. Northwest Daredevil Lake.

Figure 42. Sheared granite-bearing conglomerate overlain by felsic tuff. The felsic tuff in the photograph was probably re-deposited by water currents. Arrow indicates probable top direction to the north. Scale in centimetres. Island B, Abram Lake.

Figure 43. Felsic fragments in an agglomerate-like phase of the felsic tuff at Island B, Abram Lake. Arrow indicates probable top direction to the north. Scale in inches and centimetres.

Figure 44. Massive felsic tuff with discontinuous massive chloritic beds. Arrow indicates probable top direction to the north. Scale in centimetres. North shore of Abram Lake.

Figure 45. Well-bedded tuff at Frog Rapids. Scale in inches and centimetres.

Figure 46. Finely interbedded chloritic and cherty (?)tuff layers. Arrow indicates probable northerly top direction. Scale in inches and centimetres. East of Frog Rapids.

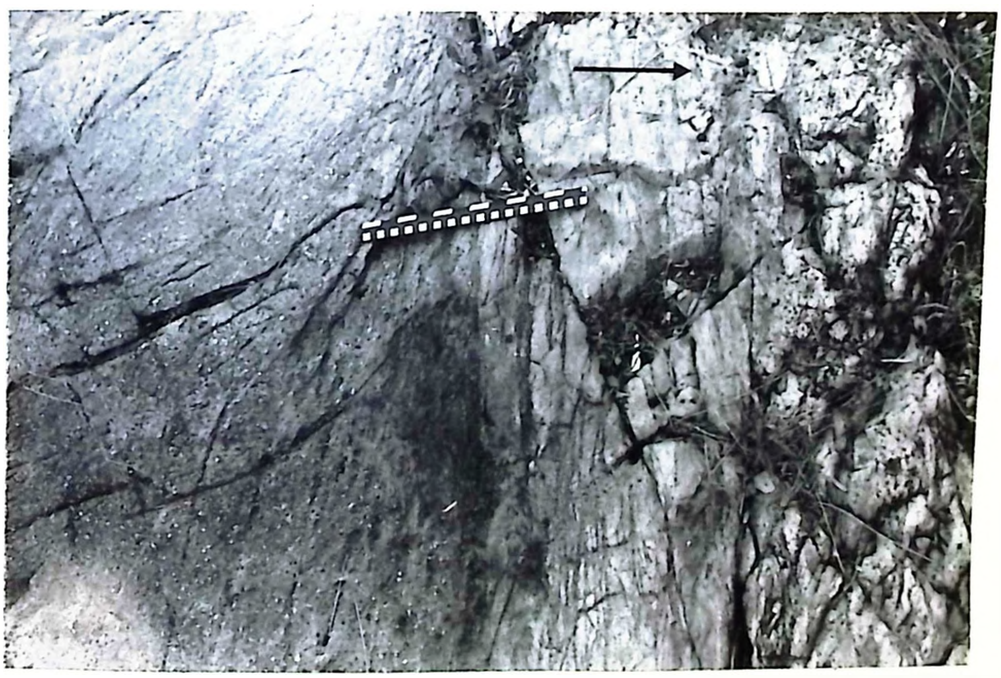


Figure 47. Quartz and feldspar porphyry body, overlain by felsic tuff. Arrow indicates probable northerly top direction. Scale in inches and centimetres. Locality D, Abram Lake.

5. Basic tuff facies

A 270 m thick unit of basic tuff occurs at the top of the Daredevil Formation at Little Vermilion Lake (1,860 to 2,130 m of figure 6). The same unit also occurs farther west, along the south shore of Daredevil Lake (fig. 2), but like the felsic tuff facies, it may die out to the west of Daredevil Lake. No basic tuff occurs at Abram Lake.

(a) Lithologies and structures

The basic tuff facies commonly has a dark, massive and fine grained appearance, but in some places it is crudely bedded, with individual units up to about 30 cm thick. These beds extend across the width of outcrops. Some outcrops contain scattered angular white feldspar grains up to 4 mm in diameter, and biotite and hornblende grains are visible in some places (see Chapter 5). No quartz grains are visible in outcrop (and there is very little quartz in the thin sections), which helps to distinguish the facies from massive greywacke.

An agglomerate-like phase is developed at a few localities in the lower part of the basic tuff unit (notably at the west end of Daredevil Lake), which has elongate blocks enclosed in a finer grained matrix (fig. 48). Some fragments are greater than 1 m in length, but most are less than 10 cm. Re-entrant angles are found on some fragments (fig. 48), which suggests that no reworking of the agglomerate has occurred. No clasts larger than about 10 cm occur within the basic tuff at Little Vermilion Lake.

Graded bedding occurs within at least 2 beds near the centre of the basic tuff facies at locality D at Little Vermilion Lake (fig. 2 and fig. 49). The beds are about 30 cm thick, have sharp bases and grade upward from coarse to very fine grained tuff. The grading indicates a southerly top direction (fig. 2). These graded tuffs may either have resulted from single volcanic eruptions, or they may have been resedimented by turbidity currents.

A few rare beds of black argillite (less than 50 cm thick) are found within the basic tuff facies, near to the occurrence of graded beds. This indicates that the basic tuffs were deposited in an environment favourable for the accumulation of shales (see Chapter 4).

The upper half of the basic tuff facies at Little Vermilion Lake has a more granular, felsic appearance than the lower half, and is also more massive. In thin section this tuff is indeed found to be more felsic than the very basic lower part (see Chapter 5).

(b) Stratigraphic relations

The basic tuff facies is in fault contact with the underlying greywacke facies (fig. 6), and has an apparently conformable boundary with the greywackes of the overlying Little Vermilion Formation. Unfortunately the actual contact between these two facies is not exposed, but massive tuff crops out about 5 m beneath the graded greywacke beds at locality E (fig. 2), and there is no reason to suspect a faulted contact.

Figure 48. Felsic fragments within an agglomerate phase of the basic tuff facies at Daredevil Lake. Note reentrant angles of the large clast. Scale in inches and centimetres.

Figure 49. Graded bed within the basic tuff facies. Note slight loading at the base (dashed line). Arrow indicates a top direction to the southeast. Scale in inches and centimetres. Locality D, Little Vermilion Lake.

6. Greywacke facies

Occurrences of the greywacke facies are found at both Abram Lake and in the Little Vermilion Lake area (fig. 2). The greatest thickness and the best exposures of the greywacke facies occur at Little Vermilion Lake. These will be described first, followed by a short section on the greywacke at Abram Lake.

A. Greywacke at Little Vermilion Lake

(a) Introduction

(i) Location of facies

At Little Vermilion Lake the greywacke facies comprises the youngest sedimentary rocks within the stratigraphic section (fig. 6), and the main outcrops of the facies occur in the south-western part of the study area (fig. 2). However, a poorly exposed 60 m section of graded greywackes occurs at about 1.8 km to the north-east of the entrance to Ament Bay, and an 80 m section of predominantly black slate occurs at Island A in the eastern part of Little Vermilion Lake (map 1). This black slate unit is problematical as it is not found at any other locality at Little Vermilion Lake, but it may represent a fault-bounded lateral equivalent of the greywacke facies.

(ii) Stratigraphic relations

One 60 m section of greywacke occurs between the felsic and basic tuffs in the Daredevil Formation (1,800 to 1,860 m of figure 6) in the south-west part of the area, and extends from Little Vermilion Lake to at least the west side of Daredevil Lake. The contacts with both the felsic and basic tuffs are apparently faulted (see Chapter 2). The basic tuffs overlying this greywacke section pass upward into the main section of greywacke of the Little Vermilion Formation (2,130 to 2,700 m of figure 6). This topmost stratigraphic unit is at least 570 m thick, but the top of the section is missing due to repitition of the lower part of the greywacke sequence in the Little Vermilion Syncline (fig. 2).

The greywacke facies was divided in the field into the following 3 subfacies:

A. Thickly bedded and graded greywacke;

B. Thinly bedded greywacke and argillite;

C. Argillite conglomerate.

The thickly bedded and graded greywacke subfacies is most prominent within the 60 m section between the felsic and basic tuffs, and also in the lower 200 m or so of the Little Vermilion Formation (fig. 6 and map 2). The thinly bedded greywacke and argillite subfacies, and the argillite conglomerate subfacies are interbedded with each other in the central 150 m (2,350 to 2,500 m of figure 6) of the Little Vermilion Formation (see below). The topmost 200 m of the Little Vermilion Formation is composed of thickly bedded and graded greywacke subfacies, together with the argillite conglomerate subfacies.

(iii) Lithologies

The greywacke facies is readily distinguishable from the arkosic sandstone facies, the most distinctive difference being the well-bedded nature of most greywacke outcrops. There is also a greater variety of sedimentary structures within the greywackes (see subfacies descriptions below). These include graded bedding, parallel lamination, convolute lamination, flame structures and load casts, soft sediment deformation structures, and one occurrence of cross-stratification (about 15 cm thick).

Most rocks in the greywacke facies are finer grained than those of the arkosic sandstone facies, and include massive argillites, siltstones, fine and medium grained greywackes, and argillite conglomerates. Most clasts in these argillite conglomerates are composed of argillite or greywacke, and are commonly less than 10 cm in length.

Petrographic and chemical analyses of greywackes and argillites are given in Chapter 5.

(b) Subfacies A - Thickly bedded and graded greywacke

This subfacies consists of alternating beds of coarse to very fine grained greywacke, and black argillite. Most sand grains within the greywackes are less than 1 mm in diameter, and few exceed 2 mm. Thickly bedded and graded greywacke beds rarely contain any large clasts, but a few angular slabs of argillite occur within the lower parts of some greywacke beds.

These clasts are aligned sub-parallel to the bedding, and the largest such clast observed has dimensions of 30 cm by 5 cm.

Most beds within this subfacies are less than 1 m in thickness, the average being about 30 cm (fig. 51), and some graded beds are as thin as 3 cm (fig. 52). All the graded greywacke beds have approximately constant thicknesses across the widths of outcrops. The greywacke beds commonly grade up into black argillite layers thinner than the underlying greywacke (fig. 51), but the tops of some graded beds consist of grey siltstone (fig. 53), and in places the greywacke beds have amalgamated to form composite beds 2 to 3 m thick. The sandstone:shale ratio for this subfacies is commonly greater than 9:1 (fig. 50).

Some detailed measured sections of greywacke beds in the thickly bedded and graded greywacke subfacies are shown in figure 50.

(i) Sedimentary structures

Soles of beds - Most greywacke beds have sharp, straight bases, and extend across the width of outcrop. Because of the vertical dips and the horizontal nature of most exposures, observation of the soles of greywacke beds is not possible. However, flame structures are present at the bases of a few greywacke beds (fig. 53), and other rare beds have load casts which are sinusoidal in profile (with amplitudes of about 1 cm).

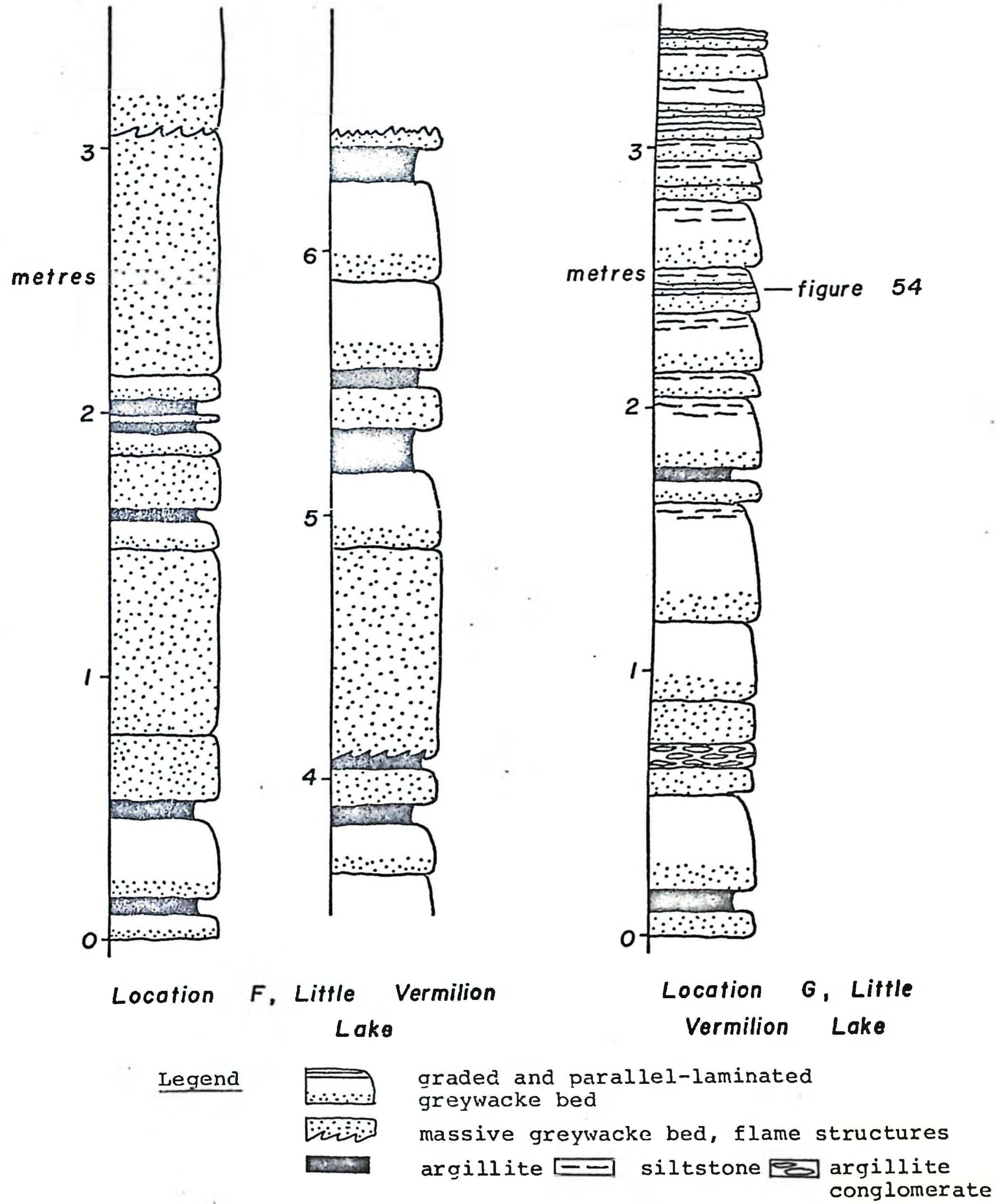


Figure 50. Stratigraphic sections from outcrops of the thickly bedded and graded greywacke subfacies. See map 2 for locations.

Grading - Some greywacke beds are massive, and contain no sedimentary structures, but most outcrops of the thickly bedded and graded greywacke subfacies contain many graded beds. This is commonly the only sedimentary structure within greywacke beds, and in places the grading is marked by refracted cleavage within the greywacke and argillite (fig, 53).

Two types of grading were recognised:

(i) The most common type occurs when a gradual decrease of the grain size takes place upward from the very base of a bed, to its boundary with an overlying bed (fig. 54). The greywacke at the bases of the beds is commonly of medium or coarse grain size (up to about 1 mm), and that at the tops of beds is commonly of very fine sand or siltstone grade.

(ii) In a few beds there is a distinct difference between the grain size in the lowest quarter or so of the bed (which is itself graded in some places), and that of the upper part (figs. 52 and 53). In these cases, the lower part of the bed is composed of coarse to fine grained greywacke, and in the finer grained upper part, grading is almost imperceptible, the top of the bed being composed of very fine greywacke or siltstone. This type of grading is rather rare.

Parallel lamination - Parallel lamination is common in graded greywackes, and normally occurs above the massive or graded bases of a greywacke bed, and dies out farther up the bed (fig. 54). These laminations are not marked by grain size changes, but are very thin physical partings parallel to the bedding. Most parallel lamination is of primary sedimentary origin, because the laminations only occur in restricted parts of greywacke beds. However, the origin of some "parallel lamination" may be tectonic, especially when the parallel partings are irregularly spaced and occur throughout graded beds (fig. 55).

Cross-stratification - No true ripple cross-lamination was found within the greywacke facies in this area. However, at one locality (I on map 2) at Little Vermilion Lake, low-angle cross-stratification is present within a 60 cm thick bed of coarse and medium grained greywacke (fig. 56). The bed is ungraded, but grain size changes are evident between individual cross-strata. This single set of cross-stratification is about 15 cm thick, and the angles of truncation of the strata at the top of the set vary from 10 to 20 degrees. The cross-strata dip in a generally westerly direction, relative to the bedding. This is the only example of cross-stratification observed within the greywacke facies.

Convolute lamination - At 2 or 3 localities around Little Vermilion and Maskinonge Lakes (fig. 2), convolute laminations occur within greywacke beds up to about 1 m thick.

Within one 50 cm thick greywacke bed at Maskinonge Lake, the lowest 15 cm consists of graded greywacke passing up into 20 cm of convolute laminations, which themselves pass upward (north-westward) into 15 cm of banded black argillite and very fine greywacke.

Another example of convolute lamination occurs at Little Vermilion Lake and is also part of a sequence of sedimentary structures within a greywacke bed (fig. 57). At this locality, a 1.5 m massive coarse grained greywacke unit containing argillite flakes (up to 4 cm long) passes up into 15 cm of parallel laminations, which become progressively more buckled upwards until a ripple form occurs (outlined on fig. 57), with an amplitude of about 1 cm. Deformation of the ripple cross-lamination gives rise to a zone of convolute lamination about 10 cm thick within very fine grained greywacke (also outlined on fig. 57).

A depositional current flowing from right to left (i.e. in a roughly north-east direction) could have produced the ripple form in this bed.

Soft sediment deformation (slumping) - Soft sediment deformation occurs within a 3.5 m thick section of graded greywackes at location J at Little Vermilion Lake (map 2). This is the only example of slumping found within this subfacies, and occurs near the top of the Little Vermilion Formation. Flat beds of graded and massive greywacke are interbedded with argillite beneath the slumped bedding (fig. 58), and an undisturbed layer of argillite conglomerate occurs above. The chaotically bedded greywacke is interfingered with irregular, discontinuous layers of massive argillite.

Penecontemporaneous slumping of the top few metres of a pile of unconsolidated greywackes and argillites could have produced this structure. However, the direction of slump movement is uncertain, as the structure can only be viewed in 2 dimensions.

(c) Subfacies B - Thinly bedded greywacke and argillite

Thin interbeds of siltstone or very fine grained greywacke, and black argillite make up this subfacies. The siltstone beds are commonly less than 5 cm in thickness, and continue for several metres across the width of outcrop (fig. 59), but some very fine grained greywacke beds are greater than 10 cm thick. In places regular inter-laminations (less than 1 mm thick) of siltstone and argillite occur over a few centimetres of sediment, but in most cases the siltstone bands are of various thicknesses. Also, some sediment consists of discontinuous siltstone streaks a few millimetres thick, enclosed within massive argillite.

(i) Sedimentary structures

The bases of the siltstone and very fine grained greywacke layers are generally sharp, but the upper boundaries with argillite are commonly less distinct.

A few siltstone beds grade visibly upward into argillite (fig. 60), but most bands appear to be massive in the field. However, polished sections of 2 apparently massive siltstone bands (about 2 cm thick), revealed upward decreases in grain sizes through the bands. Thus it is likely that grading is more common in the siltstones than would appear in the field.

Delicate parallel laminations are present in some beds of siltstone, but no cross-lamination was seen.

Small-scale soft-sediment deformation is present in at least 6 outcrops (fig. 60). The deformation is in the form of small asymmetrical folds, which only affect 5 cm or so of sediment thickness. All such folds have their crests overfolded in an easterly direction, relative to the present strike. Slumping is further discussed in Chapter 4.

(ii) Relationship with other subfacies

There is no strict stratigraphic limit to the thinly bedded greywacke and argillite subfacies, but it is best exposed and most prevalent at localities K and L at Little Vermilion Lake (map 2), in the central 200 m or so of the main greywacke section, and also at Maskinonge Lake (fig. 2). It is uncommon in the lowest 200 m or so of the Little Vermilion Formation (fig. 6).

The thinly bedded greywacke and argillite subfacies is commonly interbedded with the argillite conglomerate subfacies, with which it has sharp contacts (fig. 61), but it is only interbedded with graded greywacke at a few localities at Little Vermilion Lake. However, at Maskinonge Lake, siltstone streaks within bands of argillite up to about 50 cm thick are found in a few places, interbedded with thick massive or graded greywacke beds up to about 1 m thick.

(d) Subfacies C - Argillite conglomerate

The argillite conglomerate contains elongate clasts of argillite, siltstone and greywacke, together with small amounts of sandy matrix (fig. 62) composed of many angular quartz and white feldspar grains up to about 1 mm in diameter. The clasts are commonly less than 5 cm in length, rarely exceed 1 m, but have a maximum length of 6 m and thickness of 1.5 m at location M on map 2.

Petrographic descriptions of these clasts and the matrix, are given in Chapter 5. The sedimentary nature of the clasts indicates that the conglomerate is of intraformational origin.

(i) Sedimentary structures

Beds of argillite conglomerate vary in thickness from about 5 cm to over 10 m, but there is little regular change of clast size within any of the beds.

The bases of these beds are straight and sharp (fig. 61), and in most outcrops the clasts are aligned subparallel to the bedding. However, at one locality the clasts in an argillite conglomerate are aligned at angles of about 15° to the underlying contact with argillite (fig. 61). It is thought that this alignment represents primary imbrication, because there is no evidence of rotation of the clasts and there is little cleavage within the underlying argillite. If this structure is indeed imbrication then the depositing current flowed in a generally easterly direction.

No sedimentary structures such as cross-bedding or grading are present within these argillite conglomerates.

(ii) <u>Relationship with other subfacies</u>

This subfacies occurs only within the upper two thirds of the Little Vermilion Formation (2,350 to 2,700 m of figure 6), where it is commonly interbedded with the thinly bedded greywacke and argillite subfacies (fig. 61). In some places towards the top of the Little Vermilion Formation, argillite conglomerate beds are interbedded with massive or graded greywacke beds. No argillite conglomerate was found at Maskinonge Lake in the west.

At location L (map 2) a rock type occurs which is intermediate between typical argillite conglomerate, and the thinly bedded greywacke subfacies (fig. 63). This outcrop only exposes 1 m of section, but siltstone laminae and a few greywacke clasts within argillite pass upward into a conglomerate of irregularly shaped greywacke clasts (one of which is greater than 1 m in length). Several small-scale soft-sediment folds occur in the siltstone layers, and some of these layers extend for 1 m or so across the outcrop before being abruptly truncated. Thus, in this case the argillite conglomerate contains slabs of bedded sediment up to about 5 cm thick, and may be the result of the initiation of slumping (see Chapter 4).

B. Greywacke at Abram Lake

The only beds of greywacke at Abram Lake occur at the extreme east end of the lake (fig. 2). Here, a section of at least 300 m consists of beds of graded and massive greywacke, argillite, and two or more bands of iron formation.

The graded beds indicate that the succession youngs to the south-east. This implies that the felsic tuff to the north of the greywacke section (fig. 2) is older than the greywackes (Chapter 2). This stratigraphy is similar to that at Little Vermilion Lake (30 km to the west-south-west), and so the greywackes at the two lakes may be of similar age.

The chief differences between the greywacke facies at Little Vermilion Lake and that at Abram Lake are the absence of argillite conglomerate and the presence of lean iron formation at Abram Lake. Two bands of iron formation (each about 30 cm thick) occur within a 15 m section of greywacke beds (mostly 10-15 cm thick). The black finely banded beds of iron-rich sediment are slightly magnetic and contain fine grained magnetite and pyrite, and a considerable amount of carbonate, within a very fine matrix of quartz and chlorite (fig. 64).

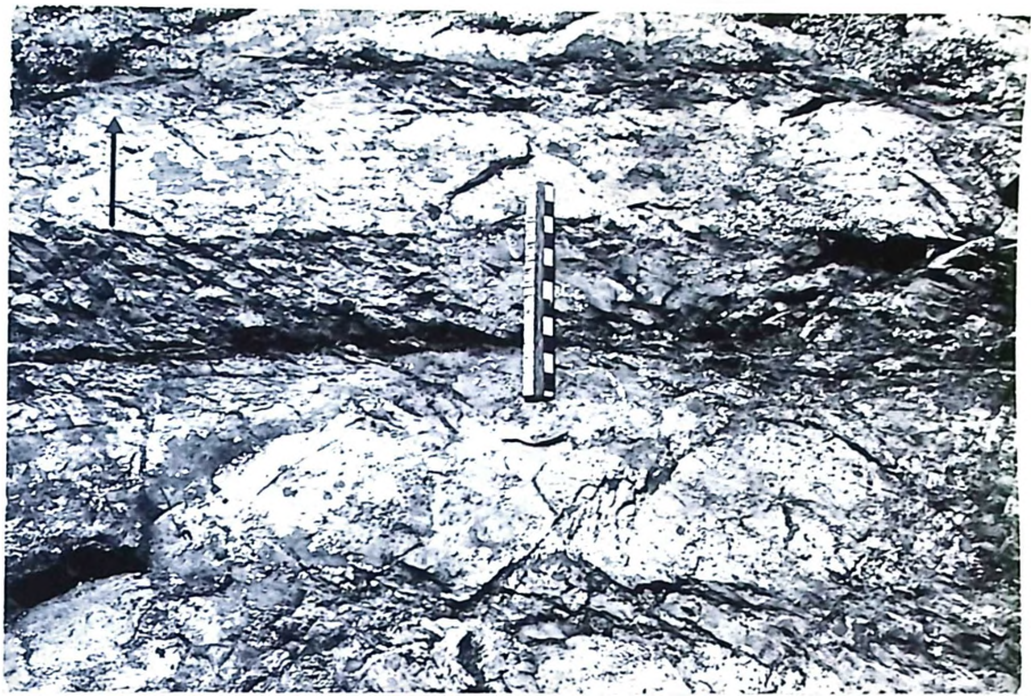


Figure 51. Interbedded greywacke and argillite within the Daredevil Formation. Note refracted cleavage and grading indicating a top direction to the southeast (arrow). Scale in inches and centimetres. Locality F, Little Vermilion Lake.

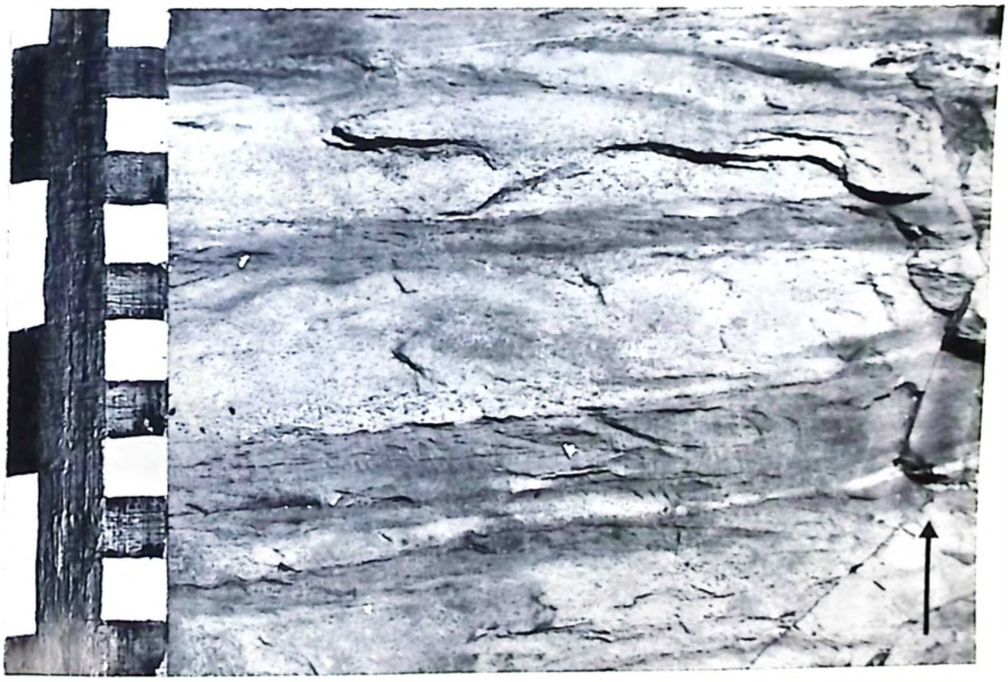


Figure 52. Thin graded greywacke beds with interbedded argillite. Note sharp bases of greywacke beds, and concentration of coarse grains near the bases. Arrow indicates top direction. Scale in inches and centimetres. Locality F, Little Vermilion Lake.

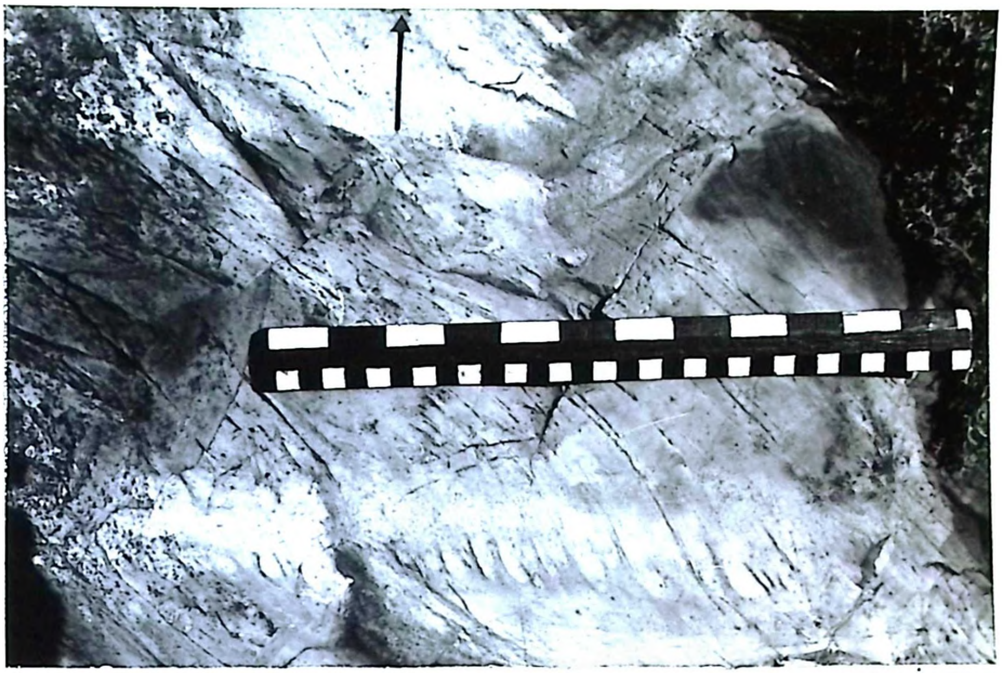


Figure 53. Graded greywacke bed, exhibiting refracted cleavage and load casts emphasized by the cleavage. Arrow indicates top direction. Scale in inches and centimetres. Locality F, Little Vermilion Lake.

Figure 54. Greywacke bed exhibiting grading and parallel lamination. Note sharp straight base of the bed, and the overall grading. Arrow indicates top direction. Scalc in inches and centimetres. Locality G, Little Vermilion Lake.

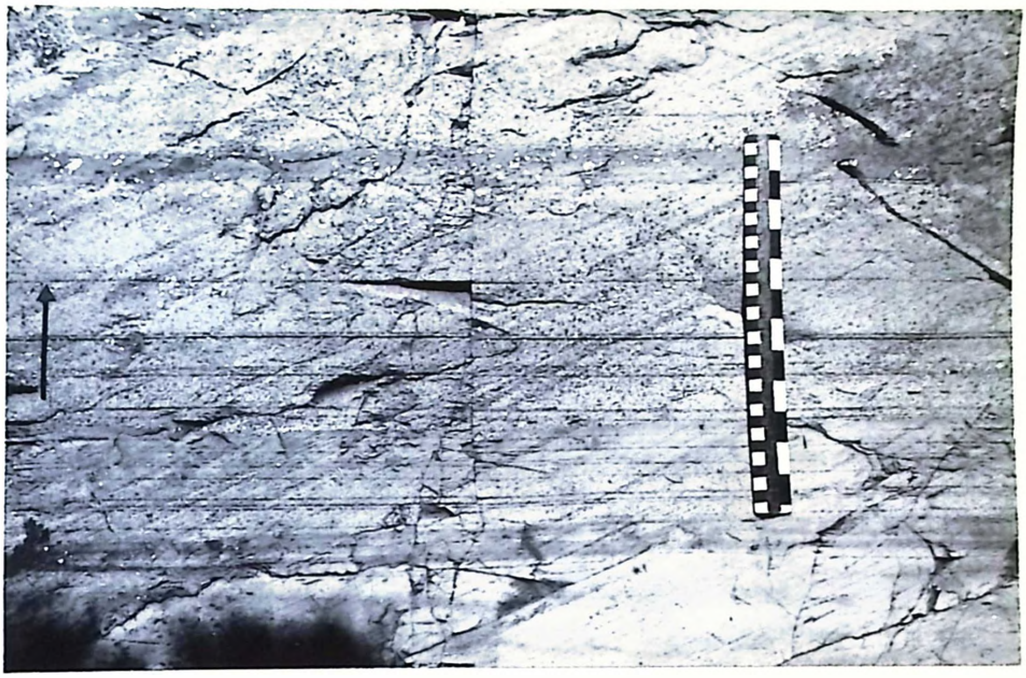


Figure 55. Graded greywacke beds with little interbedded argillite. "Parallel lamination" may be of tectonic origin. Arrow indicates top direction. Scale in inches and centimetres. Locality I, Little Vermilion Lake.

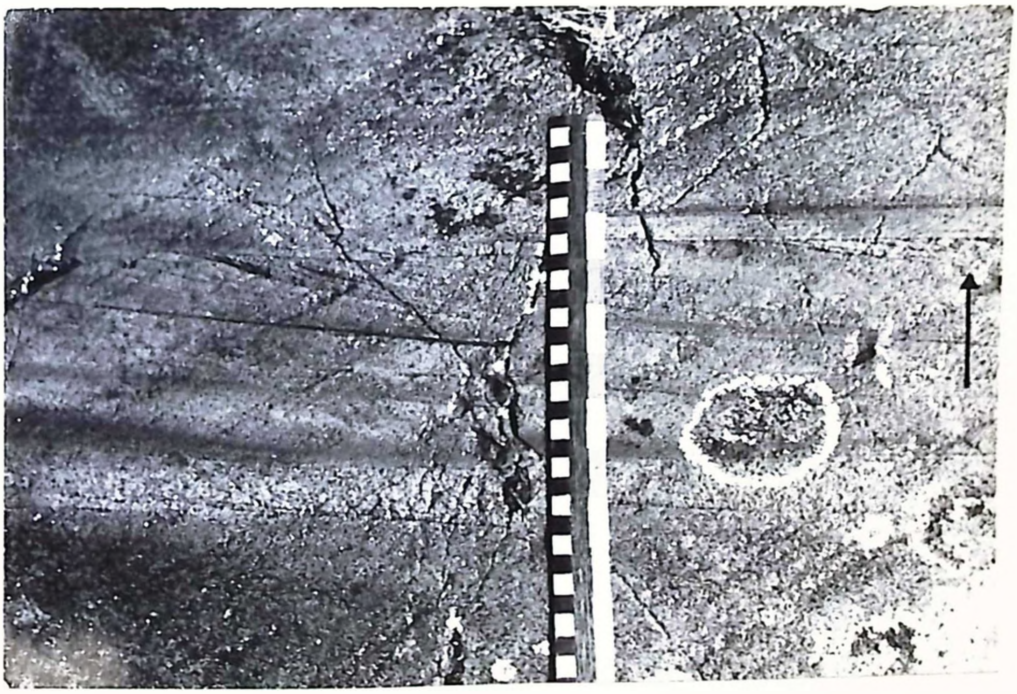


Figure 56. Cross-bedding within greywacke at locality I, Little Vermilion Lake. Arrow indicates top direction. Scale in inches and centimetres.

Figure 57. Massive greywacke passing upward into parallel laminations, then into convolute lamination (outlined). Note the light coloured argillite fragments in the lower part of the bed. Arrow indicates top direction. Scale in inches and centimetres. Locality H, Little Vermilion Lake.

Figure 58. Graded greywacke beds with argillite interbeds are replaced upward (arrow) by slumped greywacke and argillite. Scale in centimetres. Locality J, Little Vermilion Lake.

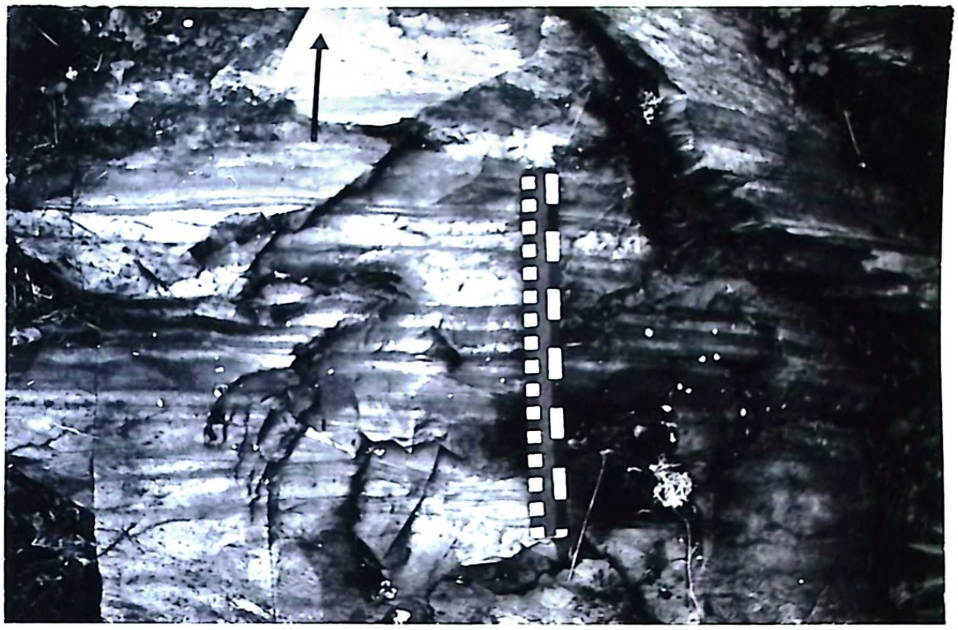


Figure 59. Thinly bedded greywacke and argillite, with an argillite conglomerate bed at the top of the photograph. Note the straight bedding surfaces. Arrow indicates top direction. Scale in inches and centimetres. Locality K, Little Vermilion Lake.

Figure 60. Thinly bedded greywacke and argillite, exhibiting grading (A) and slump folding. Arrow indicates top of sample. Scale in inches and centimetres. Locality K, Little Vermilion Lake.

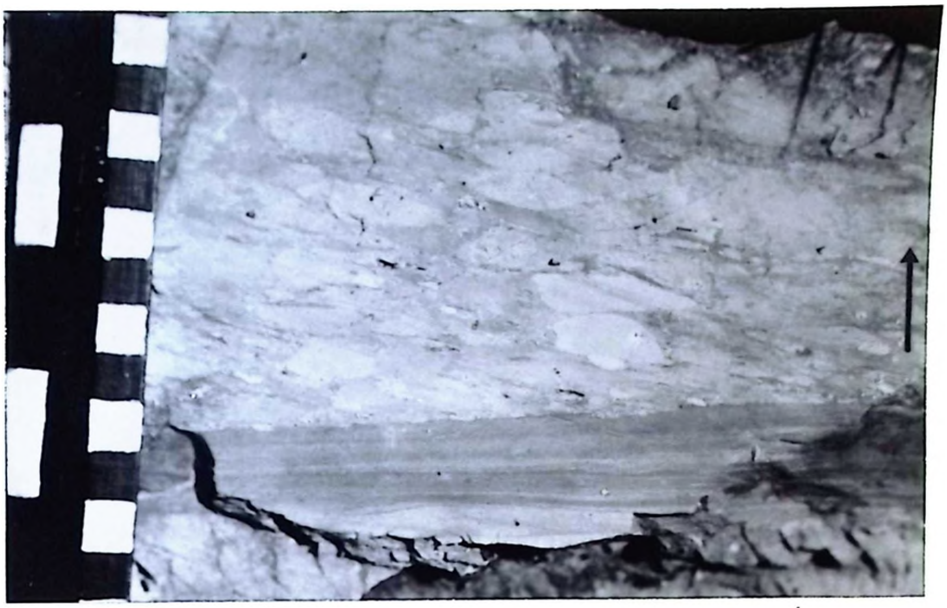


Figure 61. Imbricated argillite conglomerate, overlying thinly laminated argillite and siltstone. Note the sharp base of the argillite conglomerate bed, and the absence of cleavage from the argillite layers. An easterly component of the depositional current is indicated, as arrow indicates top direction to the southeast. Scale in inches and centimetres. Locality K, Little Vermilion Lake.

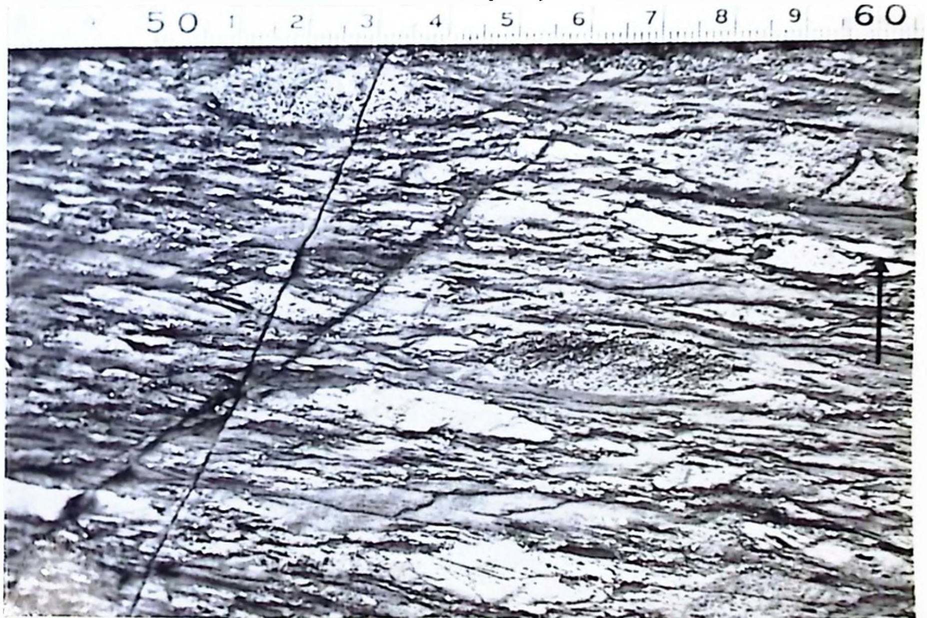


Figure 62. Argillite conglomerate. Clasts are of greywacke and argillite. White feldspar grains occur in the matrix. Arrow indicates top direction. Scale in centimetres.

Figure 63. Thinly bedded greywacke and siltstone, passing upward (arrow) into argillite conglomerate. Note the small slump folds in the greywacke layers. This outcrop may represent the initiation of slumping. Scale in inches and centimetres. Locality L, Little Vermilion Lake.

Figure 64. Iron-rich layers (A and B) interbedded with greywacke in the east of Abram Lake. Arrow indicates top direction. Scale in inches and centimetres.

CHAPTER 4

DEPOSITIONAL PROCESSES AND ENVIRONMENTS

1. Introduction

The foregoing chapters describing the stratigraphy, sedimentary lithologies and sedimentary structures of the facies of the Abram Group, permit deductions regarding the sedimentary processes and the depositional environments of the sediments. For this purpose the arkosic sandstone and granite-bearing conglomerate facies are discussed together, because their intimate field relationships indicate deposition in the same general environment, but under slightly different hydrodynamic conditions.

The discussion of the arkoses and conglomerates of the Ament Bay Formation is followed by interpretations of the depositional conditions of the felsic tuff, basic tuff and greywacke facies.

An outline of the sequence of basin filling and the changing depositional environments is given in Chapter 8.

2. Arkosic sandstones and granite-bearing conglomerates (Ament Bay Formation)

The basic facts on which the interpretation of the Ament

Bay Formation is based are as follows.

1. The Ament Bay Formation at Vermilion Lake rests unconformably on a quartz porphyry body of the underlying northern volcanic belt. Felsic tuffs of the Daredevil Formation abruptly overlie the formation.

2. Most arkose and conglomerate beds tend to be thick and massive.

3. Some strong depositional current activity is indicated by cross-bedding and parallel lamination in the arkosic sandstones, and by imbrication in the granite-bearing conglomerates. No cross-stratification occurs within the conglomerates, and no cross-lamination (sets <5cm thick) was observed in the arkosic sandstones.

4. Graded bedding is rare in the Ament Bay Formation, and repeated sequences of sedimentary structures do not occur.

5. No argillite horizons occur within the Ament Bay Formation, and no argillite clasts are present in the conglomerates.

Many Archaean sediments in recent years have been shown to be turbidites (see Chapter 1) and thus there is an obligation to consider carefully this possibility for the sediments of the Ament Bay Formation. Other environments in which massive thick conglomerates are presently accumulating include beaches, alluvial plains, braided streams, and alluvial fans. Shallow shelf areas are not presently sites of significant conglomerate deposition.

These several possibilities are considered in order below.

(a) Turbidite environment

The possibility that the Ament Bay Formation was deposited in a turbidite environment has been rejected for the following reasons. The absence of repeated sequences of sedimentary structures (Bouma sequences) and the general lack of grading do not indicate a turbidite origin. The lack of any argillite horizons suggests that deposition was in an environment distinctly unfavourable for the slow accumulation and preservation of muds. Turbidite environments, however, are generally areas below wave base where the deposition of pelagic sediment is interrupted by brief influxes of sediment-laden turbidity currents (for a review of turbidite environments, see Lajoie, 1970). In this same greenstone belt, Walker and Pettijohn (1971) demonstrated a background of mud sedimentation for the turbidites of the Minnitaki Group. A list of differences between the Ament Bay Formation and Minnitaki Group is given in table 3.

The large scale cross-bedding within the arkosic sandstones (e.g. figs. 30 and 31) is also not consistent with a deep basinal environment, and strongly suggests some form of shallow-water currents. The character of the unconformable base of the Ament Bay Formation further suggests that a deep basinal environment is unlikely.

Minnitaki Group, Facies 1 White arkose, of Walker and Pettijohn (1971)	Ament Bay Formation, Arkosic sandstone facies
1. Contain sandstones, graded conglomerates, and pebbly mudstones.	1. Contains sandstones and pebble bands. No pebbly sandstones.
2. Graded bedding fairly common.	2. Graded bedding rare.
3. Argillite interbeds common.	3. Argillite interbeds absent.
4. Argillite clasts common.	4. Argillite clasts absent.
5. Occurs in 5-15 m graded beds, or in massive beds several metres thick.	5. Normally occurs in massive beds.
6. Associated with relatively thin (10-200 cm) graded conglomerate beds.	6. Associated with massive ungraded conglomerates, up to 20 m thick.
7. Associated with slate facies.	7. Never associated with argillites.
8. Never cross-stratified on a scale >10cm. No parallel lamination, and no imbrication of pebbles.	8. Cross-stratified on a scale >10cm. Parallel lamination present, and some pebble layers are imbricated.

Table 3a. Comparison of the White arkose facies of Walker and Pettijohn (1971), with the arkosic sandstone facies of the Abram Group.

Minnitaki Group, Facies 4 Granite-bearing conglomerate, of Walker and Pettijohn (1971)	Ament Bay Formation, Granite-bearing conglomerate facies
1. Boulders occur toward the bases of very well-graded sand beds.	1. Conglomerates are massive and ungraded
2. A spectrum of facies between sandy greywackes, pebbly greywackes, and graded granite-conglomerates.	2. Conglomerates never associated with graded greywackes.
3. Boulders die out westward.	3. Boulders show no consistent size decrease along the length of Little Vermilion Lake.

Table 3b. Comparison of the granite-bearing conglomerate facies of Walker and Pettijohn (1971), with the granite-bearing conglomerate facies of the Abram Group.

(b) Beach environment

A beach origin for the conglomerates of the Ament Bay Formation was considered, but this has also been rejected as unlikely.

The thick accumulations of predominantly conglomerate with lesser amounts of interbedded arkosic sandstones at some localities (e.g. over 120 m in the north of Ament Bay - Appendix 1) are unlikely to have been built up by beaches. Beaches are likely to form only thin units in transgressive or regressive sedimentary sequences. Alternations of gravels (beaches) and sands (nearshore or continental sediments) would imply repeated transgressions and regressions in a rapidly subsiding basin; but there is no other evidence of this in the associated facies.

Also, the immature nature of both the conglomerates and the arkosic sandstones is inconsistent with a constantly shifting beach or agitated shallow marine environment. Much less feldspar might be expected in beach deposits than is present within the arkosic sandstones (Appendix 2) (Blatt, Middleton and Murray, 1972, p.282).

Studies on modern beaches have shown that a large proportion of the clasts are disc-like (Bluck, 1967; Dobkins and Folk, 1970), but this is not the case with the conglomerates of the Ament Bay Formation (see Chapter 3).

The arkosic sandstones associated with the conglomerates in the Ament Bay Formation, are commonly massive, and occasionally exhibit cross-bedding. This is also inconsistent with a shallow nearshore environment or a distal alluvial plain environment adjacent to a beach, where much cross-stratification might be expected in the arkoses (Clifton et al., 1971; Goldring, 1971; Coleman and Gagliano, 1965).

(c) Alluvial plain environment

Conglomerates and sands can be deposited in close association from streams in an alluvial plain environment (Allen, 1965). The deposits are normally cross-stratified and commonly show sequences of sedimentary structures which imply in-channel or point bar deposition of coarse sand and gravel, and fine overbank deposits. The varied deposits of alluvial plains differ considerably from the massive arkoses and conglomerates of the Ament Bay Formation, and thus an alluvial plain environment of deposition is unlikely for these Archaean deposits.

(d) Braided stream environment

Recent studies of the processes and deposits of braided streams have demonstrated that sands and gravels are intimately associated in braid bars and channels; that the sands are almost invariably cross-stratified, and that gravel bars are also commonly cross-stratified and contain many imbricate clasts (Williams and Rust, 1969; Eynon, 1972). The paucity of cross-stratification in the arkosic sandstones of the Ament Bay Formation, and its absence in the conglomerates, makes it unlikely that the bulk of these arkoses and conglomerates were deposited in braided streams.

(e) Alluvial fan environment

The arkosic sandstones and granite-bearing conglomerates of the Ament Bay Formation show similarities to younger alluvial fan deposits described in the literature, and it is with an alluvial fan origin in mind that aspects of the arkoses and conglomerates are discussed below.

(i) Basal conglomerate

The basal conglomerate of the Ament Bay Formation has a localized distribution and a different composition and texture from other conglomerates higher in this formation. The fact that the boundary of the basal conglomerate with the underlying quartz porphyry body is straight and sharp, and that angular clasts of quartz porphyry and other detritus immediately overlie the contact, indicate that erosion of the quartz porphyry body was in progress at about the same time as the deposition of the basal conglomerate was taking place. The straight nature of the bounding surface of the quartz porphyry suggests that this part of the quartz porphyry body had been eroded to a plane surface prior to the deposition of the basal conglomerate. The massive nature of this basal conglomerate, and its coarse clast sizes suggests that deposition was from a rapid water flow which carried the quartz porphyry debris for only a short distance, so that rounding of the clasts could not proceed to any degree, and also such that no sedimentary structures were produced in the basal conglomerate.

However, the flow which deposited the basal conglomerate was not so localized that it only contained quartz porphyry debris. The greenstone clasts in the basal conglomerate indicate that sediment was being supplied from other than very localized sources, but these may only have been a few kilometres from the site of deposition, as the quartz porphyry body was probably not very extensive.

The massive nature of the basal conglomerate and the small extent of the outcrop make interpretation of its origin difficult. However, the basal conglomerate may have been deposited in a similar manner to that of the massive conglomerates higher in the Ament Bay Formation. These latter conglomerates are suggested below to be alluvial fan flood surge deposits, and a similar flood surge origin is postulated for the basal conglomerates. The basal contact of the Ament Bay Formation is similar in some respects to the basal unconformity of the Torridonian sandstones and conglomerates in Scotland (Williams, 1969). The quartz porphyry body beneath the basal conglomerate at Vermilion Lake is not weathered, unlike most of the Lewisian gneiss underlying the Torridonian, but Williams (1969) suggested that areas where fresh Lewisian gneiss underlay Torridonian sediments were the result of flushing away of the weathered mantle before sediment deposition occurred. It is not known whether a weathered mantle ever existed on the quartz porphyry body at Vermilion Lake, but some form of erosion obviously occurred before the basal conglomerate was deposited.

It is tentatively suggested here that the basal conglomerate of the Abram Group represents an alluvial fan accumulation on a pediment surface.

(ii) Massive conglomerates with arkosic sandstone interbeds

The lack of sedimentary structures within the massive conglomerates and interbedded arkoses of the Ament Bay Formation is probably an original depositional characteristic. It is unlikely that cross-bedding or imbrication were originally present in most conglomerates and have since been destroyed by metamorphism or deformation, because cross-bedding is preserved in some arkoses, and imbrication is present in some conglomerates which appear to be as undeformed as the massive conglomerate units. Thus, most conglomerates were probably originally deposited in massive beds.

This implies that most conglomerates were not deposited pebble by pebble from normal streams, because this latter process might be expected to produce imbrication (Rust, 1972) and cross-stratified gravel bars (e.g. in the braided stream environment). However, strong ephemeral sediment-laden surges can deposit massive beds of conglomerate greater than 2 m thick (Scott and Gravlee, 1968). McGowan and Groat (1971) suggested that such surge deposits probably accounted for the cobble and boulder beds of the proximal facies of alluvial fan deposits in the Van Horn Sandstone of West Texas (Precambrian or lower Paleozoic in age). The stratigraphic sections of McGowan and Groat's (1971) proximal facies consist predominantly of thin interbeds of sandstone (commonly less than 30 cm thick) between massive conglomerate units (some over 10 m thick). These sections are very similar to those measured in part of the Ament Bay Formation (see Appendix 1), and the two deposits may have had similar origins.

The types of deposit which are found on alluvial fans are sheetflood deposits, stream deposits, sieve deposits, and debris flow deposits (Bull, 1972), commonly more than one type being found in an alluvial fan sequence. These types will be considered below.

It is unlikely that the conglomerates of the Ament Bay Formation were deposited by debris flows as they are not chaotically bedded, they contain little original mud, and they contain better rounded clasts than the debris flow deposits described by Hooke (1967) and Fisher (1971). A matrix of fine sediment is one of the most distinctive and essential features of a debris flow deposit (Hooke, 1967, p.452), but the conglomerates of the Ament Bay Formation have sandy matrices, and it is improbable that they could originally have been debris flows. The lack of very fine grained sediments in the Ament Bay Formation probably reflects the absence of very fine detritus in the source area.

Special conditions are required on alluvial fans to enable sieve deposits to form. Bull (1972, p.69) stated that gravel lobes (sieve deposits) may be deposited if the source area supplies little sand, silt and clay to the fan, such that the deposits will be sufficiently permeable to allow water from a flood discharge to infiltrate entirely before reaching the toe of the fan. However, most conglomerates of the Ament Bay Formation contain from 20 to 35 percent sandy matrix (fig. 7), and may thus have been too impermeable to allow sieve deposits to form. It may have been possible for the sand matrix to be emplaced after the conglomerates were deposited, in which case sieve deposits might have formed, but as arkose was the background sediment it is likely that surges which deposited conglomerates would also contain sand. Thus the sieve deposit mechanism is thought to be unlikely for the deposition of the massive conglomerates in the Ament Bay Formation.

The most probable processes for the deposition of the massive conglomerates are sheetfloods or streamfloods, and of these, streamfloods are thought to be more likely because they are capable of depositing thicker gravel units (Bull, 1972). McGowan and Groat (1971, p.36) suggested that the conglomerates of the proximal facies of the Van Horn Sandstone were deposited mostly in canyons near the head of the fan, where flow was confined; and it is here suggested that similar restricted channels may have been the sites of deposition of the thick conglomeratic sections of the Ament Bay Formation. The rounded nature of most clasts in the conglomerates suggests that the clasts have been fairly well water-worn, which is more consistent with a stream-flood origin than with any other alluvial fan process. The imbricate nature of a few outcrops of conglomerate (fig. 12) also suggests deposition from a stream flow, perhaps of lesser intensity and/or longer duration than the floods which deposited the more massive, thicker conglomerates.

The deposition of arkosic sandstone interbeds may have taken place from sheetfloods or from shallow, possibly anastomising channels on a gravelly alluvial fan surface, between the flood surges which deposited conglomerates. The cross-bedded nature of a few arkose interbeds (fig. 27) supports a stream origin, as do the lenticular geometries of some arkose interbeds (fig. 24) and the channelled nature of at least one arkose interbed (fig. 39). Other massive, parallel-sided arkose interbeds may be sheetflood deposits, or the massive deposits of wide shallow streams. These interbedded arkoses contain few pebbles, and were thus deposited from much less powerful currents than were the conglomerates. However, it is not clear whether the arkoses were deposited by surges, or whether arkose was being supplied to the fan more or less constantly, whilst the conglomerates were being deposited from violent ephemeral surges.

(iii) <u>Thick arkosic sandstone sequences</u>

The thick successions of predominantly massive arkosic sandstone are typified by the measured section along the west shore of Closs Lake (section no. 3 in Appendix 1). Besides massive arkose, these sections include cross-bedded arkose, parallel laminated arkose, a very few graded arkose beds, pebble bands and isolated pebbles.

The overriding difficulty concerning the interpretation of the arkoses is their common very massive nature. It is thought that few sedimentary structures have been destroyed, because the outcrops of massive arkose show no more signs of deformation or metamorphism than do the outcrops of finely cross-bedded arkose. However, parallel lamination may have been originally more common than is now observable, as this often rather subtle structure would be more easily obscured by post-depositional events than would the more pronounced cross-bedding. It is unlikely that much more cross-bedding was originally formed than is presently observable.

However, the sedimentary structures which do exist indicate that deposition of at least the arkoses containing these structures took place from strong currents of water transporting large volumes of predominantly coarse to fine sand, together with occasional pebbles.

By analogy with accounts of cross-stratification from modern rivers it is suggested that the large scale trough cross-bedding in the arkosic sandstones may have been produced by the infilling of scours in channel beds, perhaps by advancing dunes

(Harms and Fahnstock, 1965). It is also suggested that the parallel laminations in the arkosic sandstones were produced by flow over plane beds, probably in the upper flow regime (Simons et al., 1965). Much of the arkosic sandstone facies is composed of coarse or very coarse sand size grains, which might preclude the development of ripples (Simons et al., 1965). However, the absence of ripple-produced cross-lamination in the finer arkoses suggests that the flows which emplaced these arkoses were never low enough in the flow regime to form ripples, and also that the arkoses were not reworked in shallow standing bodies of water, as wave-produced ripples might then be expected.

At only a few localities were gradational sequences of sedimentary structures observed (fig. 29). The sequence shown in figure 29 consists of graded arkose which passes upward into trough cross-bedded arkose, and is probably the deposit from a waning flow.

The common discontinuous and continuous pebble bands, and the isolated pebbles within thick massive arkose units suggest that the flows emplacing the massive arkoses were commonly powerful enough to transport clasts up to 15 cm in diameter (fig. 19). Some of the pebble bands occur between beds of arkose (figs. 18, 19, 20) and may represent channel lag accumulations, as in these cases the bands are commonly associated with cross-bedded arkose. Also, some of the discontinuous pebble bands (e.g. fig. 19) resemble the "pebble clusters" in gravelly fluvial deposits described by Dal Cin (1968). This further suggests that the pebble bands may be lag deposits.

The massive nature of most of the arkosic sandstones (Appendix 1) suggests that deposition of the massive arkoses was not from channelized flows. If the depositional currents had been channelized then much more cross-bedding would be expected, as is the case in the sands of the mid-fan and distal-fan facies of the Van Horn Sandstone, which were probably deposited by braided streams (McGowan and Groat, 1971). It is thus suggested that the thick massive arkoses of the Ament Bay Formation were deposited as sheetfloods on the lower parts of alluvial fans, where gravel deposition was infrequent and was only in the form of very thin sheets. According to Bull (1972, p.66) sheetflood deposits on modern alluvial fans commonly consist of gravel, sand or silt, are well sorted, and may be cross-bedded, laminated or massive. This indicates that sheetfloods are indeed a possible mechanism for the deposition of massive sands.

These sheetfloods probably flowed over the lower parts of alluvial fans, possibly whilst massive gravels were being deposited in the more confined flows near the heads of the fans. Also floods which were not competent enough to transport gravel may have deposited arkoses, both on top of the head-fan gravels and also over the lower slopes of the fans.

(f) Summary and discussion of depositional processes of the Ament Bay Formation

It is suggested that the conglomerates and arkoses of the Ament Bay Formation were deposited on alluvial fans by fluvial processes. The massive nature of most conglomerates is attributed to violent ephemeral surges depositing coarse gravels near the heads of the fans, possibly in a similar manner to the deposits from the surge on the Rubican River described by Scott and Gravlee (1968).

The arkoses interbedded with the conglomerates were probably deposited by sheetfloods or in shallow anastomising streams on the gravelly upper parts of the fans. The massive and parallel laminated arkoses in the thick arkosic sandstone sequences were probably deposited by sheets of sand-laded water spreading over the lower parts of the fans. The cross-bedded arkoses within the thick arkosic sandstone sequences were deposited from better channelized flows, mostly on the lower parts of the fans.

The absence of fine grained deposits in the Ament Bay Formation suggests that no localized ponding of the floods occurred. The predominance of massive arkose in the thick arkosic sandstone sequences may be a consequence of steep gradients on the fans, which did not permit the formation of typical braided streams. In this respect the Ament Bay Formation differs from other accounts of alluvial fans, where cross-bedded braided stream deposits are commonly found in association with the coarser, more massive deposits of the head-fan region.

If steeper gradients did exist on these Archaean fans than on the younger fans described in the literature, this may be a consequence of different tectonic settings. Many present day fans are accumulations within intermontane areas, but the alluvial fans which deposited the Ament Bay Formation are visualized as bordering the main trough area of the greenstone belt, and trending from the hinterland directly into the sea without any alluvial plain. Models of greenstone belts which have been set up by some authors (e.g. Goodwin, 1969; Ridler, 1971) suggest an island arc type situation with volcanics and greywackes being deposited in the trough areas. However, it is not known what type of conditions existed in the area of the greenstone belt to the south of the Ament Bay Formation whilst the alluvial fans were forming, but the Ament Bay Formation is visualized as a transitional deposit between a rising terrain (probably to the north, see below) and a deeper trough area to the south.

The apparent absence of debris flow deposits on the fans suggests that little fine grained sediment was being produced in the source area, and thus physical weathering can be postulated to have been prevalent over chemical weathering.

(g) Paleoflow within the Ament Bay Formation

Only tentative suggestions of the dominant flow direction of the depositional currents are possible from analysis of the sedimentary structures within the Ament Bay Formation.

As was stated in Chapter 3, seven occurrences of cross-bedding have trough forms with sets of cross-strata dipping in opposite directions. This indicates that in these cases the depositional currents were probably flowing at high angles to the present outcrop sections. As the strikes of these cross-bedded units vary from about north-east to east, then by rotation of the beds to a horizontal position, the flow direction can be postulated as either roughly from the north or roughly from the south.

Seven other examples of cross-bedding indicate flow with a westerly directional component, and one other example indicates an easterly component of the flow.

Most imbrication within the granite-bearing conglomerates of the Ament Bay Formation indicate depositional flows with an easterly directional component (table 2).

The conflicting directional data from the few occurrences of cross-bedding and imbrication may be a result of predominantly oblique sections through the sedimentary structures, which only allow rough estimates of the current directions. Further study of the fabrics of the arkoses and conglomerates may help to resolve this difficulty. Also, varying directions of depositional currents might be expected if the sediments of the Ament Bay Formation were deposited on adjacent coalesing alluvial fans.

It is tentatively suggested that the sediments of the Ament Bay Formation were derived dominantly from a roughly northerly direction, because depositional currents with at least small southerly components of their flow directions could have produced the sedimentary structures within the arkoses and conglomerates.

3. Felsic tuffs

It has been stated (Chapter 3) that the unit of felsic tuff at Little Vermilion Lake probably rests conformably on the underlying Ament Bay Formation, and that felsic tuff also overlies granite-bearing conglomerates at Abram Lake. Because the sediments of the Ament Bay Formation have been interpreted as sub-aerial alluvial fan deposits it is likely that at least the basal parts of the felsic tuff facies are also sub-aerial deposits.

The field occurrences of the felsic tuffs exhibit few characteristics on which to base an interpretation. However, the massive nature of the tuffs may itself be an indication of the mechanism of deposition, as according to Ross and Smith (1961, p.19) "a principal characteristic of ash-flow tuffs is their common occurrence in thick units (tens of feet) of typically nonsorted or nonbedded materials. This characteristic is in direct contrast to ash-fall tuff deposits of comparable thickness, in which pronounced bedding is nearly always present".

The felsic tuffs at Little Vermilion Lake are unbedded except for a few chloritic bands (Chapter 3). However, they are fairly well sorted in as much as there are no clasts larger than 2 or 3 mm. This massive nature probably rules out ash-fall origins for the felsic tuffs at Little Vermilion Lake, but the finely bedded tuffs near Frog Rapids at Abram Lake (fig. 45) may be ash-fall deposits.

It is here tentatively suggested that the felsic tuffs at Little Vermilion Lake and possibly most tuffs at Abram Lake, are ash-flow deposits. If this is the case, then the tuffs are somewhat unusual as they contain few lithic fragments (Chapter 3). However, Ross and Smith (1961, p.19) state that ash-flows can consist almost entirely of ash and dust size material (<4 mm).

Microscopically, welding and distortion of pyroclastic materials is an outstanding characteristic of ash-flow materials (Ross and Smith, 1961, p.33). However, the matrix of the felsic tuffs of the Abram Group has probably been recrystallised and thus an original welded texture may have been destroyed, but the rock does retain a streaky texture (see Chapter 5 and fig. 75) which may have resulted either from flow or from flattening and stretching of glassy material. Also, the dark fine grained streaks in some thin sections of felsic tuff (Chapter 5 and fig. 75) may have originally been glass.

Another feature which is consistent with an ash-flow origin of the felsic tuffs at Little Vermilion Lake is the probable lateral equivalence of the acidic lava unit along the shore of Little Vermilion Lake, to the south of Ament Bay (map 1). This acidic lava may have formed in a more proximal position to the eruptive centre than did the felsic tuff.

4. Basic tuff facies

The basic tuffs are crudely bedded in parts, but show no obvious flow foliation and no pillow structures which would be indicative of lava flows. Also, the large angular clasts which are found in parts of the basic tuff facies (fig. 48) are good evidence of their pyroclastic origin. Because these clasts are larger at Daredevil Lake than at Little Vermilion Lake (Chapter 3) it may be that the eruptive centre was nearer to Daredevil Lake than it was to Little Vermilion Lake.

The less massive, and better-bedded nature of the basic tuffs than the felsic tuffs may indicate that they are mostly ash-fall deposits, which may in part have been resedimented by turbidity currents (producing occasional graded beds - Chapter 3).

From their stratigraphic position it is postulated that the basic tuffs are subaqueous deposits. Greywackes and argillites (interpreted as turbidites, below) occur both below and above the unit of basic tuffs (fig. 6), and a few black argillite beds are present within the basic tuff facies, which further suggests accumulation of the tuffs in a standing body of water.

5. Greywacke facies

The greywacke facies was divided in the field into 3 subfacies (Chapter 3), and these subfacies are discussed separately below.

(a) Thickly bedded and graded greywacke

The regular bedding within this subfacies, the frequent grading, the partial Bouma (1962) sequences of sedimentary structures and the interbedded argillites, indicate that the thickly bedded and graded greywackes were deposited by turbidity currents in water which was sufficiently deep to prevent the formation of shallow agitated-water structures.

The rather high sand:shale ratio of some greywacke sections (9:1, see Chapter 3), the presence of amalgamated greywacke beds, and the common massive or graded lower parts of the greywackes, may indicate that this subfacies was deposited in a fairly proximal position relative to the source, with frequent turbidity currents depositing greywacke beds and allowing little time for the accumulation of true argillites.

The presence of some slump-bedding within this subfacies (fig. 58) overlying apparently undisturbed greywacke and argillite beds, makes it likely that at least part of the subfacies was deposited on a slope steep enough to permit occasional slumping.

(b) Thinly bedded greywacke and argillite

This subfacies consists of thin siltstone and fine grained greywacke beds, some of which are graded, together with black argillite layers. These greywackes are also thought to be turbidites, because of the graded nature of some beds and their stratigraphic associations (Chapter 3). In this subfacies mud accumulation was more persistent, and influxes of turbidity currents deposited less greywacke than in the thickly bedded and graded greywacke subfacies.

The thinly bedded greywacke and argillite subfacies may represent more distal deposits than the thickly bedded and graded greywackes. However, the several horizons of this subfacies in which slump folding occurs, suggests that the thin greywackes were being deposited on slopes or on tectonically active parts of the basin floor. The close association of this subfacies with the argillite conglomerate subfacies further indicates this (see below).

(c) Argillite conglomerate

The intraformational nature of this subfacies, and its association at locality L at Little Vermilion Lake with slumped layers of the thinly bedded greywacke and argillite subfacies (fig. 63), suggests that the argillite conglomerates resulted from resedimentation of previously deposited greywacke and argillite beds.

Few argillite conglomerates were observed to be graded.

and it is not certain whether they were deposited from true turbidity currents, or whether they are more akin to slump deposits. In fact, both mechanisms may have deposited some argillite conglomerate.

Figure 63 shows fragments of siltstone and argillite layers, some of which include slump folds, overlain by irregular greywacke blocks in an argillite conglomerate. This outcrop may represent an initiation stage of slumping, which could produce fragments of sediment similar to those found in some of the coarse argillite conglomerates. The coarse chaotically bedded argillite conglomerates may be slump deposits.

However, the small size of the clasts (commonly <5 cm) in some outcrops of argillite conglomerate (fig. 62) may indicate reworking of the greywackes and argillites, and accumulation of the clasts before resedimentation, possibly by turbidity currents.

CHAPTER 5

PETROGRAPHY

1. Petrographic methods

The petrographic characteristics of the various facies throughout the Abram Group were examined with a view to determining the provenance of the sediments. In this chapter the thin sections of the facies rocks are described, and the provenance of the sediment is discussed in Chapter 6. The descriptions below include fabric and textural characteristics, compositional point counts, and percentages of the various types of quartz, feldspar and rock fragments.

Thirty-nine thin sections were point-counted, of which 14 were arkosic sandstones, 10 were granitic pebbles from the granite-bearing conglomerates, 1 was a quartz porphyry pebble from a conglomerate, 3 were felsic tuffs, 10 were greywackes, and 1 was from the quartz porphyry body at Vermilion Lake. Four hundred points were counted on each thin section at 1 mm intervals. During some of these point-counts the various types of quartz and rock fragments were also recorded. A few thin sections could not be used for this latter purpose as they were so fine grained that estimates of the percentages of quartz types would probably be biased in favour of monocrystalline quartz grains, and in some thin sections the matrix was too recrystallised to determine the rock fragments. However, the total percentage of quartz grains in these altered thin sections has probably not been changed by recrystallisation.

The types of quartz recognized were: (1) polycrystalline grains, (2) monocrystalline grains with undulatory extinctions, and (3) monocrystalline grains with sharp extinctions. The definition of a "polycrystalline grain" used here (Blatt, 1967) is a quartz grain composed of 2 or more quartz crystal units, each of which may be either strained or unstrained, with less than 90 percent of the polycrystalline grain area being occupied by one quartz crystal.

The amounts of K-feldspar and plagioclase in the various facies were determined by treating several thin sections with sodium cobaltinitrite and amaranth solutions, using the method of Boone and Wheeler (1968). K-feldspar takes up a yellow stain with the sodium cobaltinitrite, and plagioclase stains pink with amaranth solution (Laniz et al., 1964). Of the point-counted thin sections, the following number were stained: 2 arkosic sandstones, 2 granitic pebbles, 1 felsic tuff, and 2 greywackes.

A universal stage was used to determine the compositions of plagioclase grains within some of the thin sections of arkosic sandstone, granitic pebbles, felsic tuff, and greywacke.

2. Arkosic sandstone facies

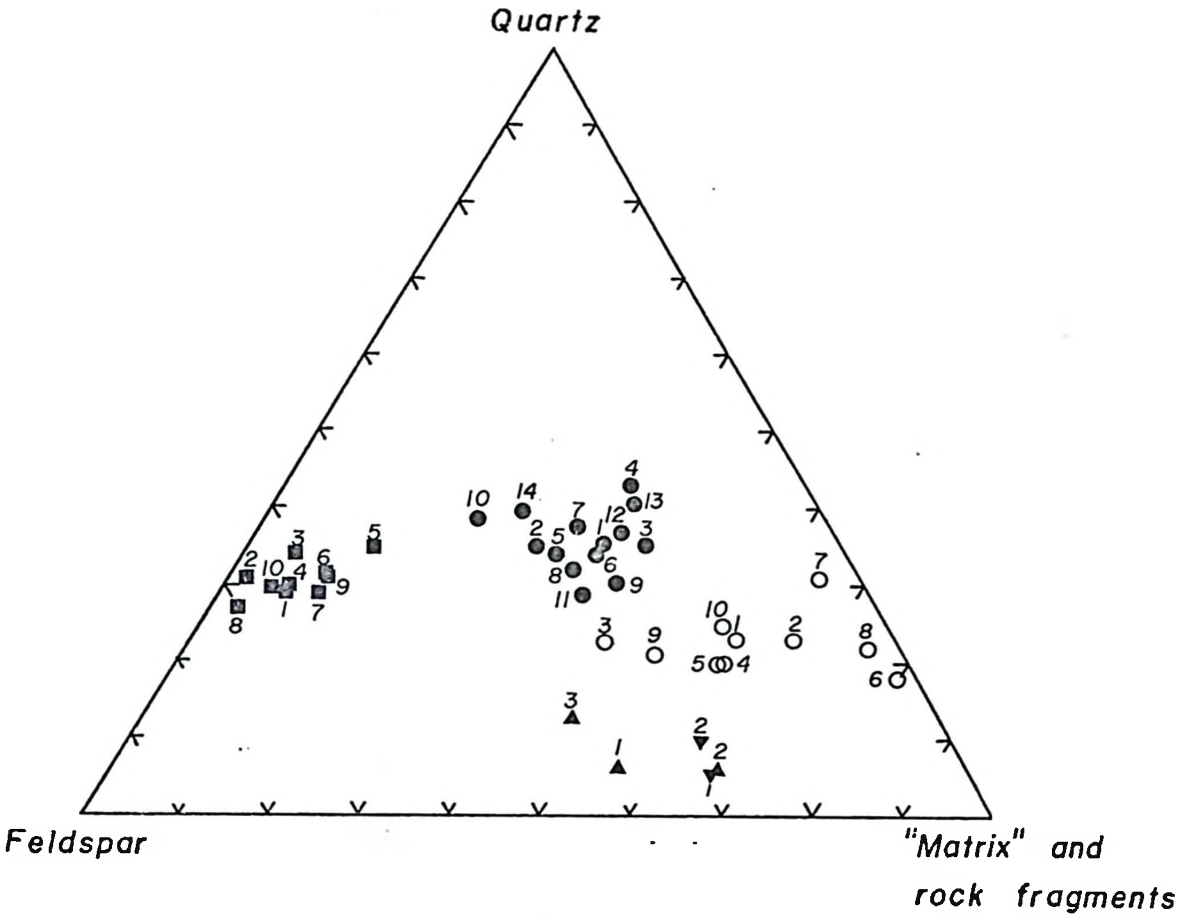
The arkosic sandstones are described first, as they not only occur as massive, cross-bedded or parallel laminated units in the field, but also form the "matrix" between clasts in the granite-bearing conglomerate facies.

In thin section, most arkosic sandstones contain between 30 and 40 percent quartz grains larger than about 0.03 mm, and between 20 and 35 percent feldspar grains. The percentage of rock fragments varies from about 2 to 28, and that of the matrix, containing sericite, quartz, chlorite, biotite, carbonate, and epidote, from 11 to 37 (fig. 65 and Appendix 2).

(a) Texture

Most of the arkosic sandstones are grain supported (fig. 66) and where large interstitial areas are filled with irregular masses of fine grained secondary minerals, it is likely that this represents a decomposed rock fragment. The maximum grain size of the quartz and feldspar fragments is about 2 mm, and most grains fall into the range 0.5 mm to 1 mm, coarse sand. Most rock fragments vary in size from coarse sand to about 5 mm in diameter in thin sections of arkosic sandstone, but in outcrops of the granite-bearing conglomerate facies all gradations exist between these sand-size grains and the pebbles and cobbles of similar compositions.

From visual estimates it is evident that most quartz grains are subangular to subrounded in shape, with only a few rounded grains (fig. 67). The edges of quartz grains are commonly sharp, but some grains have a slightly diffuse outline,



● Arkosic sandstones
○ Greywackes
■ Granodiorite clasts
▲ Felsic tuffs
▼ Quartz porphyries (1 - clast, 2 - from Vermilion Lake)

Figure 65. Modal compositions of the major lithologies of the Abram Group. Numbers refer to sample numbers in Appendix 2.

probably due to slight resorption at the rim, and no quartz grains have secondary quartz overgrowths. No alteration occurs within the quartz grains, and apart from irregular cracks, the grains are clear and free of inclusions.

The feldspar grains are much more altered than those of quartz, and they commonly have diffuse outlines. However, feldspar grains which have been only slightly altered are broken and angular in outline, and a few grains are cracked into 2 or more fragments which are separated from each other by only a thin band of matrix material.

In some thin sections, rock fragments are less clearly defined than the feldspar grains, and in places it is difficult to distinguish rock fragments from matrix. However, in most samples of arkosic sandstone, the rock fragments can be recognised and identified (fig. 66, and section below on composition).

In most thin sections of arkosic sandstone no directional fabric is present, the grains have random orientations, and most grains are recognisable (fig. 66). However, a tectonic fabric is exhibited by some thin sections, where the matrix material has a slightly schistose texture, and the clear quartz grains have their long axes aligned roughly parallel to this schistosity. In these thin sections rock fragments and feldspar grains are commonly difficult to distinguish from the matrix.

(b) Composition

(i) Quartz

Polycrystalline quartz grains constitute from 83 to 99 percent of the quartz populations of arkosic sandstone thin sections (table 4). Monocrystalline quartz with undulatory extinction makes up from 0.5 to 16 percent, and monocrystalline quartz with sharp extinction from 0 to 7 percent. The quartz crystal units of polycrystalline grains are generally not equigranular and range from about 0.1 mm to over 1 mm in diameter. Most boundaries between quartz crystal units within polycrystalline grains are sharp and commonly curved, with roughly polygonal outlines (fig. 67); and only a few quartz crystal units have straight boundaries.

(ii) Feldspar

The arkosic sandstones contain both untwinned feldspar, and plagioclase exhibiting lamellar twinning (fig. 67). However, in the 2 stained thin sections of arkosic sandstone both feldspar types took up a pink stain when treated with amaranth solution, and are hence plagioclase in composition. Small patches of yellow stain, indicating K-feldspar, were produced on one thin section, but this did not exceed 5 percent of the total thin section area, and covered interstitial matrix areas, not feldspar grains.

Composition determinations of 27 twinned plagioclase

Sample No.	Polycrystalline Quartz	Monocrystalline Quartz	
		Sharp ext.	Undulatory ext.
Ar 1	94	3	3
Ar 2	89	1.5	9.5
Ar 3	84	7	9
Ar 4	99	0.5	0.5
Ar 5	99	0	1
Ar 6	96	1	3
Ar 7	99	0	1
Ar 8	98	0	2
Ar 9	83	1	16
Ar 10	98	0	2
Ar 11	96	0	4
Ar 12	96	0	4
Ar 13	90	6	4
Ar 14	87	5	8
Gw 1	88	2	10
Gw 2	79	7	14
Gw 3	96	0	4
Gw 4	82	2	16
Gw 5	77	3	20
Gw 10	88	3	9

Table 4. Quartz type percentages within the quartz populations of arkosic sandstones and greywackes. The percentage quartz in each sample is shown in Appendix 2, and the differences between the arkosic sandstones and greywackes are discussed in Chapter 6.

grains from 12 thin sections of arkosic sandstone fell within the range An_5 to An_{14}. Thus the plagioclase of the arkosic sandstones are albite and oligoclase.

Most feldspar grains are dusky due to the presence of many extremely fine inclusions, and in some thin sections the grains are partially or almost wholly altered to a sericitic mass. It is likely that some sericitic matrix is due to the decomposition of feldspar grains.

(iii) Rock fragments

Four types of rock fragments were recognised within the arkosic sandstone facies, and are described in decreasing order of abundance below.

1. Dark chloritic fragments which contain randomly oriented feldspar laths up to about 0.2 mm in length (fig. 66). These fragments were probably originally intermediate to basic lavas, and are the most abundant rock types within the arkosic sandstones. They constitute up to 75 percent of the rock fragment populations in thin sections ("basic volcanic" in table 5), which represents up to 18 percent of the total thin section area.

2. Felsic rock fragments are next in abundance to the intermediate and basic lava fragments, and compose from about 15 to 50 percent of the rock fragment populations in individual thin sections ("acidic volcanic" in table 5), which represents up to 11 percent of the total thin section area. These fragments

Sample No.	Basic Volcanic	Acidic Volcanic	"Chert"	"Granitic"
Ar 3	75	16	8	1
Ar 4	65	33	2	0
Ar 6	56	42	2	0
Ar 9	44	49	7	0
Ar 13	63	34	2	1
Ar 14	69	28	2	1
Gw 1	24	74.5*	1.5	0
Gw 2	13	85*	0	2
Gw 3	6	91*	0	3
Gw 4	8	87*	0	5
Gw 5	35	63*	2	0

Table 5. Rock type percentages within the rock fragment populations of arkosic sandstones and greywackes. The percentage of rock fragments in each sample is shown in Appendix 2. The greywackes are from the Little Vermilion Formation, except for Gw 3, which is from the Daredevil Formation.

* - A few percents are probably sedimentary fragments (see text).

generally comprise very fine grained quartz and feldspar, with a few feldspar phenocrysts up to about 0.2 mm in length (fig. 68), and were probably originally acidic or dacitic lavas. However, a few of the slightly coarser felsic fragments could be felsic tuffs.

3. "Chert" fragments consisting of small interlocking, equigranular quartz grains, together with a small amount of sericite or carbonate, compose up to 8 percent of the rock fragment populations in thin section (up to 2 percent of a thin section area) (table 5). In these fragments the individual quartz grains are commonly about 0.05 mm in diameter, and have polygonal to sutured outlines (fig. 66).

These "chert" fragments may have originated from recrystallised true cherts, or from acidic lavas.

4. A few rock fragments (about 2 mm in size) in some thin sections comprise interlocking feldspar and quartz grains, with individual crystals about 0.5 mm in diameter. Because these fragments contain as few as 2 or 3 separate grains, their origin is uncertain, but it is likely that the fragments are granodiorite chips, as they resemble parts of larger granodiorite clasts (see section of this chapter on conglomerates). These "granitic" fragments only comprise up to 1 percent of the rock fragment populations of arkosic sandstones (table 5).

(iv) Matrix

Most matrix material consists of very small sericite laths, commonly about 0.02 mm in length, and was probably largely derived from the decomposition of feldspar grains. Some large patches of sericite contain a few fragments of feldspar, which have a common extinction and were probably originally a single grain. A thin band of matrix material commonly separates individual sand grains, but some grains are in contact with each other.

Chlorite is also a common metamorphic mineral within the arkosic sandstones, but it is much less abundant that sericite, and is commonly associated with the decomposition of rather basic rock fragments. Epidote also forms up to about 1.5 percent of some thin sections of arkosic sandstones, and occurs as irregular masses, commonly as an alteration of rock fragments.

A few of the more schistose arkosic sandstones contain secondary biotite, which is aligned roughly parallel to the foliation. Authigenic pyrite, generally less than 0.5 mm in size, is commonly found in thin sections of arkosic sandstone, but it forms less than 1 percent of the total thin section area.

3. Granite-bearing conglomerate facies

Clast types and their percentages within the granite-bearing conglomerate facies were outlined in Chapter 3 (fig. 7), but the petrographic characteristics of these clasts are set out below. Arkosic sandstone forms the "matrix" of the conglomerates, and has the same composition as typical massive arkosic sandstone.

Samples Ar6 and 7 in figure 65 and Appendix 2 are from the matrices of granite-bearing conglomerates. The 3 major clast types within the granite-bearing conglomerates are greenstone, granitic, and felsic fragments, with lesser amounts of "chert", "vein quartz", foliated granitic clasts, quartz and feldspar porphyry, and felsic tuff (fig. 7).

(a) Greenstone clasts

As defined in Chapter 3, the greenstone clasts (commonly 5 to 15 cm in length) are of very fine grain size (generally much less than 0.1 mm), and are composed predominantly of chlorite.

Nine thin sections containing 21 greenstone clasts were examined, and 3 types of fragment were distinguished.

(i) A very fine grained chloritic and sericitic mass, which contains laths of plagioclase up to about 0.5 mm in length. Some feldspar is extensively sericitized (fig. 69).

(ii) A chloritic groundmass surrounding laths of pale green amphibole and a few plagioclase laths, which are also about 0.5 mm in length.

(iii) A mass of chlorite and biotite together with a small amount of quartz, which surrounds veinlets and irregular patches of quartz and carbonate. Some irregular grains of epidote are also present.

These three types of greenstone clast resemble the textures and compositions of 3 thin sections cut from massive greenstones in the study area. It is likely that the greenstone clasts within the granite-bearing conglomerate facies were derived from intermediate to basic lavas.

(b) Granitic clasts

Twenty-four thin sections were cut from clasts designated as "granitic" in the field, and of these, 10 were point-counted for quartz, feldspar and accessory minerals (fig. 65 and Appendix 2). The accessory minerals include chlorite, chloritized biotite, sericite, calcite, epidote, and a very small amount of amphibole (Appendix 2). The 24 thin sections were further separated into 2 groups. The first comprised 13 thin sections of clasts which have undergone little textural alteration and from which the 10 sections were point-counted. The second group comprised 11 thin sections of clasts which have suffered much recrystallization of quartz.

(i) Relatively fresh granitic clasts

Texture - The granitic clasts consist of granular "mosaics" of euhedral and subhedral, clouded plagioclase crystals, together with anhedral quartz grains (fig. 70). Microcline occurs as a later stage feldspar, and surrounds the plagioclase and quartz grains. Most plagioclase and quartz grains are of roughly equal size, and are commonly from 1 to 3 mm in diameter. Quartz constitutes from 28 to 35 percent of the granitic clasts, whilst feldspar ranges from 51 to 70 percent, and the accessory minerals constitute up to 14 percent (fig. 65).

Feldspars - The types of feldspar in the point-counted sections are tabulated in table 6, where 3 divisions are recognised - (a) Twinned plagioclase, which stained pink in the 2 stained thin sections; (b) Untwinned plagioclase, which also stained pink; and (c) Late stage microcline (fig. 71), most of which stained yellow, indicating K-feldspar, but some of which stained pink, indicating that it is slightly calcic in patches. This may indicate that the microcline has been partially albitized.

The twinned and untwinned plagioclase together comprise 70 to about 88 percent of the total feldspar populations, the remainder being microcline. The plagioclase crystals are commonly very clouded, some are sericitized, some are zoned, and in some thin sections most plagioclase has narrow clear veins of untwinned, unstained albite. (Pure $NaAlSiO_3$ should not stain with either amaranth or sodium cobaltinitrite solutions).

Determinations of 6 twinned plagioclase compositions from 5 thin sections gave a range of values from An_8 to An_{15}, indicating that the plagioclase is albite in some thin sections and oligoclase in others. Most twinned plagioclase crystals exhibit albite twinning (fig. 71), but a few have both albite and pericline twins, and a very few show simple Carlsbad twinning. The interstitial microcline is clear and in

Sample No.	Twinned Plagioclase	Untwinned Plagioclase	Microcline
Gr 1	43	38.5	18.5
Gr 2	48	34	18
Gr 3	52.5	36	11.5
Gr 4	44	26	30
Gr 6	55.5	24.5	20
Gr 7	31	56	13
Gr 8	50	20	30
Gr 10	49	31	20

Table 6. Feldspar type percentages within the feldspar populations of granodiorite pebbles from granite-bearing conglomerates. The percentage of feldspar in each sample is shown in Appendix 2.

some places exhibits characteristic tartan twinning.

Quartz - Most individual quartz grains range in size from about 0.5 mm to 2 mm, but a few are as large as 3 mm. Virtually all the quartz in the granitic clasts has undulatory extinction, and the contacts between individual quartz grains are commonly sharp and straight or gently curved.

Classification - The equigranular granitic texture of these clasts, the predominance of grains from 1 to 2 mm in size, the fact that about 30 percent of the rock is quartz, and the high ratio of plagioclase to K-feldspar (table 6), suggests that the clasts can be classified as medium grained granodiorites (Williams, Turner and Gilbert, 1953).

(ii) Granitic clasts with recrystallized quartz

The second group of thin sections of granitic clasts have similar compositions to the granodiorite clasts described above, but have a different texture. Plagioclase and microcline are still recognisable, but the plagioclase is not euhedral, and the microcline is now surrounded by granular quartz with most quartz grains from 0.2 to 0.5 mm in diameter. The quartz in these thin sections has been recrystallized and in some places it fills narrow veins.

In some of the hand specimens of such granitic clasts, the clast boundaries are diffuse, and the surrounding conglomerate material is in places schistose. This suggests that localized recrystallization occurred during the metamorphic event subsequent to the deposition of the sediments. However, it is likely that originally the clasts were also granodiorite.

(c) Felsic clasts

Point counts in the field on outcrops of granite-bearing conglomerate gave estimates of the percentage of felsic fragments in the conglomerates ranging from 5 to 40 percent (fig. 7), but generally being less than 20 percent.

Five thin sections were cut from clasts which had been designated as "felsic" in the field. The groundmasses of these clasts either consist of very fine grained quartz, sericite, plagioclase, and simply twinned feldspar less than 0.1 mm in diameter, or are cryptocrystalline felsic masses. Feldspar phenocrysts up to 2 mm in length constitute up to 10 percent of the rock. Most of these phenocrysts are plagioclase with albite twinning, or are untwinned, and only a few have Carlsbad twinning. A very few quartz phenocrysts up to 0.5 mm in diameter are also present.

It is likely that these felsic clasts are fragments from dacitic or rhyodacitic lavas.

(d) "Chert" clasts

The "chert" clasts found within granite-bearing conglomerates are composed of very fine interlocking quartz grains (commonly about 0.05 mm in diameter) and are identical to those described in the previous section on arkosic sandstones. In outcrop the "chert" fragments are difficult to distinguish from fine grained felsic fragments, and are thus classed with the felsic fragments in fig. 7. However, they apparently constitute less than 5 percent of any conglomerate outcrop.

(e) "Vein quartz"

At some outcrops of granite-bearing conglomerate white granular fragments of "vein quartz" comprise up to 5 percent of the conglomerate. In thin section, the lithology consists of equant grains of quartz up to 0.5 mm in size, mostly with straight polygonal boundaries.

Donaldson and Jackson (1965) described similar lithologies from North Spirit Lake, and suggested that they were in fact fragments from recrystallised chert, and not typical vein quartz. The same explanation may be correct for the "vein quartz" of the Ament Bay Formation.

(f) Foliated granitic clasts

Foliated granitic clasts are uncommon in the granite-bearing conglomerates, and most outcrops are devoid of such clasts. However, in a few outcrops foliated granitic fragments comprise up to 3 percent of the conglomerate (fig. 7). These clasts consist of compositionally banded medium and coarsely crystalline quartzo-feldspathic rocks.

Only one sample was found which was readily removable from the conglomerate, and from this a thin section was cut. The clast has the same composition as the granodiorite clasts described above, but it contains rods of quartz surrounded by feldspar. These rods are about 2 mm in diameter and are up to 5 cm in length. This texture was obviously not produced after the conglomerate was deposited, as the quartz rods are truncated at the edge of the clast. A thin section cut perpendicular to the quartz rods revealed a "metamorphic" texture of the quartz. The elongate interlocking quartz grains have subparallel optical orientations and undulatory extinctions.

This texture probably resulted from the crystallization of the quartz under conditions of considerable stress, perhaps at the margins of granitic bodies.

(g) Quartz and feldspar porphyry clasts

Apart from the basal conglomerate of the Abram Group, very few outcrops of granite-bearing conglomerate contain clasts of quartz and feldspar porphyry, and when present, the clasts constitute less than 3 percent of the conglomerate.

Four thin sections of quartz and feldspar porphyry clasts revealed phenocrysts of plagioclase and quartz from about 0.5 mm up to 5 mm in diameter, set in an aphanitic felsic groundmass (fig. 72). From a point-count of one thin section of porphyry (Q.P.1 in fig. 65 and Appendix 2), it was found that quartz constitutes 5 percent of the porphyry, whilst feldspar makes up 28 percent. Most quartz phenocrysts are monocrystalline and have undulatory extinctions. The majority of the feldspar is plagioclase, and most contains irregular albite and pericline twinning. The twin lamellae commonly wedge out along their length (fig. 72). A few feldspars exhibit Carlsbad twinning, and some are untwinned.

(h) Felsic tuff clasts

At a few outcrops of granite-bearing conglomerate, felsic tuff fragments were distinguished on the grounds of abundant coarse angular feldspar fragments in the clasts, and the similarity of the clasts to outcrops of the felsic tuff facies. These clasts constitute less than 2 percent of the conglomerate.

A thin section of one felsic tuff clast revealed broken and bent plagioclase grains up to 2 mm in length, set in an aphanitic groundmass (fig. 73). The broken nature of plagioclase distinguishes this rock type from feldspar porphyries examined in the area.

4. Felsic tuff facies

(a) Felsic tuff at Little Vermilion Lake

In outcrop, the felsic tuffs at Little Vermilion Lake are massive and structureless, and have a granular texture.

In the 3 thin sections examined the tuffs contain from 6 to 13 percent quartz phenocrysts, and from 27 to 40 percent feldspar phenocrysts, set in a fine grained felsic groundmass (fig. 65 and Appendix 2).

The feldspar grains are up to 4 mm in diameter, most are angular in outline, some are euhedral, and some are broken with the two parts separated by felsic matrix (fig. 74). The compositions of 15 plagioclase grains from the 3 thin sections were determined, and all were found to be albite (between An_7 and An_{10}). Most feldspar grains are partially sericitized, and they commonly exhibit pericline and albite twinning (fig. 75).

The few quartz grains present in the tuffs range up to 5 mm in size, and are rounded to angular in outline. In the thin section with the least altered groundmass the quartz phenocrysts are monocrystalline with undulatory extinction, which probably represents the original quartz.

The groundmass of the felsic tuffs either consists of patches of felsic minerals separated by streaky accumulations of biotite, or is foliated and highly sericitized. These foliations are bent around the unoriented feldspar grains, and in some places irregular streaks of very fine grained dark material are present, which may originally have been volcanic glass (fig. 75).

The massive nature of the felsic tuff outcrops, the angular and broken character of some feldspar grains, and the foliated, and possibly originally glassy groundmass, suggest that these rocks are indeed crystal tuffs.

(b) Felsic tuff at Abram Lake

The tuffs at Abram Lake are generally finer grained than those at Little Vermilion Lake, and contain few phenocrysts of feldspar and quartz. The tuffs have aphanitic felsic matrices, which are commonly slightly schistose.

The tuffs which crop out immediately above the granite-bearing conglomerates at Island B in Abram Lake (fig. 2), contain an aphanitic felsic groundmass surrounding many patches of extremely fine grained sericite up to 1 mm in diameter. These sericite patches contain a few unaltered feldspar fragments, and probably represent almost completely altered feldspar grains. Apart from being more altered and finer grained, these tuffs resemble those at Little Vermilion Lake.

Eight other thin sections cut from tuffs at Island A in Abram Lake, and around the western end of this central part of the lake (fig. 2) are similar in texture and grain size to those from Island B. However, some contain a few polycrystalline quartz grains up to about 1 mm in diameter, and others contain patches of fine grained green chlorite and amphibole.

The "tuffs" around Frog Rapids (fig. 2) are extremely fine grained, and contain no phenocrysts. Some are banded, with felsic layers a few mm to a few cm in width, and other darker layers contain chlorite and amphibole.

5. Basic tuff facies

The basic tuff facies crops out at Little Vermilion Lake and also at Daredevil Lake, and can be divided into 2 parts, the lower 130 m or so being more basic in composition than the upper 140 m.

(a) The lower basic tuff

Four thin sections were cut from samples of the more northerly, stratigraphically lower 130 m of the basic tuff facies, and revealed amphibole phenocrysts with ragged outlines within a sericitic and slightly felsic groundmass.

Two types of amphibole are present, a green variety, and a colourless one which is partially replacing the green one, and both are partially replaced by biotite (fig. 76). These amphiboles are generally less than 1 mm in length.

No large feldspar phenocrysts occur in this basic tuff, but a very few small (less than 0.5 mm) feldspar grains are recognisable. Quartz constitutes a small amount of the groundmass and occurs in short veinlets.

(b) The upper basic tuff

There is no sharp, or even clear break between the two parts of the basic tuff section in the field, but 6 thin sections cut from this upper half have a much more felsic composition than those from the lower half. These thin sections are more similar to the felsic tuff facies than to those basic tuffs described above, but they do contain some amphibole and biotite grains and no quartz, thus differing slightly from the felsic tuffs.

These upper basic tuffs contain many broken and angular plagioclase phenocrysts up to 5 mm in diameter, set in an aphanitic felsic groundmass (fig. 77). In some thin sections these feldspar grains have a very fresh appearance, but in others they are very altered and occur as black masses with granular outlines.

At a few localities along the southern shore of Daredevil Lake an extremely fine grained, hard, white weathering rock crops out between tuff beds, and in thin section this consists of dark faintly polarizing material cut by small quartz veins. This may originally have been volcanic glass, and the rock could be classified as a Porcellanite. Similar glassy material is found in small patches in the thin sections from both parts of the basic tuff facies, but it is more common in the basic lower 130 m.

6. Greywacke facies

In Chapter 3 the greywacke facies was divided into 3 subfacies on the basis of megascopic textures and structures, but for the purpose of petrographic description only the coarsest sandstones were used. The argillites, and greywackes of silt-stone grade are too fine grained to be point-counted accurately. In the section below the compositions and textures of the coarsest greywackes are described, and a brief description is also given of the argillite conglomerates.

Ten point-counted thin sections contained between 18 and 31 percent quartz, 2 to 32 percent feldspar and 0 to 25 percent rock fragments (fig. 65 and Appendix 2). From 5 of the 10 thin sections, the types of quartz, feldspar, and rock fragments were determined (tables 4 and 5). The other thin sections were not used for this purpose as their matrices are highly recrystallised.

(a) Texture

In the coarsest greywackes most quartz and feldspar grains are less than 1 mm in size, but a few are as large as 1.5 mm. Rock fragments are commonly a little longer and flatter than the mineral constituents, but rarely exceed 2 mm in length.

In the least altered greywackes many grains are in contact with each other, but even these rocks contain greater than 30 percent quartz and sericite matrix (Appendix 2).

The quartz grains are angular in outline, and are commonly roughly equant in shape. The feldspar grains are also angular and broken, and are commonly highly altered. The rock fragments are the least well defined of the clastic components, and commonly have diffuse boundaries with the matrix material.

The greywackes commonly exhibit a slight foliation due to the rough alignment of clasts and matrix material parallel to the bedding (fig. 78).

(b) Composition

(i) Quartz

Seven of the 10 point-counted thin sections contained between 20 and 25 percent quartz grains (fig. 65), and 6 thin sections contained from 77 to 96 percent polycrystalline quartz, o to 7 percent monocrystalline quartz with sharp extinction, and 4 to 20 percent monocrystalline quartz with undulatory extinction (table 4). The relevance of these values to provenance of the greywackes is discussed in the following chapter.

(ii) Feldspar

Thin sections Gr 4 and 5 were stained for both plagioclase and K-feldspar, and on both, considerable areas were stained pink, indicating plagioclase. No yellow stain resulted, indicating that no K-feldspar is present in these greywackes.

Ten plagioclase compositions from 5 thin sections were determined, and all the values fell between An_4 and An_{13}. Thus, the feldspars are albite or oligoclase.

(iii) Rock fragments and matrix

The rock fragments distinguishable in the greywacke thin sections are as follows.

1. Chloritic fragments similar to those in arkosic sandstones, but with fewer included plagioclase laths. These are probably basic volcanic fragments and vary from 6 to 35 percent of the

rock fragment populations (table 5).

2. Very fine grained felsic fragments. Most of these are crystalline masses of quartz and feldspar, having no preferred orientations of the grains, and are similar to the "acidic volcanic" fragments in the arkosic sandstones. However, a few felsic fragments contain much sericite, and are slightly schistose. These fragments are similar to thin sections of greywacke of siltstone grade, and may be sedimentary fragments. As only a few such "sedimentary fragments" were recognised, and due to their uncertain origin, all fine grained felsic fragments were classified as "acidic volcanic" fragments, and constitute from 63 to 91 percent of the rock fragment populations in the greywackes (table 5).

3. A few rock fragments are composed of very fine grained quartzose "chert", similar to that found in the arkosic sandstones, and comprise 0 to 2 percent of the rock fragments.

4. From 0 to 5 percent "granitic" rock fragments (up to about 2 mm in size) are present in the greywackes, each fragment consisting of a few interlocking grains of quartz and plagioclase up to 0.5 mm in size.

Many rock fragments have diffuse borders with the sericitic matrix, and in the fine grained, more altered greywackes it is likely that rock fragments have been broken down to form much of the matrix.

(c) Argillite conglomerate

Six thin sections were cut from some of the finer grained argillite conglomerates, which consist of various sizes of angular sedimentary rock fragments surrounded by a medium to coarse matrix of greywacke. This matrix is commonly coarser than the grain sizes within the rock fragments, and individual quartz and feldspar grains range up to 2 mm in size.

Pettijohn (1935, p.1902), briefly described these conglomerates and stated that the fragments appear to be sedimentary in outcrop, but under a microscope they appear to be fine-grained acid lavas with small quartz and plagioclase phenocrysts. However, in the present study, all of the fine grained fragments studied in thin section were found to be identical to the thin sections of fine grained, bedded greywackes and argillites (fig. 79), and thus the fine grained fragments are thought to be of sedimentary origin.

None of the fragments examined microscopically from the argillite conglomerate had other than sedimentary textures.

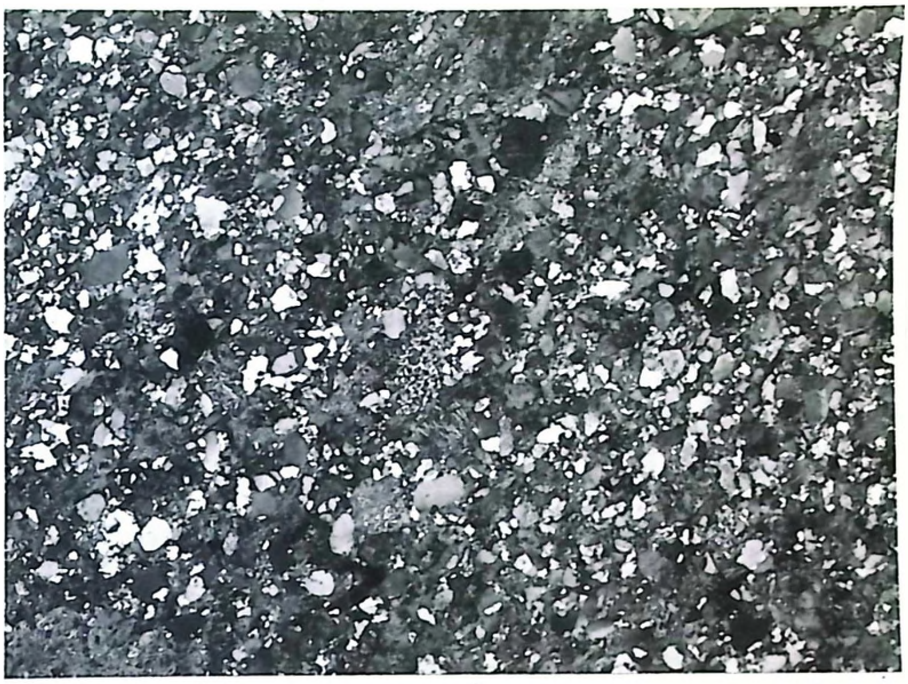


Figure 66. Projection print of a thin section of arkosic sandstone (Ar6, Appendix 2). Note polycrystalline quartz (clear), feldspar (clouded), "chert" fragment (centre), basic volcanic fragments (near centre), and the unoriented grains. Field of view is 1.4 cm wide.

Figure 67. Photomicrograph of a thin section of coarse arkosic sandstone (Ar14, Appendix 2). Note angular polycrystalline quartz, twinned plagioclase, untwinned feldspar (top left, bottom right), and fine grained sericitic matrix. Field of view is about 2 mm wide.

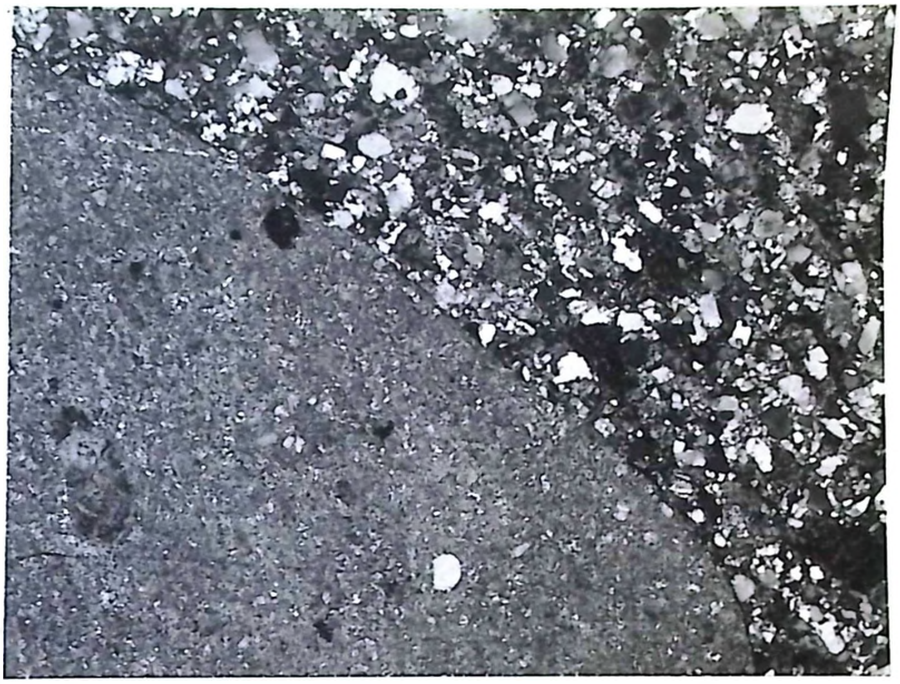


Figure 68. Projection print of a thin section of arkosic sandstone surrounding an acidic volcanic clast which contains quartz and feldspar (clouded) phenocrysts. Field of view is 1.4 cm wide.

Figure 69. Projection print of a thin section of arkosic sandstone (Ar13, Appendix 2) surrounding intermediate volcanic fragment. The feldspar laths have been sericitized. Field of view is 1.4 cm wide.

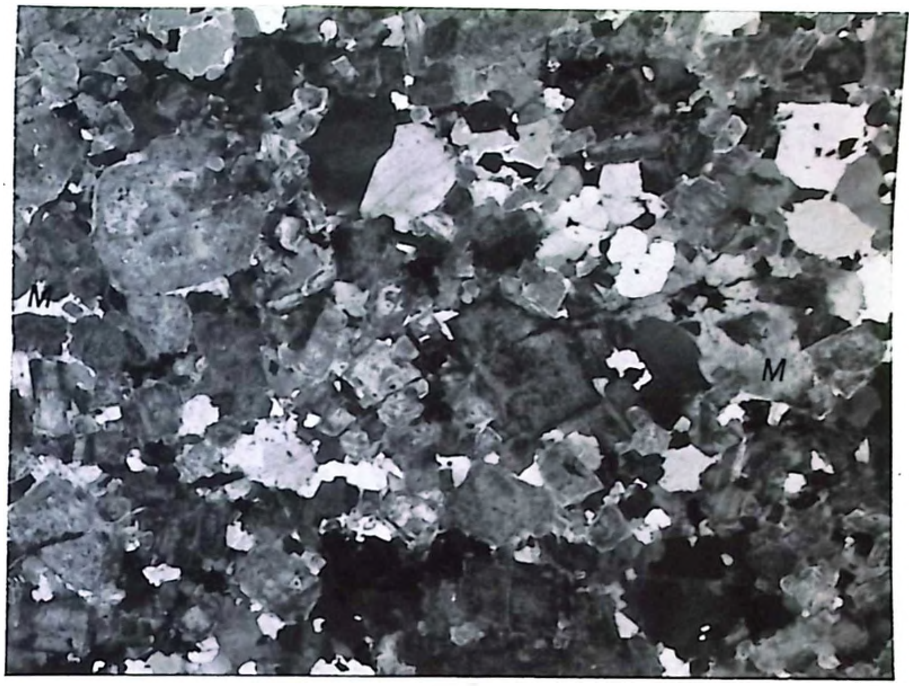


Figure 70. Projection print of a thin section of a granodiorite clast (Gr1, Appendix 2). Note subhedral clouded plagioclase crystals, quartz, and interstitial microcline (M). Lamellar twinning in plagioclase is largely masked by the alteration. Field of view is 1.4 cm wide.

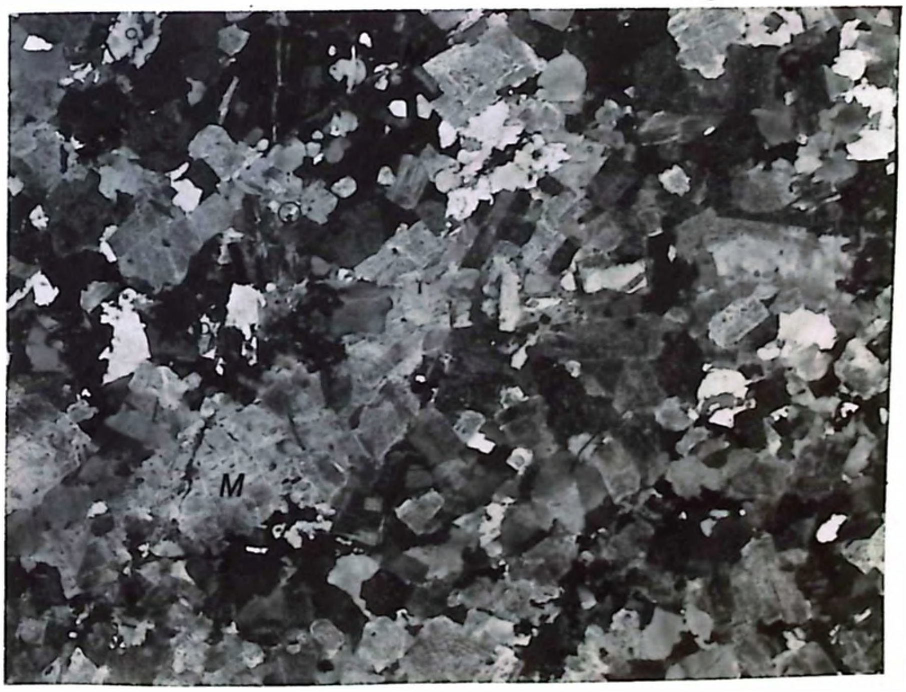


Figure 71. Projection print of a thin section of a granodiorite clast (Gr4, Appendix 2). Note interstitial microcline (M), the albite and pericline twinning in plagioclase (clouded), and quartz. Field of view is 1.4 cm wide.

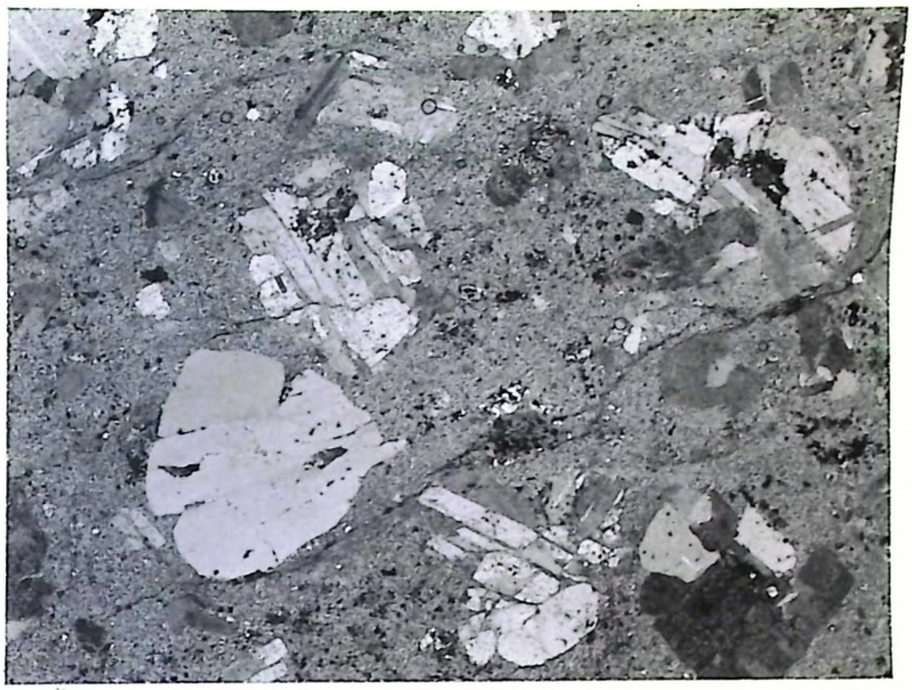


Figure 72. Projection print of a thin section of quartz porphyry from Vermilion Lake (Q.P.2, Appendix 2). Note monocrystalline quartz with embayed outlines and included felsic blebs, and plagioclase phenocrysts set in an aphanitic matrix. Field of view is 1.4 cm wide.

Figure 73. Projection print of a thin section of granite-bearing conglomerate with arkose matrix surrounding dark basic rock fragment, felsic tuff fragment, and acidic volcanic fragment (upper right). Note the cracked plagioclase grains in the tuff. Field of view is 1.4 cm wide.

Figure 74. Projection print of a thin section from felsic tuff facies (F.T.3, Appendix 2). Note cracked nature of feldspar grains showing Carlsbad and albite twinning, patches of quartz, and recrystallized matirx. Field of view is 1.4 cm wide.

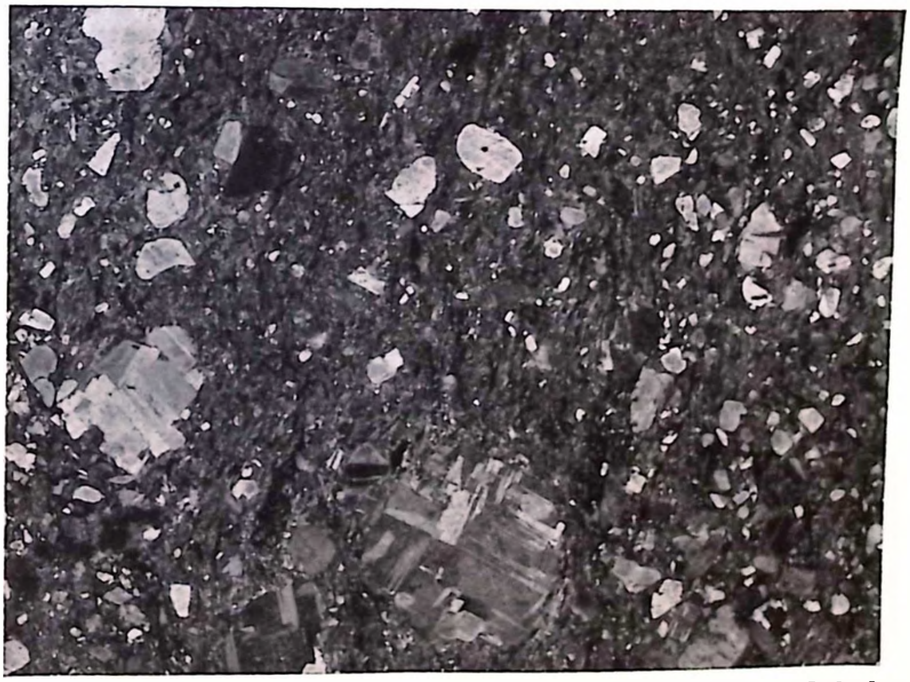


Figure 75. Projection print of a thin section from felsic tuff facies (F.T.2, Appendix 2). Note rounded outlines of some plagioclase grains, and foliated texture. Some of the dark streaks may have originally been glass. Field of view is 1.4 cm wide.

Figure 76. Photomicrograph of a thin section from the lower part of the basic tuff facies. Note amphiboles (A) being replaced by biotite (B), within a chloritic matrix. Field of view is about 2 mm wide.

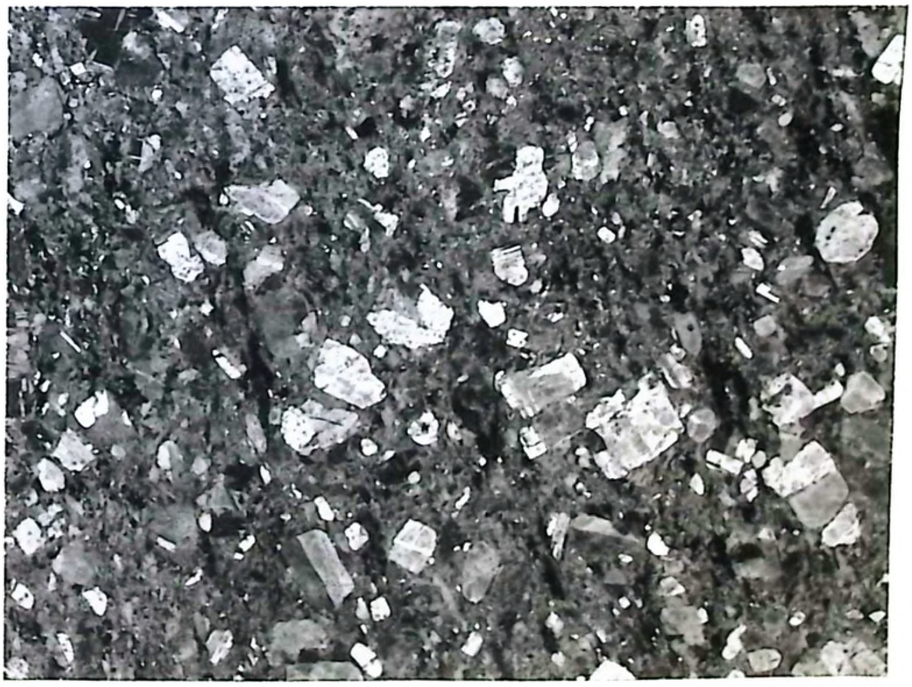


Figure 77. Projection print of a thin section from the upper part of the basic tuff facies. Note the cracked feldspar grains within a very fine grained sericitic matrix, and the absence of quartz grains. Field of view is 1.4 cm wide.

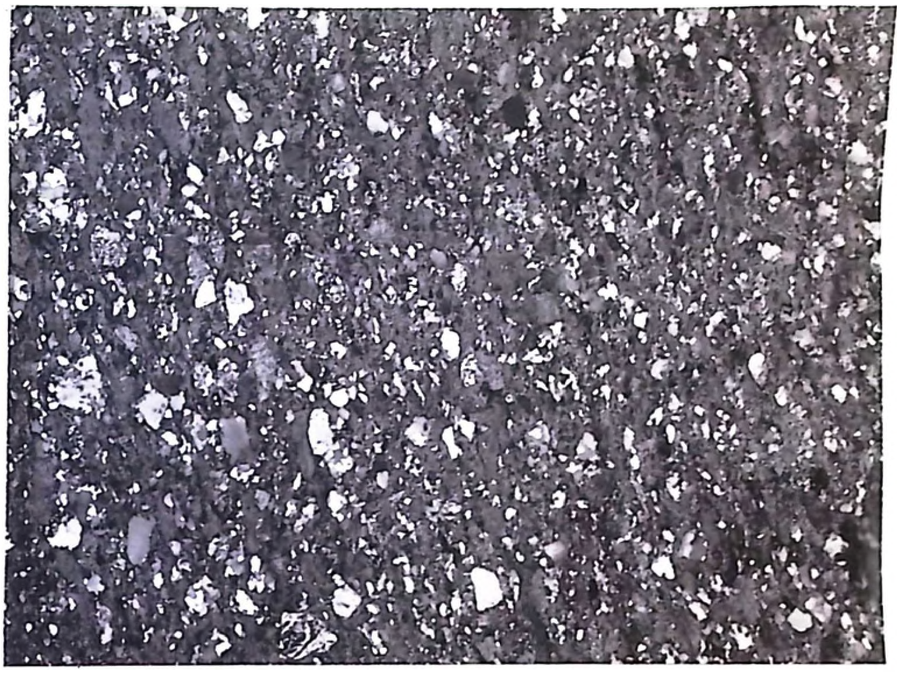


Figure 78. Projection print of a thin section of greywacke (Gw2, Appendix 2). Note foliated fabric and the many small felsic rock fragments. Field of view is 1.4 cm wide.

Figure 79. Projection print of a thin section of argillite conglomerate adjacent to finely bedded siltstone. Note the similarity of the clasts to the bedded siltstone. Field of view is 1.4 cm wide.

CHAPTER 6

INTERPRETATION OF THE PETROGRAPHY IN TERMS OF PROVENANCE

The megascopic textures and structures of the Abram Group allow interpretations of the sediments with regard to the depositional processes and environments, whilst the microscopic analyses permit deductions regarding the provenance of the sediments.

In a discussion of the sediment provenance it is impossible to separate the arkosic sandstone facies from the granite-bearing conglomerate facies, as the two are so intimately related. These two facies are first discussed, followed by the 2 tuff facies, and then the greywacke facies.

1. Arkosic sandstones and granite-bearing conglomerates

The petrographic similarity of the sandy matrix of the conglomerates and the remainder of the arkosic sandstone facies, and the similarity of the rock fragment types in both the arkoses and the conglomerates, indicate that the detritus within these two facies was supplied from the same source area. The differences in clast sizes between the facies represent differences in the hydrodynamic conditions of deposition (Chapter 4).

(a) Conglomerate clasts and arkosic sandstone rock fragments

Rock fragments commonly constitute from 10 to 25 percent of the arkosic sandstones (Appendix 2), and from 65 to 80 percent of the granite-bearing conglomerates (fig. 7). Thus, these rock fragments are very abundant throughout the 1,700 m of the Ament Bay Formation (fig. 6), and are by far the most useful components of the two facies for determining the provenance of the sediments, as they are direct representatives of the rock types which cropped out in the eroding source area.

The relative volumes of clast types in the conglomerates (fig. 7) are probably roughly correlatable with the relative abundances of the rock types in the source area, but as the percentages of the clast types vary considerably between outcrops of conglomerate it is not possible to accurately establish a single predominant rock type in the source area. Also, the percentages of rock fragment types in the arkosic sandstones (table 5) differ from those in the conglomerates (fig. 7), but it is likely that the rock fragments in both facies originated from the same rocks.

The very small numbers of "granitic" rock fragments in the arkoses are probably due to the grain sizes within the granodiorite clasts being approximately the same as the maximum clast sizes in the arkoses. Disintegration of the granodiorites into their mineral constituents would provide a feldspar and quartz sand.

The constituent grains of other rock fragments within the arkosic sandstones are much finer than those of the granodiorite clasts, and thus the rock types are recognisable even in sand-size fragments.

It is evident from figure 7 that the source area was heterogenous in nature, with the 3 most abundant rock types being intermediate to basic volcanics, granodiorite masses, and acidic lavas or tuffs. These rock types provided greenstone, granitic and felsic rock fragments to the granite-bearing conglomerates and arkosic sandstones.

The foliated granitic clasts may have been derived from the margins of granodiorite bodies, and the quartz porphyry clasts probably come from small intrusive bodies. The felsic tuffs represent more violent phases in the eruption of the volcanics of the sediment source area.

The northern belt of volcanics which underlies the Abram Group of sediments (fig. 2) is likely to have supplied volcanic debris to the Abram sediments. The unconformable contact of the basal conglomerate of the Abram Group with the underlying quartz porphyry body of the northern volcanic belt (fig. 10) suggests that these older volcanics were being eroded prior to the deposition of the Abram sediments, and that the basal conglomerates were deposited on this erosional surface. Also the orientations of the trough cross-bedding in the arkosic sandstone facies could be consistent with a roughly southern direction of transport of sediment (Chapter 4).

The source of the clasts of granodiorite within the arkoses and conglomerates is less evident, as the only granitic bodies which now exist in this area are the Kenoran granites which are younger than both the volcanics and sediments. However, in most outcrops of granite-bearing conglomerate there are from 20 to 35 percent by volume granodiorite clasts (fig. 7), and thus granodiorite must have underlain a considerable amount of the provenance area. This apparent enigma could be explained if the pre-Abram granodiorite bodies were located in part of the area now occupied by the post-Abram, Kenoran granites, and were assimilated to form these later granites.

The problem still remains of the time-relationship of the pre-Abram granodiorites and the northern volcanic belt. Were the volcanics deposited on top of a granitic terrain, implying a pre-greenstone belt sialic crust, or did the granodiorites intrude the greenstones? The only evidence (although rather meagre) for the resolution of this question, which was brought to light by this study, supports the latter interpretation. Six typical granodiorite clasts found within the granite-bearing conglomerate facies contain xenoliths of greenstone (fig. 15), which suggest that the granodiorites did in fact intrude greenstones, but it is not certain that these greenstones belonged to the northern volcanic belt.

(b) Quartz

Quartz has been used as a provenance indicator in several studies of early Precambrian sandstones. Walker and Pettijohn (1971) described the petrography of greywackes at Minnitaki Lake, and suggested that volcanic quartz could be distinguished by embayments in the crystal outlines filled with felsitic groundmass, and that the polygonised polycrystalline quartz grains were derived from gneissic marginal zones of granitic plutons. Reimer (1971) used "volcanic" quartz percentages in greywackes of the Sheba Formation of the Fig Tree Group in South Africa with some success, to show changes in the source area of the greywackes. From a study of the previous literature Reimer (1971) concluded that the only reliable quartz type indicative of source area is "volcanic" quartz, which is characterized by: "(1) the almost complete absence of monomineralic inclusions; (2) its almost exclusively non-undulatory extinction under crossed nicols; (3) its frequently idiomorphic outlines, or outlines which are influenced by crystal faces; and (4) by bulbous to fingery resorption structures".

Ojakangas (1972) also used volcanic quartz to indicate a volcanogenic source area for Archaean greywackes of the Vermilion District, northeastern Minnesota.

In the present study, the low percentages of quartz with non-undulatory extinction (volcanic quartz) in the arkosic sandstones (table 4), and the similarity of the quartz in the arkoses to that in the granodiorite clasts further support the thesis that much of the arkosic sandstone was derived from the disintegration of granodiorite bodies, and that very little quartz came from volcanic rocks. Also, the coarse nature of most quartz grains within the arkosic sandstones implies that the present rock type was of similar or coarser grain size, which excludes most volcanics. The majority of volcanic rocks in the northern volcanic belt contain only small amounts of quartz, even of fine grain size, and thus could have supplied little quartz to the arkosic sandstones.

Thin sections from quartz and feldspar porphyry clasts in the conglomerate, and from the quartz and feldspar porphyry body at Vermilion Lake revealed only 5 to 10 percent quartz (fig. 65). This quartz is noncrystalline with undulatory extinction, and thus could not have supplied the arkosic sandstones with polycrystalline quartz. However, it is likely that some of the monocrystalline quartz was supplied from quartz porphyry bodies.

Also, the felsic tuff facies contains from 6 to 13 percent quartz (fig. 65) which is commonly from 1 to 5 mm in size. Most quartz is monocrystalline with undulatory extinctions, but some is polycrystalline. However, the small number of clasts similar to the felsic tuff facies found in the granite-bearing conglomerate facies, indicates that only a small amount of the quartz in the arkosic sandstones could be derived from tuffs.

(c) Feldspar

Most feldspar grains throughout the arkosic sandstone facies are similar in grain size to the quartz grains, and constitute up to 38 percent of the arkoses (fig. 65).

If granodiorite bodies supplied most of the quartz to the arkosic sandstones, then a great deal of feldspar must also have been derived from the granodiorites. This is supported by the fact that the plagioclase grains within the arkosic sandstones have similar sodic compositions to those in the granodiorites. No calcic plagioclase grains were recorded from the arkoses, and only a small amount of K-feldspar was recognised.

Feldspar was probably also derived from felsic tuffs and from quartz and feldspar porphyries, but the low percentages of these rock types as fragments in the granite-bearing conglomerates (fig. 7) suggest that these were only minor sources of feldspar.

(d) Matrix

Both rock fragments and feldspar grains within the arkosic sandstones commonly have rather diffuse outlines with the surrounding matrix, and it is likely that much matrix was derived from the decomposition of rock fragments and feldspar grains. A small clay or silt-size fraction may have been deposited together with arkosic sandstones, but none settled out of suspension to give argillaceous layers. If there was an original very fine matrix to the arkoses, it has since been completely recrystallised, mostly to sericite.

2. Tuffs

In Chapter 4 the felsic tuffs were suggested to be ash-flow deposits. The foliated nature of some thin sections of felsic tuff and the possibility of glassy material (fig. 75) support this interpretation. The cracked feldspars with the slight displacement of the two parts (Chapter 5, fig. 74), are difficult to interpret, but may possibly be a cooling phenomenon, of perhaps be due to compaction which caused fracturing of the brittle feldspar grains.

The chemical composition of the felsic tuffs (see Chapter 7) suggests that the tuffs were derived from a magma of similar composition to that of quartz diorite. If these tuffs are indeed ash-flow deposits, then the eruptive material must have contained considerable quantities of volatiles, but less than those necessary to cause highly explosive eruptions which would have produced ash-fall tuffs.

The chemical composition of tuff in the lower part of the basic tuff facies (Chapter 7), suggests that a basaltic magma produced these basic tuffs. Their pyroclastic nature indicates explosive volcanism, which probably implies a high volatile content in the magma chamber.

The more felsic nature of the upper part of the basic tuff facies (Chapter 5) may either indicate a change in the composition of the parent magma, or it may be due to the deposition of ash from an eruptive centre other than that which was supplying the very basic tuffs.

3. Greywackes

The source rocks of the greywacke detritus are not as clearly defined as those of the arkosic sandstones and granite-bearing conglomerates. No large clasts from the provenance area are present within the greywackes, and most sediments are of very fine grain size. However, the quartz, feldspar and rock fragment grains within the coarse greywackes give some idea of the source rocks.

(a) Rock fragments

Basic and acidic volcanic clasts comprise the majority of rock fragments in the greywackes (table 5), but unlike the arkosic sandstone facies, those fragments designated as acidic volcanic now predominate over the chloritic, more basic volcanic fragments. This can be explained by the fact that 60 m of felsic tuffs and about 270 m of basic to felsic tuffs occur above the arkosic sandstones and below the greywackes (fig. 6). Thus, it is likely that much of the terrain supplying detritus to the arkosic sandstones was covered with tuff layers, before the greywackes were deposited. Erosion of the basic lavas which were supplying detritus to the depositional basin could have been largely prevented, and the acidic volcanic fragments in the greywackes could originate from the erosion of tuffs. The aphanitic groundmasses of the more felsic tuffs would provide fragments which resemble acidic lava fragments.

The few "granitic" fragments found in the greywackes (table 5), which comprise interlocking quartz and feldspar grains, may have been derived from felsic tuffs, but their equigranular character suggests that they are fragments from the granodiorite masses which provided detritus for the arkosic sandstones.

(b) Quartz

The great majority of quartz grains within the greywacke facies are polycrystalline (table 4) and have undulatory extinctions. This probably implies that most of the quartz was not derived from volcanic rocks (Blatt and Christie, 1963). The greater percentages (up to 23) of monocrystalline quartz within greywackes than in arkosic sandstones (table 4) may be a function of the slightly finer grain size of the greywackes, and could be due to the further breakdown of polycrystalline quartz grains.

The percentages of monocrystalline quartz with sharp extinctions differ only slightly between the greywackes and arkosic sandstones, and are low in both cases. Thus it is likely that volcanic quartz was an unimportant contributor to both types of sandstone. Erosion of granodiorite probably contributed to the quartz content of the greywackes, but felsic tuffs may also have supplied detritus. However, there is commonly 20 to 25 percent quartz in the greywackes, whilst felsic tuffs contain only 6 to 13 percent quartz (fig. 65), and much of this is monocrystalline in character. Also, the basic tuffs contain virtually no quartz, and could not have contributed to the quartz of the greywackes.

If most quartz in the greywackes were indeed derived from the granodiorite bodies, this implies that erosion of the granodiorite and transportation of the detritus to the basin was still continuing, but that depositional conditions within the basin had changed considerably since the deposition of the Ament Bay Formation. Erosion of the arkosic sandstones could have provided polycrystalline quartz for the greywackes, but in this case fragments from the granite-bearing conglomerates might also be expected in the greywackes.

(c) Feldspar

Most feldspar grains in the greywackes are plagioclase, as they stain pink with amaranth solution, but it is not possible to deduce the exact rock type from which they were derived. The plagioclase grains are commonly 0.5 mm to 1 mm in size, and thus were probably eroded from felsic tuffs or from the granodiorites. The compositions of the plagioclase grains provide no definitive evidence of their origin, as they are either albite or oligioclase, which is similar to the granodiorite clasts, whilst the plagioclase of the felsic tuffs is albite.

4. Summary of provenance

A flow chart indicating the probable provenance for the sediments of the Abram Group is shown in figure 80.

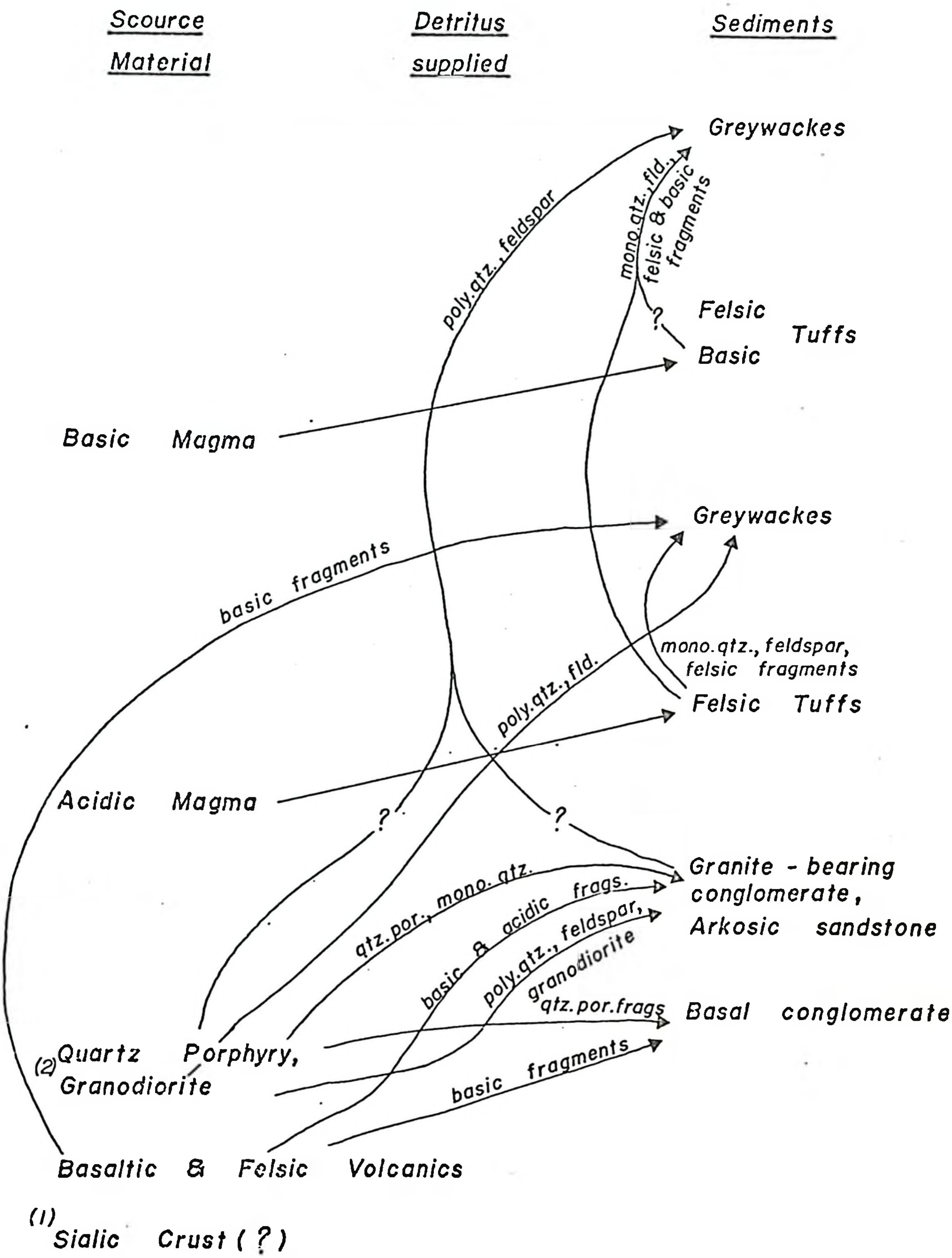


Figure 80. Flow chart showing the probable source for the detritus in the sediments of the Abram Group.

Notes: (1) Speculative crustal composition;
(2) Probably intrusive into the basaltic and felsic volcanics.

CHAPTER 7

GEOCHEMISTRY

1. Introduction

Whole rock chemical analyses were obtained for 8 samples of various lithologies from the Abram Group. These analyses were carried out by Mr. J.R. Muysson at the Geology Department, McMaster University, by using predominantly rapid wet chemical methods. The samples included 1 arkosic sandstone, 1 granodiorite clast, 1 felsic tuff, 1 basic tuff, 2 greywackes, and 2 argillites. These analyses are presented in table 7a.

2. Discussion

(a) Arkosic sandstone (sample no. 1 in table 7a)

The chemical analysis of the sandstone is very similar to that of the granodiorite clast (Gr 3 in Appendix 2), but is slightly lower in Al_2O_3, CaO, Na_2O and K_2O than the granodiorite and contains slightly more FeO and MgO. This is consistent with a dominantly granodioritic source for the arkoses, but with small amounts of added basic debris increasing the FeO and MgO percentages.

Donaldson and Jackson (1965, p.631) presented some analyses of quartz greywackes and arkoses (table 7b) from the

Sample No.*	1	2	3	4	5	6	7	8
SiO_2	71.08	72.03	65.46	49.62	69.83	64.20	59.56	64.42
TiO_2	0.43	0.30	0.44	0.69	0.34	0.47	0.70	0.48
Al_2O_3	11.86	13.28	16.81	15.57	15.05	13.94	18.52	16.44
Fe_2O_3	0.91	0.69	0.53	2.48	1.19	0.82	0.79	0.79
FeO	3.78	1.36	2.60	7.16	1.30	4.14	6.11	4.72
MnO	0.118	0.043	0.046	0.184	0.031	0.110	0.038	0.081
MgO	2.48	0.94	1.78	7.88	0.93	2.39	2.96	2.83
CaO	1.30	2.23	2.51	8.30	3.46	3.91	0.57	1.24
Na_2O	2.73	3.70	6.80	1.77	2.61	3.31	3.18	2.29
K_2O	1.88	3.37	0.94	3.09	2.74	1.41	2.64	2.86
P_2O_5	0.08	0.11	0.15	0.27	0.10	0.12	0.18	0.10
H_2O+	2.30	0.92	1.31	2.48	1.40	2.36	3.48	3.20
H_2O-	0.05	0.01	0.01	0.07	0.02	0.07	0.09	0.04
CO_2	0.66	0.39	0.58	0.04	0.42	1.56	0.37	0.03
Rb_2O	0.0050	0.0060	0.0022	0.0106	0.0101	0.0052	0.0112	0.0102
Li_2O	0.0030	0.0023	0.0038	0.0069	0.0037	0.0061	0.0101	0.0081
SrO	0.029	0.049	0.092	0.145	0.141	0.042	0.016	0.042
BaO	0.078	0.140	0.053	0.148	0.099	0.050	0.078	0.103
Total	99.77	99.57	100.12	99.91	99.57	100.03	99.30	99.63

*The samples are identified as follows:

1. Arkosic sandstone (Ar 12 in Appendix 2). Ament Bay.
2. Granodiorite clast (Gr 2 in Appendix 2). West of Kenneally Lodge.
3. Felsic tuff (F.T. 2 in Appendix 2). West of Kenneally Lodge.
4. Basic tuff. West of Kenneally Lodge.
5. Greywacke (Gw 9 in Appendix 2). From Daredevil Formation.
6. Greywacke (Gw 10 in Appendix 2). From Little Vermilion Formation.
7. Argillite. Island A, Little Vermilion Lake.
8. Argillite. Dog Island, Little Vermilion Lake.

Table 7a. Chemical analyses of lithologies from the Abram Group. For locations of samples see maps 1 and 2.

Sample No.*	9	10	11	12	13	14	15	16
SiO_2	67.8	81.9	73.4	66.5	67.4	49.83	64.67	56.30
TiO_2	0.25	0.28	0.2	0.30	0.39	0.94	0.57	0.77
Al_2O_3	17.4	9.1	14.1	17.2	17.0	14.64	13.41	17.24
Fe_2O_3	0.2	0.5	0.7	0.1	0.7	3.03	1.24	3.83
FeO	1.3	1.9	1.7	2.3	1.8	8.77	4.53	5.09
MnO	0.02	0.06	0.02	0.01	0.01	0.21	0.13	0.10
MgO	0.5	0.7	0.4	0.5	0.6	7.36	3.23	2.54
CaO	2.2	0.3	2.1	2.3	3.2	10.46	3.04	1.00
Na_2O	6.0	0.8	3.4	6.1	5.8	2.02	2.99	1.23
K_2O	2.1	1.3	3.5	1.6	1.1	0.23	2.02	3.79
P_2O_5	0.09	0.05	-	0.04	0.05	0.19	0.14	0.14
H_2O+	0.8	1.6	0.3	0.9	0.4	1.81	2.14	3.31
H_2O-			-					0.38
CO_2	0.2	0.1	-	0.1	0.1	0.33	2.15	0.84

*The samples are identified as follows:

9. Arkose from North Spirit Lake (Donaldson and Jackson, 1965, p.631 sample no. 4).

10. Quartz greywacke from North Spirit Lake (Donaldson and Jackson, 1965, p.631, sample no. 2).

11. Granodiorite (Turner and Verhoogen, 1960, p.344, sample no. 2).

12. Quartz diorite from North Spirit Lake (Donaldson and Jackson, 1965, p.635, sample no. 3).

13. "Algoman" quartz diorite (Donaldson and Jackson, 1965, p.635, sample no. 5).

14. Average of 53 Archaean basalts (Wilson et al., 1965, p.167).
15. Average Precambrian greywacke (Pettijohn, 1963).
16. Average of 33 Precambrian slates (Narz, 1953).

Table 7b. Chemical analyses of various lithologies from previous literature.

North Spirit Lake area. The arkosic sandstone of the Ament Bay Formation is fairly similar to the arkose from North Spirit Lake, but contains about 3 percent more SiO_2 and lesser amounts of Al_2O_3 and Na_2O. However, the quartz greywackes of North Spirit Lake contain up to 10 percent more SiO_2 than the arkosic sandstone of the Ament Bay Formation. Donaldson and Jackson (1965) suggested a sedimentary source area for the sandstones with very high quartz (and thus SiO_2) contents, but the much more feldspathic nature of the arkosic sandstones of the Ament Bay Formation suggests that no such second-cycle sandstones are present in the Abram Group.

(b) Granodiorite (sample no. 2 in table 7a)

The "granitic" clasts in the conglomerates of the Ament Bay Formation were classified as granodiorites on the basis of modal analyses (Chapter 5) and the chemical analysis is similar to other published analyses of granodiorites. The average composition of the granodiorite of the Southern Californian batholith is very similar to the composition of the granodiorite clast (Turner and Verhoogen, 1960; and table 7b).

The percentage of SiO_2 in sample no. 2 (table 7a) is higher than that of the quartz diorite clasts in the North Spirit Lake area (Donaldson and Jackson, 1965; and table 7b), and is also higher than that of Donaldson and Jackson's (1965) sample of "Algoman" quartz diorite (table 7b). This further emphasises the acidic nature of the granodiorite clast.

(c) Felsic tuff (sample no. 3 in table 7a)

The felsic tuff sample has less SiO_2 than the granodiorite clast, and also less K_2O and higher Na_2O (6.80 percent instead of 3.70 percent for the granodiorite clast). This high value of Na_2O and low value of K_2O reflect the albitic composition of the feldspar in the felsic tuff, and the absence of K-feldspar. This analysis is very similar to the analysis of quartz diorite from North Spirit Lake (table 7b), and the felsic tuff is more likely to have been derived from a magma of this composition, than from a magma with a composition similar to the granodiorite clast.

(d) Basic tuff (sample no. 4 in table 7a)

The truly basic nature of this tuff is demonstrated by its low value of SiO_2 (49.62 percent), and by its similarity to the average of 53 analyses of Archaean basalt samples (Wilson et al., 1965; and table 7b). However, the basic tuff differs from this basalt composition by containing the relatively high value of 3.09 percent K_2O, which suggests that it was derived from a K_2O-rich basaltic magma.

(e) Greywacke (sample nos. 5 and 6 in table 7a)

The sample from the greywacke unit in the Daredevil

Formation (sample no. 5) is rather more silicic in composition that is the sample from near the top of the Little Vermilion Formation (sample no. 6). It contains 5 percent SiO_2 in excess of that of sample no. 6. It would be inappropriate to make deductions regarding the changing provenance area based solely on 2 analyses, but the more basic composition of the greywacke from the Little Vermilion Formation is in accordance with the greater number of basic volcanic rock fragments in the greywackes of the Little Vermilion Formation than in the greywackes of the Daredevil Formation (see table 5).

Thus, it is tentatively suggested that the source area of the greywackes of the Little Vermilion Formation was slightly more basic than was the source area of the greywackes of the stratigraphically lower Daredevil Formation. However, the dissimilarities between the basic tuff analysis and the analysis of the greywacke samples suggests that the provenance areas of the Daredevil and Little Vermilion Formations were by no means composed predominantly of basic rock.

Sample no. 6 compares well with Pettijohn's (1963) average Precambrian greywacke (table 7b), but sample no. 5 has more SiO_2 than the average, and also contains slightly more K_2O than Na_2O.

(f) Argillite (sample nos. 7 and 8 in table 7a)

Both analyses of argillite contain greater amounts of

SiO_2 than does the average composition of 33 Precambrian slates (Nanz, 1953; and table 7b). These relatively high SiO_2 values are consistent with a large proportion of acidic rocks in the source areas of the argillite detritus.

CHAPTER 8

BASIN EVOLUTION

1. Discussion

In the preceeding chapters the stratigraphy and sedimentology of the Abram Group have been described, and the relationship of the Abram Group to the other deposits of the greenstone belt have also been outlined. In this chapter the changing environments within the depositional basin are traced by analysing the stratigraphic sequence.

The oldest rocks in the northern part of the greenstone belt are the volcanics of the northern volcanic belt (Chapter 2). The common occurrence of pillow lavas within this belt suggests that basic subaqueous lava flows were providing much of the volcanic material. The less common pyroclastic deposits indicate periods of more violent volcanic activity, and the Patara sediments represent a period of quiescence, when erosion and sedimentation of part of the volcanic pile took place. The Patara sediments have not been studied in sufficient detail to establish the process of deposition, but Pettijohn (1935, p.1895) suggested that they were "formed by water deposition of volcanic materials". No "granitic" detritus has been reported from these sediments.

The Abram Group of sediments occurs stratigraphically above the northern volcanic belt, but is separated from it by a large and extensive fault. However, a very different environment from the subaqueous lava flows of the northern volcanic belt is indicated by the lowest part of the Abram Group (the Ament Bay Formation). The arkoses and conglomerates of the Ament Bay Formation have been interpreted as alluvial fan deposits (Chapter 4), and a large amount of the detritus within this formation has been shown to be derived from rocks of granodiorite composition (Chapter 6). The basic rocks of the northern volcanic belt could not have supplied detritus of this composition, but it is likely that they did provide the greenstone clasts of the granite-bearing conglomerates and also the basic rock fragments within the arkosic sandstones.

The rather drastic change from conditions of subaqueous volcanism to subaerial sedimentation could be explained if considerable uplift occurred between the deposition of the volcanics and sediments, which may have been related to emplacement of granodiorite bodies. The only direct (though scanty) evidence of the intrusion of greenstones by granodiorite are the xenoliths of greenstone within some granodiorite clasts from the granite-bearing conglomerates (fig. 15). However, the interpretation of the 1,700 m thick Ament Bay Formation as being alluvial fan deposits implies a rising mountainous terrain (to the north?), and a subsiding depositional area, their movements probably being fault controlled. Much of the provenance area consisted of granodiorite, but some was basic to felsic volcanic rocks. Erosion of this terrain probably proceeded at a rapid rate, with the production of only small amounts of very fine grained detritus. Deposition of the Ament Bay Formation probably took place largely from ephemeral watery surges of sand and gravel at the mountain front. It is not known what sort of deposition (if any) was taking place in the main basinal area (i.e. farther south) at this time.

The termination of these alluvial fan conditions occurred with the deposition of the felsic tuff facies. It has been tentatively suggested that these tuffs are subaerial ash-flow deposits, which covered the alluvial fan sediments and presumably also much of the source area as well. When sedimentation resumed, greywackes were deposited from turbidity currents, which implies that the depositional area had subsided by this time such that deposition was now taking place in relatively deep water.

Another explosive eruptive volcanic episode produced the basic tuffs of the Daredevil Formation (fig. 6), which were probably deposited subaqueously. Turbidity currents were largely prevented during the deposition of the basic tuffs, but a few graded beds within the basic tuffs may have been produced by deposition from turbidity currents. Turbidite sedimentation again followed the cessation of volcanism. It is likely that some material supplied to the turbidites was still being derived from granodioritic sources, but erosion of areas of tuff may have produced much of the debris in the greywackes.

If it is accepted that the central volcanic belt is younger than the Abram Group of sediments (Chapter 2), then extensive basic volcanism again took place, which brought to an end the sedimentary conditions which existed during the accumulation of the Abram Group.

The relationship between the central volcanic belt and the southern belts of sediments and volcanics within this greenstone belt, are by no means certain, and speculations regarding their roles in the depositional evolution of the greenstone belt are not warrented here.

The later emplacement of the Kenoran granites probably remobilized the older granodiorite bodies and caused the in-folding of the volcanics and sediments of the greenstone belt.

2. Summary

(1) Extensive and prolonged subaqueous volcanism (predominantly basic) formed the oldest deposits of the greenstone belt at Sioux Lookout. This is in accordance with the oldest rocks found in other greenstone belts (see volume edited by Baer [1970] for a review of some of the Canadian greenstone belts).

(2) The Patara sediments within the northern volcanic belt are volcaniclastic deposits which contain no "granitic"

detritus.

(3) The arkosic sandstones and granite-bearing conglomerates of the Ament Bay Formation are alluvial fan sediments, which were deposited unconformably on the rocks of the northern volcanic belt.

(4) If virtually all the quartz and feldspar grains within the arkosic sandstones were derived from rocks of similar composition to the granodiorite clasts within the granite-bearing conglomerates, then over 60 percent of the Ament Bay Formation was derived from the erosion of granodiorite (see modal analyses in Appendix 2).

(5) Most "granitic" clasts within the granite-bearing conglomerates are true medium grained granodiorites, and were not derived from flows or from shallow porphyritic intrusions (both of which were suggested by Bass [1961] to account for the "granite" fragments from Archaean conglomerates in western Quebec and eastern Ontario), but came from larger bodies of granodiorite which may have intruded the greenstones of the northern volcanic belt. The sediments of the Abram Group are not simply volcaniclastic as was suggested by Ayres (1969) for the Archaean metasandstones from the Lake Superior Park area.

(6) Intrusion of granodiorite bodies into the volcanic sequence would have provided the right provenance area for the sediments of the Ament Bay Formation, and may also have provided the right tectonic conditions for alluvial fans to form (i.e. an

uprising mountainous area, bordering a depositional basin).

(7) Episodes of explosive volcanism brought to an end the alluvial fan conditions of sedimentation, and deposited the tuffs of the Daredevil Formation. Subsidence of the depositional basin took place during this time, and when sedimentation was resumed subaqueous turbidites were deposited.

(8) Turbidite sedimentation was responsible for a great deal of the sediment fill found in other greenstone belts, but this study has shown that it is not the only type of sedimentation which was operative within the Archaean depositional basins.

(9) The sedimentary sequence within the Abram Group at Sioux Lookout does not correspond to the flysch to molasse "Alpine-type" sequence which has been suggested by Glikson (1971b) to have been the trend of evolution of Archaean volcanic-sedimentary troughs. In the Abram Group, sedimentation evolved from subaerial alluvial fan deposition to subaqueous turbidite deposition.

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Appendix 1. Measured stratigraphic sections of the Ament Bay Formation, at various locations around Little Vermilion Lake (see map 1).

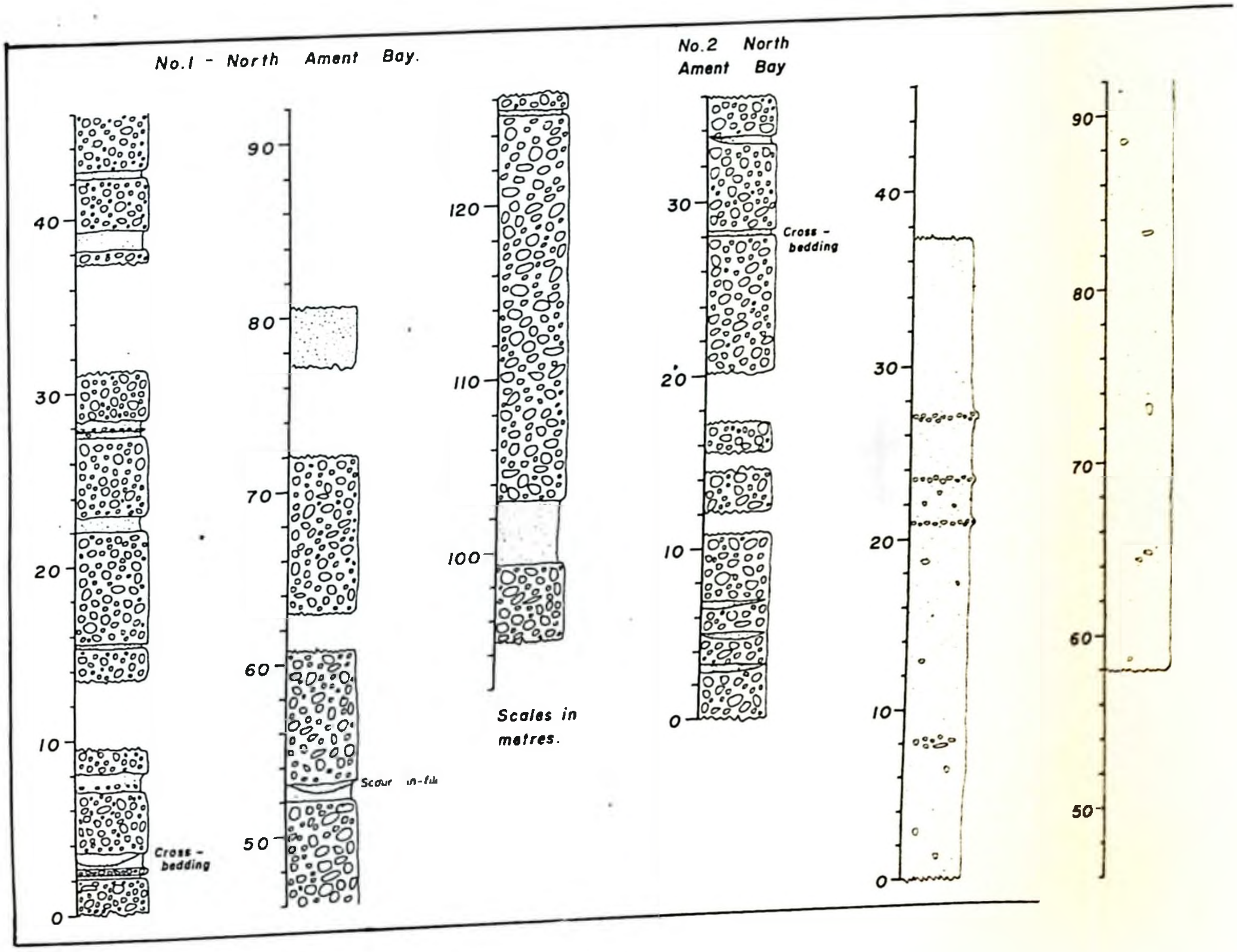
arkosic sandstone

granite-bearing conglomerate

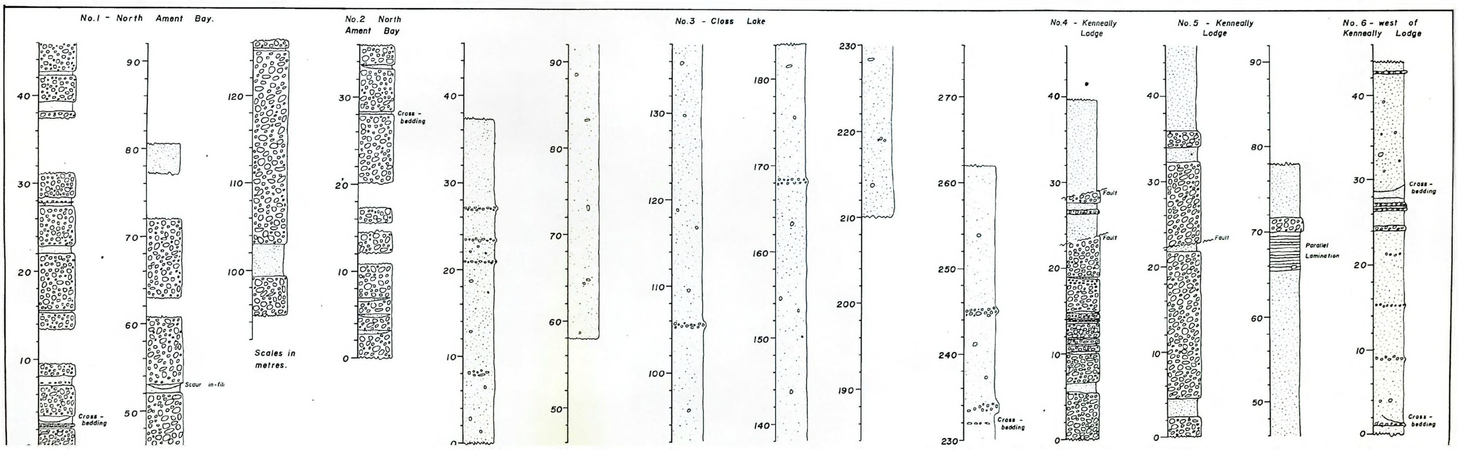
APPENDIX 2

MODAL ANALYSES

Lab. No.	Field No.	Quartz	Feldspar	Rock Fragments	Matrix
Ar 1	Ar 4	35	26	11	28
Ar 2	Ar 9/7/1	35	33	14	
Ar 3	Ar 35	35	21	25	19
Ar 4	Ar 9	43	19	25	13
Ar 5	Ar 3/7/2	34	32	11	23
Ar 6	B2-B	34	27	28	11
Ar 7	Cl 8	37	27	23	13
Ar 8	Ar 22/1	32	31	17	20



Q.F. 2- Quartz feldspar porphyry Gw - Greywacke
*Very fine grained and recrystallized matrix



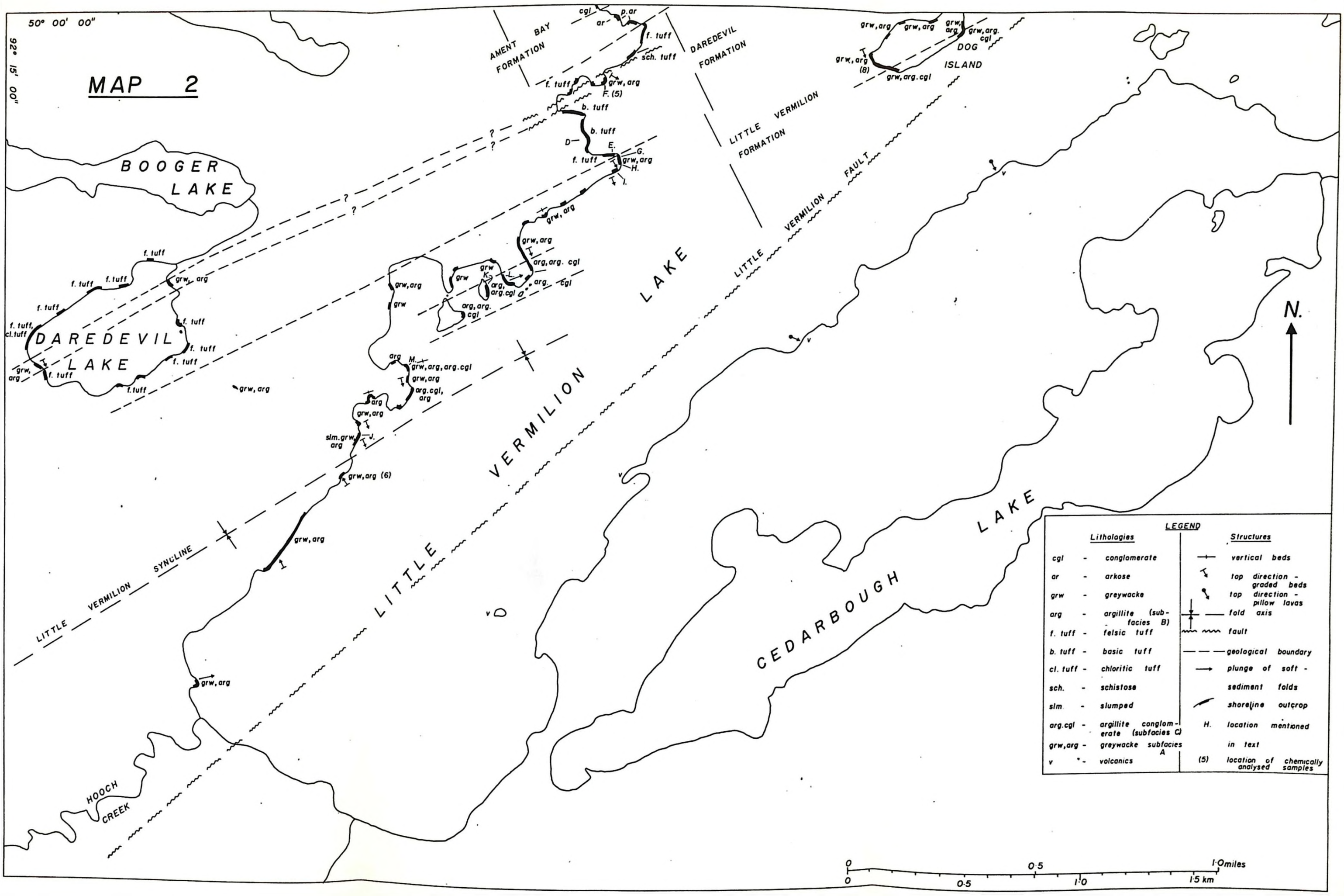
No.1 - North Ament Bay.
Cross - bedding
Scour in-fill
Scales in metres.
No.2 North Ament Bay
Cross - bedding
No.3 - Closs Lake
Cross - bedding
No.4 - Kenneally Lodge
Fault
Fault
No.5 - Kenneally Lodge
Fault
Parallel Lamination
No. 6 - west of Kenneally Lodge
Cross - bedding
Cross - bedding

APPENDIX 2

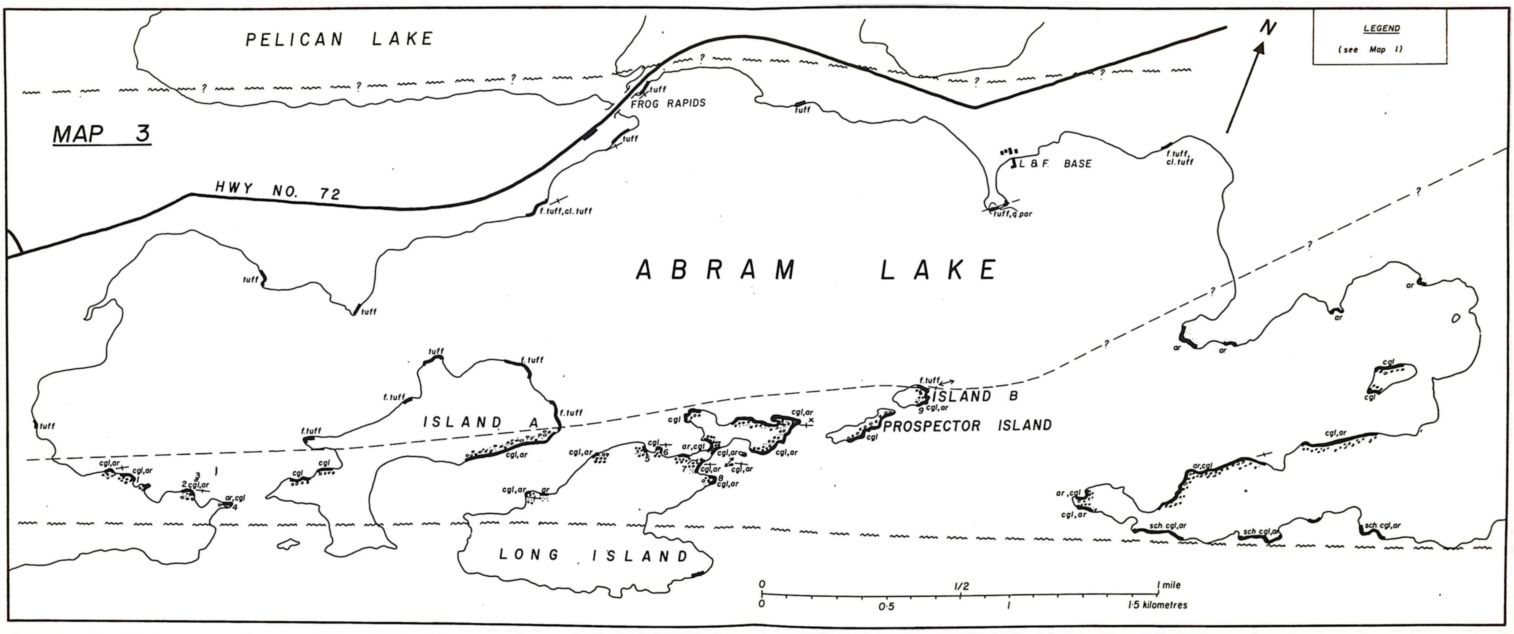
MODAL ANALYSES

Lab. No.	Field No.	Quartz	Feldspar	Rock Fragments	Matrix
Ar 1	Ar 4	35	26	11	28
Ar 2	Ar 9/7/1	35	33	14	
Ar 3	Ar 35	35	21	25	19
Ar 4	Ar 9	43	19	25	13
Ar 5	Ar 3/7/2	34	32	11	23
Ar 6	B2-B	34	27	28	11
Ar 7	Cl 8	37	27	23	13
Ar 8	Ar 22/1	32	31	17	20
Ar 9	Ar 5	30	27	22	21
Ar 10	Ar 8/7/6	38	38	2	22
Ar 11	Ar 15/8/6	28	32	3	37
Ar 12	Ar 21/8/1	37	23	19	21
Ar 13	Ar 2	40	20	18	20
Ar 14	Ar 33	37	33	13	17
Gr 1	Cl 25/7/1	29	64		7
Gr 2	Cl 5/8/2	31	67		2
Gr 3	Cl 8	34	60		6
Gr 4	Cl 21/8/1	30	63		7
Gr 5	M1	35	51		14
Gr 6	M2	32	58		10
Gr 7	M3	29	60		11
Gr 8	M4	27	70		3
Gr 9	2 Gr 4	31	58		11
Gr 10	B2-A	30	65		5
Q.P. 1	2 Gr 3	5	28		67
F.T. 1	Ar 15	6	38		56
F.T. 2	C 12	6	27		67
F.T. 3	Ar 14	13	40		47
Gw 1	Ar 23	23	17	17	43
Gw 2	D 17	23	11	24	42
Gw 3	Gw 13/7/5	23	32	13	32
Gw 4	Gw 17/7/1	20	20	25	35
Gw 5	Gw 18/7/5	20	21	21	38
Gw 6	Gw 4/8/6	18	2	4	76*
Gw 7	Gw 18/7/3	31	4	3	62*
Gw 8	Gw 26/3	22	3	0	75*
Gw 9	Gw 19/8/1	22	28	12	38
Gw 10	Ar 30	26	17	17	40
Q.P. 2	Q.P. 2	10	27		63

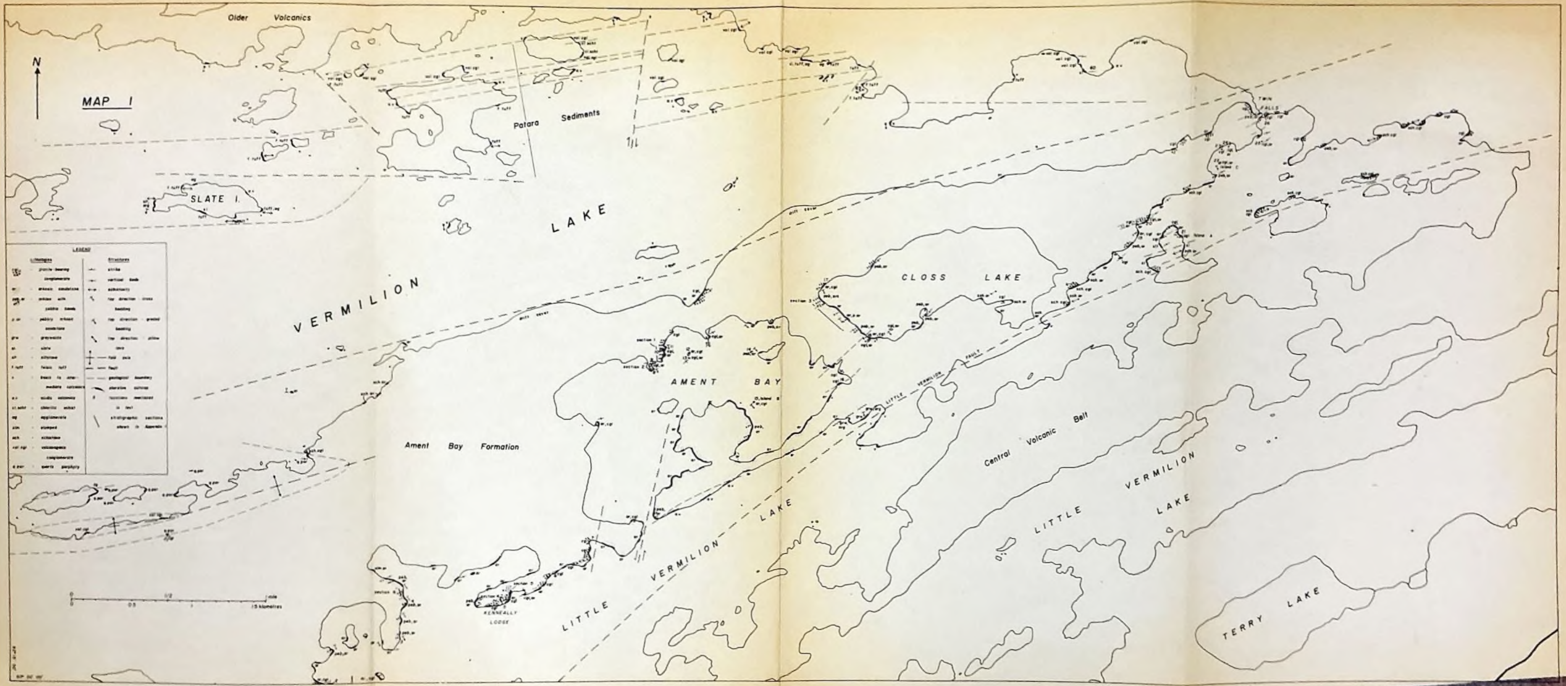
Ar - Arkosic sandstone
Q.P. 1- Quartz-feldspar porphyry clast
Q.P. 2- Quartz-feldspar porphyry
Gr - Granodiorite clasts
F.T.- Felsic tuff
Gw - Greywacke
*Very fine grained and recrystallized matrix



MAP 2
50° 00' 00"
92° 15' 00"
BOOGER LAKE
DAREDEVIL LAKE
LITTLE VERMILION LAKE
CEDARBOUGH LAKE
DOG ISLAND
HOOCH CREEK
AMENT BAY FORMATION
DAREDEVIL FORMATION
LITTLE VERMILION FORMATION
LITTLE VERMILION FAULT
LITTLE VERMILION SYNCLINE
N.
LEGEND
Lithologies
cgl - conglomerate
ar - arkose
grw - greywacke
arg - argillite (subfacies B)
f. tuff - felsic tuff
b. tuff - basic tuff
cl. tuff - chloritic tuff
sch. - schistose
slm. - slumped
arg.cgl - argillite conglomerate (subfacies C)
grw,arg - greywacke subfacies A
v - volcanics
Structures
vertical beds
top direction - graded beds
top direction - pillow lavas
fold axis
fault
geological boundary
plunge of soft - sediment folds
shoreline outcrop
H. location mentioned in text
(5) location of chemically analysed samples
0 0·5 1·0 miles
0 0·5 1·0 1·5 km



MAP 3
PELICAN LAKE
ABRAM LAKE
HWY NO. 72
FROG RAPIDS
L & F BASE
ISLAND A
ISLAND B
PROSPECTOR ISLAND
LONG ISLAND
LEGEND
(see Map 1)
N
0
1/2
1 mile
0
0·5
1
1·5 kilometres



MAP I
N
Older Volcanics
Patara Sediments
SLATE I.
VERMILION LAKE
CLOSS LAKE
AMENT BAY
Ament Bay Formation
TWIN FALLS
LITTLE VERMILION FAULT
Central Volcanic Belt
LITTLE VERMILION LAKE
TERRY LAKE
KENNEALLY LODGE
LEGEND

LEGEND
Structures
geological boundaries (?-location uncertain)
faults
fold axes (?-location uncertain)
top direction - pillow lavas
top direction - graded beds
overturned bedding
top direction - cross bedding
D locations mentioned in text
granite
greywacke/argillite
unshaded)
basic tuff
felsic tuff
conglomerate sections
Ament Bay Formation (arkoses & conglomerates)
quartz porphyry
volcanics
Patara sediments (including tuffs)
basic to intermediate volcanics
N
Lost Lake
C.N.R.
Vermilion River
HUDSON
NORTHERN
VOLCANIC
BELT
SIOUX LOOKOUT
Sturgeon R.
Highway 116
Pelican Lake
Frog Rapids
David Lake
Botham Bay
Long I.
Prospector
Patara Lake
GROUP
ABRAM
Abram Lake
Abram Chute
Alcona B.
Whitefish Island
Vermilion Lake
Close Lake
Ament Bay
AMENT BAY FORMATION
Kenneally Lodge
Little Vermilion Lake
Northeast Bay
BELT
VOLCANIC
CENTRAL
Troutfish Bay
East Bay
Ruby
Neepawa I.
Lyons Bay
Butterfly Lake
Daredevil Lake
Black Fox Lake
DAREDEVIL FORMATION
LITTLE VERMILION FORMATION
Minnitaki Lake
GROUP
MINNITAKI
Twinflower Bay
Southeast Bay
Hooch Lake
Maskinonge Lake
Pickerel Arm
Highway 72
miles
km
0 1 2 3
SOUTHERN VOLCANIC BELT