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PROTEROZOIC REDBED SEQUENCES OF CANADA

F.W. Chandler

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Critical Readers

D.F. Sangster
J.C. McGlynn

Preface

Redbeds are clastic sedimentary rocks stained red by hematite. They offer important information concerning past climates and depositional and tectonic environments, their paleomagnetic properties aid in the reconstruction of past plate motions, and – possibly of greatest significance – deposits of base metals, uranium and vanadium occur closely associated with redbeds in many parts of the world. By bringing together information on redbeds the Geological Survey is providing a tool useful in mineral exploration and is thereby meeting one of its principal objectives – determining the mineral and energy resources available to Canada.

D.J. McLaren
Director General
Geological Survey of Canada

Ottawa, September 1978

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PROTEROZOIC REDBED SEQUENCES OF CANADA

Abstract

Redbeds occur throughout the Proterozoic of Canada from the Early Aphebian (about 2.3 Ga ago). They occur in a wide range of depositional environments both shallow marine and terrestrial and are present in the thick sequences of miogeoclines and associated aulacogens as well as in epicratonic marine sediments. Few redbeds occur in flysch-related sequences. Redbeds dominate some alluvial molasse and postorogenic taphrogenic successions. Copper mineralization is common in marginal-marine, evaporite-bearing strata. Uranium mineralization is less common in Canadian Proterozoic redbed sequences and is related to alluvial redbeds and the granitic basement. This distribution may be due to the mobility of copper as chloride complexes, and to the lack of land plant-derived reducing environments in Proterozoic alluvial rocks.

Résumé

Les redbeds sont présents partout dans le Protérozoïque du Canada à partir. l'Aphébien inférieur (2.3 Ga environ). On les trouve dans des milieux géologiques très variés, continentaux, et de mer peu profonde, dans les séries épaisses des miogéosynclinaux et les aulacogènes qui leur sont associés, ainsi que dans les sédiments marins épicratoniques. Quelques bancs rouges sont présents dans les successions de flyschs apparentés. Les redbeds dominent dans la molasse alluviale et les séries taphrogéniques postorogéniques. La minéralisation en cuivre est normale dans les couches marines peu profondes à évaporites. La minéralisation en uranium est moins importante dans la série des redbeds du Protérozoïque canadien, elle est associée aux redbeds alluviaux et au socle granitique. Cette distribution est peut être due à la mobilité du cuivre sous forme de complexes de chlorure, et, à l'absence de milieux réducteurs créés par la décomposition de débris végétaux terrestres, dans les terrains alluviaux protérozoïques.

INTRODUCTION AND ACKNOWLEDGMENTS

The term redbeds as used herein signifies clastic sedimentary rocks mostly stained red by hematite, whereas a redbed-bearing sequence may contain merely a small component of redbeds. "Redbed sequence" is a broad term including one or both of the above. Though this report focuses on redbeds, redbed-bearing sequences are also discussed because the writer feels that redbeds and their mineralization are best understood if described within their geological context.

Redbeds are of interest for three reasons. Firstly, what is their paleoenvironmental significance? What do they tell us if anything about climate, and depositional and tectonic environment? Secondly, paleomagnetic studies involving the hematite of redbeds aid reconstruction of past plate motions. Thirdly, elsewhere in the world deposits of base metals (copper, lead, zinc), uranium and vanadium occur in or closely associated with redbeds but are not well represented in this country. For these reasons it was decided to collect and publish introductory information on Canadian redbeds.

The work, which is divided into an introductory part and a descriptive part, has been compiled mainly from a selection of the more important and up to date papers in the scientific literature and from federal and provincial government geological reports. In the first part current opinion on the formation of atmospheric oxygen, on the genesis of redbeds and on the mineralization associated with them is summarized. Attention is drawn to the lack of terrestrial vegetation in the Proterozoic and its effect on the redbeds of the time and their mineralization.

In the second part short descriptions of Canadian Proterozoic redbed sequences are presented. They are grouped according to their age and broad tectonic and sedimentary environments inasmuch as the uneven data that are available permit. Features of economic interest such as associated evaporites, reducing environments, unconformities on granitic rocks and evidence of mineralization are listed where available.

This report has been improved by the contributions of many federal and provincial geologists in particular by the critical reading of J.C. McGlynn who possesses an extensive knowledge of Canadian Proterozoic supracrustal rocks and by that of J.M. Franklin and D.F. Sangster who lent their wide experience in the field of economic geology. Jane Crandall, a summer student at the Geological Survey of Canada in 1976, was of considerable help in literature search and preparation of figures.

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GEOLOGY OF REDBEDS

Evolution of the atmosphere, the first redbeds

There is no evidence of redbeds having formed during the Archean (Cloud, 1976), though as the information in this paper shows they are very abundant among Proterozoic rocks. Since redbeds are mainly terrestrial sedimentary rocks which form under the influence of an oxygen-bearing atmosphere, it is instructive to review opinion on the development of the atmosphere and its effects on the sedimentary record such as the formation of extensive banded iron formation and sulphate-evaporites and later development of terrestrial vascular plants.

The earth's atmosphere contains too little of the noble gases relative to their cosmic abundances for it to have accumulated with the bodies that accreted to form the earth (Schidlowski, 1976). The earliest atmosphere, formed from volcanic emanations in equilibrium with a mantle containing metallic iron, was likely very reducing and probably contained much methane and hydrogen. After the metallic iron was removed from the mantle, volcanic emanations became less reducing and contained nitrogen and lesser amounts of carbon dioxide and water but still no oxygen (Holland, 1962). Thus atmospheric oxygen would seem to be of secondary origin. Likely mechanisms of formation include dissociation of water vapour under the influence of sunlight (photolysis), and organic photosynthesis, the latter involving formation of carbohydrates from carbon dioxide and water with the release of oxygen. Since carbon participates in the second phase of these two reactions only, rough equivalence between the amounts of organic carbon in the sedimentary record and the amount of oxygen in sedimentary rocks, the hydrosphere, and atmosphere suggest that photosynthesis was and is the more important process in generating the oxygen of the atmosphere (Rubey, 1955).

The formation of the widespread, delicate, iron-rich laminae in Archean banded iron formation is difficult to explain if the Archean atmosphere was oxidizing. Episodic marine photosynthesis would however provide widespread pulses of oxygen formation that would precipitate ferrous iron from the Archean sea (Cloud, 1976). The iron would act as a convenient scavenger for the oxygen, considered toxic for the primitive organisms of the time. Later acquisition of certain enzymes enabled these organisms to tolerate higher ambient oxygen levels. This resulted in a last burst of production of iron formation about 2 Ga ago. Following the filling of the oceanic oxygen sink, oxygen began to escape to the atmosphere (Cloud, 1972) and commence oxidation of iron in terrestrial sediments as well as allow oxidation of sulphur to sulphate (Cloud; 1968, 1972). Geological support for the timing of these events is found in the much cited detrital uraninite and pyrite in conglomerates of Blind River and the overlying redbeds of the Cobalt Group, all of early Aphebian age (Roscoe, 1973).

Absence of early sulphate evaporites (Cloud, 1968) and the low ^{34}S content of some Archean sedimentary barite suggest that the oceanic sulphate reservoir was not developed in the Archean (Perry et al., 1971). Given the relationship between redbeds and sulphate evaporites and the general tendency (see Rapitan Group, p. 20) of banded iron formation to precede redbeds in the geological record, one ought to be able to predict that a search for redbeds and associated evaporite-dependent (?) mineralization in Archean rocks would not be worthwhile. Also, undated metamorphosed sequences, in which it might be impossible to recognize redbeds because of the metamorphic destruction of hematite (Thompson, 1972) might be considered more likely sites for redbeds if banded iron formation is absent and sulphates (possibly in the form of scapolitic calc-silicates) are present for both of these rocks are refractory (Chandler, 1978b).

The above ideas should at present not be accepted uncritically. For example, Derry (1960) and Kimberley and Dimroth (1976) argued that the uraninite and pyrite at Blind River are not detrital but are diagenetic. Also Kimberley (1975) has used an actualistic aragonite-oolite replacement model to account for banded iron formation and to suggest that the oxygen content of the Archean atmosphere was not greatly different from the modern one. Further Serdyuchenko (1975) has documented sedimentary sulphate thought to be Archean.

Another consequence of the development of atmospheric oxygen was the arrival 0.7 to 0.4 Ga ago of vascular land plants. These would modify the chemistry of weathering and soil formation and the nature of fluvial regimes (Cloud, 1976). Especially relevant would be the tendency for the formation of terrestrial reducing environments. Sediment binding by plants would enhance the tendency of rivers to meander rather than braid (Schumm, 1968). These two effects increase the likelihood of intimate association of reduced and oxidized environments within fluvial sediments (Friend, 1966; Turner, 1974).

Genesis of redbeds

The long and controversial history of the subject of redbed genesis has been dealt with in some detail by Glennie (1970), Krynine (1950), Van Houten (1961, 1964, 1973) and Walker (1976) and briefly by Dunbar and Rodgers (1957) and Blatt et al. (1972). This introduction outlines the major hypotheses and discusses in more detail development of opinion in the last decade. Redbeds are essentially clastic sedimentary rocks, stained by interstitial iron hydroxide, chiefly hematite. Though mainly terrestrial, their colour may be preserved when red detritus is resedimented into the deep marine environment by turbidites (Lajoie and Chagnon, 1973; Ziegler and McKerrow, 1975). A redbed succession may contain up to 40 per cent of drab (i.e. not red) clastic units as well as carbonates and evaporites (Gary et al., 1972; Van Houten, 1964, 1973).

Until the last decade most redbeds were considered to be products of either the hot desert or the humid tropical environments. Van Houten (1961) reviewing the literature of 1834 to 1958 felt that European support of the desert hypothesis was influenced by knowledge of the Permo-Trias redbed evaporite sequence, whereas American support of the humid tropical hypothesis was in part due to the widespread occurrence (Krynine, 1950) of red lateritic soils in the humid tropics.

Evidence supporting the desert hypothesis included reports by early explorers of red desert sands, an opinion not nowadays universal; e.g. Glennie (1970). A desert environment would provide a low water table, heat, and limited vegetation, all thought to be factors favourable for the dehydration and aging of hydrated ferric oxides to hematite. This concept was supported by the many reports of association of evaporites with redbeds (Folk, 1976). Norris (1969) believed that dune sands redden with time in part due to their capacity to hold the sparse rainfall, but Folk (1976) held that the reddening of the dune sands of the Simpson Desert of Australia occurred during a Pleistocene fluvial phase, an opinion shared in the case of dune sands in general by Krynine (1950) and in the case of desert soils by Van Houten (1961). Walker (1967a) stated that red soils do form under very arid conditions but very slowly.

Krynine (1950) was impressed with the enormous area of lateritic soil in the humid tropics, attributing the colour of nearly all redbeds to hematite eroded from them. He explained the intimate association of unweathered detritus with this red, highly weathered debris in many redbeds by erosion of deep canyons into the fresh bedrock of source areas; the ideal depositional site being a well drained alluvial plain with alternating wet and dry seasons which permitted further oxidation by means of a seasonally fluctuating water table. He regarded inorganic evidence of dry climate, e.g. mudcracks and salt casts as over-emphasized.

Krynine (op. cit.) thought diagenetic redbeds volumetrically insignificant. Current experimental work required a high temperature for diagenetic formation of hematite. Such a nonselective process could hardly form interlayered the red and yellow strata observed by him amongst Cambrian strata. A bleached zone directly underlying a basalt flow in the Triassic redbeds of Connecticut was taken as proof of the diagenetic or pre-burial weathering origin of the hematite of these clastic rocks. Dunbar and Rodgers (1957) also believed in the derivation of hematite as detritus from laterites, and its deposition with the sediments of a wide range of sedimentary environments under both humid and arid climatic regimes; the reducing capacity of the environments determining whether the redness would be preserved.

Van Houten (1961) summarized the literature on redbeds published between 1830 and 1957. He too followed Krynine's (op. cit.) opinion, referring to it as "the prevailing view of the time", and considered it the only adequate explanation. Van Houten (op. cit.) classified redbeds according to their tectonic and climatic affiliation, noting that they were terrestrial clastic sediments associated with a great range of tectonic activity and with evidence of both humid (coal) and arid (thick evaporites) climate. He saw problems in the derivation of redbeds directly from lateritic soils, and wondered why the abundant kaolinite of laterites is so rare in redbeds, why modern tropical rivers carry yellow and brown mud instead of red mud, and why so little modern red alluvium is seen. He did not favour Krynine's (op. cit.) reddening by weathering of alluvium on the floodplain because of the rapidity of floodplain aggradation. He also did not accept reddening by "burial metamorphism" as he could see no correlation between burial and reddening, but concluded that diagenetic alteration of brown ferric oxide made some contribution to the formation of redbeds.

We are left at this stage with nobody having seen the Krynine model in action (Walker, 1967a). Though there was evidence of the alteration of goethite to hematite in laterites (Van Houten, 1964), there were theoretical arguments against the formation of hematite from goethite during weathering or diagenesis, for example below 130°C (Schmalz, 1959), below 170-181°C (Van Houten 1964) or at a very low activity of water (Schmalz, 1959).

In the late 1960s evidence began to accrue in support of the diagenetic origin of redbeds. Walker (1967a) documented intrastratal development of hematite from iron-bearing minerals, particularly hornblende and biotite, within a Pliocene to Holocene alluvial fan to tidal mud sequence in Baja California. Other reports of diagenetic hematite formation, from iron-bearing oxides, silicates and clays include those of Ali and Braithwaite (1977) Hubert and Reed (1978), Miller and Folk (1955), Schluger (1976), Turner (1974), Turner and Archer (1977), Van Houten (1968), and Walker (1967b). Walker (1967a) presented evidence that the regional climate of the depositional site in Baja California remained arid during the sedimentation and formation of the hematite, giving a low water table, alkaline oxidizing groundwater and sparse vegetation, all favourable for the diagenetic production and preservation of hematite. Walker (op. cit.) stressed that the climate prevented chemical weathering, allowing delivery to the depositional site of unstable iron-bearing rock forming silicates and that hematite formation occurred above and below the water table. Though desirable, aridity seems unessential, for Walker (1974) has also identified yellow and brown iron oxides in Recent and Pleistocene alluvium many tens of feet below the water table in the humid tropics in Puerto Rico. Since there the groundwater of the alluvium lay within the chemical stability field of hematite he concluded that redbeds also form diagenetically in the humid tropics. Berner (1971) considered hematite merely to be indicative of oxidative diagenesis, due to the absence of organic material, which can be brought about either by the absence of vegetation in a desert or by bacterial destruction of organic matter in moist tropical conditions.

Schmalz (1968) in discussing Walker's (1967a) paper, considered dehydration the key to formation of hematite from goethite, discounting the significance of Eh and pH and formation of hematite beneath the water table. He favoured Krynine's idea of hematite formation in the soils of the seasonally wet and dry tropical savanna since there, both the hot moist conditions for the weathering of the iron-bearing minerals to ferric hydroxides, and the dehydration environment necessary for hematite formation, would be alternately realized. Berner (1969) and Langmuir (1971) supporting Walker (1967a) used thermodynamic arguments to endorse the idea that fine grained goethite is unstable versus coarse grained hematite under practically all geological conditions, and will change to hematite under the conditions of shallow burial. Walker's (1974) suggestion that hematite will form beneath the water table is supported by Langmuir's (1971) suggestion that hematite will crystallize from water dissolving goethite. It is for kinetic reasons that goethite is more abundant than hematite among freshly precipitated iron oxides. Hematite usually forms by long aging of amorphous material or by dehydration, and for similar reasons, once formed, does not rehydrate to goethite (Langmuir, 1971).

Diagenetic reduction of redbeds, unrelated to the distribution of sedimentary facies has been documented by Thompson (1970). In general during diagenesis the formation or preservation of hematite depends on the maintenance of groundwater in the stability field of hematite (Walker, 1976). At geologically reasonable CO_2 pressures, solutions saturated with calcite and gypsum or anhydrate have fO_2 values well within the hematite stability field (Rich et al., 1977). This possible diagenetic source of redbeds may explain some redbed-evaporite associations.

In summary, the idea of widespread formation of detrital redbeds in the post-Devonian humid tropics seems unsatisfactory both because we cannot observe it nowadays and because of the reducing effect of the remains of the abundant vegetation of those regions. Widespread formation of detrital redbeds in modern deserts is also unlikely on account of the limited degree of chemical weathering. As Hubert and Reed (1978) pointed out the hydrated iron oxide stain on the surface of sand and mud particles reaching alluvial depositional environments in both wet and dry hot climates is more than enough to account for formation of diagenetic redbeds. Walker (1967a), supported by some very recent studies, has made a strong case for diagenetic formation of redbeds by alteration of iron-bearing oxides and silicates under hot desert conditions. Walker (1974) has also drawn attention to the possible diagenetic formation of hematite in Cenozoic sediments under hot humid climatic conditions. The role of bacteria, which may be important in removing organic remains under these conditions (Berner, 1971) needs further investigation.

Physical and chemical weathering and diagenesis of Proterozoic alluvial sediments in hot deserts was little different from now. Because of the absence of significant land vegetation in the Proterozoic, formation of hematite and its precursors in surface weathering and in diagenesis under wet climates, was much less restricted than at present. Under humid conditions whether or not red detritus was delivered to the depositional site would depend on local balance between very rapid rates of both chemical weathering and mechanical erosion. Nevertheless, it seems safe to conclude that in the Proterozoic under hot desert conditions redbed formation was similar to the present, but would be much enhanced under a hot humid climate.

Mineralization of redbeds

Deposits of copper, lead with minor zinc, uranium and vanadium occur in or with continental to marginal marine clastic rocks, many of which are red. These deposits have been discussed by Stanton (1972) and are reviewed in more detail in Wolf (1976). The copper deposits have been reviewed by Jacobsen (1975) and Tourtelot and Vine (1976). Canadian examples have been discussed by Kirkham (1974). Uranium deposits in sandstone have been reviewed by Adler (1974), Finch (1967) and Rackley (1976). Hydrothermal uranium deposits, now thought to have great geochemical similarities to the uranium deposits in sandstone have been reviewed by Rich et al. (1977). The purpose of this paper is not to review the literature dealing with these types of deposit, but to pinpoint aspects significant to a discussion of Proterozoic redbed sequences.

It is commonly held that before the development of free oxygen in the atmosphere it was possible, at the beginning of the Proterozoic and earlier, for detrital uraninite (Robertson, 1974) and copper sulphide (Jacobsen 1975) deposits to form. However according to Trow (1977) the formation of Blind River-type uranium deposits was not terminated by the "oxyatmoversion", for decrease of atmospheric CO_2 during major glaciations would depress the solubility of the uranyl dicarbonate complex, thereby allowing sedimentary transport of clastic uranyl oxide. But in general, with the accummulation of atmospheric oxygen and the dependent start of redbed formation these metals become mobile during weathering, in some sedimentary environments and in oxygen-bearing connate waters. This led to the formation of ores by removal of copper and uranium from oxidized solutions by redox reactions, as well as permitting formation of uranium-vanadium compounds such as carnotite in oxidized environments.

Copper deposits associated with sedimentary sequences are in many places intimately linked with redbeds (Jacobsen, 1975). Kirkham (1973) recognized at least two types of such deposit; the redbed type, formed in continental sequences and the Kupferschiefer type, formed in marginal marine-to paralic environments. The latter variety is of greater economic importance on account of its size. Evidence of the formation of redbed copper deposits in fluvial sequences is given by Woodward et al. (1974) and Dahl and Hagmaier (1974) among many others. Plant remains provide important reduction sites for fixing the redbed copper (Rose, 1976).

In the Kupferschiefer type of deposit, copper is disseminated in a very widespread thin pyritic or carbonaceous shale or marl, though other nearby sediments may be mineralized. The mineralized sediments are held generally to have been deposited in a shallow to marginal marine environment. In many cases they overlie red terrestrial clastics and are generally overlain by evaporites (Davidson, 1965; Kirkham, 1974; Renfro, 1974) and marine carbonates (Tourtelot and Vine, 1976). Though the type Kupferschiefer ores (Rentsch, 1974; Wedepohl, 1971) and those of the southwestern United States (Smith, 1974; Johnson, 1974; Johnson and Croy, 1976) are of Permian age, others in the United States (Clark, 1971; Harrison, 1972) in Australia (Rowlands, 1973) and in central Africa (Bartholomé, 1974) are Proterozoic. Hypotheses concerning the genesis of the Kupferschiefer type of deposit, reviewed by Jacobsen (1975), include replacement of earlier pyrite and iron-titanium oxides by copper sulphides, deposition of detrital copper sulphides (Binda, 1975), direct biologically induced precipitation of copper sulphides in stagnant water and precipitation of the metal in the reduced sediment after its derivation from terrestrial groundwater that has leached it from underlying redbeds (Renfro, 1974). Annels (1974) considered the copper to have been removed from marine shale.

Disseminated galena deposits with associated copper and zinc lie in continental clastics in France and Africa. The mineralized strata, themselves not red, tend to overlie granitic basement and may be associated with redbeds and evaporites (Caia, 1976). At L'Argentière, France, the lead was probably derived from the potassium feldspar of arkosic detritus during diagenesis and deposited by reduction at a geochemical interface between continental groundwater and that from an adjacent evaporitic lagoon (Samama, 1976). The African deposits may be of diagenetic origin. Deposition was as sulphides, the sulphur possibly derived from overlying evaporites (Caia, 1976).

Uranium deposits in sandstones in the western United States of mainly Mesozoic to Tertiary age (Harshman, 1974) occur in fluvial sequences derived from granitic rocks (Rackley, 1976). They are thought to have formed by the migration of neutral to alkaline (Harshman, 1974) oxygen-bearing (Adler, 1974) groundwater to reducing sites. The reducing or adsorbtion (Doi et al., 1975) sites were caused mainly by woody plant remains (Finch, 1967; Breger, 1974). Reduction was also brought about by pyrite or hydrogen sulphide derived from sulphate reduction or from petroleum. The boundary between the oxidized (hematitic and/or limonitic) sandstone and the reduced drab sandstone is an ore guide in the case of uranium roll-front deposits (Adler, 1970; Rubin, 1970).

Many Canadian vein-pitchblende deposits of Helikian age occur in fractures in granitic basement where it is overlain by red fluvial sedimentary rocks (Knipping, 1974; Robertson and Lattanzi, 1974). Rich et al. (1977) were impressed with the geochemical similarity between uranium deposits in sandstones and vein (hydrothermal) uranium deposits. They drew attention to the common field association of the latter class of deposit with competent felsic igneous and metamorphic rock and with host rocks containing ferrous silicates, pyrite or phosphate; their shallow (usually less than 300 m) vein filling habit; their simple uranium mineralogy (essentially pitchblende), wall rock alteration by hematite deposition, and gangue mineral paragenesis that implied decrease in oxidizing capacity of the ore depositing fluid during deposition. Fluid inclusion studies indicated that the ore-transporting fluids were of low to moderate salinity, contained at least one mole per cent of CO_2 and that deposition occurred at pressures beneath one kilobar* and 190°C.

These oxidizing solutions could have been of meteoric origin, having derived their oxygen from the atmosphere. Deposition of the ore could have been surficial (Robertson and Lattanzi, 1974; Barbier, 1974; Langford, 1977) or at depth. Alternatively hydrothermal solutions at depth, which are normally reducing could have passed through an oxidizing

* 1 kilobar = 10^5 kiloPascal (kPa)

aquifer and acquired the capacity to leach and transport uranium. The hematite and pyrolusite of redbeds would be a likely agent of oxidation. Further, at geologically reasonable CO_2 pressures solutions saturated with carbonate and gypsum or anhydrite have fO_2 values within the stability field of hematite.

In the absence of reducing environments uranium may be taken from solution in oxidizing environments by forming compounds, such as carnotite, with vanadium. Near exposed granite bodies in Australia recent calcrete in alluvial sequences contains economic quantities of carnotite (Langford, 1974; Premoli, 1976). Bearing in mind the common development of calcrete in alluvial sequences formed under climates with a dry season (Allen, 1974; Glennie, 1970) such mineral deposits should be expected in the more or less similar nonvegetated (Schumm, 1968) Proterozoic alluvial redbed sequences. In humid climates the chances of preservation of carnotite are reduced because of its solubility (V. Ruzicka, pers. comm., 1977).

A common thread running through the above discussion is the involvement in ore deposition of one or more of the following rock types; granite, redbeds and evaporites. A relevant point is that many of the Canadian Proterozoic redbed sequences overlie granitic areas of the Shield.

Although porphyry-copper deposits (Ahlfeld, 1967) and mafic igneous rocks are good sources of copper (see Seal Lake Group and Coppermine River Group, this report) the far greater abundance of granitic rocks (Wedepohl, 1974a) makes shield areas likely sources. Indeed many Proterozoic uranium occurrences in Canada are adjacent to granitic terrain of above average uranium content (Darnley et al. 1977). In granitic rocks biotite is an important source of copper (Wedepohl, 1974a) and zinc (Wedepohl, 1972), and potassium feldspar carries lead (Wedepohl, 1974b). A further source of zinc is magnetite, especially that of gabbroic rocks (Wedepohl, 1972). Uranium may be derived from granitic rocks (Barbier, 1974; Doi et al., 1975; Stuckless and Ferreira, 1976) where it occurs in accessory heavy minerals, themselves often present as inclusions in biotite and hornblende. It is easily leached if present interstitially (Rich et al., 1977).

Since granitic rocks provide the necessary elements for the metal deposits under discussion, arkosic debris formed from them might be thought an equally good source. Indeed some writers view the leaching of redbeds as a means of providing the lead (Helgeson, 1967), uranium (Dahl and Hagmaier, 1974) and copper (Hartmann, 1963). However uranium is very mobile in the early stages of granite weathering (Barbier, 1974). Uranium depletion from granite outcrop beneath the zone of apparent weathering has been demonstrated to a depth of 50 m in drill core (Rosholt et al., 1973) and to a depth of 390 m (Stuckless and Ferreira, 1976). Therefore by the time the bedrock is disaggregated much of the uranium should be removed. Stuckless and Ferreira (1976) state that the volcanic detritus associated with the sandstone uranium deposits of the western United States is too fresh to have been a source of the uranium. Unless weathering is very severe potassium feldspar survives as mineral grains in feldspathic sandstone, and its destruction by vigorous weathering in alluvium (Samama, 1976) or by hot subsurface chloride brines will furnish adequate lead for ore deposits (Helgeson, 1967).

Both biotite, which contains copper (Wedepohl, 1974a), and magnetite and ilmenite, which contain vanadium (Wedepohl, 1974c) usually survive sedimentary transport, but during rigorous weathering, that is laterization, these minerals would be destroyed and release their copper and vanadium. A part of these metals would be adsorbed onto newly formed ferric hydroxides (Hem and Skougstad, 1960) and incorporated into clays (Landergren, 1974; Wedepohl, 1974 a, c) and after transport could incorporate in detrital redbeds. But if the formation of a sequence of redbeds is diagenetic, (see Walker, 1976), that is, if the iron-bearing minerals are destroyed within the sediment or rock by circulating oxygen-bearing water, less of the metals is lost by surface run off. For this reason Tourtelot and Vine (1976) and Vine and Tourtelot (1976) regarded diagenetically formed redbeds to be more likely to contain copper deposits than detrital redbeds. The vanadium would be available to participate in the formation of uranium-vanadium ores. As ferric oxides formed from the destruction of iron-bearing minerals age to hematite, part of the adsorbed metals is released (Doi et al., 1975). The retention of these metals is similarly favoured by the diagenetic as opposed to the surface environment.

Groundwater in redbeds, particularly arkosic sandstones, is expected to be neutral to alkaline (Hagmaier, 1971) and to have an intermediate to oxidizing Eh (Rose, 1976). Under these conditions, that is in equilibrium with hematite, fresh or dilute groundwater is a poor transporter of copper, but the metal forms stable chloride complexes. At 25°C a chloride solution of greater than 0.01 M can be an effective copper solvent and tends to leach the metal from rock (Rose, 1976). Chloride brines are effective solvents also for lead and zinc (Helgeson, 1964; White, 1968). This is of considerable interest in view of the very common association of copper and lead deposits in sandstones with evaporites (Caia, 1976; Samama, 1976) and in view of the occurrence of modern copper-rich (Rickard, 1974, Tourtelot and Vine, 1976) and lead-rich (Carpenter et al., 1974) natural brines, some of which are escaping from redbeds. Uranium transport in hydrothermal solutions (sensu lato) is enhanced by the formation of carbonate and sulphate complexes. In hydrothermal solutions sulphide complexing is unimportant in the transport of base metals (Helgeson, 1964).

Plant remains are regarded as providing the reducing environments that fix many copper and uranium deposits in fluvial, that is terrestrial, sediments. Since land plants probably did not become established before the Devonian, the occurrence of such deposits might appear unlikely in older, especially Proterozoic, redbed sequences. However carbon has been found in the uraniferous shale at the base of the Athabasca Formation at Cluff Lake, Saskatchewan (Langford, 1977) and stromatolites have been found in Proterozoic rocks interpreted to be of terrestrial origin (Hoffman, 1976). In the context of Proterozoic terrestrial reducing sites it is important that ferrous iron-bearing rocks, not of organic origin may act as the reducing agent in hydrothermal uranium ore deposition. Examples include the biotite lamprophyre minette, chloritic pelite and perhaps hornblende gneiss (Rich et al., 1977). The ferrous iron content of basement rocks beneath alluvial sequences makes vein-type uranium deposits at unconformities relatively more important in the Proterozoic than in the post-Devonian.

DESCRIPTIONS OF PROTEROZOIC REDBED SEQUENCES IN CANADA

Aphebian miogeoclinal and associated taphrogenic sequences of the Superior Craton

Several Aphebian sequences, which formed as parts of miogeoclines on the Archean Superior Craton, contain redbeds. They include the early Aphebian Huronian Supergroup, and the Knob Lake Group, the Belcher Group and the Richmond Gulf sequence; the last three, forming parts of the Circum-Ungava Geosyncline. The Chakonipau Formation of the Knob Lake Group and the Richmond Gulf Formation are taphrogenic in as much as they are related to possible failed arms of the geosyncline. The Missi Group, overlying Amisk Group volcanics of Archean or Aphebian age does not belong within this category but has been included on account of its probable Aphebian age and because of the probable Archean age (Davidson, 1972) of much of the surrounding gneiss-greenstone terrane.

Huronian Supergroup, Early Aphebian

The Huronian Supergroup (Fig. 1) comprises of twelve formations gathered into four groups (Fig. 1b). Several review papers cover Huronian geology. Frarey and Roscoe (1970) gave a brief account of the stratigraphy. Roscoe (1969), stressed the regional stratigraphy and uranium mineralization. Young (1973) collected several papers on specific aspects of Huronian geology, and Card et al. (1972) presented a comprehensive review of the Huronian within the context of the Southern Structural Province. The Huronian outcrops in the 60 km wide Penokean Fold Belt that extends about 300 km, east-west, along the north shore of Lake Huron. It is also exposed in the Cobalt Plate, a more or less equidimensional area northeast of Sudbury about 150 km across (Card et al., 1972). The Huronian nonconformably overlies Archean rocks that are exposed to the north and west. It is overlain with angular unconformity by Hadrynian and younger strata in the south and west and truncated by the Grenville Structural Province at its eastern margin (Ayres et al., 1970). The Huronian sedimentary sequence, about 12 000 m of mainly immature sediments, was derived from a granitic source in the north and formed a southward-thickening wedge on the southern flank of the Superior Craton. Deposition in the northern part of the Huronian outcrop area was mainly fluviodeltaic and in the south was mainly deltaic-marine. This is reflected by the prominence of sandstone units in the north and the marked thickening of volcanic and pelitic units in the south as well as the broadly south-directed paleocurrents (Card et al., 1972).

The lowest group, the Elliot Lake, contains the pyritic, arenaceous Matinenda Formation that hosts the famous Blind River uranium ore, which is presumed detrital and indicative of an atmosphere devoid of free oxygen (Roscoe, 1969; Ruzicka, 1976b). This, and the presence of redbeds in some overlying Huronian formations, led Roscoe (1969) to suggest that the record of oxygenation of the atmosphere occurred during deposition of the Huronian Supergroup.

The three succeeding groups are cylic, being composed of a basal paraconglomerate, possibly of glacial origin (Young and Chandler, 1968) followed by a fine grained clastic or carbonate formation of turbidite or shallow water, possibly tidal, origin and finally by a sandy unit possibly of fluvial origin (Frarey and Roscoe, 1970). The cyclic sedimentation may have been caused by glacial fluctuations in sea level (Casshyap, 1963).

Redbeds in the Huronian Supergroup are found in the Mississagi Formation in the upper part of the Gowganda Formation (Firstbrook formation, (sic) of Thomson, 1957) and in the Lorrain, Gordon Lake and Bar River formations.

The crossbedded, subarkosic Mississagi Formation is grey to light pink and contains minor fine grained clastics especially in the lower part. It may be of beach (Palonen, 1973) or fluvial-deltaic (Card et al., 1972) or fluvial (Long, 1978) origin. It is present through most of the Huronian region and thickens southeastward to a recorded maximum of 3000 m (Roscoe, 1969). The writer has extracted pyrite from the heavy mineral fraction of samples from the western part of the Penokean Fold Belt. Card (pers. comm.; 1975, 1976) noted however that the formation is cleaved and hematite-rich in the axial zone of the MacGregor Bay anticline near Big Inlet at the northeast corner of Manitoulin Island.

The Gowganda Formation is widespread through the Huronian outcrop area and is the lowest formation in much of the northern part. Its possible glacial origin has stimulated many studies (see references in Chandler (1969) and Young (1970)). Conglomerate is common in the lower part of the formation and greywacke and argillite in the upper part. This division, well defined in the Cobalt Plate, caused Thomson (1957) to divide the Gowganda Formation into a lower Coleman formation (sic) and an upper Firstbrook formation (sic). The Firstbrook formation, a red laminated argillite, outcrops west of the town of Cobalt (MacKean, 1968) and reaches an apparent thickness of 590 m in drill core but is absent to the east and southeast (Thomson, 1966). Red and purple sandstones and argillites have also been found in the formation north of Blind River (Siemiatkowska, 1976).

The Gowganda Formation is overlapped northward by the 1500-2500 m thick arenaceous Lorrain Formation which consists of a lower arkosic part, a middle hematitic and aluminous quartz-pebbly sandstone part, and an upper white quartz-arenite part. Arguments have been presented for a fluvial and transgressive-marine origin for the formation (Card et al., 1972). Near the village of Desbarats in the western part of the Penokean Fold Belt, Frarey (1962) mapped at the base of the formation a massive red arkose 610 m thick overlain by purple siltstone and fine grained quartzite up to 110 m thick. These units, according to Roscoe (1969) may be equivalent to Thomson's (1957) Firstbrook formation. North of Blind River Siemiatkowska (1976) mapped hematitic red sandstone both at the base and toward the top of the formation.

In the quartzose sandstone and the argillite of the lower part of the succeeding shallow water, possibly tidal, Gordon Lake Formation, nodular anhydrite and gypsum may be associated with chert. Above these sulphates intra-formational breccia is present in 40 m of varicoloured siltstone and argillite. Hematite ooliths (Wood, 1975) and red sandstone and siltstone (Siemiatkowska, 1976) occur in the formation.

The possibly littoral Bar River Formation conformably overlies the Gordon Lake Formation and is at least 900 m thick near the northeast corner of Manitoulin Island. There it consists of a 370 m thick lower member of white quartz arenite overlain by a 550 m thick member of interbedded sandstone, siltstone and argillite. Up to 5 per cent of hematite is locally present in the lower member. Lenses up to 30 m thick of interbedded argillite, siltstone and fine grained quartzite, locally containing up to 20 per cent of hematite, occur in the upper member (Card et al., 1972). North of Blind River Siemiatkowska (1976) reported red and purple quartz arenite in the lower part of the formation and Wood (1975) noted hematite concentration at certain stratigraphic levels in the Bar River Formation.

The distribution of Huronian lithofacies, the southeastward thickening, and the abrupt disappearance of the Huronian at the margin of the Grenville Structural Province suggested to Dietz and Holden (1966) that the Huronian originated as a miogeocline.

In the western part of the Penokean Fold Belt strata dip gently and metamorphism is incipient. In the eastern part folds are tight, strata are highly faulted and the almandine amphibolite grade of regional metamorphism was reached. Strata of the other structural domain, the Cobalt Plate, are flat-lying and their metamorphism is incipient except in the southwest where block faulting, gentle folding and metamorphism up to middle to upper greenschist facies have occurred (Card et al., 1972).

The nonconformable relation with Archean rocks and intrusion of diabase sills into Huronian strata at 2.16 Ga (Van Schmus, 1965) limit the time of Huronian sedimentation to early Aphebian time. This age is supported by an Rb/Sr isochron based on fine grained sediments from the Gowganda Formation which yielded an age of 2.3 Ga (Fairbairn et al., 1969).

Sedimentary copper mineralization occurs in the Lorrain Formation about 30 km north of Elliot Lake, near Stag Lake, and is found in white quartzite of the upper member (Wood, 1975). Another occurrence is in hematitic feldspathic sandstone in the lower part of the formation at Desbarats (Pearson, 1978). Copper mineralization is also commonly associated with the Nipissing diabase sills in the Huronian Supergroup.

Otish Group

The Otish Group (Fig. 1), (Chown and Caty, 1973), occupies an elongate basin which strikes N60°E and which has a present area of about 5000 km^2. It overlies a rugged and probably weathered Archean surface of the Superior Province and, is cut by dykes and sills of the Otish Gabbro. The group is divided into the Indicator Formation, up to 750 m thick, and the overlying Peribonca Formation, at least 380 m thick but which has an erosional top (Chown and Caty, 1973).

The Indicator Formation has been correlated with the lithologically identical Papaskwasati Formation and the Peribonca Formation with the lower or upper Albanel Formation both of the Mistassini Group that lies 55 km to the southwest. The Papaskwasati Formation is mostly pale and drab but secondary reddish colouring, partly transgressive, is common (Chown and Caty, 1973). The Otish Group has also been correlated with the Cobalt Group (Frarey and Roscoe, 1970; McGlynn, 1970b) and with the Sakami (p. 32), and Chakonipau (p. 9) formations (Chown and Caty, 1973).

The Indicator Formation is composed of drab conglomerate and sandstone (Eade, 1966) with minor red argillite. Some hematite staining does occur at all levels particularly in the upper part (Chown and Caty, 1973). The Peribonca Formation, distinguished from the Indicator Formation by its red and purple colour (Eade, 1966), contains a lower member of laminated, uniform, well-sorted, dolomite-cemented sandstone with minor conglomerate and dolomite, and an upper member of argillaceous sandstone and conglomerate. Minor dolomite beds, one algally laminated, occur in the upper part of the lower member. In the east part of the Otish Mountains the upper member is coarse red arkose with minor polymict pebble conglomerate and minor interbeds of red argillaceous sandstone. In the west it is a silty, massive, purple sandstone (Chown and Caty, 1973).

During sedimentation the Otish Basin was bordered by high ground on the north, particularly at its west end, and was open to the south and east. A small ridge separated the Otish Basin from that of the Papaskwasati Formation. The currents that deposited the Indicator Formation conglomerate travelled to the southwest, but those of the arenaceous rocks went northeastward. Those of Peribonca Formation sandstone flowed southwestward. Clasts in clonglomerates do not become smaller in a down-current direction, but instead large clasts are confined to narrow, southwest-trending zones that probably define paleochannels. The general trend of sedimentary transport across paleoslope is most likely a function of the depositional environment (Chown and Caty, 1973).

The Otish Basin has been little deformed (Chown and Caty, 1973). Strata dip gently southeastward at 3 to 10 degrees and steepen southward. Near four parallel faults that strike north to east at the southeastern boundary of the formation, the rocks are much sheared and brecciated and dip steeply. Metamorphism is limited to development of some white mica in argillite (Eade, 1966).

Dykes cutting the probably correlative Papaskwasati Formation, have yielded K/A ages of 1.96 and 1.93 Ga from their baked margins (Fahrig in Wanless et al., 1968; in Wanless et al., 1972). Fryer (1972) derived a Rb/Sr age of 1.787 Ga from the Temiscamie Formation which overlies the probable correlative of the Peribonca Formation in the Mistassini Basin (Chown and Caty, 1973). A paleomagnetic pole position from the Otish Gabbro lies between those of the Nipissing diabase dated at 2165 Ga (Van Schmus, 1965) and the Sudbury Norite dated at 1.88 Ga (Hurst, 1975).

Knob Lake Group (Fig. 1)

The Labrador Trough is a Hudsonian fold belt, composed of Late Aphebian (Davidson, 1972) sedimentary and volcanic rocks, that forms part of the Circum-Ungava Geosyncline and extends for about 1000 km north-northwest between the Superior and the Churchill Structural Provinces in New Quebec and Labrador (Dimroth et al., 1970). The geology of the Circum-Ungava Geosyncline has been reviewed by Davidson (1972) and Dimroth et al. (1970). Dimroth (1970) has reviewed the geology of the Labrador Trough and Dimroth (1968), the central part of the Trough. Much of the Labrador Trough has been mapped by the Geological Survey of Canada on a reconnaissance scale (Frarey and Duffell, 1964; Baragar, 1967b). Nomenclature of the sedimentary rocks of the Labrador Trough has been discussed by Frarey and Duffell (1964) and by Dimroth (1968).

The sedimentary and volcanic rocks of the trough, the Kaniapiskau Supergroup, have been divided into two lithotectonic zones. One comprises the central and western parts of the Trough, and is composed of the Knob Lake Group, mainly of sedimentary rocks. The other comprises the eastern part of the Trough and is composed of the later Doublet Group, predominantly of volcanic rocks (Dimroth, 1968; Greene, 1974). Sedimentation in the western or miogeosynclinal zone happened in two, possibly three cycles. Each cycle began with an orthoquartzite-carbonate sequence and terminated with a flysch-like shale and greywacke sequence (Dimroth, 1970). The first cycle is found as far north as the central part of the Trough, approximately 50°30'N. Rocks of the second cycle overlie those of the first cycle in the southern part of the Trough, but north of 57°N overlie crystalline basement and are very thick. Parts of a third cycle have been recognized in the east-central and north parts of the Trough but the overall picture has been obscured by erosion and other factors (Dimroth, 1970).

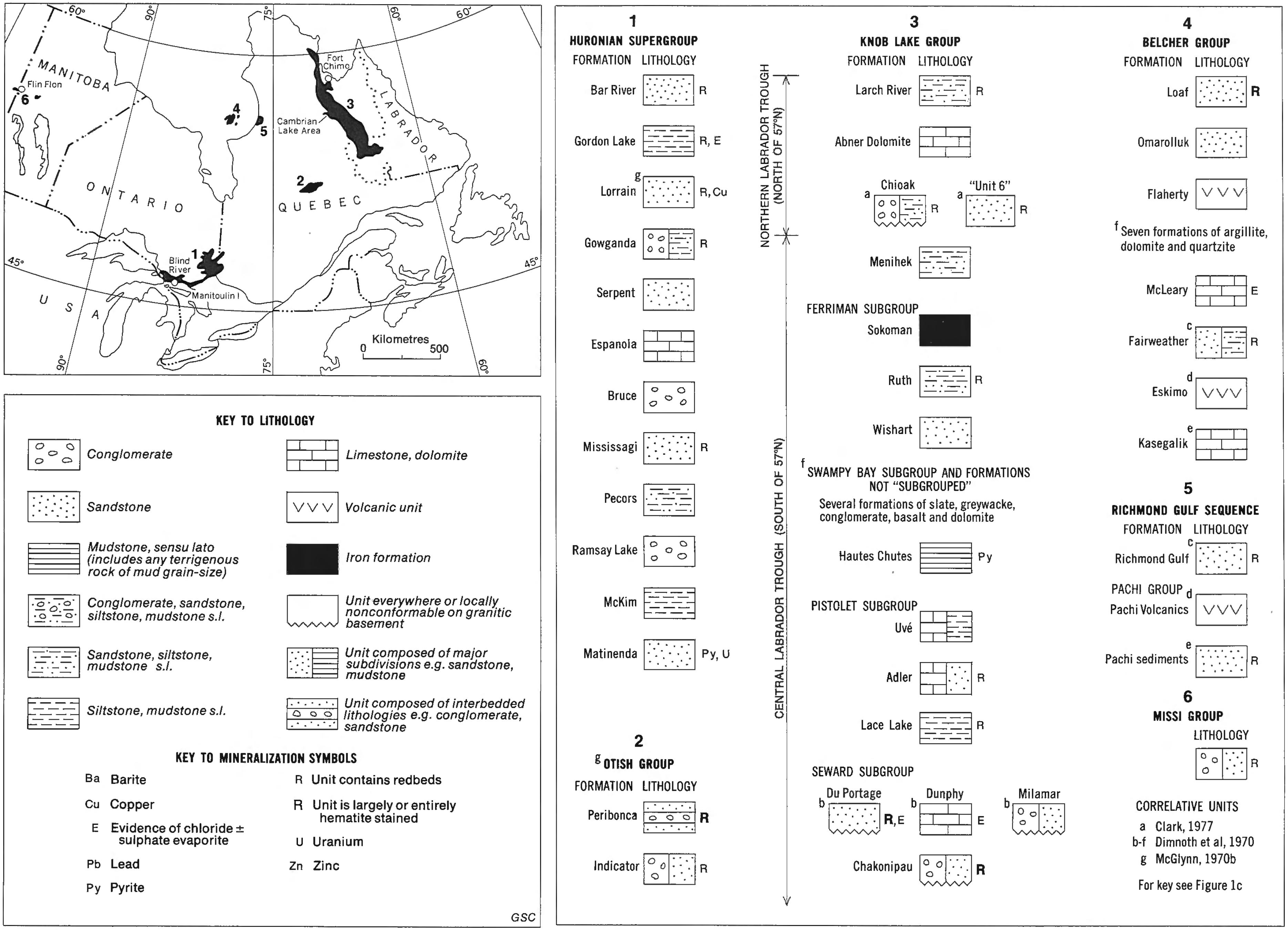


Figure 1. Aphebian miogeoclinal and associated taphrogenic sequences of the Superior Craton.

Redbeds in the central and southern parts of the Labrador Trough have been found in the lower part of the first cycle. They have been assigned to the Knob Lake Group and include the Chakonipau Formation and the overlying Du Portage Formation, both of the basal Seward Subgroup. Redbeds also occur in the Lace Lake and Adler formations, both of the overlying Pistolet Subgroup (Dimroth, 1968). The regional facies variations in the first cycie have been described in detail by Dimroth (1968). In the northern part of the Labrador Trough red clastics are found in the Ruth Formation and in the Chioak Formation and its possible correlative "unit 6" (Clark, 1977).

Redbeds of the Chakonipau Formation occur in the central part of the Trough and also underlie a westward-projecting depression in the Archean surface at about 56°30'N. This region, the Cambrian Lake area, has been mapped on a reconnaissance scale by Fahrig (1956) and Roscoe (1957) and the eastern part mapped in detail by Dimroth (1968, 1969). In the western part of this depression the rocks of the Chakonipau Formation overlie Archean granitic rocks. They are also ten times thicker (1250 m) than on the north and south margins of the depression over a distance of about 80 km (Fahrig, 1956) and have been cut by northeast and northwest-striking faults. The redbeds of the Chakonipau Formation have been likened by Dimroth (1968) to the Old Red Sandstone of Great Britan and the Rothliegendes of Germany. He distinguished a western and eastern facies lying west and east of Otelnuk Lake. In the west very fine grained arkose is overlain by interstratified arkosic pebbly conglomerate, arkosic grit and coarse and fine grained arkose. All clasts are hematite stained and some andesite fragments possess a hematite weathering crust. These rocks are interpreted as deposited rapidly in an oxidizing continental environment. A local southward facies change in the western facies led Dimroth (1968, p. 49) to infer the presence of an east-striking fault, the Cambrian-Otelnuk Fault, the granitic rocks to the south of which were a local source area and shed coarse clastics into the depression to the north. In the east the Chakonipau Formation is similar to the rocks in the western facies above the conglomerate. Brief marine transgressions in the eastern district are inferred in part from beds of stromatolitic dolomite in the redbeds.

The Du Portage, Dunphy and Milamar formations comprise the upper part of the Seward Subgroup and are mutual lateral facies equivalents. In the west the Du Portage Formation, over 300 m thick, is predominantly fine grained red arkose and subarkose and contains some beds of stromatolitic dolomite and calcareous orthoquartzite. To the east of the Du Portage Formation lies the Dunphy Formation, 350-600 m thick, of pink stromatolitic dolomite with minor fine grained arkosic sandstone. Gypsum has been noted in rocks of the Du Portage and Dunphy formations (Dressler, 1976). To the east of the Dunphy, coarse grained arkose and arkosic conglomerate, grading up into subarkose and orthoquartzite, comprise the Milamar Formation. The three formations are interpreted as deposited in a shallow marine basin (Dunphy Formation), by aeolian action on a land area of low relief to the west (Du Portage), and close to a land area that underwent strong uplift in the east (Milamar Formation).

Equivalents of the Seward Subgroup have been traced southward to latitude 50°N where they were described as "Seward (redbed) Formation" by Baragar (1967b). These rocks are conglomerate sandstone, siltstone, shale and dolomite and overlie local volcanics that indicate short periods of volcanism partly contemporaneous with or earlier than the redbed deposition. Farther south, equivalent drab medium- to coarse-grained clastics (Wynne-Edwards, 1960), some associated with silicic volcanics (Wynne-Edwards, 1961), have been mapped to 50°35'N and have been placed in the Seward Formation by Frarey and Duffell (1964).

Baragar (1967b) viewed the Seward redbeds to the southeast of the west-trending basin as derived from a nearby western source and deposited by fluvial or marine currents flowing up and down the present axis of the Labrador Trough. In the southern part of this area meagre data suggest that currents flowed to the west. The redbeds are believed by Baragar to have been reddened at the depositional site in deltaic, littoral or estuarine environments.

Facies variations, described by Dimroth (1968), are apparent also in the overlying Pistolet Subgroup. The lowest formation, the Lace Lake, a sequence of shale and siltstone with minor sandstone and dolomite, has a red and green western facies that contains minor sandstone, and a grey basinal facies to the east that does not. The succeeding Alder Formation consists of grey stromatolitic dolomite, calcareous arenite, dolomitic orthoquartzite and local pelitic interbeds. It contains submarginal and basinal calcareous facies and eastern, southwestern and northwestern marginal arenaceous facies. The northwestern marginal facies include beds of red sandstone. These rocks are succeeded by fine grained clastics and dolomite of the Uvé Formation. The Pistolet Subgroup is interpreted as having been deposited in a marine basin with the main source area to the east, and western marginal clastics derived from the west.

The Pistolet Subgroup is overlain by the Swampy Bay Subgroup, which contains large amounts of graphitic shale and quartz wacke of flysch-like appearance. At this time the area of sedimentation was divided into an eastern and western basin by a geanticline, termed by Dimroth (1968) the "central geanticline". Sedimentation in the western basin commenced with the Hautes Chutes Formation, a pyritic shale of euxinic character.

Sedimentation of the above redbeds and associated facies has been interpreted in the following way. The west-trending basin in which the Chakonipau Formation was deposited was interpreted by Fahrig (1957) to be of erosional origin. Dimroth (1968) considered it a pre-geosynclinal fault basin trending across the geosyncline. The Du Portage, Dunphy and Milamar formations he interpreted as deposited in a north-northwest trending linear basin with some instability (Baragar, 1967).

Near Fort Chimo, (Fig. 1a) among the sedimentary formations of the northern part of the Labrador Trough, hematite-bearing red sandstone occurs locally in the conglomeratic and sandy Chioak Formation (Bérard, 1965). This formation, which cannot yet be correlated with any part of the better known rocks of the Trough to the south, is currently being mapped at the scale of 1:50 000 between 57° and 58°N (T. Clark, pers. comm., 1976). The Chioak Formation is overlain by the "Abner Dolomite", which in turn is succeeded by fine grained marine clastics of the Larch River Formation. This last formation includes red sandstones (Dimroth et al., 1970).

In the Forbes Lake area, 110 km southwest of Fort Chimo, the upper part of the Ruth Formation contains red sandstone and siltstone. This formation is regarded (Clark, 1977) as transitional between the underlying Wishart (quartzite) and the overlying Sokoman (iron formation) formations.

The succeeding Menihek (quartzite, siltstone and shale) Formation is lithologically similar to the Larch River Formation. It is overlain by "unit 6" that may be correlative with the Chioak Formation. Unit 6 is a heterogeneous sandy unit containing local red siltstone and argillite. It is overlain by the dolomitic Abner Formation (Clark, 1977).

The Labrador Trough has been interpreted by Dimroth (1968) to be "unquestionably a geosyncline of Alpine type" in which were deposited three miogeosynclinal rock associations in succession. The first cycle consists of a pre-flysch sequence (upper Seward and Pistolet subgroups), followed by euxinic sediments (Hautes Chutes Formation) and then a flysch sequence (upper part of the Swampy Bay Subgroup). Occurrence of the first cycle of sedimentation only in the south and central parts of the Trough and deposition of a thick second cycle in the northern part (north of 57°N) of the Trough suggested to Dimroth (1968) northward migration of the Trough with time and he regarded the Chakonipau Formation as having been deposited in a pre-geosynclinal fault basin. Davidson (1972) drew attention to the apparent absence from the Labrador Trough of rocks characteristic of an Alpine geosyncline such as ophiolitic sequences similar to those of Cyprus, abundant granitic intrusions, the post-ophiolite flysch, and the molasse. Dewey and Burke (1973) fitted the Labrador Trough into the plate tectonic model, regarding the Circum-Ungava Geosyncline as the result of collision of continental blocks 1.75 Ga ago, a conclusion supported by Kearey's (1976) gravity studies. The west-trending depression at 50°30'N which contains the Chakonipau Formation was regarded by Burke and Dewey (1973) as an analogue of the Athapuscow Aulacogen (see Great Slave Supergroup, p.12). They assumed that this depression strikes west from a hypothetical triple junction the "Fort MacKenzie Junction". If the first cycle of sedimentation in the Labrador Trough is regarded as filling the opening southeast arm of this triple junction, the great thickness of the second cycle to the north might, as Dimroth (1968) implied, indicate later opening of the northern arm of the rift, while the western arm of the triple junction remained as a failed arm. Further, two parallel chains of outlines of redbeds of the Sakami Formation (p. 32), striking west-southwest over the Superior Province, might mark the westward continuation of the failed arm.

The redbeds of the central Trough are metamorphosed to the lower part of the greenschist facies of regional metamorphism or below (Dimroth, 1970), and lie within the western lithotectonic zone as defined by Dimroth (1970) i.e. a synclinorial zone derived from the original miogeosynclinal basin. The frontal (towards foreland) zone of this western synclinorium is characterized by a system of imbricate thrust faults dipping northeast and merging downward into a basal decollement surface at or near the surface of the Archean basement. The more internal parts of the synclinorium are characterized by narrow, doubly-plunging en-echelon isoclinal folds overturned to the west. A system of major thrust faults separates the synclinorium from an anticlinorial zone to the east that coincides with central geanticline that was active during the deposition of the Swampy Bay Subgroup (Dimroth, 1970).

The units of the northern Labrador Trough discussed above are much thrust-faulted, and folded, and dips vary from gentle to steep (Clark, 1977).

A significant uranium showing has been found in the Chioak Formation near Merchère Lake, just to the north of the Forbes Lake area. A trace of chalcopyrite was found in an arkose of "unit 6" (Clark, 1977).

The Belcher Group

The Belcher Group (Fig. 1) is exposed for more than 800 km on the Belcher Islands and other islands and on the east shore of Hudson Bay. On the mainland it overlies Archean granitic rocks. Strata correlated with those of the Belcher Group (Dimroth et al., 1970) are succeeded unconformably by lower Paleozoic rocks on the south shore of Hudson Bay (Bostock, 1969).

The Belcher Group is an internally conformable sequence thickening westward from 6100 to 9150 m and containing mainly fine grained clastics and carbonates as well as greywacke, arenite and 20 to 30 per cent of mafic volcanics (Dimroth et al., 1970). Jackson (1960) who mapped the main outcrop area, the Belcher Islands, divided the group into sixteen units which have since been given formal formational rank.

Sedimentation occurred in three cycles, the first represented only by the shallow-water stable shelf to miogeosynclinal Kasegalik Formation (unit 1 of Jackson, 1960). The second and third cycles commenced abruptly with volcanism. Following volcanism in each cycle, flysch-like sediments were deposited in relatively deep water, then after shoaling and stabilization, shallow water carbonates and orthoquartzites were deposited (Dimroth et al., 1970).

The Belcher Group may be the offshore stratigraphic equivalent of the more arenaceous Richmond Gulf sequence (p.11). The Kasegalik, Eskimo and Fairweather formations of the Belcher Group may be correlatives of the Pachi arkose, Pachi volcanics and Richmond Gulf formations respectively. The McLeary and overlying formations of the Belcher Group units 4 to 13 of Jackson (1960) are equivalent (Dimroth et al., 1970) to the Nastapoka Group that overlies the Richmond Gulf sequence (Woodcock, 1960).

The 1200 m thick Kasegalik Formation consists mainly of dolomite and limestone occur siliceous dolomite with red calcareous argillite partings containing halite casts and sulphate moulds (Bell and Jackson, 1974) interbedded near the base. Minor tuff and reddish shale and limestone occur at the top of the formation.

The second cycle commences with up to 610 m partly amygdaloidal spilitic basalt of the Eskimo Formation with interbeds of red and green argillite. The succeeding Fairweather Formation is about 350 to 600 m of varicoloured argillite with interbedded greywacke overlain by drab quartzite and contains minor dolomite. Sulphate moulds are present in the dolomite of the overlying dolomitic McLeary Formation (Bell and Jackson, 1974). The succeeding formations, units 5 to 11 of Jackson (1960), are composed mainly of argillite, dolomite and quartzite overlain by the Kipalu, an iron formation (Jackson's, 1960 unit 12).

The Flaherty Formation, which begins the third cycle, is a westward-thickening sequence of both pillowed and amygdaloidal volcanics. It is followed by over 3135 m of flysch-like, northwest-derived sediments of the Omarolluk Formation and then by molasse-like arkosic and subarkosic redbeds, the Loaf Formation (Jackson's 1960, unit 16). This unit is not widely preserved, and is thickest (210 m) on Loaf Island northeast of the Belcher Islands. Redbeds assigned to the upper part of the Omarolluk Formation may be more properly placed in the Loaf Formation (Dimroth et al., 1970).

Westward-directed paleocurrents and abundant evidence of shallow water deposition (Barrett, 1975; Dimroth et al., 1970; Ricketts, pers. comm., 1977; Stirbys, 1975) suggest that the lower part of the group represents a miogeoclinal sequence. Reversal of paleocurrents began in sediments interlayered with the Flaherty Formation (Leggett, 1974), and continued in the northwest-derived, shoaling, greywacke-turbidite Omarolluk Formation and overlying terrestrial molasse of the Loaf Formation (Dimroth et al., 1970). These events are regarded by some (Dewey and Burke, 1973; Gibb and Thomas, 1976; Gibb and Walcott, 1971) as evidence of continental plate collision but not all (Bell, 1974; Dimroth, 1972; Donaldson et al., 1976) concur.

The Belcher Group has suffered only diagenesis to low grade regional metamorphism, rising to the greenschist facies on the Hopewell Islands. Intensity of folding rises westward from the Richmond Gulf area. In the Belcher Islands the folds are closed anticlines alternating with broad open synclines which may be related to a basal décollement (Dimroth et al., 1970).

The nonconformable position of strata of the Circum-Ungava geosyncline upon Archean rocks of the Superior Province (Dimroth et al., 1970) and Fryer's (1972) 1-8 Ga Rb/Sr isochron ages for volcanics and sediments of the Belcher Group, indicate an Aphebian age for the Belcher Group.

Copper and lead-zinc mineralization occur in the Nastapoka Group (p. 11). Traces of chalcopyrite in the McLeary Formation on Belcher Islands are immediately above the stratigraphic level of the chalcopyrite mineralization in the Richmond Gulf Formation at Richmond Gulf. The lead-zinc-bearing dolomitic Mavor Formation of the Belcher Group (unit 6 of Jackson, 1960) is stratigraphically equivalent to the lead-zinc-bearing carbonates in the Nastapoka Group of the east shore of Hudson Bay (Dimroth et al., 1970).

Richmond Gulf sequence

Unmetamorphosed, chemical and fine grained clastic sediments and volcanics of the Nastapoka Group nonconformably overlie Archean granitic rocks and dip gently west along the shore and on the offshore islands on the east side of Hudson Bay (Fig. 1), for over 550 km between Cape Jones and the northern limit of the Hopewell Islands to the north (Eade, 1966, unit 13b; Stevenson, 1968, unit 9b; Woodcock, 1960, Nastapoka Group). The Richmond Gulf Sequence of coarser, mainly arkosic clastics containing redbeds and volcanics is sandwiched between the basement and the overlying fine grained sediments and volcanics only on the shores and islands of Richmond Gulf around 50°15'N. The coarser clastics extend for 65 km in a north-south direction and 35 km in an east-west direction (Eade, 1966, unit 13a; Stevenson, 1969, unit 9a). The development of the stratigraphic terminology of these rocks was given by Dimroth et al. (1970). Woodcock (1960) split the coarser clastics of Richmond Gulf into two parts, the Pachi Group and the unconformably overlying Richmond Gulf Formation. Correlation between the sequence at Richmond Gulf and the Belcher Group 130 km west (Dimroth et al. 1970) is discussed on page 10.

The Pachi Group, up to 500 m thick in the northern part of the gulf, consists mainly of feldspathic sandstone overlain by up to 50 m (Chandler, 1978a) of basaltic (Hews, 1976) and andesitic flows and some sills (Woodcock, 1960). On a regional scale it overlies a rugged Archean surface mantled by a regolith (Eade, 1966; Woodcock 1960). The Pachi sandstone is grey, green or pink and locally red and may contain quartz or feldspathic pebbles. Quartz pebble conglomerate is present locally, and granitic conglomerate at the base of the formation. The Pachi volcanic unit has features of terrestrial pahoehoe lava. The Richmond Gulf Formation where examined by the writer in the northern part of the Gulf, disconformably overlies the volcanic unit, and consists in ascending order of a 1-2 m thick coarse basal conglomerate derived from the volcanic unit, about 80 m of sandy to silty redbeds, pink feldspathic sandstone, and about 180 m of grey pyritic sandstone. Both sedimentary formations may be of braided fluvial origin. South of 56°N, that is south of the areas mapped by Woodcock and Stevenson, Eade's unit 13a (1966) consists mainly of red, buff or green arkose and its upper part contains red argillite interbeds.

The Richmond Gulf sediments were deposited in a graben that projected eastward from the main Proterozoic basin of Hudson Bay (Woodcock, 1960). Though extensive block faulting followed deposition of much or all of the Richmond Gulf Formation, a significant relief during deposition of the Pachi sediments is evidenced by marked thickness variation of that unit. Later uplift, erosion and tilting to the southwest were followed by erosional development of a hilly landscape on the Richmond Gulf Formation. After unconformable deposition of the Nastapoka Group on the Richmond Gulf sediments and on the granites to the north and south, the strata were tilted toward the centre of the Belcher Basin.

Attempts at radiometric dating of the sedimentary rocks and the volcanics of the sequence by the Rb/Sr method have yielded only Hudsonian and younger ages (Esquevin and Menendez, 1973; Hews, 1976). However, the Richmond Gulf sequence is believed to be part of the filling of the Circum-Ungava geosyncline (Dimroth et al., 1970) and consequently of Aphebian age.

Base metal mineralization, both disseminated and as veins, occurs in the Richmond Gulf Formation and in the Nastapoka Group (Chandler, 1978a). Disseminated chalcopyrite is present in fractured arkose at 56°22'N on the west shore of Richmond Gulf. The mineralized zone, some 150 m long and 7.5 m wide, is immediately below the upper unconformable contact of the Richmond Gulf Formation (Stevenson's unit 9a). A second occurrence was reported in dolomite at 56°25'N (Stevenson, 1968). Galena sphalerite mineralization has been reported from several places in a stromatolitic dolomite unit at the base of the Nastapoka Group, both on the west side of Richmond Gulf and to the south. Locations and references to earlier work on the mineralization are given by Eade (1966) and Stevenson (1968). Radioactivity in the Pachi sediments and Richmond Gulf Formation is primarily due to thorium in monazite and thorite in black sands (Miller, 1978).

Missi Group

The Missi Group (Fig. 1) outcrops in three major structural basins near Flin Flon (Stauffer, 1974) and one structural basin 130 km to the east near Snow Lake, Manitoba (Frarey, 1948). Near Flin Flon it is up to 2700 m thick and overlies the mainly metavolcanic Amisk Group. Felsic intrusives cut the Missi Group and pre-Missi felsic intrusives cut the Amisk (Stauffer, 1974). Though the Missi and Amisk groups together resemble an Archean greenstone belt (Davidson, 1972) they appear to be Aphebian in age (Mukherjee et al., 1971; Sangster, 1972). Both the Missi Group and the nearby lithologically similar Sickle Group (Bailes, 1971; Coats et al., 1972) are likely protoliths for much of the Kisseynew gneiss of northern Manitoba and adjacent Saskatchewan (Pearson, 1972).

The Missi Group consists of two formations, each conglomeratic and arenaceous, which can be further subdivided, and in which megaclasts become more mature and less abundant upward. Bedding and locally intense cross-bedding are evident in the dominantly arkosic sandstone. The pinkish grey, lithic Missi Group sandstone (Mukherjee, 1974) is largely derived from mafic volcanic rocks and is rich in black iron oxide grains, an average of oxide grains for the group being 3 per cent with some beds containing 30 per cent. At Athapapuscow Lake some almost unmetamorphosed sandstone has a red ferruginous earthy cement that indicates it was deposited as redbeds (Stauffer, 1974).

The Missi has been interpreted as an alluvial fan to alluvial plain deposit on the basis of lensing of beds, gross arkosic composition (see Mukherjee above), sedimentary structures, and texture. The sediments are inferred to have been deposited as a molasse in an active graben structure at the front of an active tectonic belt, or in an intermontane basin. The source area, which lay to the west, may have been the older Amisk island arc.

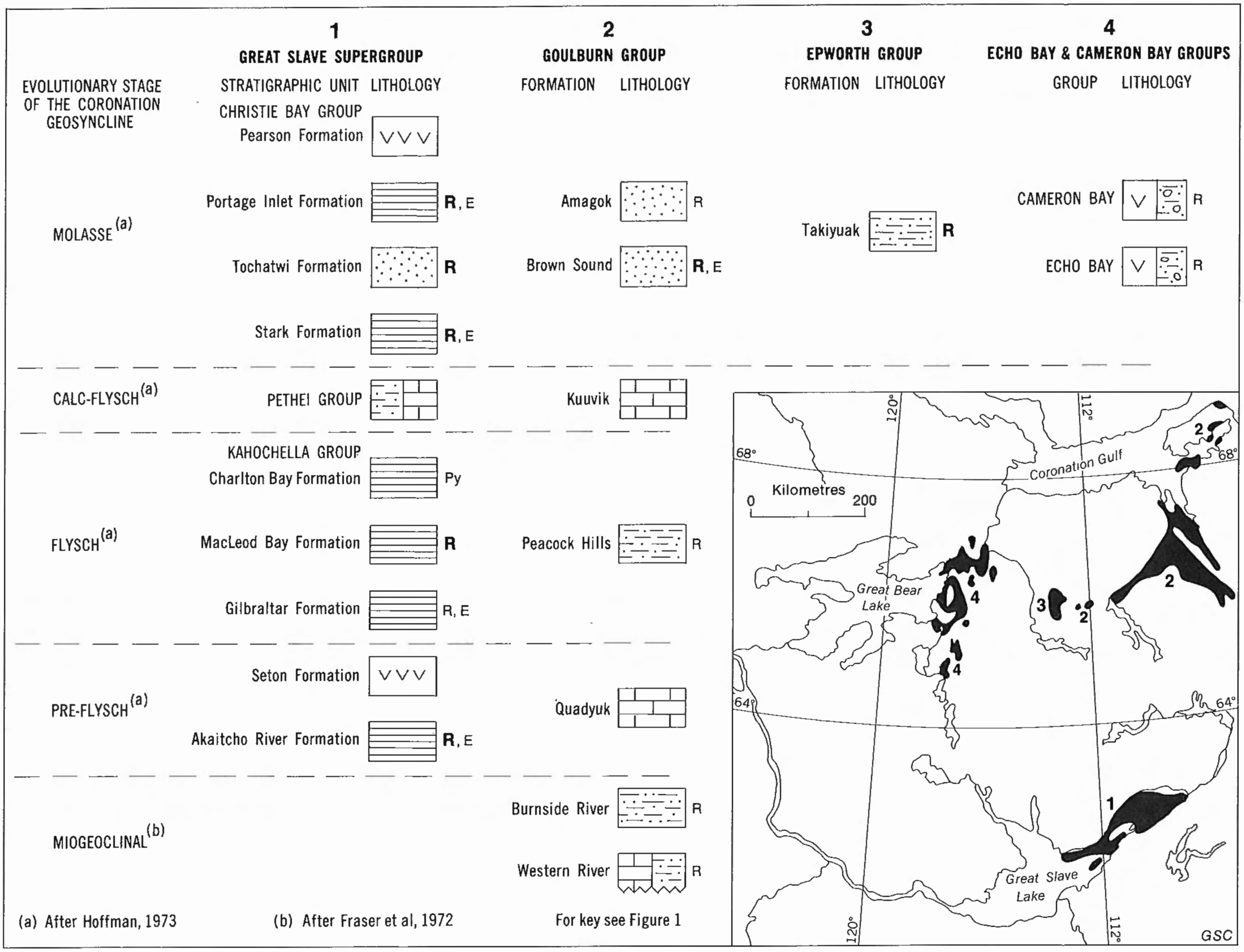


Figure 2. Aphebian sequences related to the Coronation Geosyncline.

The Missi Group has undergone polyphase deformation. Regional metamorphism of Hudsonian age varies from almandine amphibolite facies in the north near the Kisseynew gneiss domain to being almost absent in the south near Athapapuskow Lake (Stauffer, 1974). Disseminated copper mineralization has been found in the Sickle Group of the Lynn Lake-Indian Lake area (Baldwin, 1976).

Aphebian sequences related to the Coronation Geosyncline

The tectonic development of the Late Aphebian Coronation Geosyncline (Hoffman, 1973) (Fig. 2) gave rise to sedimentary rocks preserved as three broadly similar and broadly correlative successions; the Great Slave Supergroup of the East Arm of Great Slave Lake, the Epworth Group of the foreland, and the Goulburn Group of Bathurst Inlet (Fig. 2a). In addition sediments are preserved among the volcanic rocks (Echo Bay and Cameron Bay groups) that overlie the postorogenic Great Bear Batholith on the east shore of Great Slave Lake. These sequences all contain redbeds, and apart from the Great Slave Supergroup (Hoffman et al., 1974; Hoffman, 1977) those areas where red-beds are present are little deformed (Tremblay, 1968; Fraser and Tremblay, 1969; Fraser, 1974.

The Great Slave Supergroup consists of a mainly sedimentary sequence over 11 km thick underlying a fault-bounded linear depression, the Athapuscow Aulacogen that extends more than 200 km northeastward from the Paleozoic cover between the Slave and the Churchill Structure Provinces along the East Arm of Great Slave Lake. About 1500 m of equivalent strata overlie the adjacent margin of the Archean Slave Province to the northwest. Within the aulacogen the Great Slave Supergroup unconformably overlies metasediments and volcanics of the early or mid-Aphebian Union Island Group and is unconformably overlain by the red terrestrial Et-then Group (p. 15) of Late Aphebian to Early Helikian age (Hoffman, 1973).

The Goulburn Group underlies an area of at least 7000 km^2 in the north-eastern part of the Archean Slave Structural Province and in the adjacent Bathurst trench (Tremblay, 1968; Campbell and Cecile,, 1976; Fraser, 1964). The 7000 to 8000 m thick group is overlain with angular unconformity by sandstone of the Helikian Tinney Cove Formation (Campbell and Cecile, 1975). Red terrigenous clastics near Wellington Bay on Victoria Island, and the Richardson Islands (Fig. 5) were mapped as part of the Shaler

Group (Thorsteinsson and Tozer, 1962). However, near Wellington Bay two structural units may be present (Christie et al., 1972) and some of these rocks strikingly resemble the Burnside River Formation of the Goulburn Group (F.H.A. Campbell, pers. comm., 1976).

The Takiyuak Formation, the only redbed unit in the Epworth Group (Hoffman, 1973), is preserved only along 20 km of the southwest shore of Takijug (formerly Takiyuak) Lake (Fraser, 1974), underlying a south-narrowing triangular area that reaches its apex about 50 km to the south. The formation has been mapped by Fraser (1960, 1974). The Takiyuak Formation overlies the Cowles Lake Formation which consists of 500 m of grey, laminated, argillaceous limestone with minor greywacke turbidite beds whose source lay to the west. Beds of brecciated red argillaceous limestone with stromatolites and mudcracks lie at the top of the Cowles Lake Formation. The top of the Takiyuak Formation is unroofed (Hoffman, 1973).

The Echo Bay and Cameron Bay groups are parts of a continental volcanic sequence with minor amounts of alluvial sedimentary rocks. The sequence is intruded by the 1800 Ma old comagmatic Great Bear Batholith that strikes north-south along the east side of Great Bear Lake and in part overlies the Hepburn Batholith (Hoffman and McGlynn, 1977) which is about 150 Ma older.

Sedimentary uranium mineralization in the Great Slave Supergroup lies in miogeoclinal non-red, quartz-rich arenites both of the Kluziai Formation and of the lower third of the Hornby Channel Formation. Solutions of either biogenic or volcanogenic origin pyritized the ilmenite and magnetite of the heavy mineral suite of the quartzites. During later burial and metamorphism, the pyrite reacted with uraniferous solutions of uncertain origin to produce uraninite (Morton, 1974).

In the Goulburn Group disseminated chalcocite occurs along at least 90 m of strike over a thickness of 2 m in the basinal stromatolite facies of the Beechey Platform of the Western River Formation. This mineralization is spatially related to two diabase dykes. Pie- and pipe-like lithic breccia bodies mostly in the Brown Sound Formation redbeds and probably of mud volcano origin (M.P. Cecile, pers. comm., 1976) are similar to some found in the East Arm of Great Slave Lake (Reinhardt, 1972) that are associated with uranium mineralization (Campbell and Cecile, 1976). The strata of these related successions are the record of the Coronation Geosyncline and are described below in terms of its evolution.

Miogeoclinal phase

Redbeds in strata equivalent to the miogeoclinal phase of the (Coronation-Geosyncline occur only in the basal Western River and the overlying Burnside River formations of the Goulburn Group.

The westward-thinning (1200 to 100 m) Western River Formation was divided into five members including a local basal conglomerate. The conglomerate is overlain by members designated lower argillite, red siltstone, quartzite and upper argillite (Tremblay, 1968). The formation, described in detail by Campbell and Cecile (1976) contains also turbidites and stromatolitic dolomite, and is regarded (ibid.) as deltaic in origin. The overlying fluviatile Burnside River Formation thickens eastward from about 200 m to about 2500 m. It consists of red mudstone and siltstone, pink and red sandstone and minor dolomite. Sedimentary transport was westward and northwestward (Campbell and Cecile, 1976).

Pre-flysh phase

Redbeds in this phase occur only in the Great Slave Supergroup, and in that group only in the Akaitcho River Formation. This unit is of barrier-protected lagoon and nearshore open shelf origin and consists of 200 m of red mudstone with thin beds of ripple-laminated and mudcracked siltstone, white glauconitic subarkose with opposed cross-bedding, and granular hematite ironstone partly replaced by spherulitic carbonate with calcite-filled gypsum casts. In the southwest part of the aulacogen the red mudstone is complexly intercalated with mafic flows and pyroclastics that comprise the Seton Formation (Hoffman et al., 1977).

Flysch phase

Greywacke turbidites of the flysch phase were shed from the rising tectonic highlands to the west, and were presumably ponded to the west of the East Arm of Great Slave Lake by the Seton volcanics. Consequently, in the East Arm, the Akaitcho River Formation is overlain by possibly up to 1450 m of red and green mudstone of the Kahochella Group, that thins to 350 m at the northeast end of the aulacogen and on the platform where it is entirely red (Hoffman, 1973). This group was tentatively interpreted (Hoffman, 1969) to be of shallow marine origin with periodic emergence in the northeast. The lower part of the Kahochella mudstones, the Gibraltar Formation, is mainly red and contains thin beds of granular hematite ironstone intimately associated with spherulitic limestone, flat-pebble intraformational conglomerate and calcite-filled gypsum casts. The upper part of the red mudstone succession, the MacLeod Bay Formation, contains abundant calcareous concretions and lags of concretion pebble conglomerate. The red mudstones are capped by dark green pyritic mudstone of the Charlton Bay Formation that thins from 150 m in the southwest part of the aulacogen to 10 m on the platform (Hoffman, 1973).

In the Bathurst Inlet area the flysch phase was heralded by shallow water carbonates of the pre-flysch Quadyuk Formation. Then, in deepening water were deposited green and red mudstone turbidites and carbonates of the lowest of three members of the Peacock Hills Formation. Next occurred a minor clastic influx of red, green and red-brown mudstone and siltstone turbidites. Further mudstone and carbonate turbidites of the upper member of the Peacock Hills Formation and overlying stromatolitic carbonates of the Kuuvik Formation are a sign of shoaling (Campbell and Cecile, 1976).

Calc-flysch and molasse phases

The deposits of the calc-flysch phase contain no redbeds but the succeeding molasse contains redbeds in all four areas. That of the Great Slave Supergroup, the Christie Bay Group, is well preserved in the aulacogen, but almost completely eroded from the adjacent platform. In the aulacogen 650 m of red mudstone, the Stark Formation, sharply overlies the calc-flysch, and contains halite casts and a carbonate fragment megabreccia, possibly the result of salt-solution collapse (Hoffman et al., 1977). This formation was also intruded by early quartz diorite laccoliths. Above these rocks is 850 m of red laminated lithic sandstone, the Tochatwi Formation. The sandstone occurs in crossbedded units up to 5 m thick with abundant mudcracks and ripple marks. It contains a high proportion of sedimentary and silicic volcanic clasts, derived from the Great Bear Batholith, and transported along the axis of the aulacogen from the southwest (Hoffman, 1973) where it was deposited as an alluvial fan-delta complex over prodelta mudstones of the Stark Formation (Hoffman et al., 1977).

The sandstones are overlain by 230 m of red, mudcrack-bearing mudstone of the Portage Inlet Formation, with buff siltstone beds and abundant halite and gypsum casts. These rocks are overlain by 180 m of subaerial basalt of the Pearson Formation which in turn is overlain unconformably by the fanglomerates of the Et-then Group (p. 15) (Hoffman, 1973).

In the Bathurst Inlet area following the stromatolitic carbonate of the Kuuvik Formation, the equivalent of the calc-flysch, continued regression brought terrestrial conditions and the coarsening-upward redbed sequence of the Brown Sound and the overlying Amagok formations. The southward-thinning Brown Sound Formation (2300 to 800 m) is composed of a lower member of red calcareous mudstone containing salt casts, discontinuous thin sandstone and a carbonate-block olistostrome that like the Stark Formation may have originated by salt-solution collapse. The middle member is a red, muddy siltstone and the uppermost member is an arkose with minor vesicular basalts. The Brown Sound is succeeded by coarser white to mauve arkose of the Amagok Formation. The sediments of these two formations are interpreted to have been deposited during the south-to-southwestward progradation of a distant deltaic complex into a restricted marine basin (Campbell and Cecile, 1976). A detailed account of the Brown Sound and Amagok formations was written by Cecile (1976).

Redbeds in the Epworth Group are only preserved in the Takiyuak Formation, an alluvial molasse which was shed eastward onto the craton from the Great Bear Batholith (Hoffman, 1973). The formation, which has not yet been mapped in detail, consists of a lower 125 m of red mudstone with thin beds of ripple-laminated and mudcracked buff siltstone that rests with sharp contact on the red argillite of the Cowles Lake Formation (Fraser, 1974). The red mudstone is succeeded by at least 500 m of red mudstone with fining-upward units of crossbedded, friable, red lithic sandstone many metres thick. It is typically laminated, well sorted, calcite-cemented and contains abundant angular to subrounded clasts of terrigenous and carbonate sedimentary rocks, intermediate to siliceous volcanics, and subsidiary plutonic and metamorphic rocks (Hoffman, 1973).

The supracrustal rocks overlying the Great Bear Batholith have been known as the Echo Bay Group and overlying (Mursky, 1963) Cameron Bay Group (Craig et al. 1960; Feniak, 1948; Fraser et al.,1970, 1972; Hoffman et al., 1976; and Kidd, 1933). The locally derived sediments were deposited in continental to shallow marine environments (Fraser et al., 1972). In most places the rocks are only slightly deformed and metamorphism is present only at contacts with late Aphebian felsic intrusives (Fraser et al., 1972). Separation of the two groups was only possible on the shore of Great Bear Lake (Stockwell et al; 1970).

The Echo Bay Group was divided into a mainly sedimentary lower part, 1300 m thick, and a mainly felsic volcanic upper part. The lower part included tuff, conglomerate, chert, thin limy beds and mudcracked red and green argillite and arkose (Mursky, 1963; Stockwell et al. 1970), The Cameron Bay Group consisted mainly of immature sedimentary rocks "a few hundred" to 3 km thick. The sediments are a conformable sequence including conglomerate, purple and maroon arkose, siltstone and mudcracked shale (Stockwell et al., 1970) with local thin dolomite bands (McGlynn, 1964).

The Echo Bay and Cameron Bay groups are now regarded as but two of the many mappable units in the lower part of a mainly continental volcanic imbricate assemblage of maximum thickness over 40 km. On account of wide miscorrelation the original stratigraphic significance of the terms has been lost. The Lindsley Bay Rhyolite in the lower part of the informally named sequence consists of rhyolite ash flows and lava domes, and red and buff alluvial conglomerate and sandstone and local probably lacustrine mudstone (Hoffman and McGlynn, 1977).

Early post-Hudsonian taphrogenic redbed of the western Churchill Province

Several late Aphebian to Early Helikian redbed sequences overlie gneiss and folded Aphebian sediments of the western Churchill Province (Fig. 3). They are unmetamorphosed and for the most part little deformed. In several cases the basal unconformity is of great relief. Many of the basins in which the redbeds are preserved are strongly faulted and in some cases faulting markedly influenced patterns of sedimentation. The sequences generally have conglomeratic lower parts and sandy upper parts and are interpreted as molasse or of taphrogenic origin. Sedimentary interpretations stress fanglomeratic and fluviatile deposition. General correlation amongst sequences and with the Echo Bay and Cameron Bay groups (Fig. 2) have been suggested by Hoffman (1969) and Fraser et al. (1970).

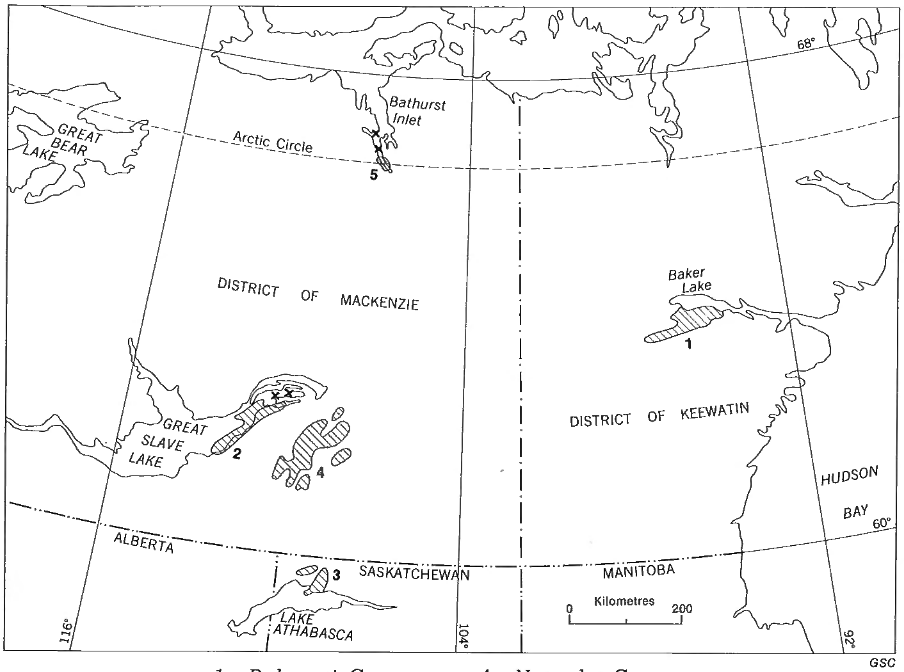
Dubawnt Group

The South Channel and Kazan redbed formations are the lowest two formations of the Dubawnt Group (Fraser et al., 1970). They are confined to only a small part of its outcrop area on the south shore of Baker Lake and in a number of small areas about 50 km west of the lake (Fig. 3). They form a thick, wedge-shaped sequence, thickening westward from a southeastern edge and bounded on the north and northwest by faults (Donaldson, 1967). The Kazan covers an area of about 1150 km^2 and the South Channel is exposed in a belt less than 8 km wide running along the east margin of the Kazan for about 40 km (Macey, 1973). These formations are at present being mapped southwest of Baker Lake at 1:250 000 scale (Eade and Blake, 1977; Le Cheminant et al., 1977).

The conglomeratic South Channel Formation lies with profound unconformity on a rugged land surface developed on Aphebian gneisses (Donaldson, 1967) and is itself overlain conformably by the fine clastics of the Kazan Formation. The intermediate volcanic Christopher Island and Pitz formations lie upon the Kazan with slight unconformity. The volcanics are in turn disconformably overlain by quartzose arenite and conglomerate of the Thelon Formation. Small amounts of an unnamed stromatolitic dolomite formation probably of shallow water origin, overlie the Thelon. These Precambrian rocks are overlain by small, unconformable outliers of Ordovician carbonates (Fraser et al., 1970).

The mainly granitoid South Channel Formation is up to 1500 m thick and contains ortho- and paraconglomerate set in a red-brown matrix. Lenses and channel fillings of sandstone and pebbly sandstone mark otherwise obscure bedding. The channel-fill sandstone is crossbedded and is intercalated with ripple marked and desiccation-cracked mudstone. The Kazan Formation is predominantly red calcareous arkose with a red mudstone and siltstone component particularly in the upper part. Sedimentary structures include abundant symmetrical and less abundant asymmetrical ripples, and current lineations. Crossbedded sandstones with pebble lenses fill scour channels. Mudstone with subordinate siltstone makes up much of the upper part of the formation also being found in the sandstone below as thin intercalations overlain by mud chip conglomerate (Donaldson, 1967).

Paleocurrents in the two formations flowed generally westward (Fraser et al., 1970). According to Donaldson (1967) the formations were deposited largely in the fluvial



1. *Dubawnt Group*
2. *Et-then Group*
3. *Martin Formation*
4. *Nonacho Group*
5. *Tinney Cove Formation*

Figure 3. Early post-Hudsonian taphrogenic redbeds of the western Churchill Province, location map.

environment, and are best interpreted as a piedmont deposit that accumulated by the growth of alluvial fans fronting a nearby granitoid source to the southeast. Donaldson also referred to the action of mudflows in producing some of the disrupted framework conglomerates in the South Channel Formation and suggested that large-scale crossbedding in the Kazan may have originated in aeolian dunes. Sedimentary structures, lithologies, and textures are suggestive of the braided fluvial environment, and siltstones and mudstones may have been deposited on a floodplain or in a lucustrine environment (Macey, 1973). The absence of evidence of the playa environment implied a humid climate. The setting of these environments was regarded by Fraser et al. (1970) as an intercratonic molasse basin.

The South Channel and Kazan formations dip gently to the northwest on the south side of Baker Lake and to the southeastward on the northwest side. Hematite is a ubiquitous matrix mineral in the Dubawnt redbeds, and predates development of quartz and feldspar overgrowths (Macey, 1973). The heavy mineral suite is mature and may have been modified by postdepositional effects including oxidation of biotite, chlorite, pyroxene, amphibole and magnetite (Macey, 1973). These iron-bearing minerals would have been a ready source for hematite.

The South Channel and Kazan formations overlie gneisses metamorphosed during the Hudsonian event and are overlain by volcanics which have yielded a good Rb/Sr isochron of 1.732 Ga (Fraser et al., 1970).

Uranium mineralization analogous to that of the volcanics in the Martin Formation (see p. this report) has been found in the volcanics overlying the Kazan Formation (Ruzicka, 1976a). Uranium is also found in the Kazan Formation (A. Miller, in prep.). On the basis of a similarity of tectonic setting, sequence and time of sedimentation to the Martin Formation, the Kazan, its basement and the overlying volcanics are considered favourable ground for uranium mineralization (Ruzicka, 1976a).

Et-then Group

The Et-then Group (Fig. 3) lies within the East Arm of Great Slave Lake. It overlies unconformably folded Aphebian molasse, the Christie Bay Group, and has an erosional upper surface.

It is up to 4.5 km thick and is divided into the Murky and overlying Preble formations (Hoffman, 1969). The group is intruded by shallow-dipping diabase dykes and sills, which in turn are cut by northwest-trending vertical dykes of the Mackenzie swarm (Hoffman, 1968).

The lower formation, the Murky, is predominantly a very massive conglomerate in units 9 to 46 m thick, separated by thin, mudcracked red shale and siltstone intervals, and minor amygdaloidal basalt flows or caliche horizons (Hoffman, 1968). Clasts up to boulder size are composed mainly of sediments from the underlying Great Slave Supergroup. The matrix of the conglomerate is a calcite-cemented lithic sandstone, composed mainly of sedimentary rock fragments. Rare granitoid fragments are present in the upper part of the sequence. The Preble Formation, composed of lithic or arkosic sandstone, is at least 3 km thick and like the underlying Murky Formation, is red or buff and friable.

The highly varied orientation of the less than 200 crossbed measurements from the Murky Formation, as well as the marked relief of the basal unconformity suggest intrabasinal derivation of some of the conglomerate. The sedimentary structures of the Preble Formation are consistent with deposition as fining-upward alluvial cycles. Sedimentary transport, based on over 1100 readings, was to the southwest along the graben axis, though material may have been derived from boundary faults (Hoffman, 1969).

On the heels of orogenesis of the Coronation Geosyncline to the west, the Christie Bay Group (p. 13), was shed northwestward into the East Arm of Great Slave Lake, the Athapuscow Aulacogen of Hoffman (1973). After folding of the molasse, strike-slip (Hoffman et al., 1977) movement of faults bounding the aulacogen, particularly the McDonald Fault System defining its southeastern margin, created the topographic setting for deposition of the southeast derived Et-then Group (Fraser et al., 1972). The greater throw along the southeast margin of this trough (taphrogeosyncline, Hoffman, 1969) created an asymmetry that controlled the lithofacies and thickness variation of the Et-then Group. Northwestward thinning of the Murky Formation, of its conglomeratic units and textural variation of the conglomerate lead to interpretation as a sequence of alluvial fans growing northwestward from scarps developed at the southeastern boundary fault system (McDonald Fault) of an asymmetric graben. In the extreme northwest, the normally thin siltstone and shale interlayers are as much as 9 m thick, rippled and mudcracked, and contain calcareous stromatolites. This fine grained northwestern facies has been interpreted as lacustrine deposits formed at the distal margins of the alluvial fans (Hoffman, 1969).

Intermediate intrusive bodies that have intruded the Great Slave Supergroup and are overlain by the Et-then Group have been dated at 1.63 to 1.845 Ga (Hoffman, 1968). Diabase dykes and sills dated at 1.3 Ga cut the Et-then Group (Fahrig and Wanless, 1963). Therefore it is likely that the Et-then Group is Late Aphebian to early Helikian in age (Hoffman, 1969).

Martin Formation

The almost entirely red Martin Formation (Fig. 3) outcrops in several basins both of structural and depositional character (Fraser et al., 1970) and it has an aggregate areal extent of 220 km^2. It overlies with profound unconformity, high grade, probably Archean, locally regolith-covered metasediments of the Tazin Group. The upper surface of the formation is erosional (Tremblay, 1972).

The Martin Formation consists of a basal conglomerate up to 750 m thick, overlain by a lower arkose unit (up to 2500 m thick), volcanics (up to 1000 m thick), an upper arkose unit (up to 2100 m thick), and lastly up to 1800 m of siltstone, giving in all a section 4000 to 5800 m thick. The basal conglomerate is a polymictic ortho- to paraconglomerate, generally massive but with some crude size sorting. Clasts are mainly granitoid, up to 1 m across. Sandstone and siltstone interbeds and sandstone channel-fillings are present in the unit. The lower arkose is medium grained with common conglomeratic interbeds and lenses. Sedimentary structures include parallel lamination and planar and trough crossbeds. The upper arkose is finer grained than the lower and has more diverse sedimentary structures including common symmetrical and asymmetrical ripples, mud-flake conglomerates and mudcracks. Transition to the overlying siltstone unit occurs over one or two hundred metres via a mixed sandy, conglomeratic silty zone. The siltstone unit itself contains common ripple marks, desiccation cracks and flaser bedding. Stromatolite-like structures have also been seen (Tremblay, 1972; Macey, 1973). Sedimentary structures indicate flow from the northeast (Fraser et al., 1970) in a braided fluvial system (Macey, 1973) on a rugged block-faulted terrain (Fraser et al., 970; Tremblay, 1972).

Hematite is a ubiquitous matrix mineral in the Martin redbeds, and predates development of quartz and feldspar overgrowths (Macey, 1973). Though much is detrital, having been derived from metamorphic hematite from the Tazin rocks (Tremblay, 1972), some may have had a diagenetic origin. For, though the mineralogy of the sequences is immature (arkose; Pettijohn, 1957) the heavy mineral suite is mature and may have been modified by post-depositional effects including oxidation of biotite, chlorite, pyroxene, amphibole and magnetite (Macey, 1973) which would be a ready source for hematite. Lower beds dip steeply at basinal margins whereas higher beds are little deformed (Fraser et al. 1970).

The age of the Martin Formation is between 1.63 and 1.65 Ga (Fraser et al., 1970) or between 1.83 and 1.65 Ga ago (Tremblay, 1972), i.e. it was deposited close to the end of the Hudsonian event (Fraser et al., 1970). Though its relation with the quartz arenitic Athabasca Formation, which lies to the south of Athabasca Lake, is uncertain (Fahrig, 1961) its greater deformation and different lithology (Tremblay 1972) and the presence of Martin-type clasts in basal conglomerate of the Athabasca Formation (Donaldson, 1968) suggest that it is older than the Athabasca.

Mineralization in the Martin Formation is restricted to pitchblende in the volcanics and the basal conglomerate. Vein-type uranium mineralization is common in the underlying Tazin Group gneisses (Tremblay, 1972).

Nonacho Group

The Nonacho Group (Fig. 3) lies mostly within a northeast-striking structural basin on a basement of granitic rocks, centred on Nonacho Lake about 80 km southeast of the East Arm of Great Slave Lake (Henderson, 1939b; Wilson, 1939; Brown, 1950). A smaller, separate basin on the Taltson River lies 6 km west of this basin (Henderson, 1939a). Though the sediments of the main basin are buff to grey and those of the smaller basin are red, the sediments of the two basins have been correlated.

There has been dispute about the age relationship between the Nonacho sediments and some of the granitoid rocks surrounding the basins. Where conglomerate was seen in contact with granite the contact was interpreted as an unconformity, but where sandstone or strongly deformed sediments are in contact with granite an intrusive relationship was inferred (Burwash and Baadsgaard, 1962; J.C. McGlynn, pers. comm., 1976). McGlynn (1966, 1970a) re-examined the field evidence and concluded that no granites had intruded the Nonacho Group.

The group comprises a conformable sequence of polymictic granitoid conglomerate, and conglomeratic sandstone, arkose, feldspathic quartzite, siltstone and in some cases mudcracked shale (McGlynn, 1970b). In the southern part of the Nonacho Basin an extensive conglomerate which overlies the group is underlain by a zone of slightly moved granitoid blocks separated by veins of muddy arkose. Elsewhere, though conglomerate is common in the lower part of the sequence, units wedge out over short distances. Also finer clastics may be prominent in the lower part of the sequence (McGlynn, 1970a). Though subdivision of the group was hampered by lack of marker horizons, there is a tendency for conglomerate to be well represented in the lower part of the sequence and for feldspathic sandstone, the bulk of the sequence, to be better represented in the upper part. Siltstone and shale may be thick in the northern part of the basin (McGlynn, 1970a, 1971a).

Hoffman (1969) regarded the Nonacho as a "taphrogeosynclinal" sequence probably correlative with the lithologicaly similar (Hoffman, 1968) Et-then Group of the East Arm of Great Slave Lake and pointed out that both basins were self-contained during sedimentation. The Nonacho Group is thrown into broad folds with northeast and east-northeast striking axes with gentle northeastward plunge. Shearing is intense where shaly rocks are prominent in the axial regions of folds. Faults are common at the margins of the basin (McGlynn, 1970a, 1971a).

The group was probably deposited in an environment similar to the lithologically similar Martin and Et-then groups (McGlynn, 1970b). McGlynn (1971a) considered the group to be entirely fluvial in origin, reporting fining upward cycles in conglomerate arkose and shale or red shale-chip conglomerate at the top of some sandstone beds. The paleocurrents pattern is not simple, being influenced by the rugged basement topography (McGlynn, 1966, 1970a, 1971a; J.C. McGlynn, pers. comm., 1976).

An older age limit for the group is established by the underlying Archean granites. Argon/argon dating of a basaltic dyke (the Sparrow Dykes) that intruded the Nonacho Group gave a younger age limit of 1.7 Ga for sedimentation of the group (McGlynn et al., 1974). No economic mineralization has been found. Uranium and copper shows are mostly in the basement in shears, mylonite or anomalous uranium values (J.C. McGlynn, pers. comm., 1976).

Tinney Cove Formation (s.s., Campbell and Cecile, 1975)

Fraser (1964) mapped grey, buff and red "arkosic" sandstone and polymictic conglomerate in the south part of Bathurst Inlet (his unit 16). Fraser et al. (1970) correlated this 600 m thick unit with the mainly grey to pink orthoquartzitic Ellice Formation of Tremblay (1971) 75 km to the southeast. Campbell and Cecile (1975) reinterpreted Fraser's unit 16 to consist of arkose and polymict conglomerate, which they designated Tinney Cove Formation (s.s.), overlain unconformably by kaolinitic quartzite and quartz pebble conglomerate of the Ellice Formation.

The Tinney Cove Formation (s.s.) overlies with marked unconformity folded Brown Sound Formation, (molasse) of the Aphebian Goulburn Group. It is divided into a basal, T_1 member of red conglomerate, sedimentary megabreccia and coarse grained polymict conglomerate and a more extensive upper T_2 member about 250 m thick of reddish pink and locally mottled arkosic grit to siltstone (Campbell, 1978).

Beneath Tremblay's (1971) Ellice Formation (p. 16) are three laterally impersistent units, possible correlatives of the Tinney Cove Formation (Fraser et al., 1970). They are, in ascending stratigraphic order a grey unit (45 m), a red unit (45 m) and a conglomerate unit (335 m). The grey unit consists of quartzite and quartzite conglomerate, the red unit of shale, carbonate and conglomerate. The conglomerate unit, possibly unconformable on the red unit, is deep red for 60-90 m above the red unit, polymictic, and coarsely stratified.

The T_1 member of the Tinney Cove Formation was deposited in fault-controlled linear (?) (Campbell, 1978) basins. Deposition of the Tinney Cove was controlled by syndepositional transcurrent faulting. The formation is interpreted as fluviatile, including fault derived fanglomerate and to have been deposited at the beginning of the Helikian Era. It is possibly a correlative of the Et-then Group of Great Slave Lake (Campbell, 1978).

Late Proterozoic miogeoclinal sequences of the Cordillera

The Precambrian history of the Canadian Cordillera (Fig. 4) is recorded in two great, dominantly sedimentary sequences, the Purcell System and the younger Windermere System, and their probable correlatives. In Canada the Purcell System is exposed along the eastern margin of the Cordillera. It outcrops in the Purcell Mountains of southern

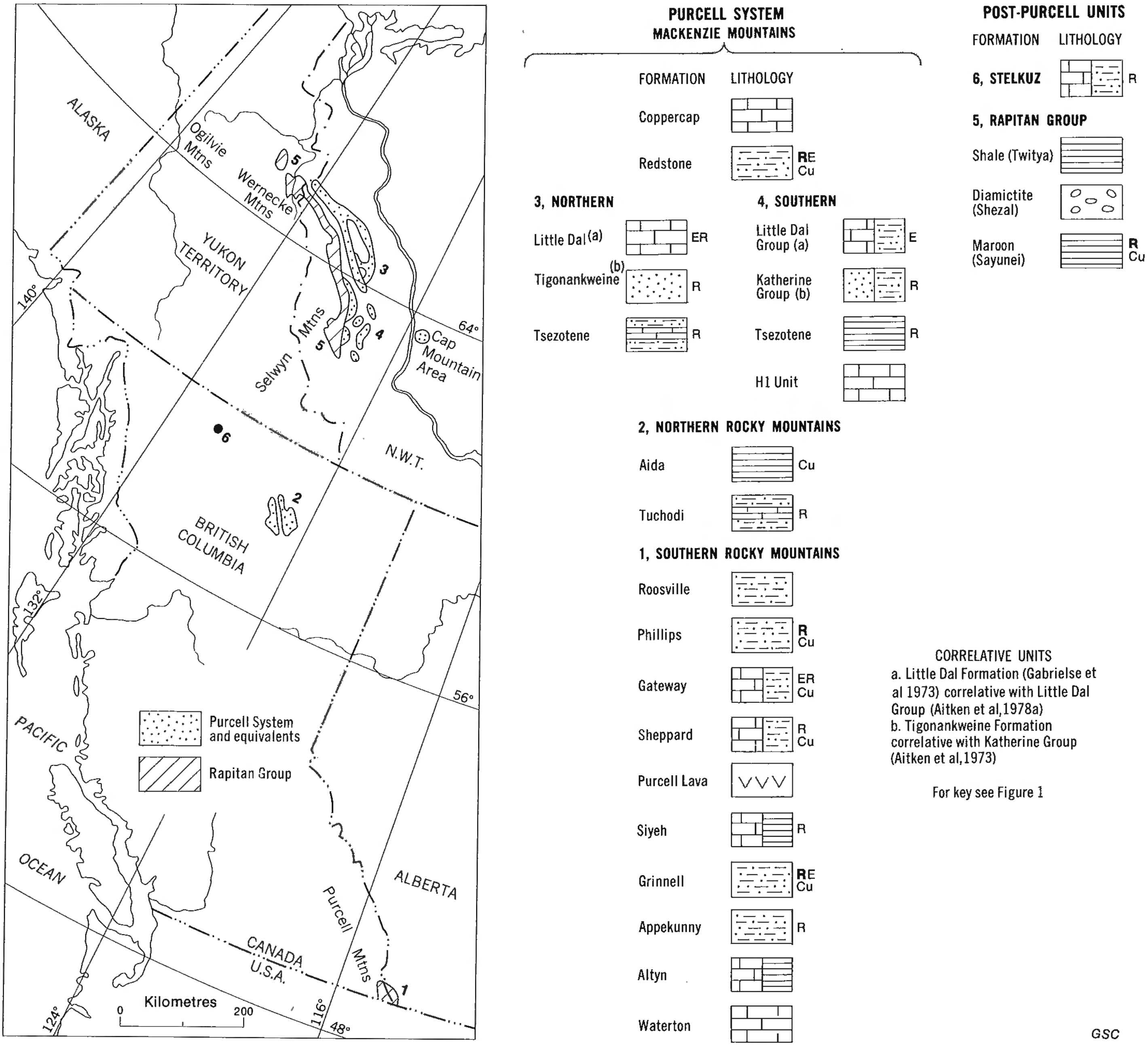


Figure 4. Late Proterozoic miogeoclinal sequences of the Cordillera.

British Columbia and in the southern Rocky Mountains to the east. To the north, similar rocks underlie parts of the northern Rocky Mountains and the Mackenzie Mountains, and underlie fairly extensive areas of the Ogilvie and Wernecke mountains (Fig. 4a) (Gabrielse, 1972b). The southern exposures in the Purcell and Rocky mountains continue southward to extensive exposures in the United States under the name Belt Supergroup (Harrison, 1972; Bishop et al., 1973). The geology of the Purcell System has been reviewed by Price (1964), Gabrielse (1972b) and Wheeler et al. (1972).

The Purcell is composed predominantly of argillite, siltstone, fine grained sandstone, carbonate, and local basalt flows, and intruded by probably related gabbroic dykes and sills. The sequence thickens from about 3 km in eastern exposures to about 9 km in those in the west. Exposures in the east display features of shallow water deposition such as mudcracks and stromatolites; those in the west have the characteristics of deep-water turbidites grading up into shallow water sediments.

The Purcell System and equivalents were probably deposited as a miogeoclinal (shelf-slope-rise) complex on the western margin of the North American continent following mid-Proterozoic rifting (Monger et al., 1972). They were probably deposited during the time interval 1.3 to 0.9 Ga. Intercalated volcanics, the Purcell Lavas, were extruded 1.1 Ga ago (Obradovich and Peterman, 1973). However, according to Stewart (1976) rifting did not occur until about 0.85 Ga ago, and the Belt-Purcell rocks were deposited in a fault-bounded trough. Deformation and low-grade metamorphism, caused by the East Kootenay Orogeny in southeastern British Columbia, and the possibly equivalent Racklan Orogeny in the Mackenzie Mountains, terminated Purcell sedimentation and gave rise to the sub-Windermere (Rapitan) unconformity (Monger et al., 1972). Accounts of the subsequent history of the Canadian Cordillera are given by Monger et al. (ibid.), and by Wheeler et al. (1972).

Unlike the Purcell System, the overlying Hadrynian Windermere is almost continuously exposed along the full length of the Cordillera (Gabrielse, 1972b). It accumulated as a miogeocline of which the source area and the site of deposition were displaced westward from those of Helikian (Purcell) time. The lower and thicker part of the Windermere assemblage, up to 6600 m thick, is poorly sorted, locally graded, rapidly deposited grit, sandstone, and shale with distinctive conglomerate or conglomeratic mudstones at or near the base. Thick sequences of basic and andesitic volcanics occur in the Windermere succession in the southern and northern parts of the Cordillera. The upper part of the Windermere assemblage is composed mainly of pelitic rocks with local distinctive green and maroon shades and contains carbonate units. The abrupt change from deposition of the mature sediments of the Purcell to the coarse, immature sediments of the Windermere assemblage suggests that during the Hadrynian, sources of significant relief become available, due either to glacial lowering of sea level or to tectonism (Wheeler et al., 1972). The Rapitan Group of the Mackenzie Mountains, and correlatives (p. 20) are probably correlatives of the Windermere Group.

Purcell System of the southern Rocky Mountains

Redbed facies are conspicuous in the southern Mackenzie Mountains and the southern Rocky Mountains (Gabrielse, 1972b). Redbed formations included in the southern exposures of the Purcell System, particularly the Clark and Lewis ranges of the southern Rocky Mountains, are in ascending stratigraphic order, the Grinnell, the upper member of the Sheppard, the lower member of the Gateway and the Phillips (Fig. 4b). These units appear broadly to be eastern equivalents of drab units in the Purcell Mountains, the Creston Formation in the case of the Grinnell, and the Dutch Creek Formation in the case of the others.

In the Clark Range the lower part of the succession consists, in ascending order, of the Waterton Formation (shelf carbonate), the Altyn Formation (a complex of carbonate and terrigeneous clastics formed in deep water in the west and shallow water in the east), and the Appekunny Formation (floodplain to brackish tidal clastics). Minor red argillite occurs at the top of the Appekunny Formation (Price, 1964). The overlying Grinnell Formation is a sequence of bright red argillite and shaly siltstone with minor white, red, or green, quartz-rich sandstones. The white quartzite units, which are commoner at the top of the formations, are lenticular and may have associated argillite-pebble conglomerate and commonly show one set of tangential crossbeds and rippled tops. Mudcracks are common and evidence of salt hoppers has been found in the formation (Collins, 1974; Price, 1962, 1965). The overlying Siyeh Formation consists of 600-900 m of drab argillite, siltstone, and dolomite, some of which is stromatolitic (Collins, 1974; Douglas, 1952; Price, 1962, 1964).

The Grinnell Formation was deposited by flash floods on a broad floodplain of low relief (Collins, 1974), and the Siyeh mostly as shallow marine shelf carbonate (Price, 1964). The andesitic Purcell Lava was extruded onto wet unconsolidated muds of the Siyeh Formation (Price, 1964).

The overlying Sheppard Formation is divided into a lower and upper part in the southern part of the Clark Range, where it is 120-180 m thick. The lower part consists of light green and fine grained clastics and dolomite. The upper part contains distinctive red siltstone and sandstone which are more abundant in the north and northeast part of the Clark Range. These "red beds" are intercalated with rocks similar to those of the lower part of the formation and are replaced progressively westward by drab, fine grained clastics.

In the Clark Range the succeeding Gateway Formation has been divided into lower and upper parts. The lower part gradationally overlies the Sheppard, consists of dark red and purplish siltstone and argillite, and contains mudcracks, oscillation ripples and abundant casts of cubic salt crystals. The upper member consists of grey and green, fine grained clastics and dolomite, some stromatolitic, and passes gradationally up to the Phillips Formation.

The Phillips Formation consists of red and purplish red, fine- to coarse-grained, quartz-rich sandstone and red siltstone, with argillite partings which pass laterally into mud chip conglomerates. Mudcracks and ripples are common. The formation passes upward into drab, fine grained clastics of the Roosville Formation.

The Sheppard, Gateway, Phillips and Roosville formations were included in the Kintla Formation (Price, 1965) and are believed to be the eastern lateral equivalent of the drab Dutch Creek Formation of the Purcell Mountains (Price, 1964).

Strata-bound copper mineralization occurs in many formations of the Belt Supergroup of the United States where it is best developed in white and green sandstones and siltstones in redbed sequences (Harrison, 1974). In the Purcell System of the Lewis and Clark ranges in Canada where many formations are similarly mineralized (Morton et al., 1973), the best mineralization is found in the Grinnell Formation (Collins, 1974). This formation is the stratigraphic equivalent of the lithologically similar and similarly mineralized Spokane Formation of the Belt Supergroup (Harrison, 1972, 1974). Both the Grinnell and the Spokane pass westward into grey-green equivalents, the Creston and Revett formations respectively. The Creston of the Purcell Mountains (Reesor, 1957) consists of greenish grey weathering, impure, fine grained quartzite and argillite. The Revett Formation (Harrison, 1972) appears to be a white quartzite which incidentally carries the only currently economic stratabound copper mineralization in the Belt Supergroup. Though these observations prompt the suggestion that similar mineralization might be found in the Creston Formation, none has yet been found (R. Kirkham, pers. comm., 1976). Two observations are relevant. Firstly the ore-bearing quartzite of the Revett Formation in Montana has not been traced northward into Canada (G.B. Leech, pers. comm., 1976) and secondly, the quartzite content of the Revett Formation increases southwestward (Harrison, 1972).

Stratabound copper mineralization in the lower Sheppard and Gateway formations is mainly in arenites (Morton et al., 1973). The Phillips Formation, like its stratigraphic equivalent, the redbed Bonner Quartzite (Harrison, 1972) in the Belt Supergroup, is unmineralized.

Purcell System of the northern Rocky Mountains

An unmetamorphosed and little deformed Helikian sequence of drab, fine grained clastics and carbonate, similar to the Purcell System of the southern Rocky Mountains, lies within the northern Rocky Mountains of British Columbia (mainly in NTS map area 94K) (Bell, 1968). The lower part of the 7600 m thick section is of shallow marine to coastal origin with detritus derived from the craton that lay to the northeast. The upper part of the sequence is a shaly flysch.

The 1500 m thick Tuchodi Formation, at the top of the shallow water sequence, is typically composed of pale feldspathic quartzite, dolomite, and siltstone. Varicoloured shale and sandstone accompanied by mudcracked layers form a minor but significant element. A local red and varicoloured shale facies is also present. The succeeding 1000-2000 m thick Aida Formation, a grey calcareous and dolomitic mudstone, is part of the shaly flysch. It contains 56 m of chamositic mudstone overlain by 60 m of carbonaceous mudstone about 200 m above the base of the formation (Bell, 1968).

Quartz-ankerite-chalcopyrite veins occur at many stratigraphic levels in the sequence but particularly in the Aida Formation. The mineralization is associated with but predates diabase dykes (Taylor and Stott, 1973). The Churchill copper deposit (Carr, 1971) lies within the Aida Formation as mapped by Taylor and Stott (1973). According to Carr (ibid.) the veins occur in grey and black argillite and limy argillite and shale, and transect bedding.

Purcell System of the Mackenzie Mountains and equivalents

A thickness of more than 5500 m of unmetamorphosed sedimentary rocks probably of Helikian age, lies within the Mackenzie Mountains south of latitude 64° (NTS 95 E,L,M). This sequence, described in ascending order after Gabrielse et al. (1973), is apparently internally conformable, and overlain by the Hadrynian Rapitan Group.

The lowermost formation, the Tsezotene, is 1100 m thick and consists of drab to purple shale, dolomite and quartzite. Stromatolites in the dolomite, and mudcracks, are present locally. The Tigonankweine Formation, 1300 to 1370 m thick, is locally variable in lithology. In the type area it is composed of pink, white and purple quartzite, and subordinate mudcracked red and purple shale and dolomite.

In its type section the Little Dal Formation contains a lower member, at least 1830 m thick and an upper member, at least 90 m thick. These may in part be correlatives. The lower member consists mainly of stromatolitic and oolitic limestone and dolomite and minor, fine grained clastics. Maroon slate containing mudcracks and salt casts is present in the basal beds in one area. The upper member, of variable thickness, is composed of dolomite and sandy dolomite as well as minor amounts of fine grained clastics.

The Redstone River Formation, 0 to 60 m thick, with a basal fluvial and alluvial fan conglomerate up to 150 m thick, unconformably overlies the Little Dal Formation (Eisbacher, 1978) and consists mainly of pink-weathering siltstone. Rain prints and mudcracks are present locally. From three to seven zones of greenish weathering, slightly dolomitic siltstone containing copper minerals are present at the top of the formation. In some places gypsum beds and gypsiferous beds are associated with the pink weathering siltstone. The Coppercap Formation, 0 to 210 m thick, is dominantly grey weathering carbonate and includes sandstone and siltstone. A thickness of 70 to 300 m of fetid limestone is present in the formation in various localities.

The thickness variation in the upper formations in the southern Mackenzie Mountains was attributed by Gabrielse et al. (1973) to an unconformity at the base of the Rapitan Formation (now Group, see p. 20) that cut down to the top of the lower member of the Little Dal Formation. On the other hand Eisbacher (1976) interpreted these thickness variations in terms of facies changes in the Helikian sediments during deposition on a complicated miogeosynclinal hinge. Eisbacher (ibid.) who measured sections in the vicinity of the Redstone River (NTS 95 M), viewed the Helikian sequence as the result of shoaling sedimentary cycles. The first cycle contained the shale and siltstone of the Tsezotene Formation and the shallow water quartzite of the Tigonankweine Formation. The second cycle included basinal carbonate of the Little Dal Formation overlain by redbeds and evaporites of the Redstone River Formation. The Coppercap Formation was the basinal carbonate phase of the cycle continuing with the overlying Rapitan Group.

The Purcell rocks of the southern Mackenzie Mountains are deformed into north- to northwest-trending folds with limbs that dip moderately steeply. The folds are broken by generally westward-dipping thrust faults (Gabrielse et al., 1973).

Mineralization in these rocks includes copper-silver sulphide in breccia of the Little Dal Formation and copper in the Redstone and Coppercap formations (Gabrielse et al., 1973). Cupriferous beds at the top of the Redstone River Formation have been examined by Coates (1964) and Eisbacher (1977) at the Redstone copper deposit about 1.5 km east of Coates Lake (Little Dal Lake of Coates 1964) in map area 95 L. According to Coates the mineralization lies at the top of a local informal unit, the "Jan Marie Formation" (Redstone River Formation of Gabrielse et al., 1973), that conformably underlies his "Cleo Formation" (Coppercap Formation of Gabrielse et al., 1973). Coates placed these units at the top of his "Redstone Formation" and suggested that either they are remnants, not removed during erosion, of the sub-Rapitan unconformity or that they are facies of restricted original distribution. Seven hundred metres of the gypsiferous "Jan Marie Formation" is exposed but this thickness may include repetition by faulting. The unit is almost entirely purple-maroon to brown siltstone with thin purple and maroon mudstone partings. Mudcracks are invariably present. The cupriferous beds lie within the uppermost 110 m, an interval that records a period of heterogeneous sedimentation transitional to the black limestone-shale facies of the "Cleo Formation" (Coppercap Formation). Coates (ibid.) reported six cupriferous beds composed of grey, silty, pure or arenaceous limestones that are generally separated from the enclosing redbeds by bleached zones.

The overlying 180 m thick "Cleo Formation" is predominantly limestone but has sandy beds at the base and the top. In ascending order it consists of 10-15 m of grey pyritic sandstone, 60 m of pyritic, calcareous black shale and silty black limestone with abundant carbonaceous remains, 90 m of fetid black limestone, and 45 m of dark and light grey limestone and calcirudite. This unit passes up into the Rapitan Formation without evidence of an unconformity (Coates, 1964).

The mineralization described above and other copper occurrences in the vicinity, have been treated briefly by Baragar and Hornbrook (1963) and Green and Godwin (1963, 1964). The mineralized Redstone River Formation has been traced from 64° to 65°N by Aitken (1977). Mineralization of this formation occurred before deposition of the Rapitan Group (Eisbacher, 1977).

In the Mackenzie Mountains, north of 64°N and east of 135°W, Aitken et al. (1973) during "Operation Norman" measured sections probably stratigraphically equivalent to those of Gabrielse et al. (1973) of the southern Mackenzie Mountains. For these northern units Aitken et al. (1973) used a partly different stratigraphic terminology (Fig. 4b).

The lowest unit, "unnamed map unit H-1", is a pale grey dolomite at least 370 m thick containing rare stromatolites (see also Aitken et al., 1978a). It is overlain by the widespread 760-1220 m thick Tsezotene Formation (Aitken et al., 1973) that is divisible into a lower member dominated by grey argillaceous rocks and an upper member mainly of varicoloured argillite. The formation is of marginal marine origin (Aitken et al., 1978a).

The Tsezotene Formation is overlain by the Katherine Group, the stratigraphic equivalent (Aitken et al., 1973) of the Tigonankweine Formation of Gabrielse et al. (1973). This predominantly arenaceous group was deposited by alternate progradation of fluviodeltaic complexes (sandstone) and marine transgressions (sandstone, shale, carbonate). Minor amounts of purple and maroon shale occur within the group.

West of the Operation Norman area in the Ogilvie and Wernecke mountains, Blusson (1974), Blusson and Tempelman-Kluit (1970) and Green (1972) mapped a "Purcell-like" Helikian sequence consisting chiefly of dolomite but containing minor maroon shale and quartzite (see also Gabrielse, 1967b). Green believed this sequence to contain part of the Katherine Group. Base metal mineralization in these rocks in the Nedaleen River map area (NTS 106 C), has been described briefly by Laznicka (1977).

Conformably overlying the Katherine Group is the extensive unnamed "H 5" unit. Characteristic rock types include dolomite and some stromatolites, limestone and drab and red shale. Aitken (1977) studied this little known 1350 m thick unit dividing it into at least five blanket formations. Among these formations the "basinal sequence of carbonate and shale contains at least one red shale interval, one of these being the "Dead End Shale", previously thought to be a regional marker unit (Aitken et al., 1973). A 530 m thick "Gypsum Subunit" has two or three thin red gypsiferous sandstone beds in the upper part and the overlying "rusty shale" subunit contains one or two red intervals. Because the "H 5" unit is correlative with the Little Dal Formation of Gabrielse et al. (1973) of the southern Mackenzie Mountains the latter is elevated to group rank. Stromatolites, cryptalgal laminites, mudcracks, gypsum, oolites and abundant carbonate in the "H 5" unit (Aitken et al., 1978a) suggest a marginal marine depositional environment.

Recent work in the northern Mackenzie Mountains (Aitken, 1977; Eisbacher, 1977) has shown that the cupriferous Redstone River and Coppercap formations extend northwestward from the southern Mackenzie Mountains to the Mount Eduni map area (NTS 106 B).

About 1760 m of fine grained clastics, much of which are hematitic, are exposed at Cap Mountain in map area 95 O, and were divided (Douglas and Norris, 1963) into four formations. The topmost formation was considered to be of Hadrynian age and to lie unconformably on three Helikian formations (Douglas et al., 1970). Aitken et al. (1973) considered the whole sequence Helikian, denying the presence of an unconformity and stressing the presence in the four formations of features characteristic of the Purcell System rather than the Windermere System.

Rapitan Group, Hadrynian

The Rapitan Group (Fig. 4) was named by Green and Godwin (1963) from the Snake River area (Wernecke Mountains) Yukon Territory (NTS 106 F). It has been mapped in the southern Mackenzie Mountains (Gabrielse et al., 1973) and sections have been measured in the northern Mackenzies (Aitken et al., 1973). Detailed examination of the group in the southern Mackenzie Mountains has been carried out by Upitis (1966), Eisbacher (1976) and Young (1976). The group has been traced from the Snake River area to the Mackenzie Mountains (Eisbacher, 1978). Discussion of controversial aspects of the sedimentation of the group may be referred to in Young (ibid.). In the southern Mackenzie Mountains the Rapitan Group overlies with marked regional unconformity equivalents of the Purcell System (Gabrielse et al., 1973), a position questioned by Eisbacher (1976). It is overlain by the Hadrynian (?) Keele Formation, (now placed within the Rapitan Group, (Eisbacher 1978; Aitken et al., 1973) or by lower Cambrian strata (Gabrielse 1972a), thus the Rapitan Group may be correlative with at least part of the Windermere System. Also the Rapitan Group may be correlated tentatively with rocks in the Selwyn Mountains on the basis of lithological similarity (Blusson, 1967, 1968). Green (1972) referred to conglomerate and red sandstone and other clastics (his unit 5) very similar to the Rapitan 130 km to the west of the Rapitan Group and the Snake River area, but considered these rocks to have been formed by erosion and redeposition of rocks correlative with the Rapitan.

The informally named "grit unit", (Gabrielse, 1973) occurs from the Alaska border to northern British Columbia (Green, 1972, Blusson, 1974) and has been partly correlated with the Windermere System (Gabrielse, 1972b). It consists mainly of grit and finer clastics including spectacular red and green shales. Red purple and green slate, 650 m thick, similar to that of the grit unit have been mapped by Gordey (1978) at a scale of 1:50 000 in the Nahanni map area (NTS 105 I) of Selwyn Mountains.

The Rapitan Group is a clastic sequence containing much pelite and conglomerate. In the southern Mackenzie Mountains the group is divided into three units by two internal unconformities (Upitis, 1966). The lowest unit, the Maroon formation of Eisbacher (1976) (Sayunei Formation, Eisbacher, 1978), is 0 to 407 m thick. Variation in thickness is mainly due to the unconformity at the base of the middle unit, the Diamictite formation of Eisbacher (1976) (Shezal Formation, Eisbacher, 1978). According to Upitis (1966) and Gabrielse et al. (1973), 70 per cent of the Maroon formation is composed of hematitic dark reddish brown or maroon weathering clayey or silty mudstone. A few beds are green. Thin lenses of fine grained sandstone and fine grained conglomerate are common. Poorly sorted conglomerate beds, with clasts of fine grained clastics or of carbonates 1 cm to 1.5 m thick, comprise 25 to 30 per cent of the lowest unit. Thirty metres of layered jasper-hematite iron formation, with a hematite content of possibly 20 per cent, are present at the top of the unit east of Little Dal (Coates) Lake. According to Eisbacher (1976) the base of the formation is marked by a striking change in rock colour. The basal member near Redstone River is a sequence of green, dark red or maroon graded siltstone, graded lithic arenite and lenticular sharpstone conglomerate. The lithic arenite displays many features typical of the turbidite mode of deposition. In the central part of the Maroon formation the most common sedimentary rock is graded rythmite consisting of siltstone laminae with dark red argillaceous partings and local thin iron formation beds. Toward the top of the formation graded lithic arenite interbedded with maroon siltstone and dark red argillite increases in abundance upward. Large clasts isolated in thinly bedded clastics of the Maroon formation, including some in the iron formation at the top of the formation, were interpreted by Young (1976) to have been dropped by floating ice.

Overlying the Maroon formation is a drab, pebbly, mudstone unit 370 to 610 m thick. The mainly dolomitic clasts are set in a matrix of pyritic dolomitic mudstone (Upitis, 1966; Gabrielse et al., 1973). Above the Diamictite formation, lies the topmost of the three units of the group, the Shale formation (Eisbacher 1976) (Twitya Formation, Eisbacher, 1978). This unit, 570 to 1170 m thick (Gabrielse, et al., 1973) and composed mainly of dark drab shale is generally conformable on the Diamictite formation. Higher in the unit crossbedded quartzose sandstone occurs. Apparent unconformity is probably due (Eisbacher, 1976) to mass flow and sliding.

Young (1976) interpreted the lower and middle units of the Rapitan Group to have been deposited under glacio-marine conditions (see also Eisbacher, 1976) and the upper unit during post-glacial marine transgression. Under this scheme the iron formation was produced by concentration of silica and iron in seawater by freezing. Gross (1973) believed the iron formation to be the result of volcanic exhalative activity and isolated boulders found in the iron formation to be the result of explosive volcanic activity rather than dropstones as thought by Young (ibid.). Gross noted that many of the boulders lay along seams of fine grained gravel and tuffaceous material. Upitis (1966) and Gabrielse et al. (1973), on the other hand, favoured mass-flow mechanisms including turbidity currents of tectonic rather than glacial origin for the genesis of much of the Rapitan Group and supported Gross's (ibid.) volcanic exhalative origin for the iron formation. Gabrielse et al. (1973) cited the large amounts of fresh ash and tuff with the iron formation and the greenstone clasts in the Diamictite formation.

In the Snake River area (Green and Goodwin, 1963) rocks of the Rapitan Group are unmetamorphosed and thrown into northwest-trending open folds with gentle dips and with northwest-trending thrust faults superimposed on them. Metamorphism of the group in the southern Mackenzie Mountains is not mentioned by Gabrielse et al. (1973). There the group has been gently folded and has suffered minor faulting except at the southern end of the Redstone Plateau.

Windermere strata, possibly correlative with the Rapitan Group, were deposited during the interval of about 0.8 to 0.6 Ga ago (Gabrielse, 1972b). Eisbacher (1978) reported copper mineralization in the Saynnei Formation and in the Keele Formation which he included in the Rapitan Group. "Copper shale"-type mineralization in Hadrynian red and argillite in the Wernecke Mountains (Laznicka, 1977) occurs in unit HS Fe which underlies the middle part of the Rapitan Group (Blusson 1974).

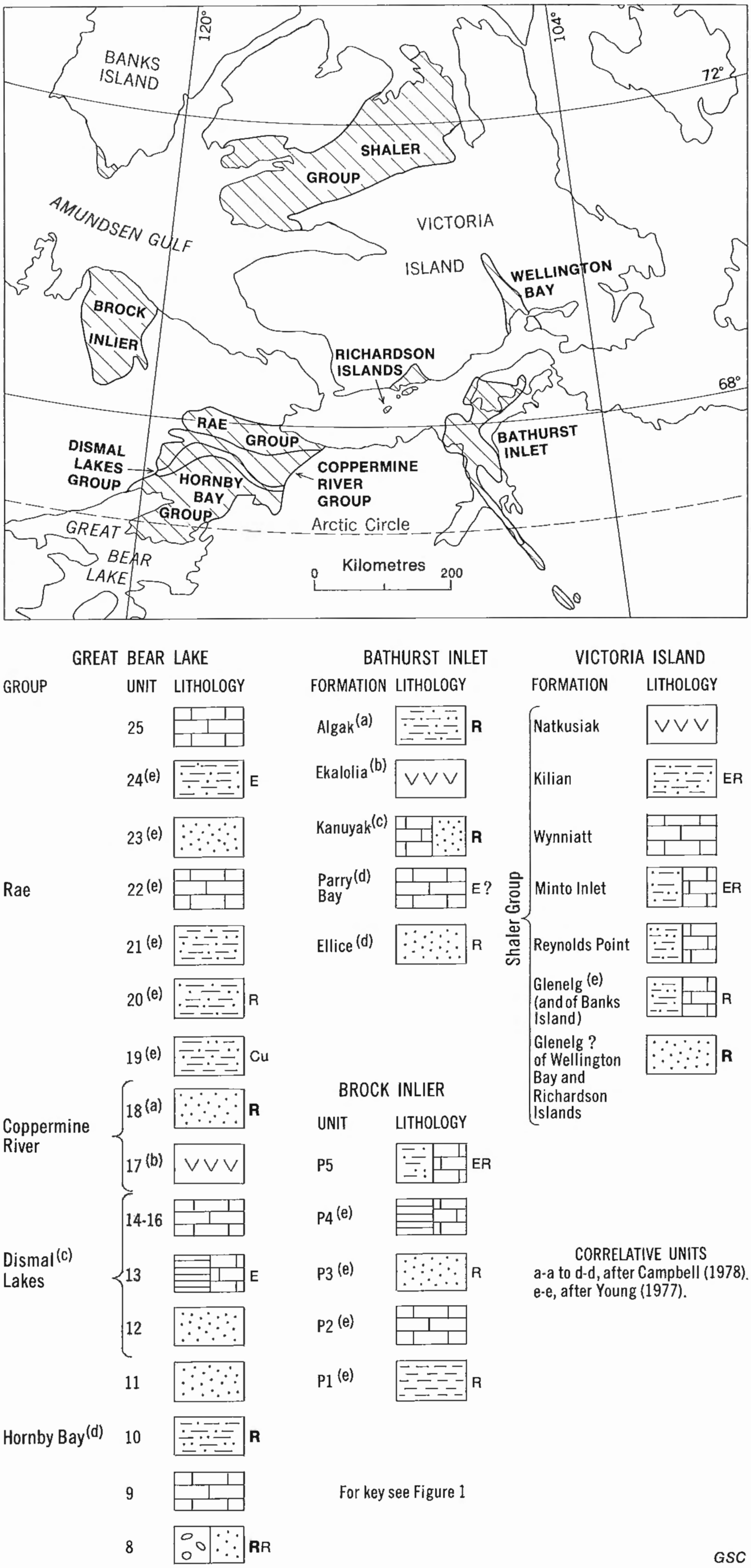


Figure 5. Late Proterozoic sequences of the Amundsen Basin.

A conspicuous regionally distributed redbed unit 80 m thick has been described from the Stelkuz Formation of the Late Proterozoic Ingenika Group of northern British Columbia (NTS 104 P). Carbonate, and shale form a large part of the enclosing sequence. The redbed unit consists of a calcareous red breccia overlain in turn by varicoloured siltstone and red slate (Mansy, 1978).

Late Proterozoic sequences of the Amundsen Basin

Christie et al. (1972) assigned all the post-Aphebian Proterozoic sequences of Victoria Island and the mainland in the Brock River, Coppermine River and Bathurst Inlet regions to a single sedimentary basin, the Amundsen Basin (Fig. 5). Most recent views on intrabasinal stratigraphic correlation are those of Campbell (1978) who correlated the ascending sequence; Hornby Bay, Dismal Lakes, Coppermine River and possibly the Rae groups with the sequence at Bathurst Inlet (Fig. 5b); Aitken et al. (1978b) who correlated the Shaler Group with the Brook Inlier sequence, and Young (1977) who correlated these two sequences with the Rae Group.

The rocks of the Amundsen Basin consist of fine grained clastics, stromatolitic carbonate and associated plateau basalts. The sediments are predominantly of stable-shelf, shallow-marine origin; some are fluviodeltaic. Redbeds are present in all sequences. Conglomeratic and sandy redbeds occur above the basal unconformity of the Hornby Bay Group and the Bathurst Inlet sequence. In all sequences except the Coppermine River Group fine grained redbeds are associated with evidence of evaporites.

Hornby Bay Group (Baragar and Donaldson, 1973)

Baragar and Donaldson (1973) split the Paleohelikian Hornby Bay Group (Fig. 5) of earlier workers into two newly defined groups, the Hornby Bay and the unconformably overlying Dismal Lakes Group. The Hornby Bay (s.s.) Group is the lower part of a more or less homoclinal, gently northward-dipping Helikian and Hadrynian succession lying between Great Bear Lake and Coronation Gulf (Fraser et al., 1970)(Fig. 5a).

The group underlies an area that extends east-west for about 250 km and north-south about 130 km. In the southeast it overlies folded and metamorphosed strata of the Aphebian Epworth Group as well as volcanic and granitic rocks of Late Aphebian Great Bear Batholith (Fraser and Tremblay, 1969) with development of a reddish saprolite on the volcanics (Hoffman and Cecile, 1974). The group is unconformably overlain by buff to white quartzose sandstone of the Dismal Lakes Group (Baragar and Donaldson, 1973).

In the Dismal Lakes area (Baragar and Donaldson, 1973) the Hornby Bay Group is a northwest-thickening (Fraser et al., 1970) sequence of 2000 m of sandstone, dolomite, shale, and minor conglomerate. The lowermost unit (unit 8) of the group consists of over 900 m of sandstone. At the base of this unit is about 6 m of deep red quartz pebble conglomerate resting on weathered felsite. The conglomerate is succeeded by deep red sandstone with scattered quartz pebbles, abundant crossbeds and ripple marks. Sandstone higher in the succession is cream, buff pink and mauve in colour.

Six hundred metres of dolomite (unit 9) with chert nodules, oolites and stromatolites apparently overlie the sandstone conformably. In some outcrops in the Dismal Lakes area the dolomite has undergone deformation of possibly soft sediment origin. Farther west (Donaldson, 1971) similar deformation in the dolomite involving red mudstone containing halite and gypsum casts may be caused by solution of evaporites. On the dolomite lie 300 m of well bedded intercalated sandstone, siltstone and shale with abundant ripple marks and desiccation structures (unit 10). These rocks are mostly maroon to deep red but some beds are locally pale green to mauve. The intercalated beds give way to unit 11, 150 m of massive white to buff quartzose sandstone with minor conglomerate interbeds similar to that of the basal unit of the group (Baragar and Donaldson, 1973).

Fraser et al. (1970) considered the Hornby Bay Group to be part of a previously much more extensive Helikian transgressive blanket, deposited over the western part of the Shield after Hudsonian stabilization. Donaldson (1973) thought the group to be part of a largely terrestrial Helikian sequence which was the lateral equivalent of the marine Belt-Purcell Strata (see Purcell System, p. 17) of the geosyncline of the continental margin to the west.

In the Dismal Lakes area the apparently unmetamorphosed Hornby Bay Group was gently folded before the deposition of the Dismal Lakes Group and has suffered minor faulting. More intense folding to the west was related to orogenic events in the Helikian eugeosyncline of the Cordilleran region (Baragar and Donaldson, 1973).

The older age limit of the Hornby Bay Group is fixed by the Aphebian age of the Epworth Group and the Hudsonian age of the Great Bear Batholith, both of which underlie the group unconformably (see Epworth, Echo Bay, and Cameron Bay Group p.14). Intrusion of the Muskox body into the group and effusion of the Coppermine River basalts (Copper Creek Formation, Baragar and Donaldson, 1973) at a higher stratigraphic level, both at 1.2 Ga (Smith in Lowdon 1961; in Wanless et al. 1965; Baragar and Donaldson, 1973) provide a younger age limit. Thus a Paleohelikian age seems appropriate for the group. The writer has seen no record in the literature of mineralization in the group.

Dismal Lakes Group

The 1100 m thick (Fraser et al., 1970) Neohelikian Dismal Lakes Group (Baragar and Donaldson, 1973)(Fig. 5) is the upper of two parts of the Hornby Bay Group of earlier writers. The group is exposed in a curvilinear east-trending strip between Great Bear Lake and Coronation Gulf. It unconformably overlies the Hornby Bay Group (s.s.) exposed to the south, is conformably overlain by the Coppermine River volcanics (Copper Creek Formation of Baragar and Donaldson, 1973) exposed to the north, and is cut by sills and north-trending dykes (Baragar and Donaldson, 1973).

The lowest of the five units of the group is a buff to white orthoquartzite, some 300 m thick, containing interlayered black carbonaceous shale and desiccation structures in its upper part. The second unit, 120 m thick, comprises red mudstone overlain by tan dolomite (unit 13 of Baragar and Donaldson, 1973). The mudstone contains ripple marks, abundant halite casts and possible gypsum pseudomorphs. The uppermost three units are a lower laminated dolomite, and a massive stromatolitic dolomite, each 180 m thick, and an upper laminated dolomite, 270 m thick. The laminated dolomite units contain abundant layers of chert nodules, intraformational conglomerates, oolites, stromatolites, molar-tooth structure and possible desiccation structures (Baragar and Donaldson, 1973; Donaldson, 1969).

Crossbeds in the Dismal Lakes Group indicate northwestward transport of detritus, and stromatolite elongation indicates bimodal onshore and offshore tidal currents in a shallow marine environment (Fraser et al., 1970). The group is unmetamorphosed, dips gently northward and has suffered minor faulting (Baragar and Donaldson, 1973). Both the Muskox intrusion which seems to predate the Dismal Lakes Group (Irvine, 1970), and the Coppermine River volcanics have been assigned an age close to 1.2 Ga (Smith in Lowdon 1961, in Wanless et al. 1965; Baragar, 1972).

Coppermine River Group (Baragar and Donaldson, 1973)

The Neohelikian Coppermine River Group (Fig. 5) of earlier workers, e.g. Fraser et al. (1970) has been redefined and divided by Baragar and Donaldson (1973) into lower and upper parts, the Coppermine River Group, and the Rae Group, respectively. The group (s.s.) stretches in an east-west belt about 240 km long and up to 30 km wide between Great Bear Lake and Coronation Gulf. In the west the belt disappears beneath flat-lying Paleozoic rocks. In the east and south it is conformably underlain by the Neohelikian Dismal Lakes Group. In the north the Coppermine River Group is overlain unconformably by the Helikian or Hadrynian Rae Group (Baragar and Donaldson, 1973).

The Coppermine River Group (s.s.) consists of a lower volcanic unit, the Copper Creek Formation, about 3 km thick, conformably overlain by a redbed unit, the Husky Creek Formation with a minimum thickness of 1200 m (Baragar and Donaldson, 1973). The Copper Creek Formation consists chiefly of saturated tholeiitic plateau basalt flows, with amygdaloidal altered reddish tops. Pyroclastics and pillows are very rare. Thin beds of red sandstone are interbedded with the lavas especially in the upper part of the sequence. The Husky Creek Formation is very poorly exposed and consists typically of red, crossbedded sandstone and siltstone and shale with subordinate interlayered basalt. The formation was interpreted as having been deposited during the waning of Copper Creek volcanism (Stockwell et al. 1970; Baragar and Donaldson, 1973). An east-west tensional phase of tectonics resulted in extrusion of the Coppermine River basalts (Irvine, 1971). Burke and Dewey on the other hand thought they were extruded on a failed arm extending eastward from the continental margin. In the Coronation Gulf region the unmetamorphosed Coppermine River Group dips north at 10 degrees or less (Baragar, 1967a) and is gently warped about a possibly northeast-trending axis.

The Copper Creek Formation has been assigned an age of 1.2 Ga by the Rb/Sr isochron method (Baragar in Wanless and Loveridge, 1972). Its paleomagnetic pole position is very close to that of the Muskox intrusion (Robertson, 1964) and to the Mackenzie dykes (Fahrig et al., 1965) both of which have radiometric ages of 1.2 Ga (Smith in Lowdon 1961, in Wanless et al., 1965; Fahrig and Jones, 1969).

Baragar and Donaldson (1973) reported that the only significant copper occurrences in the Coppermine River Group lie in the lava sequence of the Coppermine River Group and that the richest occurrences lie in the upper part of the sequence. Kindle (1969), who examined 75 occurrences, listed most as related to shears and fractures or as impregnations in dykes and flows. Rare occurrences were related to red sandstone adjoining dykes or disseminated in green sandstone. Thorpe (1970) summarized copper exploration in the group in the late sixties.

Rae Group

The term Rae Group as defined by Baragar and Donaldson (1973) is equivalent to the term "upper Coppermine River Group" of Fraser et al. (1970). Its outcrop area at the head of Coronation Gulf is an east-west belt up to 30 km wide extending for some 240 km (Fig. 5a). The group is Hadrynian in age, is about 1200 m thick, and is cut by diabase dykes and tholeiitic dolerite sills (Fraser et al., 1960; Baragar and Donaldson, 1973). In the south the group unconformably overlies the significantly truncated Coppermine River Group (s.s.) (Fraser et al., 1970) and older rocks (Baragar and Donaldson, 1973). In the north it is unconformably overlain by Paleozoic strata.

The sedimentary rocks of the group are divided by Baragar and Donaldson (1973) into seven units: (unit 19) fine grained drab sandstone interbedded with grey to black shale, some sandstone beds being glauconitic and containing sulphides including small amounts of disseminated chalcocite; (unit 20) red and green sandstone, siltstone and mudstone; (unit 21) similar to (unit 19); (unit 22), stromatolitic limestone and dolomite with grey to black siltstone and shale in the upper part; (unit 23) massive quartzose sandstone; (unit 24) red, green and rusty-weathering shale, siltstone and fine grained sandstone with desiccation cracks in some beds, stromatolitic dolomite near the base and local halite casts and massive gypsum beds; (unit 25) at least 90 m of carbonate with desiccation structures. Sedimentation was predominantly in shallow water. Scanty data suggest paleocurrents directed to the north to northwest (Fraser et al., 1970).

The group is unmetamorphosed and dips homoclinally and gently to the north and northwest. Interlayed dolerite sills have an average K/Ar age of 0.647 Ga (Wanless et al., 1966, 1970). Their paleomagnetic pole position (Robertson and Baragar, 1972) is close to that of the Franklin magnetic interval dated at 0.675 Ga (Fahrig et al., 1971). The Coppermine River basalts (Copper Creek Formation), unconformably overlain by the Rae Group are 1.2 Ga old.

Economically unimportant copper sulphide mineralization was found in at least two places in a dark shale or siltstone member within a "few feet" above the unconformity at the base of the Rae Group and the underlying Coppermine River Group (Baragar and Donaldson, 1973). These showings were examined in more detail by Kirkham (1970) and Thorpe (1970). Kirkham found the cupriferous unit at the base of the Rae Group to consist of two parts; a drab sandstone and siltstone part with stratigraphically controlled copper sulphides, and an overlying drab quartzite with quartz veins containing minor copper. This unit is overlain by a diabase sill and unconformably overlies gently folded hematitic, limy, crossbedded continental sandstone, siltstone and mudstone. The drab sediments contain desiccation cracks and possible rain imprints. The presence of the stratigraphically controlled copper sulphides in anoxic sediments overlying cupriferous flood basalts and redbeds (Coppermine River Group) reminded Kirkham of mineralization at White Pine, Michigan. Thorpe (1970) who reviewed mineral exploration in the late sixties in the Coppermine River area commented that this mineralization consisted of copper sulphides in nodules in glauconite-rich silty sandstone and laminated clayey siltstone.

Brock Inlier sequence

The Brock Inlier (Fig. 5a) located on the Arctic coast north of Great Bear Lake is Hadrynian in age, is overlain by lower Paleozoic rocks, and is very poorly exposed. It was mapped by Fraser et al. (1960) and by Balkwill and Yorath (1970) north of 69°N and Cook and Aitken (1969) south of 69°N. The sequence is at least 1800 m thick in the north and may be thinner in the south.

Both Cook and Aitken (1969) and Balkwill and Yorath (1970) divided their similar largely or entirely marine (Balkwill and Yorath, ibid.) sequences into five sedimentary units cut by gabbroic sills and dykes. They are, in ascending stratigraphic order (1) green and locally red shale and siltstone up to 900 m thick, unconformably overlain by (2), 120 to 250 m of dolomite with local stromatolites; (3) 120 to 460 m of grey to red quartzite; (4) dolomite with abundant stromatolites, minor green and maroon shale, and calcarenite, overlain with possible unconformity (Balkwill and Yorath, 1970) by (5), up to 15 m of dolomite with mudcracks, gypsum units and salt casts. Among the Cambrian rocks unconformably overlying unit 5 in many places is the evaporite-bearing Saline River Formation which has recently been discussed by Aitken and Cook (1974).

The sequence is unmetamorphosed and gently dipping. As it is younger than the Coppermine River lavas it is therefore younger than 1.2 Ga. It is intruded by gabbroic sills and dykes that yield K/Ar dates of 0.6 to 0.7 Ga (Balkwill and Yorath, 1970). They are probably correlative with those cutting the Rae Group that belong to the 0.675 Ga old Franklin paleomagnetic interval (Fahrig et al., 1971).

Shaler Group

The main exposure of the Neohelikian Shaler Group (Thorsteinsson and Tozer, 1962) lies in the Holman Syncline which trends northeast across the northwest part of Victoria Island (Fig. 5a). Sedimentary rocks near Wellington Bay on southern Victoria Island, assigned by Thorsteinsson and Tozer to the Shaler Group, may be stratigraphically equivalent to the Burnside River Formation of the Aphebian Goulburn Group (Young and Jefferson, 1975; F.H.A. Campbell pers. comm., 1976). At the eastern exposed limit of the Holman Syncline the Shaler Group rests unconformably on metasediments including reddish quartzite (Christie et al., 1972). The metasediments appear to have been intruded by Archean granite. The Shaler Group is unconformably overlain by Hadrynian volcanics, the Natkusiak Formation and by Ordovician dolomite (Thorsteinsson and Tozer, 1962). Thorsteinsson and Tozer divided the Shaler Group into five mutually conformable sedimentary formations. They are, in ascending order, the Glenelg, Reynolds Point, Minto Inlet, Wynniatt and Kilian formations. Published detailed descriptions are available for the Glenelg, and Reynolds Point formations (Young, 1974; Young and Jefferson, 1975; Miall, 1976; Young and Long, 1976) and the overlying volcanic Natkusiak Formation (Baragar, 1976). Red fluvial clastics at Wellington Bay were studied by Young (1974) but as mentioned above there is doubt about their age. An outlier of red sandstone and of quartz pebble conglomerate overlying Archean granite on strike with the Wellington High was thought to be the basal member of the Glenelg Formation (Thorsteinsson and Tozer, 1962). However, the rocks of the outlier are similar to some of those of doubtful age at Wellington Bay.

At Hadley Bay at the north end of the Holman Syncline the Glenelg Formation was divided into two members. The 600 m thick lower member consists of fine grained red, white, and grey sandstone and dark grey shale alternating in units up to 30 m thick, red and grey siltstone, and grey dolomite. At the head of Glenelg Bay a 370 m interval of the upper member consists mainly of light grey to red fine- to medium-grained quartzose sandstone overlain by stromatolitic dolomite 18 to 38 m thick (Thorsteinsson and Tozer, 1962). The upper member, which includes pyritic shale, was deposited under alternating shallow marine and deltaic conditions (Young, 1974). On the Richardson Islands and adjacent Victoria Island occur hard, red, coarse grained quartzose sandstone and quartz pebble conglomerate. About 400 m of red sandstone and interlayered gabbro sills overlying about 90 m of dolomite occur on the south coast of Banks Island. The rocks of these last three areas have been correlated with the Glenalg Formation (Thorsteinsson and Tozer, 1962, Young and Jefferson, 1975, Miall, 1976).

The overlying Reynolds Point Formation consists of marine-deltaic and distal fluvial clastics succeeded in turn by stromatolitic limestone of the marine shelf and by terrigenous fine grained clastics (Young, 1974). More than half of the succeeding, about 370 m thick, Minto Inlet Formation (Thorsteinsson and Tozer, 1962) is composed of gypsum and anhydrite. Other rock types are, in decreasing order of abundance, fine grained grey to red sandstone, limestone, shale and grey to red siltstone. Abruptly overlying the Minto Inlet Formation is the Wynniatt Formation, typically 820 m thick and composed mainly of dark limestone with stromatolitic horizons.

The 490 m thick Kilian Formation, a heterogeneous assemblage of thinly bedded varicoloured sediments overlies the Wynniatt Formation. Near Kilian Lake it is divisible into two equally thick members. The lower consists principally of gypsum and anhydrite, shale, red sandstone and siltstone and grey dolomite. The upper member is similar to the lower except that it lacks sulphates. Stromatolites have been seen in the formation. A third, overlying, member, 117 m thick, absent from the type area, is present south of Minto Inlet. It is composed of yellow to brown sandstone and conglomerate. Greyish red, fine grained sandstone is a minor constituent.

The upper contact of the Kilian Formation with the volcanic Natkusiak Formation is an erosional unconformity (Thorsteinsson and Tozer, 1962). The Natkusiak Formation and the mafic dykes and sills that cut the Shaler Group are all part of the Franklin (0.675 Ga) magmatic episode (Fahrig et al., 1971). The volcanics, which have a preserved thickness of 730 m, are significant in being the only known preserved effusive record of the episode. They are typical plateau basalts and are mostly subaerial (Baragar, 1976).

The presence of evaporites in some formations suggests that the Shaler Group was deposited at low latitude. Paleomagnetic studies (Robertson and Baragar, 1972) suggest that Victoria Island was at a low latitude during the Franklin magmatic episode. Further paleomagnetic studies by the Geological Survey of Canada are planned to determine whether the underlying Shaler Group was deposited close in time to the Franklin interval (Baragar, 1976).

The unmetamorphosed Shaler Group dips gently and its main outcrop area is thrown into two gentle folds, the Holman Syncline and the Walker Bay Anticline. The basement at the north end of the Holman Syncline has a K/Ar age of 2.4 Ga (Thorsteinsson and Tozer, 1962). A gabbro sill cutting the Shaler Group and part of the same magmatic episode as the Natkusiak Formation (Barager, 1976) has a K/Ar whole rock ages of 0.635 and 0.64 Ga (Christie, 1964).

Volcanics (Copper Creek Formation) dated at 1.2 Ga (Baragar in Wanless and Loveridge, 1972) underlie the Rae Group a possible correlative of the Shaler. Young and Jefferson (1975) suggested that stromatolites in the lower part of the Shaler Group are of latest Helikian – early Hadrynian type. Presence of the stromatolite **Boxonia** in the Wynniatt Formation suggests an age of 950 ± 50 Ma. (Aitken et al., 1978b). Thus on the above evidence a Neohelikian-Hadrynian age would seem likely for the group.

Mineralization has not been reported from the Shaler Group. The Natkusiak Formation volcanics contain copper (Thorsteinsson and Tozer, 1962) but unlike that of the Coppermine River volcanics the metal is not restricted to any stratigraphic level (Baragar, 1976).

Helikian redbeds of the Bathurst Inlet area

A sedimentary and volcanic sequence (up to 1100 m thick, Fraser and Tremblay, 1969; up to 1700 m thick, Campbell, 1978) lies on the shores and islands of Bathurst Inlet and on nearby Kent Peninsula (Fig. 5a). These rocks overlie Archean rocks of the Slave Province and folded rocks of the Aphebian Epworth Group (Campbell, 1978). Redbeds are contained in the Tinney Cove, Ellice River, Kanuyak and Algak formations (Fig. 5b). The basal Tinney Cove Formation, and correlatives underlying the Ellice River

Formation (Campbell and Cecile, 1975) are described on p.16, on account of their taphrogenic character and possible correlation with the Et-then Group.

Tremblay (1971) described a thickness of about 2200 m of Ellice Formation and associated rocks, covering an area of about 50 km^2 in the graben-like Bathurst Trench. The crossbedded Ellice Formation (Tremblay, 1971) is generally in fault contact with the surrounding Archean rocks though an unconformable relationship was seen at one place. It consists of 900-1800 m of grey to pink orthoquartzite with local red layers. According to Tremblay (1971) the Ellice Formation lies within a structural basin, was deposited generally in a fairly quiet and shallow basin, and is a probable correlative of the Tinney Cove Formation. However Campbell and Cecile (1975) and Campbell (1978) who both renamed the formation the Ellice River Formation, found it to overlie unconformably the Tinney Cove Formation, giving it a thickness of about 600 m. According to Campbell (1978) it consists of a thin basal conglomerate and a 10-20 m thick vesicular basalt unit overlain by 500 m of crossbedded reddish, pink and white quartzite (E_2 member) all of fluvial origin, and 30 m of red mudstone and siltstone (E_{2M} submember) of intertidal origin.

The succeeding Parry Bay Formation is composed mainly of dolomite and limestone with interbedded sandstone and shale. The carbonate includes chert lenses, and stromatolitic layers which contain bituminous material (Fraser, 1964). The 220 m thick formation is conformable on the Ellice River Formation and was interpreted as containing a subtidal to intertidal southern facies and a subtidal northern facies (Campbell, 1978).

The succeeding Kanuyak Formation, 0-30 m thick, overlies the Parry Bay Formation unconformably to disconformably. It consists, in ascending order, of dolomite megabreccia, red arkose, red tidal mudstone and dolomite. The megabreccia may have formed by evaporite solution-collapse or by the development of karst in the Parry Bay Formation (Campbell, 1978).

Conformably overlying massive basalt flows, named the Ekalulia Formation by Campbell (1978), are overlain conformably by black hematitic argillite and pink and white quartzite (Fraser, 1964). These sediments have been named the Algak Formation and described (Campbell, 1978) as crossbedded, reddish purple arkose and siltstone, minor mudstone and shale arranged in fining-upward cycles on a scale of usually less than a metre.

Fraser et al. (1970) related sedimentation and tectonics in the Bathurst Inlet area as follows: The redbeds preserved at the south end of the Bathurst Trench indicate tensional faults followed by strike-slip movement. These sediments are inferred to have been deposited in a fault basin in response to rapid and continued uplift in adjacent source areas. An unconformity at the top of the redbeds marks the end of the conditions of redbed deposition. Conglomerate at the base of the Tinney Cove and Ellice formations indicates renewed block faulting followed by deposition of fluvial sandstone under more stable conditions. Persistence of stable conditions in the central and northern parts of the basin is indicated by the shallow marine carbonates of the Parry Bay Formation. Renewed tension and block faulting resulted in depression and preservation of the Helikian sediments in the central part of the basin and the extrusion of the Coppermine River volcanics.

According to Campbell (1978) deposition of the Tinney Cove Formation was controlled by syndepositional transcurrent faulting. Reactivation of faults controlled the dispersal pattern of the Ellice River Formation. On the west side of Bathurst Inlet the entire Proterozoic succession has been stripped from the Archean basement.

The sequence at Bathurst Inlet is unmetamorphosed and mildly deformed. The underlying Goulburn Group has been intruded by Hudsonian granites (Fraser and Tremblay, 1969). Elsewhere, the Coppermine River basalts, equivalent to the Ekalulia Formation have been assigned an age of 1.2 Ga (Baragar and Donaldson, 1973). Native copper occurs in the Ekalulia Formation. Black mudstone and shale units in the Parry Bay Formation on Kent Peninsula contain disseminated chalcopyrite and have a radioactivity of ten times background (Campbell, 1978).

Campbell (1978), Young (1977) and Aitken et al. (1978b) have proposed stratigraphic correlations for rocks within the Amundsen Basin. The last two papers also deal with correlation of these rocks with the Proterozoic of the Mackenzie Mountains of the Cordillera.

A point of significance is that Young (1977) and Jefferson and Young (1977) correlate the cupriferous horizon at the boundary between the Redstone River and the overlying Coppercap formations of the Mackenzie Mountains (p. 19) with the boundary between the Minto Inlet and overlying Wynniatt formations of the Shaler Group of Victoria Island whereas according to Aitken et al. (1978b) the topmost sedimentary formation of the Shaler Group, the Kilian Formation, is equivalent to the Redstone River Formation. The latter view suggests that the cupriferous horizon may be absent on Victoria Island. Further, according to Eisbacher (1978) the Redstone River Formation did not extend to Victoria Island and extrapolation of the mineral potential of the Mackenzie Mountains to Victoria Island is not warranted.

Late Proterozoic redbeds of northern Baffin Island and adjacent areas

Unmetamorphosed and gently-dipping Helikian and/or Hadrynian sedimentary rocks and associated mafic intrusives and volcanics overlie granitic gneiss of the Churchill Province on northern Baffin Island. (Eqalulik and Uluksan groups; Fury and Hecla Formation) and on northern Boothia Peninsula (Aston Formation) (Fig. 6a).

Eqalulik and Uluksan groups

The Eqalulik and Uluksan groups (Fig. 6a) are widespread on Borden Peninsula, northwest Baffin Island (Blackadar, 1956, 1970; Lemon and Blackadar, 1963), and to the east and on Bylot Island, north of Baffin Island (Jackson and Davidson, 1975; Jackson et al., 1975). Further field studies of these rocks by G.D. Jackson of the Geological Survey of Canada commenced in 1977. These rocks are overlain with angular unconformity by little-deformed Cambrian and/or Ordovician sedimentary rocks (Trettin, 1969) and are transected by northwest-trending and north-northwest-trending diabase dykes of the Franklin swarm (Fahrig et al., 1971).

Because of doubts about their age, correlation of these sediments is uncertain. The Adams Sound and Arctic Bay formations of the lower part of the succession are similar to the Fury and Hecla and Autridge formations (Lemon and Blackadar, 1963; Blackadar, 1970). The Thule Group of northwest Greenland and correlatives near Bache Peninsula (Frisch et al., 1978) on the east coast of Ellesmere Island as well as the Aston and Hunting formations of Somerset Island have been correlated with the Eqalulik and Uluksan groups (Lemon and Blackadar, 1963). Rocks of the Shaler Group mainly of Victoria Island have also been correlated with these groups (R.A. Olson, pers. comm., 1976).

Formations comprising the mainly platformal (Jackson and Uluksan groups in ascending stratigraphic order, with maximum thicknesses (Blackadar, 1970) include the Nauyat Formation (600 m) and Adams Sound Formation (4500 m) in the Eqalulik Group, and the Arctic Bay Formation (300 m), Fabricius Fiord Formation (1650 m), Society Cliffs Formation (300 m), Victor Bay Formation (490 m), Athole Point Formation (1500 m), Strathcona Sound Formation (1200 m) and Elwin Formation (1500 m) in the Uluksan Group. Redbeds are present in the Adams Sound, Society Cliffs, Strathcona Sound and Elwin formations.

The Nauyat Formation is a sequence of thin flows of amygdaloidal, massive or pillowed basalt, andesite and thin tuff beds (Blackadar, 1970). The first of four phases of uplift during deposition of the sequence caused removal of the Nauyat Formation from the eastern part of the Borden Peninsula allowing the succeeding Adams Sound Formation to overstep the Nauyat volcanics onto the Aphebian basement. In the western part of the peninsula, pale orange to reddish brown sandstone of the Adams Sound Formation lies conformably on the Nauyat Formation. The sandstone is quartz-rich and contains interlayers of quartz pebble conglomerate and shale. Though in many sections red iron oxide coats the quartz grains (Blackadar, 1970) the formation is not everywhere red, for example at Tay Sound about 200 km to the east (Jackson et al., 1975). The redness of the formation on Bylot Island appears to be related to proximity to the Byam Martin High, a local source area composed of Aphebian gneisses (Jackson and Davidson, 1975). Crossbeds and ripples are locally common in the formation (Blackadar, 1970). Trough-crossbed orientations reflect fluviatile transport of detritus from the southeast (Geldsetzer, 1973a).

Sediments deposited during transition to the conformable Arctic Bay Formation reflect a floodplain environment with decreasing influence of the source area to the southeast, and southeastward transgression of the subtidal to intertidal black shale of the Arctic Bay Formation. Following a second minor phase of uplift, this time in the western part of Bordon Peninsula, an algal carbonate platform formed and is represented by the Society Cliffs Formation.

Blackadar's (1970) stratigraphic sequence has in part been revised by Geldsetzer (1973a) who interpreted the lower part of the Fabricius Fiord Formation, a black shale, as a lateral equivalent of the Arctic Bay Formation. He interpreted the arkosic upper part of the Fabricius Fiord Formation, found in central Borden Peninsula, as a facies of the Society Cliffs Formation derived from a local basement source area.

In Society Cliffs time a possible easterly source of clastics is reflected in westward-thinning tongues in the carbonates up to 45 m thick of varicoloured, fine grained sandstone and of siltstone with mudcracks and associated gypsum beds. These tongues were interpreted as the sediments of evaporitic pans on a westward-prograding floodplain (Geldsetzer, 1973a). Similar red and varicoloured clastics and gypsum beds on southwest Bylot Island and on the west side of Tay Fiord were reported by Jackson and Davidson (1975) and Jackson et al. (1975) respectively.

A third phase of uplift, again more pronounced in the western part of the area, led to development of karst in the Society Cliffs Formation carbonates (Geldsetzer, 1973b). Renewed subsidence permitted deposition in subtidal to intertidal environments of sediments of the Victor Bay Formation that consisted of fine grained, black, clastics and calcareous lutite grading up into algal carbonates.

Carbonate sedimentation was interrupted by a fourth regional uplift and subsequent deposition on a slight angular unconformity of a thick (1500 to 2100 m) molasse wedge. In the northern part of Borden Peninsula, the lower part of this wedge is the Strathcona Sound Formation and the upper part, the Elwin Formation. The eastern thin lateral equivalent of the wedge is composed of fine clastics of the Athole Point Formation (Geldsetzer, 1973a). The Strathcona Sound Formation was divided into a sequence composed of dark red mudstone and shale and a generally overlying sequence composed of grey sandstone and siltstone (Blackadar, 1970). Geldsetzer (1973a) recognized in the formation a lower sequence of cycles each consisting of basal carbonate-conglomerate grading up into coarse grained sandstone and an upper sequence of cycles in which coarse grained sandstone graded up into fine grained sandstone. The more brightly coloured upper part of the molasse sequence, the Elwin Formation, is also cyclic but contains increasing amounts of finer grained clastics in the upper part of individual cycles. Well developed trough crossbedding, indicating a western provenance, ripples, and fining-upward cycles and quartzofeldspathic composition of the sandstones in the thick western part of the Strathcona-Elwin redbed sequence, point to its deposition on a floodplain with derivation from a granitoid source.

Toward the east the clastic wedge decreases in thickness to 600 m, and grades into the Athole Point Formation, a possibly subtidal succession of commonly flaser-bedded, fine grained sandstone and siltstone and some thin carbonate layers (Geldsetzer, 1973a).

The Eqalulik and Uluksan groups are preserved in grabens between northwest-trending horsts (Trettin, 1969) that may comprise an aulacogen related to the Franklin Geosyncline to the northwest (Olson, 1977). Faulting in the rift zone occurred during and after deposition of the two groups (Jackson and Davidson, 1975).

The group is horizontal to gently inclined (Blackadar, 1970) or gently folded about northwest-trending axes except where steeply dipping adjacent to faults at the margins of grabens (Jackson et al., 1975). Gravity studies (Berkhout, 1973) suggest that the group is 14 000 m thick in a trough north of Trettin's (1969) central Borden fault zone.

Since the Proterozoic sediments overlie Hudsonian gneisses of the Churchill Province nonconformably and are overlain unconformably by Cambrian and/or Ordovician sedimentary rocks, they are of Helikian and/or Hadrynian age. A K/Ar whole-rock age from basalt from the basal Nauyat Formation of 0.903 Ga (Blackadar, 1970) indicates a Hadrynian age for the succeeding sediments. Potassium/Argon ages of 1.14 and 0.915 Ga from a diabase dyke cutting the sediments which were regarded by Geldsetzer (1973a) as minimum ages due to the possibility of argon loss, indicate that the sediments might be Helikian. Though the dyke, redated by the K/Ar method, yielded an age of 0.439 and 0.437 Ga Fahrig et al. (1971) considered it part of the Franklin diabase dyke swarm, intruded at a mean age of 0.675 Ga.

Paleomagnetic properties of the Franklin diabases suggest that the northern part of the Canadian Shield, lay at a low latitude when they were intruded into it. The presence of gypsum and other sedimentary criteria in parts of the Eqalulik and Uluksan groups indicate deposition of these sediments at low latitude also. Thus Fahrig et al. (1971) concluded that they may have been deposited close in time to the intrusion of the diabase, that is during the Hadrynian Era.

According to Jackson et al. (1975) the Arctic Bay and Society Cliffs formations merit examination as favourable strata for kupferschiefer-type copper mineralization. Mississippi Valley-type lead-zinc mineralization occurs in the karsted carbonates of the Society Cliffs Formation (Geldsetzer, 1973b; Olson, 1977).

Fury and Hecla Formation

These rocks are exposed mainly on the shores and islands of Fury and Hecla Strait (Fig. 6a) which separated Baffin Island from Melville Peninsula (Blackadar, 1958). The outcrop belt extends mainly along the north shore for about 200 km where it is up to 40 km wide. It is overlain locally and with gradational contact by the Autridge Formation which consists of black shale, slate and ferruginous dolomite. The basal Nauyat (volcanic) Formation of Eqalulik Group to the north, is not present.

The Fury and Hecla Formation is dominantly greyish orange, pink to pale red "quartzite and quartzitic sandstone" (Blackadar, 1970). Conglomerate layers with sedimentary clasts at the base of and within the formation, and grit layers are of minor importance. Other minor lithologies include dark red and purple-red quartzite and red and black shale, some of which is mudcracked.

Based on an average 10 degree dip over the outcrop area the formation exceeds 4600 m in thickness. Dip of the strata toward Fury and Hecla Strait suggests that they are preserved in an east-striking basinal structure (Blackadar, 1970). Two K/Ar dates from a gabbroic sill overlying the Autridge Formation of 0.639 Ga and 0.475 Ga lie within the range of those of the Franklin diabases (Fahrig et al., 1971). Northwest orientation of associated dykes, similar to that of the Franklin Dykes suggests that this sill is of Franklin age.

Aston Formation

This formation (Fig. 6a) is exposed mainly on Somerset Island, on Prescott Island, and on Pandora Island immediately south of Prescott Island. (Blackadar and Christie, 1963; Brown et al., 1969; Tuke et al., 1966.) The area underlain by the formation is about 250 km^2. The Aston is conformably overlain by the dolomitic Hunting Formation and both are cut by dykes and sills of diabase. The maximum thickness of the Aston Formation is 800 m of which about a third is diabase sills. Its outcrop width varies up to 3 km (Tuke et al., 1966).

Basal breccia of the formation lies with angular discordance on granitic gneiss and is overlain by sandstone grading up into sandy dolomite. Scarcity of this breccia implies that the surface of the unconformity is of low relief (Tuke et al., 1966). In the southern part of the outcrop area at the north end of Somerset Island locally pebbly white quartzite is the basal unit. A few metres above the base is a 10 m - thick stromatolitic dolomite unit which is traceable throughout the Aston outcrop. Except for a few metres of shale, quartz-rich, silica-cemented, red sandstone is the only rock type above the dolomite. The hematite, related to bedding, was presumably detrital, and derived from lateritic weathering of mafic units of the basement. The red sandstone contains rare bands of pebbles of quartz, feldspar and granitic gneiss. Layers of mud flakes are locally abundant but generally have been scoured away. Sedimentary structures in the red sandstone include lamination, crosslamination, units of planar crossbeds up to 30 cm thick, and less abundant trough crossbeds. Symmetrical ripple marks with northeast-trending axes, desiccation cracks and parting lineations occur locally.

The Aston sandstones lack features of nonmarine deposition and contain ripples, desiccation cracks and stromatolites and therefore were interpreted as having been deposited in a shallow marine environment (Tuke et al., 1966). Crossbed azimuths indicated transport northwestward from the Boothia Peninsula gneiss (Tuke et al., 1966), and Kerr and Christie (1965) believed the Aston to have been deposited over the present Boothia Uplift. The Aston Formation grades up through 70 m of interbedded sandstone, shale and dolomite into the 2250 m thick dolomitic Hunting Formation that contains abundant stromatolites in its upper part (Tuke et al., 1966)

Though the Aston and Hunting formations are lithologically similar to part of the Helikian or Hadrynian Eqalulik Group of Baffin Island (Blackadar, 1967) Tuke et al. (1966) suggested a lower Paleozoic age for the Aston Formation because of apparent conformity between the Aston, Hunting and overlying "undoubtedly" Ordovician rocks and because of lithological similarity to a Cambro-Ordovician succession directly overlying the gneiss of Boothia Peninsula. More

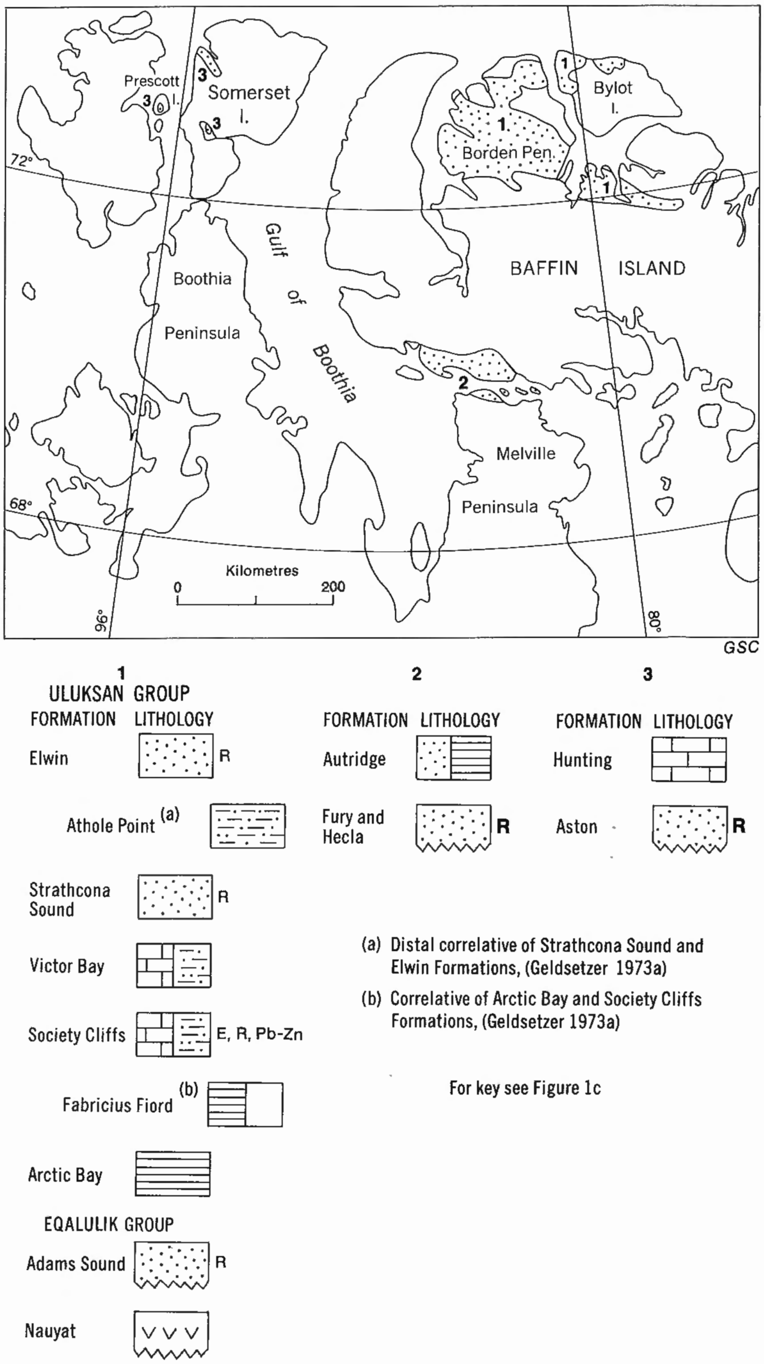


Figure 6. Late Proterozoic redbeds of northern Baffin Island and adjacent areas.

recently Dixon (1974) found an unconformity above the Hunting Formation as well as Cambrian fossils in the above "undoubtedly Ordovician rocks". He also reported a 0.702 Ga K/Ar age from a biotite-hornblende separate from a diabase sill in the Aston Formation.

Aulacogen-filling sequences of the Grenville-Superior Rift Zone

Two Late Helikian sequences, the mainly continental Seal Lake Group and the marginal marine Sibley Group, have been interpreted as having been deposited in failed arms associated with a continental rift system of mid-Helikian age (Fig. 7a). They consist mainly of fine grained clastics and some carbonates and are associated with plateau basalts.

Sibley Group

The 150 m (Card et al., 1972), (300 m (Coates, 1972)) thick Sibley Group underlies an area of 145 km². Its distribution is shown by Pye (1968), Pye et al. (1966), Pye (1970) and Pye and Fenwick (1964). On Sibley Peninsula in Lake Superior it overlies Aphebian Animikie Group rocks disconformably to unconformably. Northward it overlies Archean rocks nonconformably. To the southeast on islands off the north shore of Lake Superior it is overlain by the Osler volcanics, an amygdaloidal tholeiitic plateau basalt sequence, that forms part of the Keewatin volcanics (Card et al., 1972; McIlwaine et al., 1974; Stockwell et al., 1970).

McIlwaine et al. (1974) have informally applied the names Pass Lake, Red Rock and Kama Hill to the three formations comprising the group. The Red Rock Formation has been renamed Rossport, and the three names formalized by Franklin et al. (in prep.) whence the following descriptions of type-sections are taken. The Pass Lake Formation thins rapidly northward, and consists of a buff to pink quartz arenite with minor red interbeds, underlain by 3 m thick lenses of basal conglomerate. Towards the west clasts within it are larger and units thicken. Bedding thins upward and directional structures are limited to crossbeds and ripples.

The red dolomitic Rossport Formation, generally about 100 m thick lies disconformably to unconformably on the Pass Lake and nonconformably on Archean rocks. The overlying Kama Hill Formation is an interlayered mudstone and siltstone with up to 50 m preserved thickness. It is distinguished from the Red Rock Formation by its lack of carbonate and by its deep red to purple colour. Stromatolites have been found in the formation north of the town of Nipigon.

The conglomerate of the Pass Lake Formation was deposited in fans along active fault scarps and the quartz arenite was deposited as shelf sands. The Rossport Formation reflects marine conditions. The Kama Hill Formation was deposited on quiet saline mud flats at low latitude (Franklin et al., in prep.). Paleocurrent data are limited. Coates (1972) reported southward (170°) transport in rippled mudstone. Card et al. (1972) illustrated paleocurrents in several directions. Minor redbeds among the overlying Osler Volcanics contain sedimentary structures indicative of currents flowing southeastward (McIlwaine and Wallace, 1970).

Card et al. (1972) regard the Sibley Group as a postorogenic "typical molasse". They consider the area underlain by the Sibley Group and the overlying Osler Volcanics a graben or half-graben structure and probably a subsidiary basin of the Lake Superior structure. The Lake Superior structure itself has been likened to part of a continental rift structure (Kumarapeli and Saull, 1966). Franklin et al. (in prep.) describe the Sibley "graben" as a failed arm. This term was also implied by McIlwaine et al. (1974). According to Franklin et al. (in prep.) it is uncertain whether the graben formed before or after deposition of the Sibley Group and the Osler Volcanics.

The Rove Formation of the Animikie Group, which underlies the Sibley Group, has yielded an Rb/Sr isochron age of 1.550 Ga (Wanless and Loveridge, in prep.). Keweenawan volcanics, of which the Osler Volcanics are a part, were extruded 1.0-1.2 Ga ago (Goldich, 1968). Geological, paleomagnetic and radiometric data suggest deposition close to 1.4 Ga (Card et al., 1972). An Rb/Sr isochron from the Kama Hill and Redrock formations (Wanless and Loveridge, in prep.) yielded an age of 1.294 Ga.

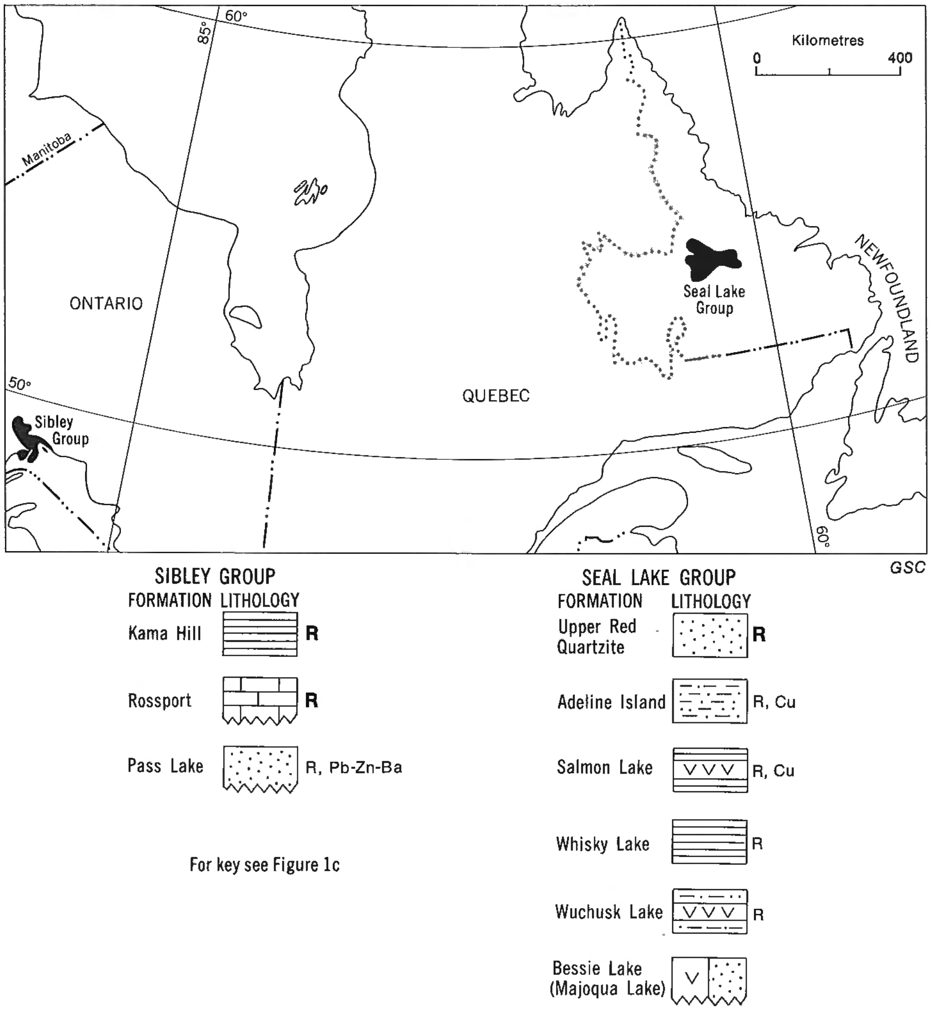


Figure 7. Aulacogen-filling sequences of the Grenville-Superior rift zone.

Lead-zinc-barite veins are associated with the feather edge of the Pass Lake Formation against the Archean basement. The metals are derived from clays in the Kama Hill Formation or from dolomite in the Rossport Formation. They were transported within the Pass Lake quartz arenite, which acted as an aquifer, and were deposited at the boundary between the Pass Lake and Red Rock formations as the result of a pH change (Franklin and Mitchell, 1977).

Seal Lake Group

The Seal Lake Group (Fig. 7a) underlies an area of about 2100 km^2 (Brummer and Mann, 1961) and its general geology is shown on maps by Fahrig (1959), Emslie (1964) and Roscoe and Emslie (1973). On the south the group is separated by south-dipping thrust faults from granites and gneisses of the Grenville Structural Province which were derived in part from the Seal Lake Group. In the north and west the group overlies anorthosite and granitic gneiss. On the east it is faulted against the older Croteau Group volcanics and at its southwest margin it overlies the virtually identical Letitia Lake Group volcanics (Ghandi and Brown, 1975) (Baragar, pers. comm., 1976). No younger rocks overlie the eroded top of the Seal Lake Group. The thickness of the group has been estimated at 10 400 m (Brummer and Mann, 1961) and 6400 m (Baragar, 1969).

Brummer and Mann (1961) divided the sedimentary and volcanic rocks of the Seal Lake Group into six formations: (1) The Bessie Lake and its northern equivalent, the Majoqua Lake, (2) the Wuchusk Lake, (3) the Whisky Lake, (4) the Salmon Lake, (5) the Adeline Island and (6) the Upper Red Quartzite. The Bessie Lake Formation overlies peneplaned older rocks. It is about 900 m thick, a blue, white and pink quartzite with a gritty, feldspathic and locally conglomeratic base. Amygdaloidal basalts are interlayered with the sediments. The Wuchusk Lake Formation, about 6100 m thick, consists of closely spaced dolerite sills with minor interlayered red and green siltstone and shale, red quartzite and arkose, rare black shale, limy shale and carbonate layers and two stromatolitic horizons (Brummer and Mann, 1961; Baragar, 1969). The Whisky Lake Formation, about 900 m thick, is composed of red and green shale, phyllite and minor quartzite and lacks the sills of the Wuchusk Lake Formation. The Salmon Lake Formation, also about 900 m thick, is lithologically very similar to the Whisky Lake except that it contains widespread basalt flows and minor diabase sills. It grades by decreasing abundance of basaltic flows into the 430 m thick (170 m according to Ghandi and Brown, 1975), Adeline Island Formation, that consists of green, grey, black, red and purple shale and slate interbedded with white and pink quartzite. The Upper Red Quartzite Formation, 800 m thick, consists of massive to well bedded quartzite with little variation (Brummer and Mann, 1961; Ghandi and Brown, 1975).

But for a thin zone of dolomite and black shale in the middle of the group (Wuchusk Lake Formation?) all other units indicate that sedimentation was continental. Volcanism during Bessie Lake-Majoqua and Salmon Lake time was amygdaloidal, pillows and tuffs being rare (W.R.A. Baragar, pers. comm., 1976). Mann (1959) viewed the lowest and uppermost formations, the Bessie Lake and the Upper Red Quartzite, as strandline deposits. Some deepening and fluctuation of sea level occurred during Wuchusk Lake time. Terrestrial deposition commenced again at the beginning of Whisky Lake time. The sills of the Wuchusk Lake Formation are probably genetically related to the chemically similar Salmon Lake volcanics (Mann, 1959).

The Seal Lake volcanics are typical plateau basalts and as such are chiefly characteristic of fault of rift-controlled volcanism in cratonic regions. In support of this the similarity of the upper and lower sequences indicates a source of very large dimensions (W.R.A. Baragar, pers. comm., 1976). The sediments of the group may have been deposited in a rift (failed arm?) striking northwest from an obliterated triple junction (phantom junction) lying on a southweststriking continental rift now closed within the Grenville Structural Province (Burke and Dewey, 1973). The Red Wine alkalic complex, of similar age to the Seal Lake Group (Baragar in Wanless and Loveridge, in prep.) lies within gneisses immediately south of the Seal Lake Group. Its genetic connection to the group, if valid, would support association of the Seal Lake Group with rifting (W.R.A. Baragar, pers. comm., 1976). The Sibley Group and associated Keweenawan volcanics as well as the Eriksfjord Formation of continental sandstone and associated Gardar plateau basalts of Greenland (Van Breemen and Upton, 1972) might have similar geneses along this Grenville rift system (Baragar in (Baer et al., 1974)).

The Seal Lake Group has been folded into an arcuate synclinorium and overturned northward in the south against the Grenville Front. Metamorphism rises southward to greenschist facies at the south margin of the synclinorium (Gandhi and Brown, 1975).

The Harp Lake anorthosite, which on field evidence predates the Seal Lake Group (R. Emslie, pers. comm., 1976) was intruded 1.46 Ga ago (U/Pb method, Krogh and Davis, 1973). Basalt and gabbro within the Seal Lake Group have been dated by Rb/Sr Isochron at 1.278 Ga (Wanless and Loveridge, in prep.). The Bruce River Group (Upper Croteau Group volcanics), unconformable beneath the Seal Lake Group (Marten and Smyth, 1975), have a Rb/Sr isochron age of 1.474 Ga (Wanless and Loveridge, 1972).

Many small copper occurrences in the Salmon Lake and Adeline Island formations are associated with basaltic lavas or diabase sills or the sedimentary rocks adjacent to them (Brummer and Mann, 1961). Copper is mainly in the form of native copper in the Salmon Lake Formation and in the form of sulphides in grey sediments immediately overlying a quartzite in the Adeline Island Formation (Ghandi and Brown, 1975).

Hadrynian redbeds of the southeastern Appalachians and the Jacobsville Formation of Lake Superior

The late Proterozoic to Carboniferous rocks of Newfoundland (Fig. 8a, 8b) have been divided into four parallel northeast-trending structural belts that have been traced farther along the Appalachian Orogen. The easternmost of these, the Avalon Zone consists of Hadrynian volcanic and sedimentary rocks cut by granite that is overlain by Cambrian shale.

The Avalon Zone in Newfoundland consists of three assemblages of rocks. (Strong et al., 1978). The lowest assemblage consists of marine and terrestrial volcanics of the Harbour Main Group (McCartney, 1967) at least 0.715 Ga old (Anderson, 1972) and the possibly correlative Love Cove Group (Jenness, 1963). These rocks are overlain by turbidites of the Conception Group and the broadly lithologically similar Connecting Point Group (McCartney, 1967). The middle assemblage consists mainly of coarse grained, red, sedimentary rocks derived from the lowest assemblage and belonging to the Cabot Group (term now discarded, see Williams and King, 1976), the Musgravetown, Hodgewater and the Rencontre groups. The uppermost assemblage consists of Infra-Cambrian to Lower Cambrian orthoquartzite and Lower Cambrian to Lower Ordovician sediments.

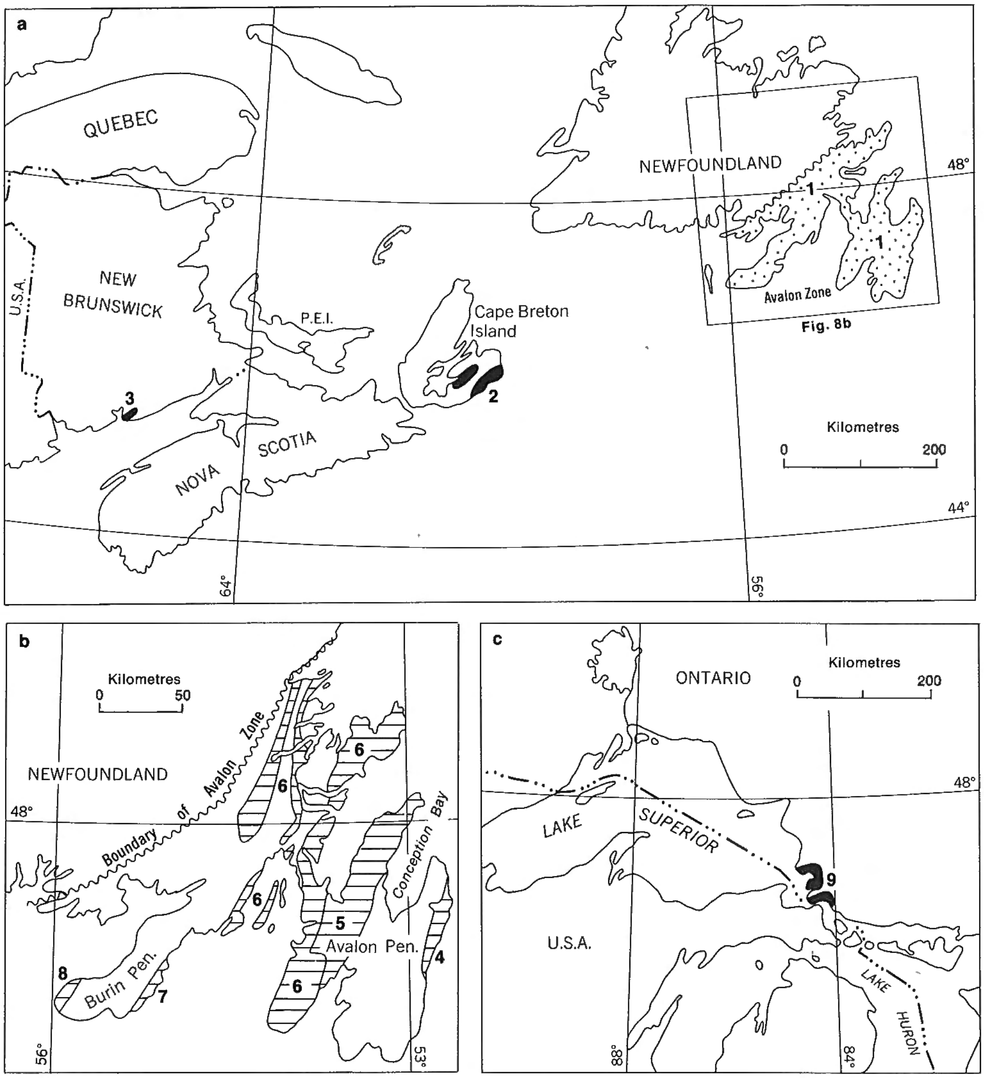
The rocks of the Avalon Zone are associated with granitoids, some associated with volcanism during the deposition of the middle assemblage and others related to the Late Paleozoic Acadian Orogeny. The lowermost assemblage has developed a cleavage or schistosity as a result of a Late

Precambrian "Avalonian Orogeny". Acadian deformation dies out eastward across the Zone (Strong et al., 1978). There is much speculation about the tectonic environment of the Avalon Zone (e.g. Hughes and Brückner, 1971; Poole, 1976; Schenk, 1971; Strong et al., 1978). Hughes and Brückner (1971) considered the redbed-bearing middle assemblage to be detritus eroded from volcanic islands during slow post-volcanic isostatic rise. Poole (1976) envisaged it to have been deposited in taphrogenic troughs.

Redbeds of the Avalon Zone are present in the lowest assemblage in the Conception Group where they comprise the 160 m thick Hibbs Hole Formation that consists of red and purple laminated slate and siltstone. Similar rocks occur elsewhere in the group on Conception Bay (McCartney, 1967).

Redbeds are well represented in the mutually correlative groups of the middle assemblage. The Cabot Group[1], some 4290 m thick, extends for about 80 km along the eastern coast of Avalon Peninsula. It consists of three mutually conformable formations. The lowest, the St. John's[2], 400 to 730 m thick (McCartney, 1969) consists of black, gritty, locally pyritic slate and argillite. It passes up into the Signal Hill[2] Formation, about 2300 m thick, drab sandstone overlain by dark red hematitic sandstone and red conglomerate. The overlying Blackhead Formation is about 1680 m thick and similar to the Signal Hill. It is arkosic and contains a basal reddish member, a middle greenish grey member and an uppermost reddish sandstone member (Rose, 1952). Brückner (1969) interpreted the Cabot Group as having been deposited in a prograding deltaic-alluvial environment.

The Hodgewater Group lies west of the Cabot Group and is separated from it by the rocks of the Holyrood Horst (Williams in Poole et al., 1970). It consists in ascending stratigraphic order of the following mutually conformable formations: the Carbonear, 980 to 1220 m thick, the Halls Town 860 to 1530 m thick, the Whiteway, 100 m thick and the Snow's Pond 950 to 2170 m thick (McCartney, 1967). The Carbonear Formation transitionally overlies the Conception Group and consists of dark grey slate and minor siltstone. The Halls Town is a massive grey to green-grey arkose, with a dull red member at the base, locally up to 60 m thick. The Whiteway Formation is characterized



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INFRACAMBRIAN ? QUARTZITES (for locations see Figures 8a, 8c)

ONTARIO (LAKE SUPERIOR) Formation	Lithology	NEW BRUNSWICK Formation	Lithology	CAPE BRETON ISLAND Formation	Lithology	NEWFOUNDLAND Formation	Lithology
Jacobsville[9]	R	Ratcliffe Brook[3]	R	Morrison River[2]	R	Random**	R

HADRYNIAN RED BEDS OF NEWFOUNDLAND* (for locations see Figure 8b)

BURIN PENINSULA		MUSGRAVETOWN GROUP[6]		HODGEWATER GROUP[5]		CABOT GROUP[4]	
		Crown Hill	R	Snows Pond	R	Blackhead	R
Rencontre Formation[8]	R	Trinny Cove	R	Whiteway	R	Signal Hill	R
Marystown Group[7]		Maturin Ponds	R	Halls Town	R	St. John's	Py
		Big Head	R	Carbonnear			
For key see Figure 1		Bull Arm					

* These sequences broadly correlative (Strong et al, 1978)
** Overlies rocks 4-6 of Avalon Zone[1] (see Figures 8a, 8c)

a – Location map, southeastern Appalachians; *c – Location map of the Jacobsville Formation;*
b – The Avalon Zone of Newfoundland; *d – Table of formations.*

Figure 8. Hadrynian redbeds of the southeastern Appalachians and the Jacobsville Formation of Lake Superior.

[1] term now discarded
[2] elevated to group rank (Williams and King, 1976)

by purple-red slate, siltstone, or sandstone but may be drab locally. Redbeds form about one third of the thickness of the formation. Also red and drab strata are commonly interlayered. The Snows Pond Formation consists of a lower part, 340 m thick, of red arkose and siltstone and drab arkose, siltstone and slate, and an upper part, 610 to 1830 m thick, of red and drab arkose and siltstone. The formation thins northward by truncation at the angular unconformity beneath the Random Formation. Some crossbed measurements and westward thickening of conglomerate indicate derivation of detritus from the west.

The Musgravetown Group overlies the Connecting Point Group and has been divided into the basal, volcanic Bull Arm Formation, 2440 m thick and four probably conformable (Strong et al., 1978) overlying sedimentary formations as follows, in ascending stratigraphic order: the Big Head, 460 to 2140 m thick, the Maturin Ponds, 330 to 610 m thick, the Trinny Cove 670+ m thick, and the Crown Hill, 0 to 300 m thick, as well as an undivided red and drab sedimentary unit 360+ m thick. The mainly subaerial (Strong et al., 1978) mafic to felsic Bull Arm contains subordinate amounts of sediments, some red (McCartney, 1967). The Big Head Formation has a basal division of either red conglomerate and arkose or red basalt and siltstone 30 to 370 m thick. Dominant rock types in the formation are drab siltstone, argillite and arkose (McCartney, 1967). The redbed Maturin Ponds Formation consists of arkose, siltstone and conglomerate. Paleocurrents travelled from the northwest and the Isthmus Horst may have been a source area. The fluviatile Trinny Cove Formation consists of irregularly bedded red and green conglomerate to coarse arkose. Direction of increase of the proportion of conglomerate as well as a few crossbed measurements indicate a sedimentary source to the west or the northwest.

The Crown Hill Formation consists of 60 to 150 m of red siltstone and arkose. Red pebble conglomerate comprises the rest of the formation. The formation can be traced at least 65 km over a north-northeast direction but appears to have a limited extent in an east-west direction. This linear distribution is common in the Hadrynian sediments and volcanics of the southeast Newfoundland area. The group was deposited in a dominantly deltaic depositional environment (McCartney, 1969).

Fluvial conglomerate to siltstone both red and green, occur as lenses and isolated patches within subaerial volcanics of the Marystown Group of the southeast Burin Peninsula amongst rocks considered by Williams (1967) to be equivalent to the Musgravetown Group (Strong et al., 1978). At the southwest tip of Burin Peninsula the Marystown Group is overlain disconformably by about 130 m of red sandstone and mudstone of the Rencontre Formation and the drab and red siltstone and grey limestone of the 740 m thick Chapel Island Formation, both of Upper Hadrynian (?) to Middle Cambrian age (O'Brien et al., 1977).

The redbeds of the middle assemblage are overlain conformably to unconformably (Strong et al., 1978) by outliers of the upper assemblage, scattered across the Avalon Zone (Williams, 1967). The lowest formation of the uppermost assemblage, the Random, is characteristically white orthoquartzite but in the western part of the Avalon Zone where it is up to 240 m thick it contains 210 m of sandy redbeds. The age of the Random Formation has been the subject of controversy but it is now thought (Butler and Greene, 1976) to be of early Cambrian age. The formation is a littoral deposit of a late Precambrian transgressive sea (Hughes and Brückner, 1971).

Quartzites similar in lithology and stratigraphic position to the Random occur in the Avalon Zone of Cape Breton Island and New Brunswick (Butler and Greene, 1976). In Cape Breton Island the 40 to 730 m thick Morrison River Formation consists of a basal hematitic crossbedded quartz sandstone unit with interlayered pebble conglomerate overlain by thin units of shale and white quartzite (Weeks, 1954). The redbeds of the formation may be correlative with the Middle assemblage of Newfoundland (Weeks, 1954). In southeastern New Brunswick immediately under the orthoquartzitic Glen Falls Formation lies the 60 to 600 m thick Ratcliffe Brook Formation that consists mostly of dark purple to red, fine grained sandstone and micaceous shale with drab interbeds of similar lithologies (Hayes and Howell, 1937). Patel (1973) thought deposition of the formation reflected a change from a continental regime at the base to a deltaic to shallow marine one at the top.

Jacobsville Formation

Although it is found at least 2000 km to the west of the Avalon Zone the Jacobsville Formation (Fig. 8c) is lithologically similar to and of the same age as the "Infracambrian" quartzites of that zone. It also lies within a broadly similar stratigraphic sequence. For these reasons it is convenient to discuss it with those rocks.

Sandstone of the Jacobsville Formation outcrops mainly along the south shore of Lake Superior in Michigan, underlies most of Lake Superior (Halls and West, 1971) and is exposed in the vicinity of, and north of, Sault Ste. Marie, Ontario (Frarey, 1978). In Ontario it overlies unconformably rocks as young as Keweenawan volcanics (Annels, 1973) and in Michigan it is unconformably overlain by upper Cambrian rocks (Hamblin, 1958).

In Ontario the up to 210 m thick (Frarey, 1978), ferruginous formation comprises basal pebble to boulder conglomerate overlain by red and mottled sandstone that appears to grade up into a grey sandstone. Interbeds of red and grey shale and thin discontinuous pebble bands occur in the lower part of the sandstone sequence (McConnell, 1927). The sandstone is quartz-rich, medium grained, and consists of a lower red portion, mottled and streaked by leaching, that grades up into a white or buff sandstone. Sand grains are mostly quartz, ten per cent feldspar, and lie in a sparse matrix of hematite and quartz cement.

The sedimentology of the formation has been studied from the exposure in Michigan where it is up to 550 m thick and consists of a crossbedded "lenticular" sandstone facies and a "conglomerate" facies, both of fluviatile origin and a "massive" sandstone facies and a "red siltstone" facies both of lacustrine origin. The sediments were transported by northward-flowing streams from a chemically-weathered and rugged source area which stretched east-west across the central part of northern Michigan (Hamblin, 1958). Babcock (1975) regarded the formation as fluvial-deltaic-lacustrine, the conglomerates being deposited in alluvial fans in a rapidly subsiding rift zone, and the sediments and interbedded and laterally equivalent volcanics, a manifestation of mid-Keweenawan continental rupture.

In Canada the formation is flat-lying or dips gently away from basement highlands (Frarey, 1978). The age of the formation is not established. Some more recent opinions are those of Hamblin (1958), who considered it lower to middle Cambrian, Dubois (1962) who considered it upper Keweenawan, Babcock (1974) who considered it lower to middle Keweenawan and Frarey (1978) who considered it Hadrynian. The stratigraphic assignment of the outcrop in Ontario is also in dispute. Hamblin considered some on the east shore of Lake Superior to be equivalent to the Keweenawan Freda sandstone of the west side of the Keweenawan Peninsula, whereas Frarey considered it to belong to the Jacobsville Formation.

Outliers of northern Quebec and Labrador

Small outliers of clastic rocks, mostly undeformed and unmetamorphosed and each with a significant redbed component are scattered over the northeastern part of the Canadian Shield (Fig. 9). Their ages and stratigraphic correlation are uncertain.

Double Mer Formation, Hadrynian or younger

Rocks of the Double Mer Formation (Fig. 9) outcrop in several areas near Lake Melville, Labrador (Greene, 1972). These are shown in more detail by the following writers: Lake Melville area (Christie et al., 1953; Eade, 1962), Churchill River area (Stevenson, 1967a), Kenamu River area (Stevenson, 1967b), and Double Mer area (Stevenson, 1970). The formation may underlie an area of 1800 km^2, but this figure is uncertain because of very sparse outcrop.

The basal unconformity with the underlying Grenville gneisses is drift-covered (Kindle, 1924; Stevenson, 1967a) and is in some cases interpreted as a fault (Stevenson, 1967b). North of Double Mer adjacent to Lake Melville, a "band" of polymict conglomerate is exposed intermittently and parallels the faulted contact of the formation with the Grenville gneisses (Stevenson, 1970). Younger rocks have not been reported to overlie the Double Mer.

Lithologies in the formation include granitic or chert conglomerate, arkose and pebbly arkose, and shale. The finer grained rocks and the matrix of the conglomerate are red. Kindle (1924) considered that the 45 m thickness of the formation exposed at the eastern head of Double Mer constituted a better exposure than the 23 m exposed at the type locality on the north shore of Mulligan Bay on Lake Melville. Exposure in the Double Mer area indicated a probable formational thickness of greater than 150 m. Both Kindle (1924) and Stevenson (1967a, 1970) mentioned the abundance of crossbeds.

The unmetamorphosed sediments may be preserved in down-faulted basins. They are flat-lying to gently-dipping (Stevenson, 1967a, b) except near faults bounding the outcrop area (Stevenson, 1970). The stratigraphic position of the Double Mer Formation on Grenville gneisses indicates a Hadrynian or younger age (Greene, 1974). Stevenson (1970), citing similarity to rocks overlying Cambrian rocks on the north shore of the Strait of Belle Isle, about 100 km south, (Bradore Formation), tentatively assigned the formation to the Cambrian while stressing the lack of fossils and of radiometrically datable material.

Gilbert Formation, Hadrynian or younger

The Gilbert Formation (Fig. 9) was found as one small outcrop at 56°21'W, 52°42'N. It overlies mylonite probably derived from quartzofeldspathic gneiss and consists of heavily hematite-stained, well-rounded conglomerate mainly of quartz pebbles with some of pink feldspar, in an arkosic matrix. Bedding is distinct and the grain size varies from bed to bed. Dips of the strata are steep to vertical. The rock is regarded as a remnant preserved on a down-faulted block (Eade, 1962). Since it is unmetamorphosed and lies within a terrain of gneisses of Grenville age the formation is Hadrynian or younger (Greene, 1974).

Sakami Formation, Aphebian or younger

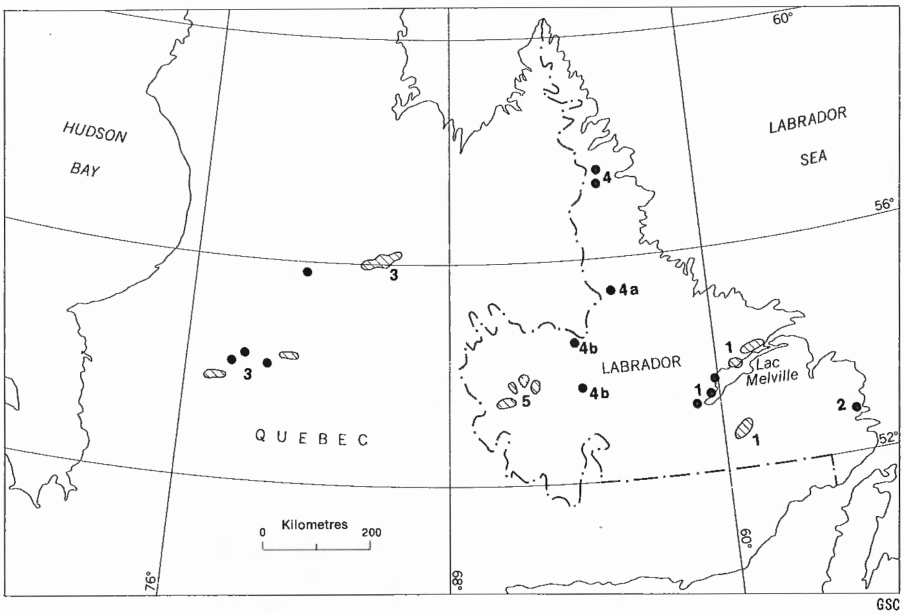
The Sakami Formation (Fig. 9) consists of isolated outliers arranged in two east-trending belts. Most of the outcrop areas are shown by Eade (1966). The southern belt of outcrops will in part be submerged as a result of imminent dam construction (1978) on the La Grande (Fort George) River as part of a hydroelectric project. In order to record in more detail the geology of the area to be submerged Sharma (1974a, b, 1976) has published maps at the scale of one inch to the mile.

The formation underlies an area of about 1200 km^2 and is about 500 to 550 m thick. It is usually either faulted against or unconformable on Archean rocks. In one place a regolith was observed (Eade, 1966). The only younger rocks associated with the formation are northwest-trending mafic dykes which transect it in the southern belt of outliers (Chown, 1977).

The internal stratigraphy of the formation, where it can be determined, is remarkably consistent from outlier to outlier. The lower part consists of continental redbeds containing commonly graded quartzofeldspathic conglomerate with sandstone lenses, abundant arkose, siltstone and shale, including some drab units. The shale and siltstone are commonly mudcracked and may occur as thin layers in sandstone. The upper part of the formation consists of buff, grey, pink or white, relatively pure quartzite (Eade, 1966; Sharma, 1976), in which large scale crossbeds were noted at a few localities.

The lower, redbed, part of the Sakami Formation may have been deposited in fault-bounded continental basins, and the upper quartz-rich sandstone may be small remnants of sandstone deposited in a much more widespread epicontinental environment (Eade, 1966; Stockwell et al., 1970).

In the southern belt of outliers Chown (1977) divided the formation into a basal unit of coarse- to fine-grained red alluvial graben-filling, a middle pink sandstone unit of similar origin with south and southeastward-directed paleocurrents



1. *Double Mer Formation*
2. *Gilbert Formation*
3. *Sakami Formation*
4. *Siamarnekh Formation*

4a. *Harp River Arkose*
4b. *Michikamau Arkose*
5. *Sims Formation*

Figure 9. Redbed outliers of northern Quebec and Labrador.

and an uppermost unit of fine grained, orange, quartz arenite largely of aeolian origin. He has some doubts (pers. comm., 1978) about the aeolian origin of the uppermost unit.

Away from marginal faults, where dips may be steep, the unmetamorphosed formation is horizontal to gently dipping. The steep dips may be due to faulting or to deposition on a rugged erosional surface (Eade, 1966).

For lack of evidence to the contrary Eade (1966) assumed the formation to be Proterozoic in age and has suggested on the basis of marked lithological similarity a similar age to the Aphebian Chakonipau Formation 65 km to the northeast. On the west side of Chakonipau Lake (Robertson, 1974) drab sediments overlie Archean rocks, resemble the Early Aphebian Matinenda Formation of the Elliot Lake area, and carry pyritic uraniferous conglomerate. The drab rocks are overlain by redbeds of the Sakami Formation. On the basis of this stratigraphic association he likened the Sakami Formation of the lower Aphebian Lorrain Formation (p. 6) of the Huronian Supergroup of Ontario.

Sharma (1977) reported radioactivity in pyritiferous green shales in the Sakami Formation and in shear zones at the contact with the granitic basement.

Siamarnekh Formation, Neohelikian or Hadrynian?

The formation lies in a north-trending series of scattered outliers (Fig. 9). The northernmost of these, which first received the formational name (Wheeler, 1964), and another outlier, both less than 4 km long, about 18 km to the south-southeast, have also been mapped by Taylor (1970). The Harp River Arkose (Greene, 1974), a patch of erratics and some probable outcrop of similar lithology covers an area of about 100 km^2 (Taylor, 1972). Two other outcrop areas, probably also belonging to the Siamarnekh Formation, lie in the vicinity of Michikamau Lake; (a) on the northwest shore (Emslie, 1963, 1970), and (b) on the south shore (Stevenson, 1969). Both are termed Michikamau Arkose by Greene (1974).

The outliers rest nonconformably on high grade metamorphic rocks of the Churchill Structural Province and Elsonian adamellite, granite and anorthosite (Greene, 1970). Northeast of Michikamau Lake intermediate tuff may overlie the Siamarnekh Formation (Emslie, 1964). Limestone, some bearing organic structures like "Stromatopora", were reported by Low (1896) to overlie sandstone of the Siamarnekh Formation and overlie granite near the south end of Michikamau Lake. R.F. Emslie (pers. comm., 1976) however felt that this limestone, which he only observed as loose fragments, resembled rocks from the Labrador Trough and were glacial erratics older than the Siamarnekh Formation.

The 110 m thick (Wheeler, 1964) Siamarnekh Formation is composed mainly of hematitic, partly friable, feldspathic sandstone with lesser amounts of red and green siltstone and conglomerate (Emslie, 1963, 1970). Quartzite-bearing conglomerate (Wheeler, 1964) and granite wash passing up into arkosic conglomerate with quartzite, felsic volcanic and granitic clasts (Emslie, 1970) were observed near the base of the formation. Brief references to sedimentary structures included some to rare crossbeds and common ripple marks (Wheeler, 1964; Taylor, 1972).

Except locally on the south shore of Lake Michikamau (Stevenson, 1969) the unmetamorphosed formation is horizontal to gently dipping. The anorthositic and associated felsic plutonic rocks in Labrador that underlie the Siamarnekh Formation are assumed to have been intruded in the order of 1.4 to 1.5 Ga ago (Emslie, 1975). This figure restricts the age of the formation to Neohelikian or younger and is in accord with observations of felsic volcanic pebbles possibly from the Letitia Lake Group (p. 29) and quartzite pebbles possibly (Wheeler, 1964) from the Late Aphebian Ramah Group (Morgan, 1975) both of which have been reported in conglomerate of the Siamarnekh Formation. An intermediate tuff associated with the Siamarnekh Formation, but of uncertain relative stratigraphic position, has been dated by the K/Ar whole rock method of 0.843 Ga (Wanless et al. 1967). Though Stevenson (1969) believed the formation to be "apparently much younger" than the 1.3 Ga old Seal Lake Group (p. 29) Wanless et al. (1967) noted the similarity between the Siamarnekh Formation and the Seal Lake Group and considered the likely correlatives.

Sims Formation, Paleohelikian

The Sims Formation outcrops in a cluster of four equiareal patches centred about 120 km south of Schefferville, and covers an area of about 500 km (Fig. 9). Where best observed, near Muriel Lake (Fahrig, 1967), it rests unconformably on folded metasediments of the Aphebian Kaniapiskau Supergroup. It is cut by olivine gabbro (Wynne-Edwards, 1961), and is not overlain by younger strata (Fahrig, 1967).

Though the lower boundary of the formation is not exposed, polymict conglomerate near the base, northwest of Muriel Lake, consists of boulders of sedimentary rock like unmetamorphosed to low grade metamorphic rocks along the west side of the Labrador Trough. The matrix of the conglomerate is red and poorly sorted. Six kilometres southwest of Muriel Lake the base of the formation is not exposed and the lowest rocks of the formation are over 200 m of locally crossbedded, medium grained, dark red, hematitic feldspathic quartzite overlain by 6 m of quartz and granite pebble conglomerate. The mineralogy of the redbeds suggests their derivation from high grade metamorphic and igneous rocks. These redbeds are overlain by an unknown thickness of similar but very coarse and friable redbeds (Fahrig, 1967).

Locally rippled and crossbedded pink or white quartz arenite, containing lens-shaped chert-quartz pebble conglomerate layers probably represents the upper part of the Sims Formation near Muriel Lake. Quartz arenite, similarly coloured but foliated, slightly metamorphosed and cut by metagabbro sills, about 25 km south toward the Grenville Front may be stratigraphically equivalent to the upper quartz arenitic part of the Sims Formation at Muriel Lake. Two hundred metres of quartz arenite containing jasper-bearing grit and conglomerate beds were mapped by Wynne-Edwards (1961) up to 40 km east of the occurrences mapped by Fahrig (1967).

Strong preferred orientation of crossbeds, lithology of the clasts in the basal conglomerate and the mineralogy of the redbeds of the Sims Formation suggests uplift of the Kaniapiskau Supergroup and granitized equivalent, or Archean granitic rocks as a source area in the southeast (Fahrig, 1967).

Northern outcrops of the Sims Formation are horizontal or dip gently and are unmetamorphosed. Southward, toward the Grenville Front, dips increase, foliation is present, and secondary muscovite may indicate some metamorphism (Fahrig, 1967).

An unconformable relationship to the Kaniapiskau Supergroup suggests a post-Aphebian age for the Sims Formation. Gabbro cutting the Sims was assigned a Grenville age by Wynne-Edwards (1961), however gabbro to the west also cutting the Sims Formation was considered (Fahrig, 1967) to be related to the Elsonian (Emslie, 1975) anorthosites and related rocks so widespread in and adjacent to the Grenville Structural Province. Thus it would appear that the Sims Formation is Helikian and possibly Paleohelikian in age. References in the literature to mineralization in the Sims Formation have not been found by the writer.

SUMMARY AND CONCLUSIONS

Earlier opinion favoured a detrital origin for the hematite of redbeds but was divided on whether it was formed under a humid or an arid climate. There is now abundant evidence that much of the hematite of redbeds has resulted from diagenesis in the depositional environment under oxidizing conditions. In the humid tropics where laterization produces abundant hematite, the presence of vegetable material is a very important influence both on whether hematite reached the depositional environment and on whether it will form there after deposition. Similarly the absence of plant remains is cited as a key factor in permitting diagenetic formation of hematite in alluvial sediments under hot desert conditions. Some redbeds interpreted as marine, could have formed due to the diagenetic chemical environment provided by sulphate and carbonate ions from associated evaporites (Rich et al., 1977).

The development of an oxygen-bearing atmosphere in the Early Proterozoic markedly increased the solubility of copper and uranium in surficial environments. Redbeds formed under the oxygen-bearing atmosphere brought groundwater passing through them into the stability field of hematite so enhancing uranium and copper in the subsurface.

Whereas other rock types may contain higher concentrations of base metals and uranium, granitic rocks by virtue of their abundance, especially where high in uranium are as likely a source. Consequently the nonconformable position of many of the sequences listed in this report on the granitic gneisses of the Canadian Shield may be of some importance (see Darnley et al., 1977). Diagenetic formation of redbeds from sediments weathered from these gneisses would release some base metals from oxide and silicate clasts, though uranium is likely to have been lost during early weathering of the source granites. Some of the base metals could be temporarily stored in the hematite and clays formed during the diagenesis.

Formation of sedimentary uranium, redbed copper and other base metal deposits by redox reactions in terrestrial sequences was less common in the Proterozoic than in the post-Devonian because of the relative lack of organic remains derived from land plants. However some uranium deposits appear to have been fixed inorganically by ferrous silicate-bearing rocks. Formation of uranium deposits in the oxidizing environment by vanadate formation seems feasible. Kupferschiefer-type copper deposits on the other hand are to be expected in Proterozoic marginal marine sequences where redbeds, evaporites and sediments bearing the remains of marine organisms are juxtaposed.

Starting with the Early Aphebian, Huronian Supergroup redbeds were deposited on the mainly granitic rocks of the Canadian Shield in a wide range of depositional environments from continental to shallow marine and in a wide range of tectonic environments. Miogeoclinal marine to alluvial sequences which include some of the thickest redbed-bearing sections are represented by the Belcher Group, Huronian Supergroup, Knob Lake Group and Purcell System. In possible aulacogens associated with these continental-margin prisms lie the Chakonipau Formation, the Richmond Gulf Sequence, the Seal Lake and Sibley groups the thicker part of the Eqalulik and Uluksan groups and the thick Great Slave Supergroup. Mainly marine sequences deposited on stable platforms and in some cases interpreted as stratigraphic equivalents of the miogeoclinal sequences include the Goulburn Group, the deposits of the Amundsen Basin and some of the redbeds of northern Baffin Island and adjacent areas. Few redbeds are attributed to the flysch-related phases of geosynclinal evolution. Among these are the Akaitcho River Formation and Kahochella Group of the Great Slave Supergroup and the Peacock Hills Formation of the Goulburn Group. Late orogenic redbeds are more common. They were deposited in mainly alluvial environments and a greater proportion of each sequence is red. Some sequences interpreted as molasse include the Christie Bay Group, Takiyuak, Western River and Amagok formations all related to evolution of the Coronation Geosyncline, the Loaf Formation of the Circum-Ungava Geosyncline and possibly the largely metamorphosed Missi Group that lies within the Churchill Province in Manitoba. Other late orogenic redbeds include the taphrogenic Dubawnt Group, Et-then, and Nonacho groups and the Martin Formation that immediately follow the Hudsonian event. A number of small fault-bounded outliers overlie rocks of the Labrador Trough and gneissic granitic rocks of northern Quebec and Labrador.

Amongst the redbed-bearing sequences described in this report copper mineralization is commonly associated with those of miogeoclinal and shallow marine-supracratonic origin which, as might be expected, contain evaporites and marine sediments that have the potential to furnish reducing environments (Chandler, 1978a; Adler, 1974). Modern and later Phanerozoic hot deserts and evaporites formed in the dry tropical climate zones about latitudes 30° North and South. Thus if Proterozoic atmospheric circulation was similar to the present (Chandler and Musk, 1976) in the above respect, one should expect that Proterozoic redbed sequences in which evaporites have not been found, but which yield paleomagnetic latitudes of the dry tropics would be a likely source for these deposits.

Canadian Proterozoic terrestrial redbeds on the other hand contain little copper but some uranium, which occurs as vein mineralization in the basement granites as well as in the alluvial sediments. These observations tend to support the view that chloride complexing is important in Kupferschiefer-type mineralization but not in uranium mineralization and that Proterozoic sedimentary reducing environments are more common in shallow marine than in alluvial environments.

SELECTED BIBLIOGRAPHY

Adler, H.H.
1970: Interpretation of colour relations in sandstone as a guide to uranium exploration and ore genesis; p. 331-344 in Uranium exploration geology., Proceedings of the International Atomic Energy Agency, Vienna 13-17, April 1970, 384 p.

1974: Concepts of uranium-ore formation in reducing environments in sandstones and other sediments; p. 141-168 in Symposium of the Formation of uranium ore deposits Athens, 6-10 May 1974. International Atomic Energy Agency, 748 p.

Ahlfeld, F.
1967: Metallogenic epochs and provinces in Bolivia, Mineralism Deposita, v. 2, p. 291-311.

Aitken, J.D.
1977: Redstone River Formation (upper Proterozoic) in Mount Eduni and Bonnet Plume map areas, District of Mackenzie; p. 137-138 in Report of Activities, Part. A., Geological Survey of Canada, Paper 77-1A, 527 p.

Aitken, J.D., and Cook, D.G.
1974: Caronjou Canyon map area, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 74-13, 28 p.

Aitken, J.D., Long, D.G.F., and Semikhatov, M.A.
p. 481-484 in Current Research, Part A,.

1978a: Progress in Helikian Stratigraphy, Mackenzie Mountains; Geological Survey of Canada, Paper 78-1A 588 p.

1978b: Correlation of Helikian strata, Mackenzie Mountains – Brock Inlier – Victoria Island; p. 485-488 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 588 p.

Aitken, J.D., MacQueen, R.W., and Usher, J.L.
1973: Reconnaissance studies of Proterozoic and Cambrian stratigraphy, lower Mackenzie River area (Operation Norman), District of Mackenzie; Geological Survey of Canada, Paper 73-9, 178 p.

Ali, A.J. Braithwaite, C.J.R.
1977: Penecontemporaneous weathering of the Old Red Sandstone of the Midland Valley of Scotland;, Scott. Journal of Geology, v.13, p. 305-312.

Allen, J.R.L.
1974: Studies in fluviatile sedimentation: Implications of pedogenic carbonate units, Lower Old Red Sandstone, Anglo-Welsh outcrop; Geological Journal, v. 9, p. 181-208.

Anderson, M.M.
1972: A possible time span for the late Precambrian of the Avalon Peninsula, southeastern Newfoundland, in the light of worldwide correlation of fossils, tillites and rock units within the succession; Canadian Journal of Earth Science, v. 9, p. 1710-1726.

Annels, A.E.
1974: Some aspects of the stratiform ore deposits of the Zambian copper belt and their genetic significance; p. 235-254 in Gisements stratiformes et provinces cuprifères; Centenaire de la Société Géologique de Belgique. Liège 1974, Bartholomé Paul (Ed.) 427 p.

Annells, R.N.
1973: Proterozoic flood basalts of eastern Lake Superior: the Keweenawan volcanic rocks of the Mamainse Point area, Ontario; Geological Survey of Canada, Paper 72-10, 51 p.

Ayres, L.D., Lumbers, S.B., Milne, V.G., and Robeson, D.W.
1970: East central sheet; Ontario Department of Mines and Northern Affairs, Geological Map 2198.

Babcock, L.L.
1974: Jacobsville sandstone: a lower-middle Keweenawan redbed sequence; p. 1-2 in Institute of Lake Superior Geology, 20th Annual Meeting, Sault Ste. Marie, Ontario; Giblin, P.E., Bennett, G., and Leahy, E.J. (Ed.)

1975: The Jacobsville sandstone: evidence for a lower-middle Keweenawan age; p. 87-123 in Institute of Lake Superior Geology 21st Annual Meeting, Marquette Michigan; Earney, F.C.F. and Hughes, J.D. (Ed.) 261 p.

Baer, A.J. (Ed)
1970: Symposium on basins and geosynclines of the Canadian Shield; Geological Survey of Canada, Paper 70-40, 165 p.

Baer, A.J., Emslie, R.F., Irving, E., and Tanner, J.G.
1974: Grenville geology and plate tectonics; Geoscience Canada, v. 1, no. 3, p 54-61.

Bailes, A.H.
1971: Preliminary compilation of the geology of the Snow Lake - Flin Flon-Sheridan area; Manitoba Department of Mines and Natural Resources, Mines Branch, Geological Paper 1/71, 27 p.

Baldwin, D
1976: Sickle group-Laurie River substudy; the evaluation of disseminated base metal environments; p. 62-92 in Canada, Department of Energy, Mines and Resources, and Manitoba, Department of Mines and Natural Resources, Open File Report 77/1 Non-renewable resource evaluation program, First Annual Report 1975/6, 150 p.

Balkwill, H.R. and Yorath, C.J.
1970: Brock River map-area, District of Mackenzie (97D); Geological Survey of Canada, Paper 70-32, 25 p.

Baragar, W.R.A.
1967a: Volcanic studies: Coppermine River basaltic flows; p. 26-28 in Report of Activities, Part A, Geological Survey of Canada, Paper 67-1A, 221 p.

1967b: Wakuach Lake map-area, Quebec-Labrador; Geological Survey of Canada, Memoir 344, 174 p.

1969: Volcanic studies in the Seal Lake area, Labrador, (13K,-L); p. 142146 in Report of Activities, Part A, Geological Survey of Canada, Paper 69-1A, 263 p.

1972: Coppermine River basalts - District of Mackenzie; p. 21-24 in Rubidium - Strontium isochron age studies, Report 1; Geological Survey of Canada, Paper 72-23, 77 p.

1974: Seal Lake and Croteau volcanic rocks of Labrador; p. 5, in Program with abstracts and 3rd circular, Geological Association of Canada/Mineralogical Association of Canada, Annual Meeting, St. John's, Newfoundland, 101 p.

1976: The Natkusiak basalts, Victoria Island, District of Franklin; p. 347-352 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 552 p.

Baragar, W.R.A. and DonaldsonJ.A.
1973: Coppermine and Dismal Lakes map-areas; Geological Survey of Canada, Paper 71-39, 20 p.

Baragar, W.R.A. and Hornbrook, E.H.
1963: Mineral industry of District of Mackenzie, 1962; Geological Survey of Canada, Paper 63-9, 44 p.

Barbier, M.J.
1974: Continental weathering as a possible origin of vein-type uranium deposits; Mineralism Deposita, v. 9, p. 271-288.

Barrett, K.R.
1975: A paleocurrent, paleoscope and provenance study of the Belcher Group; unpublished B. Sc. thesis, Brock University, St. Catherines, Ontario (Canada), 49 p.

Bartholomé, Paul (Ed)
1974: Gisements stratiformes et provinces cuprifères; Centenaire de la Société Géologique de Belgique, Liège, 1974, 427 p.

Bell, R.T.
1968: Proterozoic stratigraphy of northeastern British Columbia; Geological Survey of Canada, Paper 67-68, 75 p.

Bell, R.T. and Jackson, G.D.
1974: Aphebian halite and sulphate indications in the Belcher Group, Northwest Territories; Canadian Journal of Earth Sciences, v. 11, no. 5, p. 722-728.

Bérard, Jean
1965: Bérard Lake area, New Quebec; Quebec Department of Natural Resources, Geological Report 111, 139 p.

Bergeron, Robert
1957a: Proterozoic rocks of the northern part of the Labrador Geosyncline, the Cape Smith Belt, and the Richmond Gulf area; p. 101-111, in The Proterozoic in Canada; J.E. Gill (Ed.), Royal Society of Canada, Special Publication no. 2, 191 p.

1957b: Late Precambrian rocks of the north shore of the St. Lawrence River and of the Mistassini and Otish Mountains areas, Quebec; p. 124-131 in The Proterozoic in Canada, J.E. Gill (Ed.). Royal Society of Canada, Special Publication no. 2, 191 p.

Berkhout, A.W.J.
1973: Gravity in the Prince of Wales, Somerset, and northern Baffin islands regions; p. 63-79, in Symposium on the Geology of the Canadian Arctic, Proceedings, Saskatoon, May 1973, 368 p.

Berner, R.A.
1969: Goethite stability and the origin of red beds; Geochimica et Cosmochimica Acta, v. 33, p. 267-273.

1971: Principles of chemical sedimentology; p. 197-199, McGraw Hill (Pub.) 240 p.

Binda, P.L.
1975: Detrital bornite grains in the late Precambrian B. graywacke of Mufulira, Zambia; Mineralium Deposita, v. 10, p. 101-107.

Bishop, D.T., Powell, J.D., Reid, R.R., Savage, C.N., Siems, P.L., and Williams, G.A.
1973: Belt Symposium; University of Idaho, Idaho Bureau of Mines and Geology, v. 1, 322 p.

Blackadar, R.G.
1956: Geological reconnaissance of Admiralty Inlet, Baffin Island, Arctic Archipelago, Northwest Territories; Geological Survey of Canada, Paper 56-6, 25 p.

1958: Fury and Hecla Strait, District of Franklin, Northwest Territories; Geological Survey of Canada, Map 3-1958.

1967: Precambrian geology of Boothia Peninsula, Somerset Island, and Prince of Wales Island, District of Franklin; Geological Survey of Canada, Bulletin 151, 62 p.

1970: Precambrian geology of northwestern Baffin Island, District of Franklin; Geological Survey of Canada, Bulletin 191, 89 p.

Blackadar, R.G. and Christie, R.L.
1963: Geological reconnaissance, Boothia Peninsula and Somerset, King William and Prince of Wales islands, District of Franklin; Geological Survey of Canada, Paper 63-19, 15 p.

Blackadar, R.G. and Fraser, J.A.
1960: Precambrian geology of Arctic Canada, a summary account; Geological Survey of Canada, Paper 60-8, 24 p.

Blatt, H. et al.
1972: Origin of sedimentary rocks; Prentice-Hall Inc. (Publ.), 634 p.

Blusson, S.L.
1966: Geology, Frances Lake, Yukon Territory and District of Mackenzie; Geological Survey of Canada, Map 6-1966.

1967: Sekwi Mountain, Nahanni and Frances Lake map-areas; p. 44-45 in Report of Activities, Part A, Geological Survey of Canada, Paper 67-1A, 221 p.

1968: Sekwi Mountain and Nahanni map-areas, District of Mackenzie and Yukon Territory; p. 14 in Report of Activities, Part A, Geological Survey of Canada, Paper 68-1A, 230 p.

1974: Geology, Mount Eduni, Bonnet Plum, Nedaleen River, Lansing and Niddery Lake map areas, District of Mackenzie and Yukon Territory; Geological Survey of Canada, Open File 205.

1976: Selwyn Basin, Yukon and District of Mackenzie; p. 131-132 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 522 p.

Blusson, S.L. and Tempelman-Kluit, D.J.
1970: Operation Stewart, Yukon Territory, District of Mackenzie; p. 29-32 in Report of Activities, Part A, Geological Survey of Canada, Paper 70-1A, 251 p.

Bostock, H.S.
1947: Mayo, Yukon Territory; Geological Survey of Canada, Map 890A.

1969: Precambrian sedimentary rocks of the Hudson Bay lowlands; p. 206-214 in Earth Science Symposium on Hudson Bay; P.J. Hood, (Ed.), Geological Survey of Canada, Paper 68-53, 386 p.

Breger, I.A.
1974: The role of organic matter in the accumulation of uranium; p. 206-214 in Symposium on the formation of uranium ore deposits, Athens 6-10, May 1974; International Atomic Energy Agency, 748 p.

Brown, I.C.
1950: Preliminary map, Reliance, Northwest Territories; Geological Survey of Canada, Paper 50-15.

Brown, R.L., Dalziel, I.W.D., and Rust, B.R.
1969: The structure, metamorphism and development of the Boothia Arch, Arctic Canada; Canadian Journal of Earth Sciences, v. 6, p. 533-543.

Bruckner, W.D.
1969: Geology of eastern part of Avalon Peninsula, Newfoundland-A summary; p. 130-138 in North Atlantic - geology and continental drift; Marshall Kay (Ed.); American Association of Petroleum Geologists, Memoir 12, 1082 p.

Brummer, J.J. and Mann, E.L.
1961: Geology of the Seal Lake area, Labrador; Geological Society of America, Bulletin, v. 72, p. 1361-1382.

Burke, K. and Dewey, J.F.
1973: Plume-generated triple junctions: key indicators in applying plate tectonics to old rocks; Journal of Geology, v. 81, p. 406-433.

Burwash, R.A. and Baadsgaard, H.
1962: Yellowknife-Nonacho age and structural relations; in Tectonics of the Canadian Shield; J.S. Stevenson (Ed.), Royal Society of Canada, Special Publication no. 4, p. 22-29.

Butler, A.J. and Greene, B.A.,
1976: Silica resources of Newfoundland; Mineral Development Division, Department of Mines and Energy, Province of Newfoundland, Report 76-2, 68 p.

Caia, J.
1976: Paleogeographical and sedimentological controls of copper, lead, and zinc mineralizations in the Lower Cretaceous sandstones of Africa; Economic Geology, v. 71, p. 409-422.

Campbell, F.H.A.,
1978: Geology of the Helikian rocks of the Bathurst Inlet area, Coronation Gulf, Northwest Territories; p. 97-106 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 588 p.

Campbell, F.H.A. and Cecile, M.P.
1975: Report on the geology of the Kilohigok Basin, Goulburn Group, Bathurst Inlet, N.W.T.; p. 297-306 in Report of Activities, Part A, Geological Survey of Canada, Paper 78-1A, 602 p.

1976: Geology of the Kilohigok Basin, Goulburn Group, Bathurst Inlet, District of Mackenzie; p. 369-377 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 522 p.

Card, K.D.
1976: MacGregor Bay - Bay of Islands area, District of Sudbury and Manitoulin; Ontario Division of Mines, Ministry of Natural Resources, Geoscience Report, 138, 63 p.

Card, K.D., Church, W.R., Franklin, J.M., Frarey, M.J., Robertson, J.A., West, G.F., and Young, G.M.
1972: The southern provinces; p. 336-380 in Variations in Tectonic Styles in Canada; R.A. Price and R.J.W. Douglas (Ed.); Geological Association of Canada, Special Paper 11, 688 p.

Carpenter, A.B., Trout, M.L., and Pickett, E.E.
1974: Preliminary report on the origin and chemical evolution of lead-and zinc-rich oil field brines in central Mississippi; Economic Geology, v. 69, p. 1191-1206.

Carr, J.M.
1971: Geology of the Churchill copper deposit; Canadian Institute of Mining and Metallurgy Bulletin, v. 64, no. 710, p. 50-54.

Casshyap, S.M.
1968: Petrology of the Bruce and Gowganda Formations and its bearing on the evolution of Huronian sedimentation in the Espanola -Willisville area, Ontario (Canada); Palaeogeography, Palaeoclimatology, Palaeoecology, v. 6, p. 5-36.

Cecile, M.P.
1976: Stratigraphy and depositional history of the upper Goulburn Group, Kilhihok Basin, Bathurst Inlet, N.W.T., unpublished Ph.D. thesis, Carleton University, Ottawa (Canada), 210 p.

Chandler, F.W.
1969: Geology of the Huronian rocks, Harrow Township area, Ontario; unpublished Ph.D thesis, University of Western Ontario, London, Ontario, 327 p.

Chandler, F.W. (cont.)
1978a: Geological environment of Aphebian red beds of the north half of Richmond Gulf, New Quebec; p. 107-110 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 588 p.

1978b: Geology of part of Wollaston Lake Fold Belt, Saskatchewan; Geological Survey of Canada, Bulletin 277.

Chandler, T.J. and Musk, L.F.
1976: The atomosphere in perpetual motion; Geographical Magazine, v. 49, no. 2, p. 93-102.

Chown, E.H.
1977: Origin of the Sakami Formation, Lagrande River, Quebec; p. 12, in Program with Abstracts, v. 2, Geological Association of Canada, Annual Meetings, April 25-27, 1977, Vancouver, B.C., 68 p.

Chown, E.H. and Caty, J.L.
1973: Stratigraphy, petrography and paleocurrent analysis of the Aphebian clastic formation of the Mistassini-Otish Basin; p. 49-71 in Huronian Stratigraphy and Sedimentation; G.M. Young, (Ed.) Geological Association of Canada, Special Paper, no. 12, 271 p.

Christie, A.M., Roscoe, S.M., and Fahrig, W.F.
1953: Central coast of Labrador; Geological Survey of Canada, Paper 53-14, 3 p.

Christie, R.L.
1964: Diabase-gabbro sills and related rocks of Banks and Victoria islands, Arctic Archipelago; Geological Survey of Canada, Bulletin 105, 13 p.

Christie, R.L., Cook, D.G., Nassichuk W.W., Trettin, H.P., and Yorath, C.J.
1972: The Canadian Arctic Islands and the Mackenzie region; 24th International Geological Congress, Guidebook, A66, 146 p.

Clark, A.L.
1971: Strata-bound copper sulphides in the Precambrian Belt Supergroup, Northern Idaho and Northwestern Montana; p. 261-267 in Proceedings IMA-IAGOD Meeting 1970, IAGOD Volume; Society of Mining Geology, Japan, Special Issue No. 3, 500 p.

Clark, T.
1977: Geology of the Forbes Lake area, New Quebec Territory; Quebec Department of Natural Resources, Preliminary Report DPV-452, 16 p.

Cloud, P.E., Jr.
1968: Atmospheric and hydrospheric evolution on the primitive Earth; Science, v. 160, p. 729-735.

Cloud, Preston
1972: A working model of the primitive earth; American Journal of Science, v. 272, p. 537-548.

1976: Beginnings of biospheric evolution and their biogeochemical consequences; Paleobiology, v. 2, p. 351-387.

Coates, C.J.A., Quirke, P.T., Bell, C.K., Cranstone, D.A., and Campbell, F.H.A.
1972: Geology and Mineral deposits of the Flin Flon, Lynn Lake and Thompson areas, Manitoba and the Churchill-Superior Front of the western Precambrian Shield; 24th International Geological Congress Guidebook A31-C31, D.J. Glass, (Ed.), Montreal, Quebec, 96 p.

Coates, J.A.
1964: A redstone bedded copper deposit and a discussion on the origin of red bed copper deposits; unpublished M.Sc. thesis, University of British Columbia, Vancouver, B.C., 77 p.

Coates, M.E.
1972: Geology of the Black Sturgeon River area, District of Thunder Bay; Ontario Department of Mines and Northern Affairs, Geological Report 98, 41 p.

Collins, J.A.
1974: The Sedimentary Copper Universal; from sedimentologic and stratigraphic synthesis of the Proterozoic of Icon, Quebec; Grinnell Formation, Alberta; Nonesuch shale, Michigan; and the Mississippian Horton-Windsor formations of Nova Scotia; unpublished Ph.D. thesis, Queen's University, Kingston, Ontario, 259 p.

Cook, D.G. and Aitken, J.D.
1969: Geology, Erly Lake, District of Mackenzie; Geological Survey of Canada, Map 5-1969.

Cormier, R.F.
1969: Radiometric dating of the Coldbrook Group of southern New Brunswick, Canada; Canadian Journal of Earth Sciences, v. 6, p. 393-398.

Craig, B.G., Davison, W.L., Fraser, J.A., Fulton, .J., Heywood, W.W., and Irvine, T.N.
1960: Geology, north-central District of Mackenzie; Geological Survey of Canada, Map 18-1960.

Dahl, A.R. and Hagmaier, J.L.
1974: Genesis and characteristics of the southern Powder River Basin uranium deposits, Wyoming, U.S.A.; p. 201-218 in Symposium on the formation of uranium ore deposits, International Atomic Energy Agency, Athens, 6-10 May, 1974, 748 p.

Darnley, A.G., Charbonneau, B.W.; and Richardson, K.A.
1977: Distribution of uranium in rocks as a guide to the recognition of uraniferous regions; p. 55-86 in Recognition and evaluation of uraniferous areas, Proceedings, Technical Committee Meeting, International Atomic Energy Agency, Vienna, 1975, 295 p.

Davidson, A.
1972: The Churchill Province; p. 381-433 in Variations in Tectonic Styles in Canada; R.A. Price, and R.J.W. Douglas (Ed.), Geological Association of Canada, Special Paper 11, 688 p.

Davidson, C.F.
1965: A possible mode of origin of stratabound copper ores; Economic Geology, v. 60, p. 942-954.

Davies, J.C., Pryslak, A.P., and Pye, E.G.
1970: Sioux Lookout-Armstrong sheet, Kenora and Thunder Bay districts; Ontario Department of Mines, Compilation Map 2169, Scale 4 miles = 1 inch.

Derry, D.R.
1960: Evidence of the origin of the Blind River uranium deposits; Economic Geology, v. 55, p. 906-927.

Dewey, J.F., and Burke, K.C.A.
1973: Tibetan, Variscan and Precambrian basement reactivation; products of continental collision; Journal of Geology, v. 81, p. 683-692.

Dietz, R.S. and Holden, J.C.
1966: Miogeoclines (miogeosynclines) in space and time; Journal of Geology, v. 74, p. 566-583.

Dimroth, E.
1968: The evolution of the central segment of the Labrador Geosyncline; Part 1: stratigraphy, facies and palaeogeography; Neues Jahrbuchfuer Geologie und Palaeontologie, Abhandlungen, v. 132, no. 1, p. 22-54.

1969: Geology of the Castignon Lake area, New Quebec Territory; Quebec Department of Natural Resources, Preliminary Report 571, 54 p.

1970: Evolution of the Labrador Geosyncline; Geological Society of America Bulletin, v. 81, p. 2717-2742.

1972: The Labrador Geosyncline revisited; American Journal of Science, v. 272, p. 487-506.

Dimroth, E., Baragar, W.R.A., Bergeron, R., and Jackson, G.D.
1970: The filling of the circum-Ungava geosyncline; p. 45-142 in Symposium on Basins and Geosynclines of the Canadian Shield; Geological Survey of Canada, Paper 70-40, 265 p.

Dixon, J.
1974: Revised stratigraphy of the Hunting Formation (Proterozoic) Somerset Island, Northwest Territories; Canadian Journal of Earth Sciences, v. 11, p. 635-642.

Doi Kazumi, Hirono Shuichiro, and Sakamaki Yukio
1975: Uranium mineralization by groundwater in sedimentary rocks, Japan; Economic Geology, v. 70, p. 628-646.

Donaldson, J.A.
1967: Two Proterozoic clastic sequences; a sedimentological comparison; Geological Association of Canada, Proceedings, v. 18, p. 33-54.

1968: Proterozoic sedimentary rocks of northern Saskatchewan; p. 131 in Report of Activities, Part A, Geological Survey of Canada, Paper 68-1A, 265 p.

1969: Stratigraphy and sedimentology of the Hornby Bay Group, District of Mackenzie; p. 154-157 in Report of Activities, Part A, Geological Survey of Canada, Paper 69-1A, 263 p.

1971: Sedimentologic and stratigraphic study of the Hornby Bay and Coppermine River Groups, District of Mackenzie; p. 123-125 in Report of Activities, Part A, Geological Survey of Canada, Paper 71-1A, 259 p.

1973: Possible correlations between Proterozoic strata of the Canadian Shield and North American Cordillera; p. 61-75 in Belt Symposium, University of Idaho, Idaho Bureau of Mines and Geology, Moscow, Idaho, v. 1, 322 p.

Donaldson, J.A., Irving, E., McGlynn, J., and Tanner, J.
1976: Stockwell Symposium on the Hudsonian Orogeny and plate tectonics, conference report; Geoscience Canada, v. 3, no. 4, p. 285-291.

Douglas, R.J.W.
1952: Preliminary map, Waterton, Alberta; Geological Survey of Canada, Paper 52-10.

Douglas, R.J.W. (Ed.)
1970: Geology and economic minerals of Canada; Geological Survey of Canada, Economic Geology Report 1, 5th Ed., 838 p.

Douglas, R.J.W., Gabrielse, H., Wheeler, J.O., Stott, D.F., and Belyea, H.R.
1970: Geology of Western Canada; p. 367-488 in Geology and Economic Minerals of Canada; Douglas, R.J.W. (Ed.), Geological Survey of Canada, Economic Geology Report 1, 838 p.

Douglas, R.J.W., and Norris, D.K.
1963: Dahadinni and Wrigley map areas, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 62-63, 34 p.

Dressler, B.
1976: Geology of the north-central Labrador Trough, unpublished abstract; in The Stockwell Symposium: the Hudsonian Orogeny and Plate Tectonics, Ottawa, Ontario.

Dubois, P.M.
1962: Palaeomagnetism and correlation of Keweenawan rocks; Geological Survey of Canada, Bulletin 71, 75 p.

Dunbar, C.O. and Rodgers, John
1957: Principles of Stratigraphy; John Wiley & Sons Inc. (Pub.) 356 p.

Eade, K.E.
1962: Geology, Battle Harbour-Cartwright, coast of Labrador, Newfoundland; Geological Survey of Canada, Map 22-1962.

1966: Fort George River and Kaniapiskau River (west half) map areas, New Quebec; Geological Survey of Canada, Memoir 339, 84 p.

Eade, K.E. and Blake, D.H.
1977: Geology of the Tulemalu Lake map area, District of Keewatin; p. 209-211 in Report of Activities, Geological Survey of Canada, Paper 77-1A, 527 p.

Eisbacher, G.H.
1976: Proterozoic Rapitan Group and related rocks, Redstone River area, District of Mackenzie; p. 117-125 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 522 p.

1977: Tectono-stratigraphic framework of the Redstone copper belt, District of Mackenzie; p. 229-234 in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, 527 p.

1978: Two Major Proterozoic unconformities, Northern Cordillera; p. 53-58 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 588 p.

Emslie, R.F.
1963: Michikamau Lake, east half, Quebec-Newfoundland; Geological Survey of Canada, Paper 63-20, 4 p.

1964: Kasheshibaw Lake, Newfoundland-Quebec; Geological Survey of Canada, Map 3-1964.

1970: The geology of the Michikamau intrusion, Labrador; Geological Survey of Canada, Paper 68-57, 85 p.

1975: Nature and origin of anorthosite suites; Geoscience Canada, v. 2, no. 21, p. 99-104.

1978: Anorthosite massifs, rapakivi granites and late Proterozoic rifting of North America; Precambrian Research, 7, p. 61-98

Esquevin, J. and Mendendez, (Pau), R.
1973: Geochronology of Aphebian sediments from eastern Hudson Bay; Fortschritte der Mineralogie, v. 50, p. 68-70.

Fahrig, W.F.
1956: Cambrian Lake (west half) New Quebec; Geological Survey of Canada, Paper 55-42.

1957: Geology of certain Proterozoic rocks in Quebec and Labrador; p. 112-123 in The Proterozoic in Canada; J.E. Gill (Ed.), Royal Society of Canada, Special Publication, no. 2, 191 p.

1959: Snegamook Lake, Labrador, Newfoundland; Geological Survey of Canada, Map 1079A.

1961: The geology of the Athabasca Formation; Geological Survey of Canada, Bulletin 68, 41 p.

1967: Shabogamo Lake map-area, Newfoundland-Labrador and Quebec; Geological Survey of Canada, Memoir 354, 23 p.

Fahrig, W.F., Gaucher, E.H., and LaRochelle, A.
1965: Palaeomagnetism of diabase dykes of the Canadian Shield; Canadian Journal of Earth Sciences, v. 2, p. 278-298.

Fahrig, W.F.,Irving, E., and Jackson, G.D.
1971: Paleomagnetism of the Franklin diabases; Canadian Journal of Earth Sciences, v. 8, p. 455-467.

Fahrig, W.F. and Jones, D.L.
1969: Paleomagnetic evidence for the extent of Mackenzie igneous events; Canadian Journal of Earth Sciences, v. 6, p. 679-688.

Fahrig, W.F. and Wanless, R.K.
1963: Age and significance of diabase dyke swarms of the Canadian Shield; Nature, v. 200, p. 934-937.

Fairbairn, H.W., Hurley, P.M., Card, K.D., and Knight, C.J.
1969: Correlation of radiometric ages of Nipissing diabase and Huronian metasediments with Proterozoic orogenic events in Ontario; Canadian Journal of Earth Sciences, v. 6, p. 489-497.

Feniak, M.
1948: MacAlpine Channel, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 1011A.

Finch, W.I.
1967: Geology of the epigenetic uranium deposits in sandstone in the United States; U.S. Geological Survey, Professional Paper No. 538, 121 p.

Fischer, A.G.
1963: Essay Review of "Descriptive Paleoclimatology" Nairn, A.E.M. (Ed.) 1961 Interscience (Pub.), 380 p.; American Journal of Science, v. 261, p. 282-293.

Folk, R.L.
1976: Reddening of desert sands: Simpson Desert N.T., Australia; Journal of Sedimentary Petrology, v. 46, p. 604-615.

Franklin, J.M., McIwaine, W.H., Poulsen, K.H., and Wanless, R.K.
Stratigraphy and sedimentation of the Sibley Group. (in prep.)

Franklin, J.M. and Mitchell, R.H.,
1977: Lead-zinc-barite veins of the Dorion area, Thunder Bay District, Ontario; Canadian Journal of Earth Sciences, v. 14, p. 1963-1979.

Frarey, M.J.
1948: Preliminary map, Crowduck Bay, Manitoba; Geological Survey of Canada, Paper 48-22.

Frarey, M.J. (cont.)
1962: Bruce Mines, Ontario; Geological Survey of Canada, Map 32-1962.

1978: Geology of the Hermian belt between Sault Ste. Marie and Blind River, Ontario; Geological Survey of Canada, Memoir 383.

Frarey, M.J. and Duffell, S.
1964: Revised stratigraphic nomenclature for the central part of the Labrador Trough; Geological Survey of Canada, Paper 64-25, 13 p.

Frarey, M.J. and Roscoe, S.M.
1970: The Huronian Supergroup north of Lake Huron; p. 143-158 in Symposium on Basins and Geosynclines of the Canadian Shield, Geological Survey of Canada, Paper 70-40, 265 p.

Fraser, J.A.
1960: Geology, north-central District of Mackenzie; Geological Survey of Canada, Map 18-1960.

1964: Geological notes on northeastern District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 63-40, 46 p.

1974: The Epworth Group, Rocknest Lake area, District of Mackenzie; Geological Survey of Canada, Paper 73-39, 23 p.

Fraser, J.A., Craig, G.B., Davidson, W.L., Fulton, R.J., Heywood, W.W., and Irvine, T.N.
1960: North-central District of Mackenzie, N.W.T.; Geological Survey of Canada, Map 18-1960.

Fraser, J.A., Donaldson, J.A., Fahrig, W.F., and Tremblay, L.P.
1970: Helikian basins and geosynclines of the northwestern Canadian Shield; p. 213-238 in Symposium on Basins and Geosynclines of the Canadian Shield; Geological Survey of Canada, Paper 70-40, 265 p.

Fraser, J.A., Hoffman, P.F., Irvine, T.N., and Mursky, G.
1972: The Bear Province; p. 451-501 in Variations in Tectonic Styles in Canada; R.A. Price, and R.J.W. Douglas, (Ed.), Geological Association of Canada, Special Paper 11, 688 p.

Fraser, J.A. and Tremblay, L.P.
1969: Correlation of Proterozoic strata in the northwestern Canadian Shield; Canadian Journal of Earth Sciences, v. 6, p. 1-9.

Friend, P.F.
1966: Clay fractions and colours of some Devonian redbeds in the Catskill Mountains, U.S.A.; Quarterly Journal of the Geological Society of London, v. 122, p. 273-292.

Frisch, T., Morgan, W.C., and Dunning, G.R.
1978: Reconnaissance geology of the Precambrian Shield on Ellesmere and Coburg Islands, Canadian Artic Archipelago; p. 135-138 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 588 p.

Fryer, B.J.
1972: Age determinations in the Circum-Ungava Geosyncline and the evolution of Precambrian iron-formations; Canadian Journal of Earth Sciences, v. 9, p. 652-663.

Gabrielse, H.
1967a: Tectonic evolution of the northern Canadian Cordillera; Canadian Journal of Earth Sciences, v. 4, p. 271-298.

Gabrielse, H. (cont.)
1967b: Geology, Watson Lake, Yukon Territory; Geological Survey of Canada, Map 19-1966.

1972a: Operation Finlay; p. 26-29 in Report of Activities, Geological Survey of Canada, Paper 72-1A, 243 p.

1972b: Younger Precambrian of the Canadian Cordillera; American Journal of Science, v. 272, p. 521-536.

Gabrielse, H., Blusson, S.L., and Roddick, J.A.
1973: Geology of Flat River, Glacier Lake and Wrigley Lake map-areas, District of Mackenzie and Yukon Territory; Geological Survey of Canada, Memoir 366, 153 p.

Gary, Margaret, McAfee, Robert Jr., and Wolf, C.L. (Eds.)
1972: Glossary of Geology; American Geological Institute (Pub.), 805 p.

Gastil, G., Blais, R., Knowles, D.M., and Bergeron, R.
1960: The Labrador Geosyncline; p. 21-38 in Precambrian stratigraphy and correlations; 21st International Geological Congress Report, Part 9.

Geldseltzer, H.
1973a: The tectono-sedimentary development of an algal-dominated Helikian succession on northern Baffin Island, N.W.T.; p. 99-126, in Symposium on the Geology of the Canadian Arctic Proceedings, Saskatoon, May, 1973, 368 p.

1973b: Syngenetic dolomitization and sulphide mineralization; p. 115-127 in Ores in sediments, G.C. Amstutz, and A.S. Bernard, (Ed.), Springer (Pub.), New York, 350 p.

Gandhi, S.S. and Brown, A.C.
1975: Cupriferous shales of the Adeline Island Formation, Seal Lake Group, Labrador; Economic Geology, v. 70, p. 145-163.

Gibb, R.A. and Thomas, M.D.
1976: Gravity signature of fossil plate boundaries in the Canadian Shield; Nature, v. 262, p. 199-200.

Gibb, R.A. and Walcott, R.I.
1971: A Precambrian suture in the Canadian Shield; Earth and Planetary Science Letters., v. 10, p. 417-422.

Glennie, K.W.
1970: Desert Sedimentary Environments; Development in Sedimentology 14; Elsevier (Pub.), 222 p.

Goldich, S.S.
1968: Geochronology in the Lake Superior region; Canadian Journal of Earth Sciences, v. 5, p. 715-724.

Gordey, S.P.
1978: Stratigraphy and structure of the Summit Lake area, Yukon and Northwest Territories; in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 588 p.

Grandstaff, D.E.
1976: A kinetic study of the dissolution of uraninite; Economic Geology, v. 71, p. 1493-1506.

Green, L.H.
1972: Geology of Nash Creek, Larsen Creek, and Dawson map-areas, Yukon Territory; Geological Survey of Canada, Memoir 364, 157 p.

Green, L.H. and Godwin, C.I.
1963: Mineral industry of Yukon Territory and southwestern District of Mackenzie, 1962; Geological Survey of Canada, Paper 63-38, 71 p.

Green, L.H., and Godwin, C.I. (cont.)
1964: The mineral industry of Yukon Territory and southwestern District of Mackenzie, Northwestern Territories, 1963; Geological Survey of Canada, Paper 64-36, 94 p.

Green, L.H., Roddick, J.A., and Blusson S.L.
1968: Geology, Nahanni, District of Mackenzie and Yukon Territory; Geological Survey of Canada, Map 8-1967.

Greene, B.A.
1970: Geological map of Labrador; Mineral Resources Division, Department of Mines, Agriculture and Resources, Provinces of Newfoundland and Labrador; Scale 1:1 000 000.

1974: An outline of the geology of Labrador; Department of Mines and Energy, Province of Newfoundland, Information Circular, no. 15, 64 p.

Gross, G.A.
1973: The depositional environment of principal types of Precambrian iron-formations; p. 15-21 in Genesis of Precambrian iron and manganese deposits, Proceedings, Kiev Symposium, 1970, U.N.E.S.C.O. Earth Science, v. 9, 382 p.

Hagmaier, J.L.
1971: The relation of uranium occurrences to groundwater flow systems; Wyoming Geological Association, v. 4, p. 19-24.

Halls, W.C. and West, G.F.
1971: A seismic refraction survey in Lake Superior; Canadian Journal of Earth Sciences, v. 8, p. 610-630.

Hamblin, W.K.
1958: The Cambrian Sandstones of northern Michigan; Michigan Geological Survey Publication 51, 146 p.

Harrison, J.E.
1972: Precambrian Belt Basin of northwestern United States: its geometry, sedimentation, and copper occurrences; Geological Society of America Bulletin, v. 83, p. 1215-1240.

1974: Copper mineralization in miogeosynclinal clastics of the Belt Supergroup, northwestern United States; p. 353-366 in Gisements stratiformes et provinces cupriferes, Centenaire de la Société Géologique de Belgique, 427 p.

Harshman, E.N.
1974: Distribution of elements in some roll-type uranium deposits; p. 169-183 in Symposium on the Formation of uranium ore deposits, Athens, 6-9 May, 1974, International Atomic Energy Agency, 748 p.

Hartmann, M.
1963: Einige geochemische Untersuchungen an Sandsteinen aus Perm und Trias; Geochimica et Cosmochimica Acta, v. 27, p. 459-499. (Abstract in English.)

Hay, R.E.
1963: The geology of the Sault Ste. Marie area; unpublished Ph.D. thesis, McGill University, Montreal, Canada 325 p.

Hayes, A.O. and Howell, B.F.
1937: Geology of Saint John, New Brunswick; Geological Society of America, Special Paper 5, 146 p.

Helgeson, H.C.
1964: Complexing and hydrothermal ore deposition; International Series of Monographs in Earth Science, v. 17; MacMillan, 128 p.

1967: Silicate metamorphism in sediments and the genesis of hydrothermal ore solutions; p. 333-342 in Genesis of stratiform lead-zinc barite-fluorite deposits in carbonate rocks (the so-called Mississippi valley type deposits) in Symposium of Society of Economic Geologists, U.N.E.S.C.O., International Union of Geological Sciences, The John Lindsley Foundation; J.S. Brown (Ed.), 443 p.

Hem, J.D. and Skougstad, M.W.
1960: Coprecipitation effects in solutions containing ferrous, ferric, and cuprous ions, chemistry of iron in natural waters; U.S. Geological Survey, Water Supply Paper 1459E, p. 95-110.

Henderson, J.F.
1939a: Taltson Lake, District of Mackenzie; Geological Survey of Canada, Map 525A.

1939b: Nonacho Lake, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Map 526A.

Hews, P.C.
1976: A Rb-Sr whole rock study of some Proterozoic rocks, Richmond Gulf, Nouveau, Quebec; unpublished M. Sc. thesis, Carleton University, Ottawa, Canada, 64 p.

Hoffman, P.
1976: Environmental diversity of middle Precambrian stromatolites; p. 599-611 in Stromatolites Developments in Sedimentology, M.R. Walter (Ed.), v. 20, 795 p.

Hoffman, P., Dewey, J.F., and Burke, K.
1974: Aulacogens and their genetic relation to geosyncline, with a Proterozoic example from Great Slave Lake, Canada; in Modern and ancient geosynclinal sedimentation; R.H. Dott Jr., and R.H. Shaver, (Ed.), Society of Economic Paleontologists and Mineralogists, Special Publication, no. 19, 380 p.

Hoffman, P.F.
1968: Stratigraphy of the lower Proterozoic (Aphebian) Great Slave Supergroup, East Arm of Great Slave Lake, District of Mackenzie; Geological Survey of Canada, Paper 68-42, 93 p.

1969: Proterozoic paleocurrents and depositional history of the East Arm fold belt, Great Slave Lake, Northwest Territories; Canadian Journal of Earth Sciences, v. 6, p. 441-462.

1973: Evolution of an early Proterozoic continental margin: the Coronation geosyncline and associated aulacogens of the northwestern Canadian Shield; Philosophical Transactions of the Royal Society of London, v. 273, p. 547-581.

Hoffman, P.F., Bell, I.R., Hildebrand, R.S., and Thorstad, L.
1977: Geology of the Athapuscow Aulacogeh, East Arm of Great Slave Lake, District of Mackenzie; p. 117-129 in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, 527 p.

Hoffman, P.F., Bell, I.R., and Tirrul, R.
1976: Sloan River map-area, Great Bear Lake, District of Mackenzie; p. 353-358 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 522 p.

Hoffman, P.F. and Cecile, M.P.
1974: Volcanism and plutonism, Sloan River map-area (86R), Great Bear Lake, District of Mackenzie; p. 173-176 in Report of Activities, Part A, Geological Survey of Canada, Paper 74-1A, 396 p.

Hoffman, P.F., Fraser, J.A., and McGlynn, J.C.
1970: The Coronation geosyncline of Aphebian age, District of Mackenzie; p. 201-212 in Geological Survey of Canada, Paper 70-40, 265 p.

Hoffman, P.F. and McGlynn, J.C.,
1977: Great Bear Batholith: A volcano-plutonic depression; p. 169-192 in Volcanic Regimes in Canada, Geological Association of Canada, Special Paper 16, 476 p.

Holland, H.D.
1962: Model for the Evolution of the Earth's Atmosphere; p. 447-477 in Petrological Studies, a volume to honor A.F. Buddington; A.E.J. Engel, H.L. James, and B.F. Leonard (Ed.), Geological Society of America, 660 p.

Hubert, J.F. and Reed, A.A.
1978: Red-bed diagenesis in the East Berlin Formation, Newark Group, Connecticut Valley; Journal of Sedimentary Petrology, v. 48, p. 175-184.

Hughes, C.J. and Bruckner, W.D.
1971: Late precambrian rocks of eastern Avalon Peninsula; Newfoundland – a volcanic island complex; Canadian Journal of Earth Sciences, v. 8, p. 899-915.

Hurst, R.
1975: Geochronological studies in the Precambrian Shield of Canada; unpublished Ph.D. thesis, University of California, 130 p.

Irvine, T.N.
1970: Geological age and structural relations of the Muskox Intrusion; p. 149-153 in Report of Activities, Part A, Geological Survey of Canada., Paper 70-1A, 251 p.

1971: Emplacement of the Muskox Intrusion; p. 112-114 in Report of Activities, Part A, Geological Survey of Canada, Paper 71-1A, 259 p.

Irving, E. and Hastie, J.
1975: Catalogue of paleomagnetic directions and poles, second issue, Precambrian results 1957-1974; Earth Physics Branch, Energy, Mines and Resources, Canada, Geomagnetic Series 3, 42 p.

Jackson, G.D.
1960: Belcher Islands, Northwest Territories; Geological Survey of Canada, Paper 60-20, 13 p.

Jackson, G.D. and Davidson, A.
1975: Bylot Island map-area, District of Franklin; Geological Survey of Canada, Paper 74-29, 12 p.

Jackson, G.D., Davidson, A., and Morgan, W.C.
1975: Geology of the Pond Inlet map-area, Baffin Island, District of Franklin; Geological Survey of Canada, Paper 74-25, 33 p.

Jacobsen, J.B.E.
1975: Copper deposits in time and space; Mineral Science Engineering, v. 7, p. 337-371.

Jefferson, C.W. and Young, G.M.
1977: Use of stromatolites in regional lithological correlations of Upper Proterozoic successions of the Amundsen Basin and Mackenzie Mountains, Canada; Program with Abstracts, v. 2 Geological Association of Canada, p. 26.

Jenness, S.E.
1963: Terra Nova and Bonavista map-areas, Newfoundland; Geological Survey of Canada, Memoir 327, 184 p.

Johnson, K.S.
1974: Permian copper shales of southwestern United States; p. 383-393 in Gisements stratiformes et provinces cupriferes, Bartholemé, Paul (Ed.), Centenaire de la Société Géologique de Belgique, Liège, 1974, 427 p.

Johnson, K.S. and Croy, R.L.
1976: Stratiform copper deposits of the mid continent region, a symposium; Oklahoma Geological Survey Circular 77, 99 p.

Kearey, P.
1976: A regional structural model of the Labrador Trough, northern Quebec, from gravity studies, and its relevance to continent collision in the Precambrian; Earth and Planetary Science Letters, v. 28, no. 3., p. 371-378.

Kerr, J.W. and Christie, R.L.
1965: Tectonic history of Boothia Uplift and Cornwallis Fold Belt, Arctic Canada; American Association of Petroleum Geologists Bulletin, v. 49, p. 905-926.

Kidd, D.F.
1933: Great Bear Lake area, Northwest Territories; Geological Survey of Canada, Summary Report, 1932, Part C, p. 1-36.

Kimberley, M.M.
1975: Archean iron-formation and the early atmosphere; 1975 Geotraverse Symposium, University of Toronto, 5 p.

Kimberley, M.M. and Dimroth, E.
1976: Basic similarity of Archean to subsequent atmospheric and hydrospheric compositions as evidenced in the distributions of sedimentary carbon, sulphur, uranium and iron; p. 579-585 in The Early history of the Earth, Proceedings of N.A.T.O. Advanced Study Institute Leicester University, 5-11 April 1975, B.F. Windley, (Ed.), 619 p.

Kindle, E.D.
1969: The nature of the Coppermine River copper deposits (86O and 86N); p. 112-113 in Report of Activities, Part A, 1968, Geological Survey of Canada, Paper 69-1A, 263 p.

Kindle, E.M.
1924: Geography and geology of Lake Melville district, Labrador Peninsula; Geological Survey of Canada, Memoir 141, 105 p.

Kirkham, R.V.
1970: Some copper occurrences in younger sedimentary rocks of the Coppermine River area, Northwest Territories; p. 57-63 in Report of Activities, Part B, Geological Survey of Canada, Paper 70-1B, 99 p.

Kirkham, R.V. (cont.)
1973: Environments of formation of concordant and peneconcordant copper deposits in sedimentary sequences; Canadian Mineralogist, v. 12, p. 145-146.

1974: A synopsis of Canadian stratiform copper deposits in sedimentary sequences; p. 367-382 in Gisements stratiformes et provinces cuprifères, Centenaire de la Société Géologique de Belgique, Liège, Bartholomé, Paul (Ed.), 1974, 427 p.

Knipping, H.D.
1974: The concepts of supergene versus hypogene emplacement of uranium at Rabbit Lake, Saskatchewan, Canada; p. 531-549 in Symposium on the Formation of uranium ore deposits, Athens 6-10 May 1974, International Atomic Energy Agency Vienna, 748 p.

Krogh, T.E. and Davies, G.L.
1973: The significance of inherited zircons on the age and origin of igneous rocks - an investigation of the ages of the Labrador adamellites; p. 610-613 in Carnegie Institution of Washington, Yearbook 72, 777 p.

Krynine, P.D.
1950: Petrology, stratigraphy and origin of the Triassic sedimentary rocks of Connecticut; Connecticut State Geology and Natural History Survey, Bulletin 73, 239 p.

Kumarapeli, P.S. and Saull, V.A.
1966: The St. Lawrence Valley System: a North American equivalent of the East African Rift Valley System; Canadian Journal of Earth Sciences, v. 3, p. 639-658.

Lajoie, Jean and Chagnon, André
1973: Origin of red beds in a Cambrian flysch sequence, Canadian Appalachians, Quebec; Sedimentology, v. 20, p. 91-103.

Landergren, S.
1974: Vanadium; Section 23 B-O in Handbook of Geochemistry, Wedepohl, K.H., (Ed.), v. II-2, Springer (Pub.)

Langford, F.F.
1974: A supergene origin for vein-type uranium ores in the light of the western Australian calcrete-carnotite deposits; Economic Geology, v. 69, p. 516-526.

1977: Surficial origin of North American pitchblende and related uranium deposits; American Association of Petroleum Geologists Bulletin v. 61, p. 28-42.

Langmuir, Donald
1971: Particle size effect on the reaction goethite = hematite + water; American Journal of Science, v. 171, p. 147-156.

Laznicka, Peter
1977: Geology and mineralization in the Dolores Creek area, Bonnet Plume Range, Yukon; p. 435-439 in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, 527 p.

LeCheminant, A.N., Blake, D.H., Leatherbarrow, R.W. and de Bie, L.
1977: Geological Studies: Thirty Mile Lake and MacQuoid Lake map areas, District of Keewatin; p. 205-208 in Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, 527 p.

Leggett, S.R.,
1974: A petrographic and stratigraphic study of the Flaherty Formation, Belcher Islands, Northwest Territories; B.Sc. thesis (unpublished) Brock University, St. Catherines, Ontario, Canada, 66 p.

Lemon, R.R.H. and Blackadar, R.G.
1963: Admirality Inlet area, Baffin Island, District of Franklin; Geological Survey of Canada, Memoir 328, 84 p.

Long, D.G.F.,
1978: Depositional environments of a thick Proterozoic sandstone: the (Huronian) Mississagi Formation of Ontario, Canada; Canadian Journal of Earth Sciences, v. 15, p. 190-206.

Low, A.P.
1896: Report on explorations in the Labrador Peninsula along the Eastmain, Koksoak, Hamilton, Manikuagan and portions of other rivers in 1892-93-94-95; Geological Survey of Canada, Summary Report 1895, v. 8, part L, 387 p.

Lowdon, J.A.
1961: Age determinations by the Geological Survey of Canada, Report 2 - isotopic ages; Geological Survey of Canada, Paper 61-17, 127 p.

Macey, G.
1973: A sedimentation comparison of two Proterozoic red bed successions (the South Channel and Kazan Formations of Baker Lake, Northwest Territories and the Martin Formation at Uranium City, Saskatchewan); unpublished M.sc. thesis, Carleton University, Ottawa, Canada, 175 p.

MacKean, B.E.
1968: Geology of the Elk Lake area; Ontario Department of Mines, Geological Report 62, 62 p.

Mann, E.L.
1959: The geology of the Seal Lake area, central Labrador; unpublished Ph.D. thesis, McGill University, Montreal, Canada, 252 p.

Mansy, J.L.
1978: The stratigraphy and structure of Proterozoic rocks near Good Hope Lake, McDame map area, British Columbia; p. 5-6 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 558 p.

Marten, B.E. and Smyth, W.R.
1975: Uranium potential of the basal unconformity of the Seal Lake Group, Labrador; p. 106-115 in Report of Activities, 1974, Newfoundland Department of Mines and Energy, Mineral Development Division, Report 75-1, 133 p.

McCartney, W.D.
1967: Whitbourne map-area, Newfoundland; Geological Survey of Canada, Memoir 341, 135 p.

1969: Geology of Avalon Peninsula, southeast Newfoundland; p. 115-129 in North Atlantic – geology and continental drift; M. Kay, (Ed.), American Association of Petroleum Geologists, Memoir 12, 1082 p.

McConnell, R.G.
1927: Sault Ste. Marie area, District of Algoma; Ontario Department of Mines, 35th Annual Report, Part 2, 1926, p. 1-52.

McGlynn, J.C.
1964: Grant Lake area; p. 14 in Summary of Activities: Field 1963, Geological Survey of Canada, Paper 64-1, 92 p.

1966: Thekulthili Lake area; p. 32-33 in Report of Activities, Geological Survey of Canada, Paper 66-1, 210 p.

1968: Chart I. Geotectonic correlation chart for the Canadian Shield; in Geology and Economic Minerals of Canada; R.J.W. Douglas, (Ed.), Geological Survey of Canada, Economic Geology Report, no. 1, 5th Ed., 838 p.

1970a: Study of the Nonacho Group of sedimentary rocks, Nonacho Lake, Taltson and Reliance areas, District of Mackenzie; p. 154-155, in Report of Activities, Part A, Geological Survey of Canada, Paper 70-1A, 251 p.

1970b: Churchill Province; p. 85-101 in Geology and Economic Minerals of Canada; R.J.W. Douglas, (Ed.), Geological Survey of Canada, Economic Geology Report no. 1, 5th Ed., 838 p.

1971a: Stratigraphy, sedimentology and correlation of the Nonacho Group, District of Mackenzie; p. 140-142 in Report of Activities, Part A, Geological Survey of Canada, Paper 71-1A, 259 p.

1971b: Metallic mineral industry, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 70-17, 194 p.

1974: Geology of the Calder River map area, District of Mackenzie; p. 383-385 in Report of Activities, Part A, Geological Survey of Canada, Paper 74-1A, 396 p.

McGlynn, J.C., Hanson, G.N., Irving, E., and Park, J.K.
1974: Paleomagnetism and age of Nonacho Group sandstones and associated Sparrow Dikes District of Mackenzie; Canadian Journal of Earth Sciences, v. 11, p. 30-42.

McIllwaine, W.H. and Wallace, H.
1970: Black Bay Peninsula and vicinity, District of Thunder Bay; p. 35-38 in Summary of Field Work, 1970, Ontario Department of Mines, Miscellaneous Paper 43, 96 p.

McIllwaine, W.H., Wallace, H., Franklin, J.M., and Poulsen, K.H.
1974: Stratigraphy and tectonic setting of the Late Precambrian (Helikian) of northwestern Ontario (abstr.); p. 60-61 in Geological Association of Canada - Mineralogical Association of Canada, Annual Meeting, St. John's, Newfoundland, Program with Abstracts, 101 p.

Miall, A.D.
1976: Proterozoic and Paleozoic geology of Banks Island, Arctic Canada; Geological Survey of Canada, Bulletin 258, 77 p.

Miller, A.R.
1978: Uranium reconnaissance of the northern Richmond Gulf area, New Quebec; p. 111-117 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 558 p.

Miller, D.N. Jr. and Folk, R.L.
1955: Occurrence of detrital magnetite and ilmenite in red sediments-a new approach to significance of red beds; American Association of Petroleum Geologists, Bulletin, v. 39, 338-345.

Milligan, G.C.
1960: Geology of the Lyn Lake District; Manitoba Department of Mines and Natural Resources, Publication 57-1, 317 p.

Monger, J.W.H., Souther, J.G., and Gabrielse, H.
1972: Evolution of the Canadian Cordillera: a plate-tectonic model; American Journal of Science, v. 272, p. 577-602

Morgan, W.C.
1975: Geology of the Precambrian Ramah Group and basement rocks in the Nachvak Fiord - Saglek Fiord area, North Labrador; Geological Survey of Canada, Paper 74-54, 42 p.

Morton, R., Goble, E., and Goble, R.J.
1973: Sulphide deposits associated with Precambrian Belt - Purcell strata in Alberta and British Columbia, Canada; p. 159-179 in Belt Symposium, University of Idaho, Idaho Bureau of Mines and Geology, Moscow, Idaho, v. 1.

Morton, R.D.
1974: Sandstone-type uranium deposits in the Proterozoic strata of northwestern Canada; p. 255-273 in Symposium on the formation of uranium ore deposits, Athens, 6-10 May 1974, International Atomic Energy Agency, 748 p.

Mukherjee, A.C.
1974: Some aspects of sedimentology of the Missi Group and its environment of deposition; Canadian Journal of Earth Sciences, v. 11, p. 1018-1019.

Mukherjee, A.C., Stauffer, M.R., and Baadsgaard, H.
1971: The Hudsonian orogeny near Flin Flon, Manitoba: a tentative interpretation of Rb/Sr and K/Ar ages; Canadian Journal of Earth Sciences, v. 8, p. 939-946.

Murksy, G.
1963: Mineralogy, petrology and geochemistry of Hunter Bay area, Great Bear Lake, Northwest Territories, Canada; unpublished Ph.D. thesis, Stanford University, 144 p.

1973: Geology of the Port Radium map-area, District of Mackenzie; Geological Survey of Canada, Memoir 374, 40 p.

Norris, D.K., Price, R.A., and Mountjoy, E.W.
1963: Geology, northern Yukon Territory and northwestern District of Mackenzie; Geological Survey of Canada, Map 10-1963.

Norris, R.M.
1969: Some reddening and time; Journal of Sedimentary Petrology, v. 39, p. 7-11

Obradovich, J.D., and Peterman, Z.E.
1973: A review of the geochronology of Belt and Purcell rocks; p. 89 in Belt Symposium; University of Idaho, Idaho Bureau of Mines and Geology, v. 1, 322 p.

O'Brien, S.J., Strong, P.G. and Evans, J.L.,
1977: Geology of the Grand Bank (1M/4) and Lamaline (1L/13) map areas; p. 35-42 in Report of Activities, 1976, Mineral Development Division, Department of Mines and Energy, Province of Newfoundland, Report 77-1, 113 p.

Olson, R.A.
1977: Geology and genesis of zinc-lead deposits within a Late Proterozoic dolomite, northern Baffin Island, N.W.T.; unpublished Ph.D. thesis, University of British Columbia, 371 p.

Palonen, P.A.
1973: Paleogeography of the Mississagi formation and lower Huronian cyclicity; p. 157-167 in Huronian stratigraphy and sedimentation; G.M. Young, (Ed.), Geological Association of Canada, Special Paper no. 12, 271 p.

Patel, I.
1973: Sedimentology of the Ratcliffe Brook Formation (Lower Cambrian?) in southeastern New Brunswick; p. 206-207 in 8th Annual Meeting Northeastern Section, Geological Society of America, Abstracts with Program, v. 5, no. 2, 246 p.

Pearson, D.E.
1972: The location and structure of the Precambrian Kisseynew gneiss domain of northern Saskatchewan; Canadian Journal of Earth Sciences, v. 9, p. 1235-1249.

Pearson, W.N.,
1978: Copper Metallogeny, Lake Huron area, Ontario; p. 263-268 in Current Research, Part A, Geological Survey of Canada, Paper 78-1A, 558 p.

Perry, E.C., Jr., Monster, Jan, and Reimer, Thomas
1971: Sulfur isotopes in Swaziland system Barites and the evolution of the Earth's atmosphere; Science, v. 171, p. 1015-1016.

Pettijohn, F.J.
1957: Sedimentary rocks; 2nd Ed., Harper and Brothers (Pub.), New York, 718 p.

Poole, W.H.
1976: Plate tectonic evolution of the Canadian Appalachian region; p. 113-126 in Report of Activities, Part B, Geological Survey of Canada, Paper 76-1B, 380 p.

Poole, W.H., Sanford, B.V., Williams, H., and Kelley, D.G.
1970: Geology of Southeastern Canada; p. 229-304 in Geology and Economic Minerals of Canada; R.J.W. Douglas, (Ed.), Geological Survey of Canada, Economic Geology Report 1, 838 p.

Premoli, Camillo
1976: Formation of, and prospecting for, uraniferous calcretes; Australian Mining, v. 68, p. 13-16.

Price, R.A.
1962: Fernie map-area, east half, Alberta and British Columbia; Geological Survey of Canada, Paper 61-24, 65 p.

1964: The Precambrian Purcell System in the Rocky Mountains of southern Alberta and British Columbia; Bulletin of Canadian Petroleum Geology, v. 12, p. 399-426.

1965: Flathead map-area, British Columbia and Alberta; Geological Survey of Canada, Memoir 336, 221 p.

Price, R.A., and Douglas, R.J.W. (Ed.)
1972: Variations in tectonic styles in Canada, Geological Association of Canada, Paper no. 11, 688 p.

Pye, E.G.
1968: Nipigon Schreiber Sheet, Thunder Bay District; Ontario Department of Mines, Compilation Map 2137, scale 4 miles = 1 inch

1970: Sioux Lookout - Armstrong sheet, Kenora and Thunder Bay District; Ontario Department of Mines, Compilation Map 2169, scale 4 miles = 1 inch.

Pye, E.G. and Fenwick, K.G.
1964: Atikokan-Lakehead sheet, Kenora and Thunder Bay districts; Ontario Department of Mines, Compilation Map 2065, scale 4 miles = 1 inch.

Pye, E.G. Harris, F.R., Fenwick, K.G., and Baillie, J.
1966: Tashota-Geraldton sheet, Thunder Bay and Cochrane Districts; Ontario Department of Mines Compilation Map 2102, scale 4 miles = 1 inch.

Rackley, R.I.,
1977: Origin of western-states type uranium mineralization; p. 89-156 in Wolf K.H., (Ed.), Handbook of stratabound and stratiform ore deposits v. 7, Au, U, Fe, Mn, Hg, Sb, W and P deposits, 656 p.

Rankama, Kalervo and Sahama, Th.G.
1949: Geochemistry; University of Chicago Press, 912 p.

Reesor, J.E.
1957: The Proterozoic of the Cordillera in southeastern British Columbia and southwestern Alberta; p. 150-177 in The Proterozoic in Canada; J.E. Gill, (Ed.); Royal Society of Canada, Special Publication, no. 2, 191 p.

Reinhardt, E.W.
1972: Occurrences of exotic breccias in the Petitot Islands and Wilson Island map-areas, East Arm of Great Slave Lake, District of Mackenzie; Geological Survey of Canada, Paper 72-25, 43 p.

Renfro, A.R.
1974: Genesis of evaporite-associated stratiform metalliferous deposit – a sabkha process; Economic Geology, v. 69, p. 33-45.

Rentsch, J.
1974: The Kupferschiefer in comparison with the deposits of the Zambian copperbelt; p. 395-418 in Gisements stratiformes et provinces cupriferes, Centenaire de la Société Géologique de Belgique, Liège, Bartholomé, Paul (Ed.), 1974, 427 p.

Rich, R.A., Holland, H.D., and Peterson, Ulrich,
1977: Hydrothermal uranium deposits; Developments in Economic Geology, 6, Elsevier (Pub.) 264 p.

Rickard, D.T.
1974: Low temperature copper geochemistry: gitological aspects; p. 134 in Gisements stratiformes et provinces cupriferes, Centenaire de la Société Géologique Belgique, Liège, Bartholomé, Paul, (Ed.), 1974, 427 p.

Robertson, D.S.
1974: Basal Proterozoic units as fossil time markers and their use in uranium prospection; p. 495-512 in Symposium on the Formation of uranium ore deposits, Athens 6-10 May 1974, International Atomic Energy Agency 748 p.

Robertson, D.S. and Lattanzi, C.R.
1974: Uranium deposits of Canada; Geoscience Canada., v. 1, no. 2, p. 9-20.

Robertson, J.A.
1962: Iron Bridge area, District of Algoma, Ontario; Ontario Department of Mines, Map 2012.

Robertson, J.A., Card, K.D., and Frarey, M.J.
1969: The Federal-Provincial Committee on Huronian stratigraphy; progress report, Ontario Department of Mines, Miscellaneous Paper, no. 31, 26 p.

Robertson, W.A.
1964: Palaeomagnetic results from northern Canada suggesting a tropical Proterozoic climate; Nature, v. 204, p. 66-67.

Robertson, W.A., and Baragar, W.R.A.
1972: The petrology and paleomagnetism of the Coronation sills; Canadian Journal of Earth Sciences, v. 9, p. 123-140.

Robinson, B.W. and Morton, R.D.
1972: The geology and geochronology of the Echo Bay area, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 9, p. 158-171.

Roddick, J.A., and Green, L.H.
1961: Geology, Sheldon Lake, Yukon Territory; Geological Survey of Canada, Map 12-1961.

Rodgers, J.
1972: Latest Precambrian (post-Grenville) rocks of the Appalachian region; American Journal of Science, v. 272, p. 507-520.

Roscoe, S.M.
1957: Cambrian Lake (East Half), New Quebec; Geological Survey of Canada, Paper 57-6, 13 p.

1969: Huronian rocks and uraniferous conglomerates in the Canadian Shield; Geological Survey of Canada, Paper 68-40, 205 p.

1973: The Huronian Supergroup, a Paleoaphebian succession showing evidence of atmospheric evolution; p. 31-47 in Huronian stratigraphy and sedimentation; Geological Association of Canada, Special Paper no. 12, G.M. Young, (Ed.), 271 p.

Roscoe, S.M. and Emslie, R.F.
1973: Geology Kasheshibaw Lake, Newfoundland, Quebec; Geological Survey of Canada, Map 1342A.

Rose, A.W.
1976: The effect of cuprous chloride complexes in the origin of redbed copper and related deposits; Economic Geology, v. 71, p. 1036-1048.

Rose, E.R.
1952: Torbay map-area, Newfoundland; Geological Survey of Canada, Memoir 265, 64 p.

Rosholt, J.N., Zartman, R.E., and Nkomo, I.T.
1973: Lead isotope systematics and uranium depletion in the granite mountains, Wyoming; Geological Society of America Bulletin, v. 84, p. 989-1002.

Rowlands, N.J.
1973: The Adelaidean System of South Australia; a review of its sedimentation, tectonics and copper occurrences; p. 80-112 in Belt Symposium University of Idaho, Idaho Bureau of Mines and Geology, vol. 1, D.R. Bishop and J.D. Powell (Ed.), 332 p.

Rubey, W.W.
1955: Development of the Hydrosphere and Atmosphere with special reference to probable composition of the early atmosphere; p. 631-650 in Crust of the Earth, (A. Poldervaart, Ed.) Geological Society of America, Special Paper no. 62, 762 p.

Rubin, Bruce
1970: Uranium role front zonation in the southern Powder River Basin, Wyoming; Wyoming Geological Association, v. 3, p. 5-12.

Ruitenberg, A.A., Venugopal, D.V., and Giles, P.S.
1973: "Fundy Cataclastic Zone" New Brunswick: evidence for post-Acadian penetrative deformation; Geological Society of America, Bulletin, v. 84, p. 3029-3044.

Ruzicka, V.
1976a: Assessment of some Canadian uranium occurrences; p. 341-342 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 522 p.

1976b: Evaluation of uranium resources in the Elliot Lake-Blind River area, Ontario; p. 127-129 in Report of Activities, Part B, Geological Survey Canada, Paper 76-1B, 380 p.

Samama, J.C.
1976: Comparative review of the genesis of the copper-lead sandstone-type deposits; p. 1-20 in Handbook of strata-bound and strataform ore deposits, Wolf, K.H., (Ed.), v. II, Regional studies and specific deposits, 585 p.

Sangster, D.F.
1972: Isotopic studies of ore-leads in the Hanson Lake-Flin Flon-Snow Lake Mineral Belt; Canadian Journal of Earth Sciences, v. 9, p. 500-513.

Schenk, P.E.
1971: Southeastern Atlantic Canada, northwestern Africa, and continental drift; Canadian Journal of Earth Sciences, v. 8, p. 1218-1251.

Schidlowski, Manfied
1976: Archaean-Archrean atmosphere and evolution of the terrestrial oxygen budget; p. 525-535 in The early history of the earth; Proceedings of N.A.T.O. Advance Study Institute Leicester University, 5-11 April, 1975; B.F. Windley, (Ed.), 619 p.

Schluger, P.R.
1976: Petrology and origin of the red beds of the Perry Formation, New Brunswick, Canada and Main, U.S.A.; Journal of Sedimentary Petrology, v. 46, p. 22-37.

Schmalz, R.F.
1959: A note on the system Fe_2O_3-H_2O, Journal of Geophysical Research, v. 64, p. 575-579.

1968: Formation of red-beds in modern and ancient deserts; Discussion, Geological Society of America Bulletin, v. 79, p. 277-280.

Schumm, S.A.
1968: Speculations concerning paleohydrologic controls of terrestrial sedimentation; Geological Society of America Bulletin, v. 79, p. 1573-1588.

Serdyuchenko, D.P.
1975: Some Precambrian scapolite-bearing rocks evolved from evaporites; Lithos, v. 8, p. 1-7.

Sharma, K.N.M.
1974a: La Grande River area (1973 Project); Quebec Department of Natural Resources, D.P. 221, 37 p.

1974b: La Grande Rivière area (1974 Project); Quebec Department of Natural Resources, D.P. 275, 21 p.

1976: La Grande Riviere area (1975 Project); Quebec Department of Natural Resources, D.P. 345, 23 p.

1977: La Grande Rivière area, Nouveau-Québec Territory; p. 44-49; in Travaux sur le Terrain 1977, Ministère des Richesses Naturelles, Québec, 82 p.

Siemiatkowska, K.M.
1976: Geology of the Endikai Lake area, District of Algoma; Ontario Division of Mines, Geological Branch, Open File Report 5212, 104 p.

Smith, G.E.
1974: Depositional systems, San Angelo Formation (Permian), North Texas-Facies control of red-bed copper mineralization; Report of Investigations No. 80, Bureau of Economic Geology, University of Texas, Austin, 58 p.

Stanton, R.L.
1972: Ore petrology; McGraw-Hill (Pub.), 713 p.

Stauffer, M.R.
1974: Geology of the Flin Flon Area: a new look at the sunless city; Geoscience Canada, v. 1, no. 3, p. 30-35.

Stevenson, I.M.
1967a: Goose Bay map area, Labrador; Geological Survey of Canada, Paper 67-33, 12 p.

1967b: Geology, Minipi Lake, Newfoundland; Geological Survey of Canada, Map 6-1967.

1968: A geological reconnaissance of Leaf River map area, New Quebec and Northwest Territories; Geological Survey of Canada, Memoir 356, 112 p.

1969: Lac Brulé and Winokapau Lake map areas, Newfoundland and Quebec; Geological Survey of Canada, Paper 67-69, 16 p.

1970: Rigolet and Groswater Bay map areas, Newfoundland (Labrador); Geological Survey of Canada, Paper 69-48, 24 p.

Stewart, J.H.
1976: Late Precambrian evolution of North America; plate tectonics implication; Geology, v. 4, p. 11-15.

Stirbys, A.S.
1975: A petrographic and geochemical study of the Eskimo Formation, Belcher Islands, Northwest Territories; B.Sc. Thesis unpublished, Brock University, St. Catherines, Ontario, Canada, 72 p.

Stockwell, C.H., McGlynn, J.C., Emslie, R.F., Sanford, B.V., Norris, A.W., Donaldson, J.A., Fahrig, W.F., and Currie, K.L.
1970: Geology of the Canadian Shelf; p. 44-150 in Geology and economic minerals of Canada; R.J.W. Douglas, (Ed.), Geological Survey of Canada, Economic Geology Report 1, 838 p.

Strong, D.F., O'Brien, S.J., Taylor, S.W., Strong, P.G., and Wilton, D.H.,
1978: Aborted Proterozoic rifting in eastern Newfoundland; Canadian Journal of Earth Sciences, v. 15, p. 117-131.

Stuckless, J.S. and Ferreira, C.P.
1976: Labile uranium in granitic rocks; International Atomic Energy Agency, SM-208/17, p. 717-730.

Taylor, F.C.
1970: Reconnaissance geology of a part of the Precambrian Shield, northeastern Quebec and northern Labrador, Part II; Geological Survey of Canada, Paper 70-24, 10 p.

1972: Reconnaissance geology of a part of the Precambrian Shield, northeastern Quebec and northern Labrador, Part III; Geological Survey of Canada, Paper 71-48, 14 p.

Taylor, G.C. and Stott, D.F.
1973: Tuchodi Lakes map area, British Columbia; Geologial Survey of Canada, Memoir 373, 37 p.

Thompson, A.M.
1970: Geochemistry of color genesis in red bed sequence, Juanita and Bald Eagle formations, Pennsylvania; Journal of Sedimentary Petrology, v. 40, p. 599-615.

Thompson, J.B., Jr.
1972: Oxides and sulfides in regional metamorphism of pelitic schists; 24th International Geological Congress, section 10, p. 27-35.

Thomson, Robert
1957: The Proterozoic of the Cobalt area; p. 40-45 in The Proterozoic in Canada; J.E. Gill, (Ed.), Royal Society of Canada, Special Publication no. 2, 191 p.

1966: Geology of Henwood Township, District of Timiskaming; Ontario Department of Mines, Miscellaneous Paper MP5, 48 p.

Thorpe, R.I.
1970: Geological exploration in the Coppermine River area, Northwest Territories; Geological Survey of Canada, Paper 70-47, 150 p.

Thorsteinsson, R. and Tozer, E.T.
1962: Banks, Victoria and Steffanson islands, Arctic Archipelago; Geological Survey of Canada, Memoir 330, 85 p.

Tourlelot, E.B. and Vine, J.D.
1976: Copper deposits in sedimentary and volcanogenic rocks; U.S. Geological Survey, Professional Paper 907-C, 33 p.

Tremblay, L.P.
1968: Preliminary account of the Goulburn Group, Northwest Territories, Canada; Geological Survey of Canada, Paper 67-8, 37 p.

1971: Geology of Beechey Lake map area, District of Mackenzie: a part of the western Canadian Precambrian Shield; Geological Survey of Canada, Memoir 365, 56 p.

1972: Geology of the Beaverlodge mining area, Saskatchewan; Geological Survey of Canada, Memoir 367, 468 p.

1973: Volcanic rocks of Beaverlodge area; p. 14-27 in Volcanism and Volcanic Rocks; Geological Survey of Canada, Open File 164, 182 p.

Trettin, H.P.
1969: Lower Palaeozoic sediments of northwestern Baffin Island, District of Franklin; Geological Survey of Canada, Bulletin 157, 70 p.

Trow, James
1977: Uraniferous quartz-pebble conglomerates and their chemical relation to CO_2– deficient atmospheres synchronous with glaciations of almost any age; Geological Society of America, Abstracts with programs, v. 9, no. 7, p. 1206.

Tuke, M.F., Dineley, D.L., and Rust, B.R.
1966: The basal sedimentary rocks in Somerset Islands, N.W.T.; Canadian Journal of Earth Sciences, v. 3, p. 697-711.

Turner, Peter
1974: Origin of Red beds in the Ringerike Group (Silurian) of Norway; Sedimentary Geology, v. 12, 215-235.

Turner, P. and Archer, R.
1977: The role of biotite in the diagenesis of red beds from the Devonian of Northern Scotland, Sedimentary Geology, v. 19, p. 241-251.

Upitis, U.
1966: The Rapitan Group, southeastern Mackenzie Mountains, Northwest Territories; unpublished M.Sc. thesis, McGill University, Montreal, Canada, 70 p.

Van Breemen, O. and Upton, B.G.J.
1972: Age of some Gardar intrusive complexes, south Greenland; Geological Society of America, Bulletin, v. 83, p. 3381-3390.

Van Houten, F.B.
1961: Climatic significance of red beds; p. 89-139 in Descriptive Palaeoclimatology, A.E.M. Nairn (Ed.)., Interscience (Pub.), 380 p.

1964: Origin of red beds - some unsolved problems; p. 647-672 in Problems in Palaeoclimatology, Proceedings N.A.T.O. Paleoclimatology Conference, University of Newcastle upon Tyne, January 7-12 1963, A.E.M. Nairn (Ed.); Interscience (Pub.) 705 p.

1968: Iron oxides in red beds; Geological Society of America Bulletin, v. 79, p. 399-416.

1973: The origin of red beds a review - 1961 - 1972; p. 39-61 in Annual Review of Earth and Planetary Science, v. 1, 1973, F.A. Donath, F.G. Stehli, and G.W. Wetherill (Ed.), 350 p.

Van Schmus, W.R.
1965: The geochronology of the Blind River-Brice Mines area, Ontario, Canada; Journal of Geology, v. 73, p. 775-780.

Vine, J.D., and Tourtelot, E.B.
1976: Red beds and copper deposits - why the association? Abstr. p. 97 in stratiform copper deposits of the midcontinent region, a symposium; K.S. Johnson, and R.L. Croy, (Ed.), Oklahoma Geological Survey, Circular 77, 99 p.

Walker, T.R.
1967a: Formation of red beds in modern and ancient deserts; Geological Society of America, Bulletin, v. 78, p. 353-368.

1967b: Colour of recent sediments in tropical Mexico: A contribution to the origin of red beds: Geological Society of America, Bulletin, v. 78, p. 917-920.

Walker, T.R.
1974: Formation of red beds in most tropical climates: a hypothesis; Geological Society of America, Bulletin, v. 85, p. 633-638.

1976: Diagenetic origin of continental red beds; p. 240-282 in the Continental Permian in central, west and south Europe; Proceedings N.A.T.O. Advance Study Institute, Johannes Gutenburg University, Mainz, F.A.G., 23 Sept. - 4 Oct. 1975, H. Falke, (Ed.), Reidel (Pub.) 352 p.

Wanless, R.K. and Loveridge, W.D.
1972: Rubidium-strontium isochron age studies, Report 1; Geologial Survey of Canada, Paper 72-23, 77 p.

Rubidium isotopic age report No. 2 (Canadian Shield) in prep.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and Delabio, R.N.
1970: Age determinations and geological studies, K-Ar isotopic ages, Report 9; Geological Survey of Canada, Paper 69-2A, 78 p.

1972: Age determinations and geological studies, K-Ar isotopic ages, Report 10; Geological Survey of Canada, Paper 71-2, 96 p.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and Edmonds, C.M.
1967: Age determinations and geological studies, K-Ar isotopic ages, Report 7; Geological Survey of Canada, Paper 66-17, 120 p.

1968: Age determinations and geological studies, K-Ar isotopic ages, Report 8; Geological Survey of Canada, Paper 67-2A, 141 p.

Wanless, R.K., Stevens, R.D., Lachance, G.R., and Rimsaite, J.Y.H.
1965: Age determinations and geological studies, Part 1 - Isotopic ages, Report 5; Geological Survey of Canada, Paper 64-17, 126 p.

1966: Age determinations and geological studies, K-Ar isotopic ages, Report 6; Geological Survey of Canada, Paper 65-17, 101.

Wedepohl, K.H.
1971: "Kuspferschiefer" as a prototype of syngenetic sedimentary ore deposits; Society of Mining Geology, Japan, Special Issue 3, p. 268-273.

1972: Zinc: Section 30-B-0, in Handbook of Geochemistry; K.H. Wedepohl, (Ed.). v. II-3, Springer, (Pub.).

1974a: Copper; Section 29-B-0 in Handbook of Geochemistry, K.H. Wedepohl, (Ed.), v. II-4, Springer (Pub.).

1974b: Lead; Section 82C-O, in Handbook of Geochemistry, K.H., Wedepohl, (Ed.), II-4, Springer Pub. 898 p.

1974c: Vanadium; Section 23-B-0, in Handbook of Geochemistry, K.H. Wedepohl, (Ed.), v. II-4, Springer, (Pub.).

Weeks, L.J.
1954: Southeast Cape Breton Island, Nova Scotia; Geological Survey of Canada, Memoir 277, 112 p.

1965: Geological map of the province of Nova Scotia; Nova Scotia Department of Mines.

Wheeler, E.P., II
1964: Unmetamorphosed sandstone in northern Labrador; Geological Society of America, Bulletin, v. 75, p. 569-570

Wheeler, J.O., Aitken, J.D., Berry, M.J., Gabrielse, H., Hutchison, W.W., Jacoby, W.R., Monger, J.W.H., Niblett, E.R., Norris, D.K., Price, R.A., and Stacey, R.A.
1972: The Cordilleran Structural Province; p. 1-82 in Variations in Tectonic Styles in Canada; R.A. Price and R.J.W. Douglas, (Ed.), Geological Association of Canada, Special Paper 11, 688 p.

White, D.E.
1968: Environments of generation of some base-metal ore deposits; Economic Geology, v. 63, p. 301-335.

Williams, H.
1967: Geology, Island of Newfoundland; Geological Survey of Canada, Map 1231A.

Williams, Harold and King, A.F.
1976: Southern Avalon Peninsula, Newfoundland; Trepassey map-area (1K); p. 179-182 in Report of Activities, Part A, Geological Survey of Canada, Paper 76-1A, 522 p.

Wilson, J.T.
1939: Preliminary maps, Fort Smith area, Northwest Territories; Geological Survey of Canada, Paper 39-11.

Wolf, K.H.
1976: Ore genesis influenced by compaction, p. 475-675 in Compaction of coarse-grained sediments, II, developments in sedimentology 18B; G.V. Chilingarian, and K.H. Wolf, (Ed.), Elsevier (Pub.), 808 p.

Wood, J.
1973: Stratigraphy and depositional environments of Upper Huronian rocks of the Rawhide Lake-Flack Lake area, Ontario; p. 73-95 in Huronian Stratigraphy and sedimentation; G.M. Young, (Ed.), Geological Association of Canada, Special Paper no. 12, 271 p.

1975: Geology of the Rawhide Lake area, District of Algoma; Ontario. Department of Mines and Natural Resources., Report 129, 67 p.

Woodcock, J.R.
1960: Geology of the Richmond Gulf area, New Quebec; Geological Association of Canada Proceedings p. 21-40.

Woodward, L.A., Kaufman, W.H., Schumacher, O.L., and Talbott, L.W.
1974: Strata-bound copper deposits in Triassic sandstone of Sierra Nacimiento, New Mexico; Economic Geology, v. 69, p. 108-120.

Wynne-Edwards, H.R.
1960: Geology, Michikamau Lake (west half), Quebec-Newfoundland; Geological Survey of Canada, Map 2-1960.

1961: Geology, Ossokmanuan Lake (west half), Newfoundland; Geological Survey of Canada, Map 17-1961.

Young, G.M.
1970: An extensive Early Proterozoic glaciation in North America?; Palaeogeography, Palaeoclimatology, Palaeoecology, v. 7, p. 85-101.

1973: Huronian stratigraphy and sedimentation; Geological Association of Canada, Special Paper G.M. Young, Ed., 271 p.

1974: Stratigraphy, paleocurrents and stromatolites of Hadrynian (Upper Precambrian) rocks of Victoria Island, Arctic Archipelago; Precambrian Research, v. 1, p. 13-41.

1976: Iron-formation and glaciogenic rocks of the Rapitan Group, Northwest Territories, Canada; Precambrian Research v. 3, p. 137-158.

1977: Stratigraphic correlation of Upper Proterozoic rocks of northwestern Canada; Canadian Journal of Earth Sciences, v. 14, p. 1771-1787.

Young, G.M. and Chandler, F.W.
1968: Possible glacial origin for three Precambrian (Huronian) conglomerates north shore of Lake Huron (abstr.); p. 42-43 in 14th Annual Institute of Lake Superior Geology, Superior, Wisconsin, 60 p.

Young, G.M. and Jefferson, C.W.
1975: Late Precambrian shallow water deposits, Banks and Victoria Islands, Arctic Archipelago; Canadian Journal of Earth Sciences, v. 12, p. 1734-1748.

1975: Late Precambrian stratigraphy and sedimentation; Banks and Victoria Islands, Arctic Archipelago, Canada (abstr.); Geological Society of America, Abstracts with Program, v. 7, p. 885.

Young, G.M. and Long, G.F.
1976: Stromatolites and basin analysis: an example from the upper Proterozoic of northwestern Canada; Paleogeography, Paleoclimatology, Paleoecology, v. 19, p. 303-319.

Ziegler, A.M. and McKerrow
1975: Silurian Marine Red beds; American Journal of Science, v. 275, p. 31-56.

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