GEOLOGY OF THE ATHAPUSCOW AULACOGEN, EAST ARM OF GREAT SLAVE LAKE, DISTRICT OF MACKENZIE

Project 660009

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Introduction

This project was originally undertaken in 1966-67 as a study of Proterozoic sedimentary rocks in the East Arm of Great Slave Lake. From this work came a detailed description of the stratigraphy (Hoffman, 1968), a reconstruction of the depositional history (*idem*, 1969), an interpretation of the regional tectonics (*idem*, 1973), and a facies analysis of the important stromatolite-bearing formations (*idem*, 1974). The area is interpreted as an aulacogen, or failed rift arm, genetically related to the Coronation Geosyncline of late Aphebian age (Hoffman and others, 1974).

Many aspects of the structural and magmatic development of the aulacogen were still poorly understood, however, because of the lack of systematic large-scale mapping. The best maps, based on the canoe-reconnaissance work of Stockwell (1936), were published at a scale of four miles equals one inch. To improve them, the project was reactivated and, during the 1976 field season, all the Proterozoic rocks were remapped for publication at 1: 50 000 scale. Most of the Archean rocks in the East Arm had been mapped in 1967-68 by the late E.W. Reinhardt, and his meticulous but largely unpublished work is being incorporated in the maps now in preparation. It is expected that the new 1: 50 000 scale maps will be placed on open file early in 1977.

This report summarizes only new data and concepts and, for a more general description of the area, the reader is referred to the earlier publications mentioned above. Among the many highlights of the new mapping project, the following have especially intriguing regional implications:

(1) There are many huge nappes, up to 70 km in length, of recumbently-folded sedimentary rocks that moved, at the end of Great Slave Supergroup deposition,

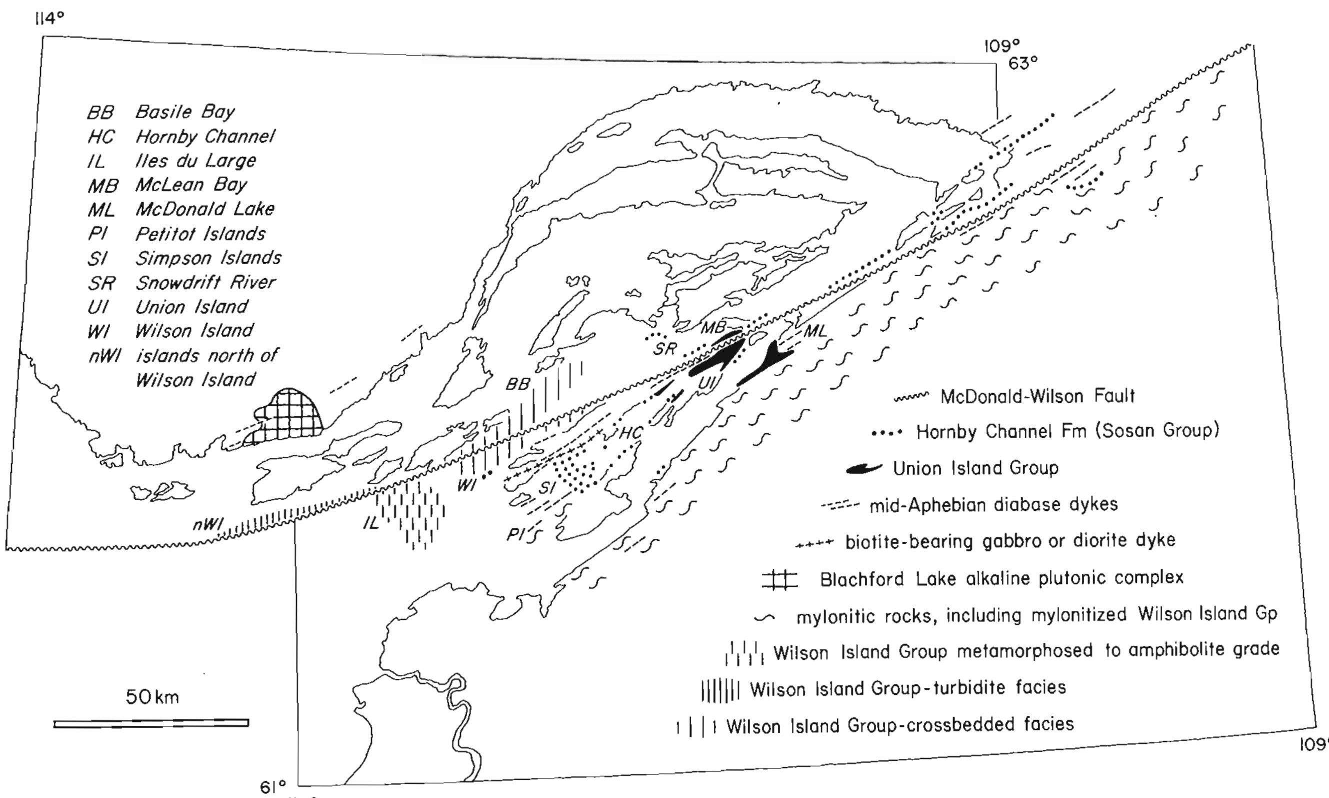

Figure 25.1.

Distribution of various pre-Kahochella Group rocks with the effects of strike-slip on the McDonald-Wilson Fault removed.

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northwestward into the East Arm from the region of tectonic denudation south of the McDonald-Wilson Fault. The nappes contain stratigraphic units that are correlative with, but of more basinal facies than, the subjacent autochthon. They indicate that the axial trough of the aulacogen originally lay south of the McDonald-Wilson Fault.

(2) The chaotic megabreccia, or olistostrome, of the Stark Formation is now believed to result from solution of a thick sequence of salt, deposited above the basinal carbonates of the Pethei Group. Solution collapse occurred in advance of the prograding fan-delta that deposited the overlying red clastic sediments of the Christie Bay Group. The internal structure of the megabreccia was subsequently complicated by its involvement in nappe tectonics, and later by the forceful intrusion of giant laccoliths, some more than 20 km in diameter, of diorite and monzonite.

(3) More than 35 volcanic centres have been located and are found in every formation of the Sosan and Kahochella groups. Most are basaltic and range from simple breccia pipes, through small (less than 1 km in diameter) submarine pyroclastic cones, to larger (more than 5 km) multicycle complexes or reworked tuffs and lavas. Many centres in the Kahochella Group occur along old fault lines, where displacements occurred during deposition of the underlying Sosan Group. A line of silicic centres extends from Seton Island to near Sachowia Point, where large rhyolite domes and ash-flow tuffs occur. In general, volcanic rocks are more extensive in the autochthon than in the nappes, indicating that volcanism was concentrated at the margin of the aulacogen, not in the axial trough.

(4) The Wilson Island Group, possibly of early or middle Aphebian age, records a stage of rifting, beginning with bimodal volcanic rocks and fanglomerate, that predates the main development of the aulacogen in the late Aphebian. The group was metamorphosed, locally to amphibolite grade, and involved in regional mylonitization of mid-Aphebian age, before a second cycle of rifting occurred, which initiated the aulacogen.

(5) An important swarm of diabase dykes, of mid-Aphebian age, extends the entire length of the East Arm. The dykes intrude the Archean basement rocks only, but are presumed to postdate the deformation of the Wilson Island Group. They pass beneath the sedimentary rocks of the aulacogen and are probably a manifestation of the cycle of rifting that initiated the aulacogen. An unusual biotite-bearing dyke, previously dated at 2170-2200 million years, is cut by the dyke swarm and provides a maximum age for the aulacogen.

(6) There is about 75 km of dextral strike-slip, estimated from displacements of Wilson Island and Union Island Group rocks, on what is here termed the McDonald-Wilson Fault. West of the Snowdrift River, the main locus of movement diverges from the McDonald Fault and skirts the north shore of Simpson Island and Wilson Island. The often photographed fault-line scarp on the south shore of McDonald Lake marks a mere splay fault of little strike-slip displacement. The extensive belt of mylonites south of the East Arm is not genetically related to the McDonald-Wilson Fault. Most of the Archean granites along the fault are relatively undeformed and, where mylonites do occur, they are clearly truncated by, and therefore older than, the fault.

Archean Basement Rocks and the southward extent of the Slave Province

Archean basement underlies all the Proterozoic rocks of the aulacogen. Most of it consists of massive adamellite and granite but, in the western Simpson Islands, there is a northwest-trending belt of gneisses and migmatite. The gneisses are an extension of the Archean metasedimentary belt exposed on the north shore of Hearne Channel near Narrow Island. Foliation in the gneisses dips gently to the northeast, projecting beneath the massive granitic rocks. Dips are about the same as those in the overlying Sosan Group (Table 25.1) and, therefore, the gneisses must have been nearly horizontal at the time of Sosan deposition.

Most of the Archean rocks along the McDonald-Wilson Fault are remarkably little deformed and the massive granitic rocks extend southward, as much as 10 km, to a northeast-trending belt of intensely lineated mylonites (Fig. 25.1). North of this belt, Archean radiometric ages have been obtained from micas even using the K-Ar method. So far as deformation of the basement is concerned, it is useful to consider that the Slave Province extends as far south as the north edge of the mylonite belt.

Wilson Island Group and the First Cycle of Rifting

This is an important group, over 6 km in thickness, of early or middle Aphebian sedimentary and volcanic rocks (Table 25.1). The oldest rocks, exposed south of Wilson Island, are an intimately mixed assemblage of basalt and rhyolite flows, and conglomerate containing volcanic, granitic and gneissic clasts. The granitic and gneissic clasts are derived from the underlying Archean basement, although no basal unconformity is exposed. On Wilson Island itself, the volcanic rocks are overlain by a thick, homoclinal, vertically dipping, northward facing sequence of cross-bedded quartzite, feldspathic quartzite, and dolomitic quartzite. Eliminating the displacement on the McDonald-Wilson Fault, the sequence on Wilson Island is juxtaposed against the other main outcrop belt of Wilson Island Group, that around Basile Bay (Fig. 25.1). There, stratigraphically above the sequence on Wilson Island, a sequence of impure dolomite, argillaceous quartzite, and argillite with local flows of porphyritic basalt is preserved in the core of a major synclinorium. Using the same pre-fault reconstruction, the quartzose turbidites exposed on the islands north of Wilson Island, a facies quite distinct from that on Wilson Island itself, are placed far to the west. The transition from the eastern cross-bedded facies to the western turbidite facies is consistent with the group's westerly directed paleocurrents (Yeo, 1976).

The Wilson Island Group is everywhere significantly more metamorphosed than the adjacent Great Slave Supergroup. At all of the localities mentioned above, the group is of greenschist grade. On the Iles du Large (Fig. 25.1), amphibolite grade is attained and the pelitic rocks contain prominent metacrysts tentatively identified as andalusite and staurolite. On some of the islands, notably Butte Island, the Wilson Island Group is intruded by deformed stocks and dykes of pink biotite adamellite, which have been sampled for geochronological studies.

On the Petitot Islands, a fault-bounded sliver of Wilson Island Group has suffered severe cataclastic deformation, as have the adjacent Archean gneisses. This area is probably an extension of the mylonite belt south of the McDonald Fault (Fig. 25.1). The Great Slave Supergroup, however, is not mylonitized in this area. Furthermore, on an island at the western end of Inconnu Channel, a conglomerate near the base of the Sosan Group, which there lies unconformably on Archean gneiss, contains mylonitized (and nonmylonitized) clasts of both Archean rocks and Wilson Island Group quartzite.

Despite the fact that the Wilson Island Group is nowhere exposed in unfaulted contact with either the Archean basement or the Great Slave Supergroup, the following conclusions can be made in light of the relations outlined above (Table 25.1):

(1) The Wilson Island Group is younger than the Archean and older than the Great Slave Supergroup.
(2) Mylonitization postdates the Wilson Island Group and predates the Great Slave Supergroup.
(3) Metamorphism of the Wilson Island Group probably also predates the Great Slave Group, although there is some uncertainty on this point because the juxtaposition of metamorphosed Wilson Island Group and unmetamorphosed Great Slave Supergroup may be the result of younger nappe tectonics. If, however, metamorphism can be proved to predate mylonitization, then it must also predate the Great Slave Supergroup.

Yeo (1976) has suggested, on the basis of paleocurrent data, that the Wilson Island Group was deposited in a precursor trough coincident with the Athapuscow Aulacogen. That this trough originated by rifting is indicated by the occurrence, at the base of the Wilson Island Group, of the classic rift-valley assemblage of compositionally bimodal volcanic rocks and conglomerate. This cycle of rifting is separated, however, from the main development of the aulacogen by a period of intense mylonitization and, probably, thermal metamorphism of mid-Aphebian age. The earlier cycle of rifting may not bear the same genetic relationship to the Coronation Geosyncline as does the later but, by providing a zone of crustal weakness, it may ultimately have predetermined the site of the aulacogen.

Mid-Aphebian Diabase Dyke Swarm and the Beginnings of the Second Cycle of Rifting

An important swarm of mid-Aphebian diabase dykes extends, on an east-northeast trend, the entire length of the East Arm (Fig. 25.1). The dykes intrude only the Archean basement rocks, and are most numerous at the northeast end of Simpson Island and south of the McDonald Fault east of the Snowdrift River. On the southern Simpson Islands, they are overlain by the Sosan Group and, southwest of McDonald Lake, they pass beneath the Union Island Group (Table 25.1). Dykes of the same trend intrude the Blachford Lake alkaline plutonic complex (Davidson, 1972) north of Hearne Channel and, south of Hornby Channel, they intrude mylonitic gneisses (Reinhardt, 1972).

In the west-central Simpson Islands, there is an unusual dyke, almost 30 km long, of biotite-bearing gabbro or diorite (Fig. 25.1). The biotite, which is of igneous origin, has been dated radiometrically by the K-Ar method at 2200 (Burwash and Baadsgaard, 1962, p. 28) and 2170 (Leech *et al.*, 1963, p. 61) million years. There has long been confusion, however, concerning the relation of this dyke to the Sosan Group, with which, in its eastern part, it is in fault contact. Reinhardt's unpublished maps clearly show, and we have confirmed, that this dyke is cut by numerous diabases of the swarm that passes beneath the Sosan Group and must, therefore, be older. Part of the confusion arose from reports that a "trachytic" border phase of the dyke closely resembles other dykes, associated with diatreme breccias (*see* Reinhardt, 1972) occurring along old fault lines, that do intrude the Sosan Group. Mapping shows that the "trachyte" occurs only where the dyke is faulted against the Sosan Group and is not a normal border phase of the dyke. The "trachyte", if it is related to the diatreme breccias, probably intrudes the fault and is, therefore, younger than the biotite-bearing dyke.

If we have interpreted these relations correctly, it can be concluded that:

(1) The diabase dyke swarm is younger, although probably not very much younger, than 2170-2200 million years.

(2) The dyke swarm is older, although probably not very much older, than the Union Island Group and Great Slave Supergroup.

(3) The dyke swarm apparently post-dates the mylonite belt south of the McDonald Fault and, therefore, should be younger than the Wilson Island Group. There is uncertainty on this point, however, because the dyke swarm is not known to intrude the Wilson Island Group itself. The group is cut by basic dykes, but they are metamorphosed and of more northerly trend. They could be local feeders to the upper Wilson Island Group basalt flows. Whether involvement of the group in nappe tectonics may explain this anomaly we cannot say, but there remains some ambiguity concerning the relative ages of dyke emplacement, mylonitization, metamorphism, and the Wilson Island Group. For now, we consider the dyke swarm, including the dated biotite-bearing dyke, to mark the beginnings of the second cycle of rifting in the East Arm (Table 25.1).

Union Island Group: Internal Stratigraphy and Age Relations

This is an interesting group of stagnant basin sediments, subaqueous basalts and intrusive gabbros. They are the oldest supracrustal rocks in the aulacogen that can be correlated with rocks in the Coronation Geosyncline and were the first to be deposited in the second cycle of rifting. It is exposed almost exclusively in the vicinity of Union Island and, previously, its internal stratigraphy and relation to the Great Slave Supergroup were uncertain.

The group was deposited on an irregular surface of Archean granitic rocks and is not cut by the mid-Aphebian diabase dyke swarm. Five mappable stratigraphic units are recognized (Fig. 25.2):

(1) The lower dolomite is unbedded and characteristically has complex vein systems of alternating quartz and radiaxial dolomite. Some veins are filled in part by clastic sediments. Ancient hills of granite, flanked by narrow aprons of fossil talus breccia, stick up into the dolomite and, in places, even into the overlying black shale. Beds of quartzite and quartz-pebble conglomerate occur in depressions on the granite surface at the base of the dolomite.

(2) The black shale is fissile, highly carbonaceous, and contains abundant iron sulphide nodules. In the lower part, there are many beds of black dolomite, commonly with shale rip-up clasts. In the upper part, there are quartzitic turbidites. Southwest of McDonald Lake, on the south shore of the East Arm, the black shale is intruded by two gabbro sills, the lower of which is glomero-porphyritic.

(3) The volcanic rocks, which do not everywhere occur at the same stratigraphic horizon and which are missing altogether north of Union Island, consist of wedge-shaped units amygdaloidal flows and flow breccia, pillow lavas and pillow breccia, coarse poorly bedded aquagene tuff, and laminated tuffaceous sediments. All are basaltic in composition, save for rare tuff beds which may contain more silicic pyroclasts. Many of the pillows have cement-filled centres, and were presumably lava tubes.

(4) The upper dolomite is distinguished from the lower by being well bedded and, for the most part, laminated. Slump breccia beds are common but sedimentary structures resulting from traction currents are rare. There are many reddish mudstone beds and, in places, the dolomite contains lenses of pyritic quartz-pebble conglomerate.

(5) The upper red and green mudstone is laminated, contains abundant slump scars, and has soft-sediment thrust and extension faults. There is no evidence of subaerial exposure.

The Union Island Group is overlain unconformably by the Sosan Group (Table 25.1). On an island northeast of Union Island, the basal Sosan is exposed in contact with the red and green mudstone unit. On north-central Union Island, although the contact is covered, the Sosan truncates the black shale and lower dolomite. On the south side of Simpson Island, across Hornby Channel from central Union Island, the basal Sosan overlies the upper dolomite, although here also the contact is covered. Nevertheless, we believe the stratigraphic position of the group to be established and it will be formally subdivided into formations.

Sosan Group: Facies Relations and Contemporaneous Block Faulting

The Sosan Group is formally subdivided into four formations, which occur in simple stratigraphic order at the type section north of Lac Duhamel, from the top:

(4) Akaitcho River Formation: red micaceous sandstone,
(3) Kluziai Formation: white to pink quartzite,
(2) Duhamel Formation: stromatolitic dolomite, and
(1) Hornby Channel Formation: pebbly feldspathic granulestone.

Locally, there is a newly discovered unit, possibly correlative with the Quadyuk Formation of the Goulburn Group at Bathurst Inlet (Campbell and Cecile, 1976), of laminated argillaceous dolomite that overlies the Akaitcho River Formation.

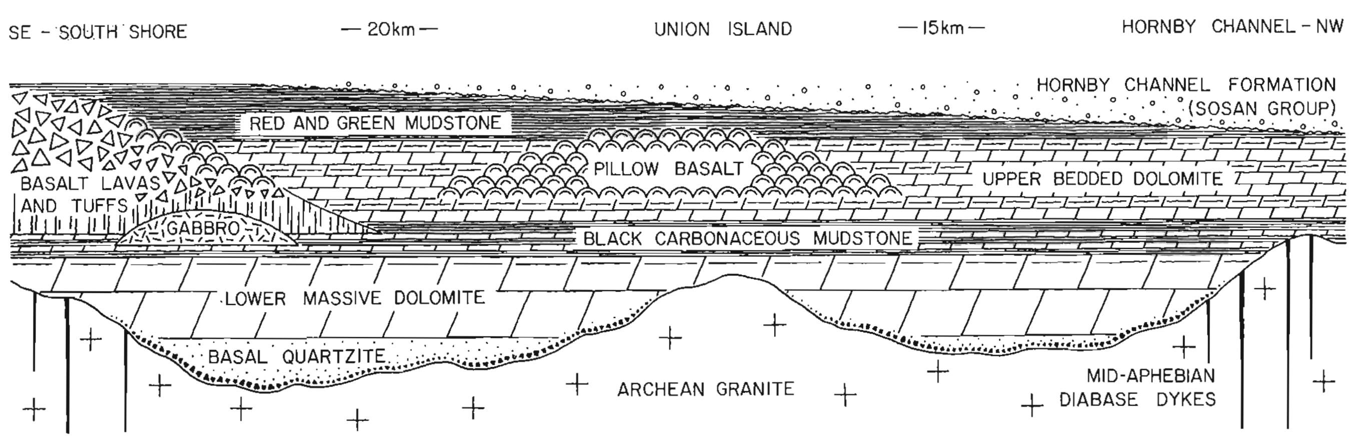

Figure 25.2. Stratigraphy and facies relations in the Union Island Group.

Table 25.1

Summary of Proterozoic sedimentation, magmatism and tectonics in the East Arm of Great Slave Lake

	diabase dykes (Mackenzie Swarm)	
	funnel-shaped gabbro intrusion	
Et-then Group	minor basalt	strike-slip faulting
	unconformity	
		folding
	diorite-monzonite laccoliths	
		movement of nappes
Christie Bay Group	basalt	
Stark megabreccia	minor basalt	
Pethei Group		
	gabbro intrusions	
Kahochella Group	basalt and rhyolite	
Sosan Group	basalt and felsic porphyry	block faulting
	gabbro intrusions	block faulting
Union Island Group	basalt	
	unconformity	
	diabase dyke swarm	
	biotite-bearing gabbro or diorite dyke	
		mylonitization
	adamellite stocks	metamorphism
Wilson Island Group	basalt and rhyolite	
	unconformity	
	Archean basement rocks	

Regional facies analysis is greatly complicated by four factors. Firstly, there is considerably more intertonguing of the principal facies that define each formation than is apparent in the type section. Kluziai-like tongues are common near the top of the Akaitcho River Formation, Hornby Channel-like tongues occur in the Kluziai Formation and *vice versa*, both Hornby Channel and Kluziai-like tongues occur in the Duhamel Formation, which is of limited areal distribution. The facies relations are especially complex around Charlton Bay, at the extreme northeast end of the East Arm, where the group as a whole is relatively thin. Nevertheless, it has been decided to retain the formational nomenclature and, so far as is possible at 1: 50 000 scale, to map the intertonguing relations. The intertonguing is a reflection of the complexity and instability of depositional environments that range from braided rivers (Hornby Channel facies), barrier islands and distributary-mouth bars (Kluziai facies), interdeltaic tidal flats (Duhamel facies), to barrier-protected lagoons and the nearshore open shelf (Akaitcho River facies).

Secondly, there are old faults, apparently active during Sosan deposition, across which there are abrupt changes in stratigraphy. For example, there is a northeast-trending line of volcanic centres between Lac Duhamel and McLean Bay north of which, the Hornby Channel Formation is more than 200 m thick. Only 2 km south of Lac Duhamel, however, on a ridge where Archean granite is exposed, the formation is as little as 2 m thick. South of the ridge, the formation is again relatively thick. The Duhamel Formation undergoes comparable changes in thickness over this paleotopographic high. Another old fault, also marked by a line of volcanic centres, occurs at Taltheilei Narrows and bounds a paleotopographic high east of the narrows (Fig. 25.3), against which the Sosan Group must abut in the subsurface. The complexities of the block-fault topography that probably existed during Sosan deposition, especially the lower Sosan, is only hinted at in the limited areas in which the group is now exposed.

Thirdly, although most of the Sosan exposures are autochthonous, the 20-km-long outcrop belt that passes between Bunting Lake and Moose Lake is a nappe (Fig. 25.4). A 2-km-long klippe of Sosan quartzite, possibly once part of the same nappe, rests on the Stark megabreccia south of Meridian Lake. The northwestward transport of these allochthonous masses must be considered in making paleogeographic reconstructions of the Sosan Group.

Fourthly, the effects of strike-slip displacement on the McDonald-Wilson Fault must be accounted for. Palinspastic restoration of the fault effectively moves the important Sosan exposures of the Hornby Channel region much closer to those of the type area around the Snowdrift River (Fig. 25.1). It also explains why the Sosan sediments around Seton Island, almost overwhelmed by volcanic products, are so different from the volcanic-free Sosan rocks located only a short distance away on the other side of the fault.

Kahochella Group and the Seton Volcanic Centres

The Kahochella Group is formally subdivided into three sedimentary formations, from the top:

(3) Charlton Bay Formation: green concretionary shale,
(2) McLeod Bay Formation: red concretionary shale, and
(1) Gibralter Formation: red shale and siltstone.

Volcanic rocks, occurring in both this and the Sosan Group, have collectively been referred to the Seton Formation. Most of the new information concerns the distinction in facies between allochthonous and autochthonous sections, and the areal and stratigraphic distribution of volcanic rocks.

The difference in sedimentary facies between the autochthonous rocks and those in the nappes is not great. In general, the McLeod Bay Formation is thinner in the nappes, and both the Charlton Bay and Gibralter formations are thicker and contain siltstone turbidites. The turbidites in the Gibralter Formation are especially numerous in the 25-km-long nappe that is best exposed between Hair Lake and Bunting Lake, near the east end of the aulacogen (Fig. 25.4).

The regional distribution of volcanic centres is significant (Fig. 25.3). Most are concentrated either along the northern margin of the axial trough of the aulacogen, or along the old fault line that bounds the basement high east of Taltheilei Narrows. Rocks of the axial trough itself, whether in the autochthon south of the McDonald-Wilson Fault or in the nappes to the north, are relatively deficient in volcanic products. In general, the volume of volcanic rocks decreases from west to east, that is, away from the mouth of the aulacogen, and there are no centres at all on the platform north of Christie Bay, east of Taltheilei Narrows.

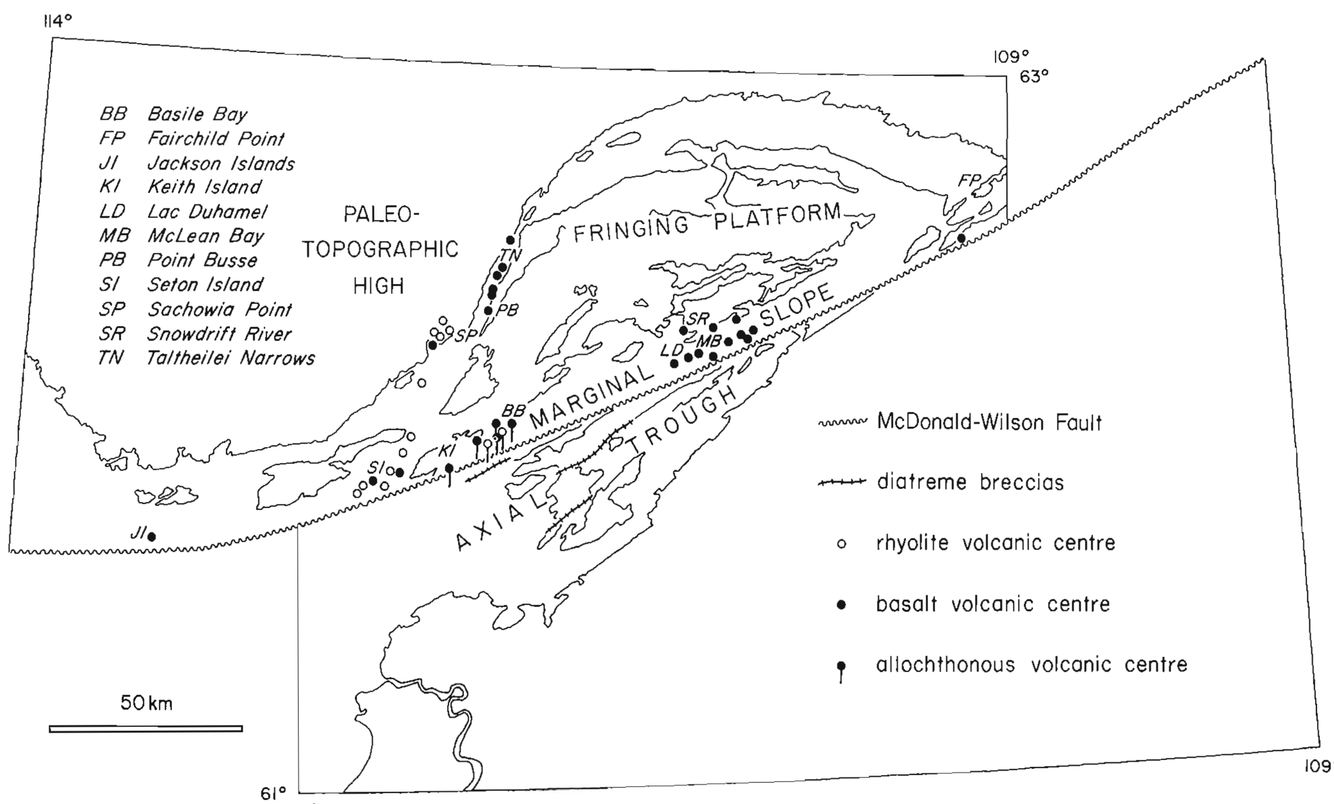

Figure 25.3. Regional distribution of Seton volcanic centres and their relation to major tectonic elements in the aulacogen.

Several clusters of volcanic centres are worth describing individually. In the Snowdrift River area (Fig. 25.3), there are no less than twelve volcanic breccia pipes and eight submarine pyroclastic cones, not to mention numerous tuffaceous beds of unknown derivation. All are basaltic in composition. Half of the pipes intrude the Akaitcho River Formation but their total stratigraphic range is from the Kluziai to McLeod Bay Formation. Of the cones, four occur in the Charlton Bay Formation, all located north of McLean Bay, three are in the lower part of the Gibralter Formation, and one occurs just above the base of the Akaitcho River Formation. Most of the cones are less than 1 km in diameter. A dark green concretionary shale, lithologically not unlike the Charlton Bay Formation, occurs at the base of the Gibralter Formation, and may be related to volcanism.

A line of six, possibly seven, basaltic breccia pipes occur between Taltheilei Narrows and Point Busse (Fig. 25.3). The pipes are irregular in detail and the southern four all intrude the Gibralter Formation. Basalt flows occur at the base of the formation and beds of lapilli tuff, some with large-scale cross-bedding, are associated with green shale and beds of red granular ironstone at the top of the formation. The pipes may well have fed the upper tuffs. The most northerly pipes, those on the west side of Taltheilei Narrows, are mineralized and very poorly exposed. They intrude the Akaitcho River Formation and, if they are the same age as those to the south, are more deeply eroded. This suggests that the southern pipes may be mineralized at depth.

A unique line of silicic, probably rhyolitic, centres extends from near Sachowia Point to Seton Island (Fig. 25.3). West of Sachowia Point, a complex of flow-banded rhyolite domes and intimately-related welded ash-flow tuffs cover an area of almost 35 km^2. The rhyolite overlies basalt flows, probably correlative with those near Taltheilei Narrows, and are intruded by basic dykes and sills. Basic and silicic rocks are also intimately associated on Seton Island and nearby islands, the thickest volcanic complex in the East Arm. Five funnel-shaped bodies of albite porphyry intrude an assemblage of basalt flows, tuffs and their erosion products, which are intercalated with sediments of the Kluziai and Akaitcho River formations. The felsic porphyry bodies also contributed detritus and must, therefore, have been unroofed during volcanism or had extrusive equivalents.

Nappes bearing volcanic rocks deposited in the axial trough of the aulacogen occur in a structurally complex belt between Keith Island and Basile Bay (Fig. 25.4). There are two major nappes, both of which contain locally thick piles of submarine volcanic rocks in the upper Gibralter and McLeod Bay formations. Amygdaloidal pillow basalts and pillow breccias are by

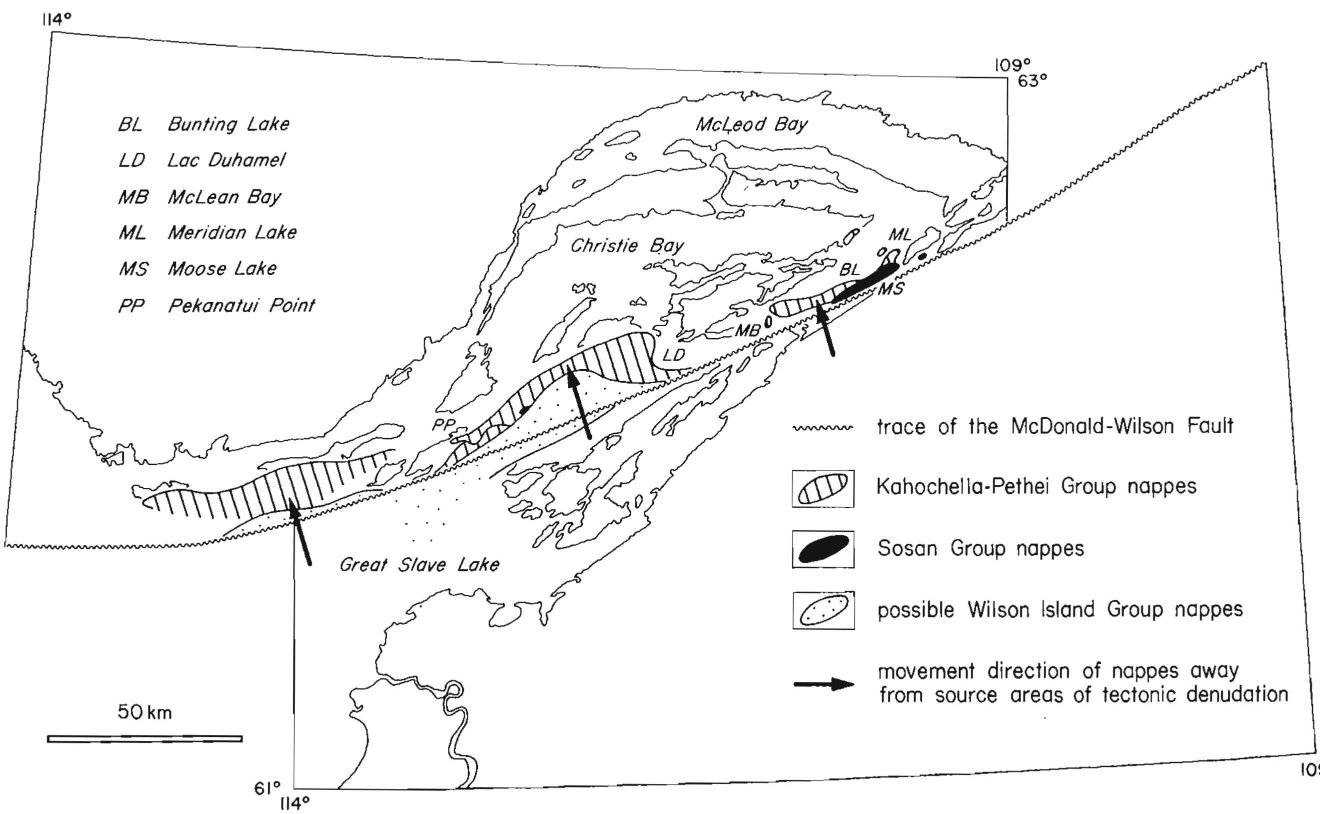

Figure 25.4. Distribution of major nappes in the East Arm and their source areas of tectonic denudation.

far the most voluminous volcanic rocks, but felsic, probably rhyolitic, domes and their erosion products also occur. Most of the shales in the vicinity of the volcanic piles are green, rather than the usual red, and beds of red granular ironstone, commonly associated with red and white chert lenses, are spatially related to the volcanic piles.

Intrusive gabbros, of irregular shape and variable texture, are closely associated with Seton volcanic rocks, both in the nappes and the autochthon. They are prominently exposed on the Jackson Islands, near the west end of the East Arm, where they intrude basalt flows, tuffs and a breccia pipe in the Duhamel and Kluziai formations; and on Fairchild Point, near the east end of the East Arm, where they intrude tuffaceous sandstones of the Kluziai and Akaitcho River formations. They are cut by many high-angle oblique-slip faults, which serves to distinguish them from the much younger dykes and sills of the Mackenzie Swarm. They constitute an important set of basic intrusions, probably closely related in age and origin to the Seton volcanic rocks.

Pethei Group and the Paleobathymetric Zonation of the Aulacogen

Facies analysis of this group led to the recognition of three paleobathymetric zones (Fig. 25.3), from the south:

(1) An axial trough, where tongues of greywacke turbidites (Blanchet Formation), derived from the west, are interbedded with deepwater stromatolitic mudstone (McLean Formation) and rhythmically laminated limestone (Pekanatui Point Formation),

(2) A marginal slope, where the McLean and Pekanatui Point formations occur without greywacke turbidites and are more calcareous than in the axial trough,

(3) A fringing platform, where two formations of shallow-water stromatolitic dolomite and limestone occur at the top of the marginal slope. The Taltheilei Formation, the older, occurs on the flanks of the old structural high east of Taltheilei Narrows but the Wildbread Formation, the younger, extends almost the entire length of the East Arm.

An appreciation of this zonation is essential in interpreting the relative movements of the various nappes. The largest nappes have transported Pethei of the axial trough northward onto the autochthonous rocks of the marginal slope. Smaller nappes of marginal slope facies have moved onto the fringing platform. Movement of the nappes has telescoped the facies changes, such that the axial trough was originally some 20 km south of the platform edge, not 10 km as previously reported. Mapping has also revealed that greywacke turbidites in the axial trough extend almost to Meridian Lake, 60 km farther to the northeast than previously known.

Stark Formation: Salt Solution-Collapse Megabreccia?

The Stark megabreccia, or olistostrome, is the most difficult formation in the East Arm to interpret, and is one of the most important because of the occurrence of a very similar megabreccia at a precisely correlative stratigraphic level in the Goulburn Group at Bathurst Inlet, 500 km to the north (Campbell and Cecile, 1976). The megabreccia rests sharply on the Pethei Group and consists of chaotically dispersed blocks of stromatolitic limestone and dolomite, containing abundant shallow-water sedimentary structures, in a brecciated matrix of red mudstone. The carbonate blocks, which are less than 50 m thick but range up to 1 km in length, are indigenous to the Stark Formation. None of the blocks is derived from the underlying Pethei Group, nor does the brecciation extend downwards into the Pethei. The carbonate of the blocks was intimately interbedded with the mudstone before brecciation. In a few places, notably 1.5 km south of the Snowdrift townsite, there are rare, large blocks of pillow basalt and basalt breccia.

The top of the megabreccia is generally indistinct and varies considerably in stratigraphic level. Overlying the main part of the megabreccia, that which contains the carbonate blocks, is an interval of thin bedded, red mudstone. The mudstone contains ripple marks and mudcracks, and becomes much more sandy toward the western end of the East Arm. In many places, notably in Tochatwi Bay, it contains numerous beds of carbonate-pebble conglomerate and granulestone, clearly derived from carbonate blocks in the underlying megabreccia, some of which project as pinnacles into the mudstone from below. The mudstone is overlain conformably by cross-bedded, lithic and feldspathic red sandstone of the Tochatwi Formation, which also contains beds of carbonate granulestone derived from blocks in the megabreccia. Both the mudstone and the sandstone are themselves extensively brecciated in places, notably at the east end of Tochatwi Bay and on the Caribou Islands.

It is difficult to measure the thickness of the megabreccia, but the Stark Formation as a whole ranges from about 500 to 700 m. There are no complete sections of the formation in which the megabreccia is not at least a prominent lower member. Thus, the extent of the megabreccia in the East Arm alone is 250 km.

Salt crystal casts occur both above and below the megabreccia. The uppermost bedding surface of the Pekanatui Point Formation, at the top of the Pethei Group, is littered with hopper-shaped salt casts and the red mudstones, above the megabreccia, contain highly skeletal salt casts, up to several centimetres in size, the poikilotopic nature of which suggests *in situ* precipitation from hypersaline groundwater.

Another unique feature is the presence of travertine-like coatings, up to 1 m thick, that were precipitated during brecciation. They are mostly well laminated

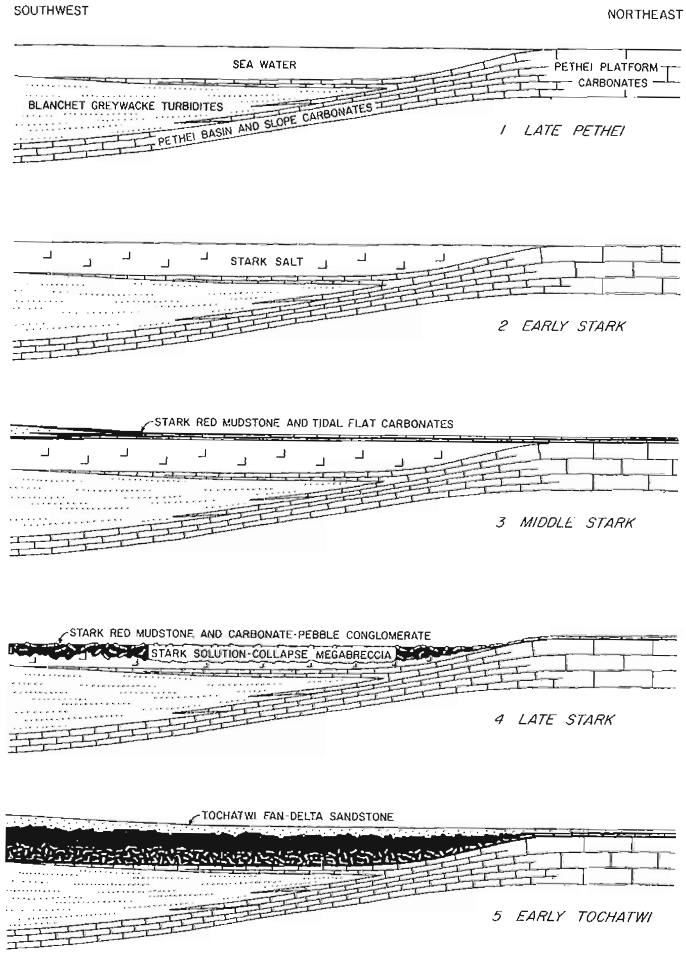

Figure 25.5. Genetic model for development of the Stark megabreccia by salt solution collapse.

and contain breccia layers that are draped by succeeding laminations. They are commonly riddled with small prismatic crystals of low-temperature quartz. The travertine occurs as detached blocks within the megabreccia, as discordant coatings that accreted downwards from the undersides of other blocks, and as a more or less continuous encrustation on the upper surface of the Pethei Group at the base of the megabreccia.

Origin of the megabreccia by salt solution-collapse was previously discounted because of the abundance of contorted and overturned carbonate blocks indicative of lateral transport by slumping. Much of this is probably related, however, to subsequent involvement of the megabreccia in nappe tectonics and the forceful intrusion of igneous laccoliths (*see* below). The absence of blocks derived from older formations, which make up the nappes, indicate that it is not a type of "precursory olistostrome" (*see* Elter and Trevisan, 1973) related to nappe formation. That the laccoliths are not ultimately responsible for the megabreccia is indicated by the fact that they do not control its distribution. Finally, neither nappe formation nor the intrusion of laccoliths, as possible causes of the megabreccia, can explain its occurrence in the Goulburn Group, whereas the deposition of salt may be a regional phenomenon.

According to the genetic model we now favour (Fig. 25.5), the end of Pethei carbonate deposition was marked by the precipitation of probably several tens of metres of salt, a thickness sufficient to nearly fill the axial trough of the aulacogen. Red mudstone and shallow-water carbonates of the Stark Formation were deposited on top of the salt in front of the alluvial Tochatwi fan-delta that was prograding into the East Arm from the west. Downward percolation of relatively fresh water from the fan-delta began to dissolve the salt at depth, resulting in subsidence by solution collapse in advance of the fan-delta. This produced a karst-like surface, on which detritus derived locally from tilted blocks of resistant carbonate was mixed with far-travelled sand and mud of the fan-delta. Solution of salt produced ephemeral caves, in which travertine was precipitated. As the caves were destroyed by downward collapse of the overlying sediments, hypersaline solutions were displaced upwards and, far in front of the fan-delta, precipitated poikilotopic salt crystals in the red mudstones above the main part of the megabreccia. As solution of salt and progradation of the fan-delta were contemporaneous, the upper Stark and lower Tochatwi were involved, to varying degrees, in the collapse breccias. Rotation of very large blocks during the final stages of solution collapse may be responsible for the aberrant paleomagnetic results obtained even from seemingly unbrecciated outcrops of these formations (*see* Bingham and Evans, 1976; Evans and Bingham, 1976).

Christie Bay Group: A new formation above the Pearson Basalt

This group consists of red nonmarine detrital sediments and subaerial basalt flows. Previously, the basalt flows, named Pearson Formation, were believed to be the youngest unit. On the south side of Tochatwi Bay, at latitude 62°33', longitude 110°14', about 200 m of buff feldspathic sandstone, with an interval of

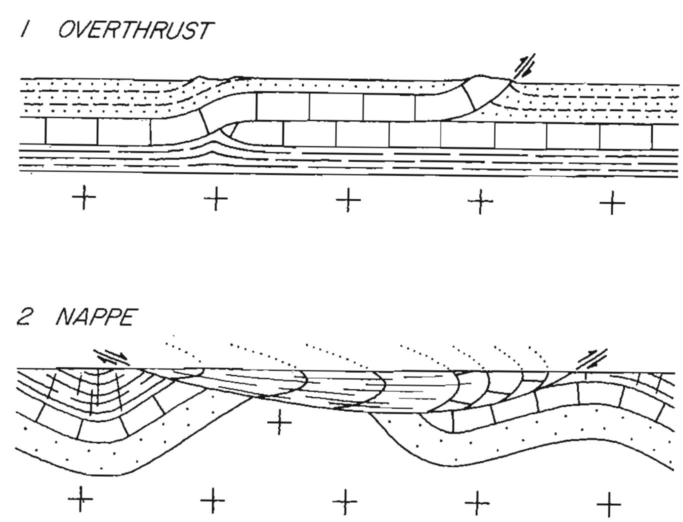

Figure 25.6. Distinction between a typical listric thrust fault and a nappe of the East Arm type.

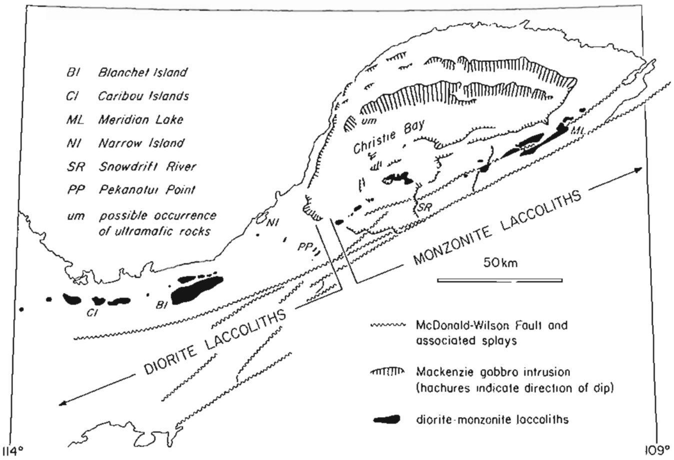

Figure 25.7. Distribution of diorite and monzonite laccoliths and the funnel-shaped gabbro intrusion centred at Christie Bay.

laminated mudstone at its base, overlies the Pearson Formation. This new formation may be correlative with the Amagok or uppermost Brown Sound Formation of the Goulburn Group at Bathurst Inlet (Campbell and Cecile, 1976).

Nappe Tectonics and the Tectonic Denudation of the Axial Trough of the Aulacogen

Nappes are the most spectacular structural feature of the East Arm, although their presence was unsuspected when mapping began. Huge rock masses, up to 70 km in length and 10 km in width, have been transported northwestward out of what was the axial trough of the aulacogen, now a zone of tectonic denudation. The transported rocks extend stratigraphically from Sosan Group to Stark Formation, and the Wilson Island Group may be involved as well. The nappes were emplaced before intrusion of the diorite-monzonite laccoliths, deposition of the Et-then Group, or strike-slip on the high-angle McDonald-Wilson Fault. There are important differences between the geometries of the East Arm nappes and such belts of listric thrust faults as occur, for example, in the foreland of the Coronation Geosyncline (Fig. 25.6). The nappes are unrooted at the present level of erosion, which means that they have a trailing edge as well as a leading edge. Rocks within the nappes are steeply homoclinal or recumbently folded, and their facing and vergence is invariably in the direction of transport. The detachment surface, on which the nappe has moved, is generally discordant with the strata both above and below. The nappes have either or both younger-over-older and older-over-younger relations with the rocks beneath them. In cases of younger-over-older superposition, the missing rocks are presumed to have been removed in an earlier nappe, moving in front of the nappe in question. This is unlike the situation in most listric thrust belts, where older-over-younger relations are the rule.

The three largest nappes (Fig. 25.4) are composed of Kahochella and Pethei Group rocks. The Pethei is of the axial trough facies, whereas that in the underlying autochthon is of the marginal slope. The Pethei generally occurs at the front of the nappes, where the crumpling is most severe, and the Kahochella trails behind. Folds are mostly recumbent, commonly overturned, and cleavage is gently dipping. This contrasts with the upright folds and weak high-angle cleavage in the autochthon. Of the three largest nappes, the central one is the most interesting. Near its eastern end, around Lac Duhamel, it transgresses autochthonous rocks ranging stratigraphically from Archean granite to Stark megabreccia. Near its western end, around Pekanatui Point, there is a stacked pair of nappes, each with a complete Kahochella-Pethei succession and both having internal thrust faults. Rocks both above and below the basal detachment surface of large nappes may be highly altered, and this zone is commonly complicated by many smaller nappes, torn from either the nappe or the autochthon. Intense alteration of autochthonous rocks north of McLean Bay may be related to a western extension, now removed by erosion, of the 35-km-long nappe between McLean Bay and Meridian Lake, the easternmost of the three largest nappes. Using reconstructions eliminating strike-slip on the McDonald-Wilson Fault system, it can be seen that the easternmost of the three largest nappes is derived from the region south of McDonald Lake, the central one from the Simpson Islands area, and the westernmost from the area covered by Great Slave Lake west of the Outpost Islands.

Overlying the Kahochella-Pethei nappes, and in places overlapping onto the autochthon, are a few smaller nappes made up of Sosan Group. The nappe that angles between Bunting Lake and Moose Lake is by far the largest of these (Fig. 25.4). It is at least 20 km in length, 27 km if the small klippe south of Meridian Lake is included, and contains a very thick section of the Hornby Channel Formation. Most of the other Sosan exposures, however, are autochthonous.

It is possible that the Wilson Island Group is allochthonous. This would help explain why it is nowhere seen in unfaulted contact with the Archean basement or the younger Sosan Group. If so, the fact that the grade of metamorphism is higher in the Wilson Island Group than in the adjacent Great Slave Supergroup cannot be used to determine the age of metamorphism. It is difficult to prove that the Wilson Island Group is involved in nappe tectonics because most of the critical contacts are either underwater or obscured by high-angle faults unrelated to nappe formation.

The age of nappe formation seems tightly constrained. Although the Stark megabreccia is the youngest formation carried in the nappes, their age is believed to post-date the entire Great Slave Supergroup. It seems improbable that the nappes could have been emplaced during Great Slave sedimentation and leave no evidence in the form of precursory fanglomerates or olistostromes. The Stark megabreccia itself contains no blocks of pre-Stark age and is very unlikely, therefore, to be derived from the nappes. The sandstones above the Stark are generally far too feldspathic to have a sedimentary

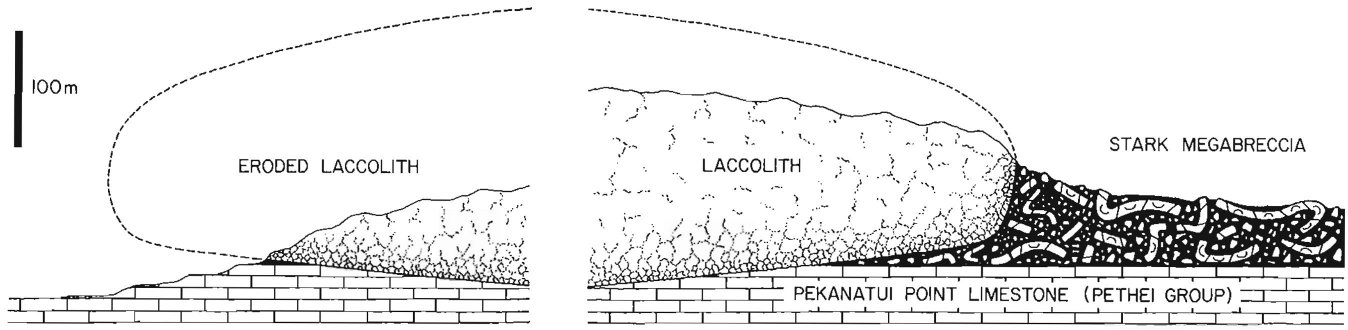

Figure 25.8. Intrusive relations of a typical East Arm laccolith, showing the distinction between a true lateral margin (right) and an eroded margin (left).

provenance as would be required if the nappes were being emplaced at that time. The minimum age of nappe development is constrained by the intrusion of diorite-monzonite laccoliths (*see* below), many of which override contacts between allochthonous and autochthonous rocks.

It is too soon for us to speculate on the mechanics of nappe formation, especially on the contentious issue of the extent to which the nappes were gravity driven. However, it can be concluded that they are fundamentally related to uplift and tectonic denudation of what had previously been the axial trough of the aulacogen.

Diorite-Monzonite Laccoliths: Composition, shape and age of intrusion

There is a string of more than twenty igneous intrusions, the largest 25 km in length, that extends almost the entire length of the East Arm (Fig. 25.7). Compositionally, there are two distinct sets. Those west of longitude 111°30' are hornblende-biotite diorites although, at their borders, they contain more acidic dykes and veins. Those to the east are mostly porphyritic monzonites with phenocrysts, in rapidly decreasing order of abundance, of plagioclase, hornblende and quartz. Both sets are compositionally and texturally similar to certain synvolcanic intrusions in the Great Bear Batholith (*see* Hoffman and McGlynn, in press), and may be about the same age.

The vast majority of intrusions are localized by the Pethei-Stark contact (Fig. 25.8). They tend to have flat bottoms that are nearly concordant with limestone of the Pekanatui Point Formation. In places, thin thrust slices of limestone have been driven out from beneath the laccoliths into the overlying Stark megabreccia. The laccoliths themselves have steep-walled lateral margins that abut against, and have pushed aside, the megabreccia. The tops of the laccoliths are rarely exposed but contain numerous large blocks of highly altered megabreccia. The margins of the laccoliths are characterized by autoclastic breccias and detached igneous blocks are locally abundant in the Stark megabreccia within a few metres of the contact. Such contacts are especially well exposed along the west shores of Blanchet Island and the Caribou Islands. The contrast between steep contacts with Stark megabreccia and flat contacts with Pethei limestone serve to distinguish the original lateral margins of laccoliths from merely erosional margins. The small sill-like diorite intrusions in the McLeod Bay and Charlton formations south of Pekanatui Point, and an intrusion in the Seton Formation on a solitary small island in Hearne Channel south of Narrow Island, are exceptional in their stratigraphic positions.

The fact that the laccoliths are so strongly localized stratigraphically tempts one to presume that they predate the nappes. However, several of the laccoliths override contacts between allochthonous and autochthonous rocks, thus necessitating that the laccoliths postdate nappe formation. This is especially well shown in the area west of Meridian Lake. In many areas, steep dips in the Pethei Group flatten beneath the laccoliths, suggesting that folding, as opposed to napping, continued after intrusion of the laccoliths.

McDonald-Wilson Fault: Strike-slip displacement and relation to Et-then sedimentation

A complex system of dextral strike-slip faults became active after intrusion of the laccoliths and was accompanied by nonmarine sedimentation of the Et-then Group. Although some of the old fault lines, active during Sosan sedimentation, were reactivated at this time, there is no evidence of strike-slip movement during Great Slave Supergroup deposition.

The bulk of strike-slip displacement occurred along a fault line that skirts the north shore of Wilson Island and Simpson Island, and the north shore of McDonald Lake, joining the McDonald Fault itself near the Snowdrift River (Fig. 25.8). The famous scarp on the south side of McDonald Lake is a splay fault that extends, west of the lake, through the main outcrop area of the Union Island Group. In common with other splays, it is probably an oblique-slip fault and has relatively little net strike-slip displacement. The only other exposures of Union Island rocks occur in windows through the Et-then Group south of McLean Bay. Matching these with the area west of McDonald Lake requires 70-90 km of strike-slip on the McDonald-Wilson Fault (Fig. 25.1). Displacement of this magnitude nicely resolves otherwise puzzling relations between dissimilar outcrop belts of the Sosan and Wilson Island groups (*see* above), and agrees with the more conservative of two estimates based on the matching of aeromagnetic anomalies northeast of the East Arm (Thomas *et al.*, 1976).

The massive Archean granites exposed along much of the length of the McDonald-Wilson Fault are chloritized but not penetratively deformed. The belt of mid-Aphebian mylonites (Fig. 25.1) is clearly truncated by, and not genetically related to the fault.

The alluvial fan sediments of the Et-then Group were deposited in down-dropped fault blocks during strike-slip movement. Although the conglomeratic Murky Formation is cut by the main fault and its splays, there are great changes in its thickness across the fault lines. Even more impressive is the way sandstones high in the Preble Formation overstep several splay faults, proving that movement on those splays ceased before the end of sedimentation. These relations are especially well shown north of McDonald Lake.

Basic Intrusions of the Mackenzie Swarm and the Christie Bay Gravity Anomaly

These are the only rocks not cut by strike-slip faults. Centred at Christie Bay is a tiered, crudely funnel-shaped, intrusion of gabbro, 125 km in maximum diameter (Fig. 25.8). The intrusion takes the form of anastomosing sills on the north side and inward-dipping dykes elsewhere. Deeply weathered outcrops of a possibly ultramafic phase of the intrusion occur on Pethei Peninsula at latitude 62°37'30", longitude 111°16'. If ultramafic rocks occur more extensively at depth, they may account for the 30 mgal positive gravity anomaly associated with the intrusion (Hornal and Boyd, 1972). The intrusion is cut by northwest-trending dykes of the Mackenzie Swarm.

Regional Tectonics and implications for the Aulacogen Hypothesis

The original reasons for hypothesizing the existence of an aulacogen – the longitudinal paleocurrents, the unusual ubiquity of volcanic rocks, the evidence of early rifting in the form of old faults and dyke swarms, and the stratigraphic correlations with other basins in the foreland of the Coronation Geosyncline – have all been strengthened by the new mapping. Nevertheless, certain newly discovered features may require modification of earlier interpretations of the paleogeography and time of initial rifting of the aulacogen.

Discovery of the nappes led to the realization that the axial trough of the aulacogen originally lay south of the McDonald-Wilson Fault (Fig. 25.3), perhaps coincident with the densest part of the mid-Aphebian dyke swarm. This means that the aulacogen was a somewhat larger structure than previously shown and that the East Arm is merely part of its northern margin. The nappes were presumably triggered by uplift south of the McDonald-Wilson Fault but whether this uplift is genetically related to the aulacogen or a manifestation of the so-called Hudsonian Orogeny that affected the entire Churchill Province is not known. However, medial uplifts of this type seem to be a common feature of old aulacogens, those in the Wichita Aulacogen of Oklahoma (Walper, 1976) being especially well known.

An important aspect of the aulacogen hypothesis is that it be cogenetic with the Coronation Geosyncline. The interpretation of the Wilson Island Group as being the result of a first stage of rifting in the East Arm complicates the question of the origin of the aulacogen. No rocks correlative with the Wilson Island Group are known in the geosyncline and, therefore, cogenecity cannot be argued until the second stage of rifting in the East Arm, that beginning with the mid-Aphebian dyke swarm. Whether the Wilson Island Group is truly part of the aulacogen, or the product of an entirely earlier cycle of rifting that merely provided a zone of crustal weakness later occupied by the aulacogen, is hard to say, especially so as the structural position and age of metamorphism of the Wilson Island Group is uncertain.

The possibility that the diorite-monzonite laccoliths may be related to the Great Bear Batholith has interesting implications. The batholith (*see* Hoffman and McGlynn, in press), is believed to have been generated above an east-dipping subduction zone at the old continental margin. Are the laccoliths what happens when a subduction zone passes beneath an aulacogen?

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