EVOLUTION OF THE CANADIAN ARCTIC ISLANDS — A TRANSITION BETWEEN THE ATLANTIC AND ARCTIC OCEANS

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Chapter Evolution of the Canadian Arctic Islands — A Transition Between the Atlantic and Arctic Oceans, by J. Wm. Kerr.

I. INTRODUCTION

A. Geographic Limits of Study

The Canadian Arctic Islands and nearby mainland areas (Fig. 1) form a physiographic transition between the Atlantic and Arctic Oceans. Thus, the margins of two ocean basins are involved in this study.

The Canadian Arctic Islands are bordered on the southeast by Baffin Bay and by the Labrador Sea, which together form a major arm of the Atlantic Ocean. The relationship of both these features to the Arctic Islands is discussed.

The Arctic Islands are bordered on the northwest by the Arctic Ocean, where the study area partly overlaps with the northern mainland and Beaufort Shelf, treated in this volume by Norris and Yorath (1979). On the northeast the study connects with northern Greenland, which is treated in this volume by Dawes and Peel (1979).

B. Objectives and Format

This paper outlines geological relationships between the Canadian Arctic Islands and the two oceanic areas that border it on the southeast and northwest, which respectively are parts of the Atlantic and Arctic Oceans. An attempt is made to show how these three regions evolved. The Canadian Arctic (Fig. 1) is unique among lands surrounding the Arctic Ocean Basin because it is a fragmented and drowned part of a continent that now forms a shallow transitional connection between two deep ocean basins, both of which probably have an origin in rifting.

Evolution of the region extended from Precambrian to Holocene, culminating in the fragmentation of a continent and formation of ocean basins.

This paper concentrates on the evolution and on the interrelationship of the main geological features. Details and documentation are sometimes not given, but appropriate sources are cited. Much documentation is contained in earlier syntheses (Douglas *et.al.*, 1963; Thorsteinsson and Tozer, 1970; Trettin *et al.*, 1972; Kerr, 1977a; Balkwill, 1978).

Evolution of the Canadian Arctic Islands occurred in two main phases that overlapped somewhat in time. First was a constructional phase, the evolution of the continent as a substructure. During this phase the continent developed by formation of a crystalline crust, and later by development of large depositional basins upon it. By the end of this phase almost the entire area shown by Figure 1 was underlain by continental crust of approximately normal thickness (37 km). This crust was the substructure beside which and within which the ocean basins and their branches were to form later. The internal structure of that crust in large measure controlled the shapes and history of the sedimentary basins, and the present seaways.

Second was the fragmentation phase, the evolution of the ocean basins and their branches within the continent. This period extends from the earliest signs of continental breakup through the time of ocean formation to the present day.

C. Relationship of Sequences to Major Geological Provinces

Major geologic provinces in northern Canada are shown in plan view (Fig. 2), and in cross-section (Fig. 3). They make up seven stratigraphic-structural sequences.

Sequence 1 is the Precambrian crystalline basement, a gneissic crust that lies on the mantle. It is the foundation of the continent, upon which younger tectonic units lie, and where exposed at the surface forms the Canadian Shield (Douglas, 1970).

The history of the Arctic region was dominated by the Innuitian Mobile Belt. This is a continental margin-type mobile belt, made up of sedimentary sequences that are designated by numbers 2 to 8 (Fig.3). The temporal and spatial relationships of these sequences are shown in a tectonic time diagram (Fig. 4). They are separated from each other by widespread angular unconformities, each related to a tectonic event.

Sequence 2 is Proterozoic rock, remnants of a former much more extensive basin. This basin was fragmented and its parts became isolated from each other before deposition of Sequence 3. Sequence 3 is a broad depositional basin of Proterozoic to Devonian age that consists of the Arctic Platform and the Franklinian Geosyncline (Figs. 2 and 3). The geosyncline includes a clastic belt in the northwest (formerly called the eugeosyncline), and a miogeocline on the south and east (formerly called the miogeosyncline).

The Pearya Geanticline (Trettin *et al.*, 1972) has a core of gneissic rocks, that was intruded, uplifted, and eroded at various times. It was primarily a positive feature and a major source of sediments, although sediments also were deposited upon the geanticline at times during Sequence 3 deposition. These sediments are assigned to the clastic belt of the geosyncline. The geanticline may possibly be an extension of the Shield, connected with it beneath the geosyncline (Fig. 3).

Sequences 4, 5 and 6 were deposited in several basins, the best known and best exposed being the Sverdrup Basin (Thorsteinsson and Tozer, 1970; Balkwill 1978). The Sverdrup Basin is a successor basin of late Paleozoic and Mesozoic age that is unconformable on the deformed Franklinian Geosyncline and Pearya Geanticline. The southeastern limit of thick sedimentation in the basin was approximately at the southeastern limit of exposure (Fig. 3). On the northwest, the Sverdrup Basin was bordered throughout its history by the recurrently positive Sverdrup Rim, which lies above the southwestward extension of the Pearya Geanticline. Rocks of Sequences 4, 5 and 6 that are coeval with those of the Sverdrup Basin probably thin northwestward onto the Sverdrup Rim, but may thicken from there northward into the sedimentary wedge beneath the Canada Basin.

Sequences 5 and 6 are present in the Baffin Bay Basin and are exposed on its margins (Fig. 2).

Sequences 7 and 8 include Cenozoic (mid-Tertiary) and younger rocks (Fig. 3). These sequences occur as thick bodies of sediment in the Canada Basin and Baffin Bay, and as thinner columns in the channels between the Arctic islands. Northwest of the Arctic islands these sequences form a continental terrace wedge that is exposed in the Arctic Coastal Plain (Fig. 2), and thickens northwestward beneath the continental shelf and continental slope (Fig. 3). They form a more confined but thick wedge of sediments in Baffin Bay, but there the rocks of Sequences 7 and 8 are beneath the water.

II. CONSTRUCTIONAL PHASE: EVOLUTION OF THE CONTINENT AS A SUBSTRUCTURE

A. Sequence 1: Precambrian Crystalline Basement

The Precambrian crystalline basement forms the foundation of the North American Continent and rests on the underlying mantle (Figs. 2 and 3). It crops out as the Canadian Shield, and is widely exposed in the southeastern part of the study area, including much of Greenland. The Canadian Shield is subdivided into structural provinces (Stockwell, 1964; Stockwell *et al.*, 1970, based mainly on differences in structural trend, style of folding, and radiometric age.

A narrow belt of gneissic rocks on northwestern Ellesmere Island has metamorphic ages of 800 m.y. to 1000 m.y., indicating an intense metamorphism of about Grenvillian age (Sinha and Frisch, 1975, 1976). It is part of the Pearya Geanticline (Trettin *et al.*, 1972), and may connect with the Shield beneath the sedimentary sequences (Fig. 3). The thickness of the crust in the Arctic Islands is about 37 km (Sobczak and Weber, 1973), which is approximately normal for continental crust.

B. Innuitian Mobile Belt

The history of the Arctic continental margin of Canada from Ellesmere Island to Banks Island has been dominated by the Innuitian Mobile Belt from late Proterozoic time to the present (Figs. 2 and 3). This is a continental margin-type mobile belt that trends northeastward, and includes features extending from the Proterozoic basins northwestward to the modern basin that trends along the margin of the Arctic Ocean. Several major depositional basins that make up the mobile belt lie one above the other. They are: a Proterozoic basin (Sequence 2), the Franklinian Geosyncline (Sequence 3), the Sverdrup Basin and related troughs (Sequences 4, 5 and 6), and a mid-Tertiary to modern basin (Sequences 7 and 8). The northern part of the continent was undergoing a long lasting constructional phase until the end of Sequence 3 (Fig. 4). From that time to the present day it has been in a fragmentation phase.

C. Sequence 2: Proterozoic Basins

These oldest sediments of the Innuitian Mobile Belt are Proterozoic (Helikian and possibly Hadrynian), and are exposed in several areas that are partly or completely disconnected from each other (Fig. 2). They were subjected to intermittent tectonic activity including dyke intrusion and faulting. A remnant of these Sequence 2 rocks that occurs on northern Baffin Island is shown in cross-section (Fig. 3). It is isolated on this line of cross-section (Fig. 3), but on its west side probably connects with other parts of an original larger unnamed western basin. These rocks form the Milne Inlet Trough (Geldsetzer, 1973; Jackson and Davidson, 1975; Jackson *et al.*, 1975; Jackson *et al.*, 1978). This trough plunges northwestward, and probably connects in the subsurface with equivalent rocks of an originally very large Proterozoic basin farther west. The Nanisivik Lead-Zinc Mine occurs in a carbonate formation in the Milne Inlet Trough (Olson, 1977).

An isolated outcrop area of Proterozoic rocks on northwestern Greenland (Fig. 2) is called the Thule Basin (Davies *et al.*, 1963; Kerr, 1967b; Dawes, 1976). Tectonism occurred there during Proterozoic time, but much of the isolation of the Thule Basin apparently resulted from Cretaceous-Tertiary faulting.

Thick Proterozoic rocks of Sequence 2 occur in northern Somerset and Prince of Wales Islands (Fig. 2). They form a north-plunging crescent-shaped body wrapped around the core of the north-plunging Boothia Horst (Kerr, 1977a). At least two pre-Paleozoic tectonic events affected these rocks. Each appears to have involved uplift, erosion, and diabase intrusion. The partial isolation of these rocks from equivalent rocks to the east and west was mainly achieved before Paleozoic deposition.

On Victoria Island (Fig. 2) a thick inlier of Sequence 2 occurs in the

Minto Uplift (Thorsteinsson and Tozer, 1962; Young and Jefferson, 1975). A broad belt of the large unnamed western basin making up Sequence 2 occurs on the Canadian mainland southwest of Victoria Island (Fig. 2), trending eastward, and dipping partly northwest (Baragar and Donaldson, 1973).

The rocks of Sequence 2 in the study area are everywhere very thick and contain unconformities. They also were intruded at one or more times by diabase dykes (Fahrig and Jones, 1969; Fahrig *et al.*, 1971). Their present distribution suggests that they were deposited in a single basin that trended generally northeastward. The present partial isolation of parts of this sequence resulted from a combination of syndepositional, pre-Paleozoic, and younger tectonism. The several remnants exposed in the west may largely connect in the subsurface beneath the lower Paleozoic cover as an unnamed western basin (Fig. 2).

D. Sequence 3 : late Proterozoic to Late Devonian

1. Foreward

The late Proterozoic to Late Devonian history of the Innuitian Mobile Belt is mainly the history of the Franklinian Geosyncline, the Arctic Platform, and the magmatic stage of the Pearya Geanticline (Fig. 4). The geosyncline was an orthogeosyncline in the sense of Kay (1951). The geosyncline was linear and lay between the Precambrian Shield and a geanticline (Figs. 2 and 3). Its history included five depositional phases (Fig. 5), that correspond to the three phases described from the miogeocline of Ellesmere Island (Kerr, 1967a, 1968a, 1976a). The youngest of the original three is now itself divided into three phases, so that five are now recognized.

The stratigraphic section comprising Sequence 3 ranges in age from late Proterozoic to late Late Devonian. It overlies older Proterozoic rocks unconformably, and is truncated at the top by a major unconformity related to the Ellesmerian Orogeny (Fig. 4). The sequence is not interrupted internally by major orogenic

episodes, although some minor angular unconformities are present within it.

Sequence 3 was the most extensive sequence of the Innuitian Mobile Belt (Fig.3). It formed an extensive, continuous depositional basin that at times covered the entire region (Fig. 1). The geosyncline in the Canadian Arctic first began to take shape in late Proterozoic or Early Cambrian time. It began as a narrow basin and became progressively wider and deeper for more than 200 million years. It reached its maximum width and depth in Late Ordovician to Early Silurian time. From then until its termination in Late Devonian time, the geosyncline became progressively narrower and shallower. This resulted because uplifts adjacent to and within the geosyncline became active and filled it with sediments. Deposition ended in the geosyncline when the Late Devonian to early Mississippian Ellesmerian Orogeny deformed and uplifted it.

¶ The Franklinian Geosyncline was a wide basin with a thick sedimentary column. It was bounded on the southeast by the Arctic Platform, a stable area where a thin succession of rocks was deposited from time to time on the Precambrian Shield. Northwest of the geosyncline was the Pearya Geanticline, a linear positive feature with a crystalline Precambrian core (Fig.2). At times during Sequence 3 the geanticline was stable and provided a narrow shelf on the northwest side of the geosyncline. At other times it was an active sedimentary source providing sediments to the geosyncline.

For some years the entire lower Paleozoic mobile belt was called the Franklinian Geosyncline, and within it two divisions were recognized (Thorsteinsson and Tozer, 1960; 1970; Kerr, 1967a; Trettin, 1969a). On the southeast was a belt called the miogeosyncline, containing mainly carbonates. It was separated from the Arctic Platform by a hinge-line across which there is marked change in thickness but little change in lithology (Fig. 3). Those parts of the mobile belt north and west of the miogeosyncline were called the eugeosyncline, following the terminology of Kay (1951). Trettin *et al.* (1972) suggested changes in this terminology, and these are employed in this report (Figs. 2 and 3). The name miogeocline replaces miogeosyncline, but the feature is unchanged. Four belts exist within the area of the former eugeosyncline as follows. The Pearya Geanticline is the partly exposed metamorphic belt on the extreme northwest; since it was mainly an area of uplift it is no longer considered part of the geosyncline. The remaining part of the former eugeosyncline is called the clastic belt of the geosyncline. It includes three divisions; from northwest to southeast, these are a coastal plain, a shelf, and the Hazen Trough.

2. Metamorphism

Metamorphism of the Innuitian Mobile Belt was related to the magmatic stage of the Pearya Geanticline (Fig. 4). It was restricted to early Paleozoic and earlier time and therefore affected the rocks of Sequence 3. Metamorphism affected only the Pearya Geanticline and nearby parts of the clastic belt, on northernmost Ellesmere and Axel Heiberg Islands (Figs. 2 and 3). Exposed gneisses of northern Ellesmere Island exhibit metamorphism that falls into two and possibly three episodes (Sinha and Frisch, 1975, 1976).

Major regional metamorphism about 1000 m.y. ago, and about 800 m.y. ago may or may not be the same event. Both are Proterozoic, and one or both may be the metamorphic event reported by Trettin *et al.* (1972) as Cambrian or earlier. There was a subsequent retrograde metamorphism of the Cape Columbia complex (Fig. 1) between 500 and 600 m.y. ago (Sinha and Frisch, 1976) that may relate to Middle Ordovician or earlier cataclastic deformation and metamorphism (Frisch 1974; Trettin *et al.*, 1972). Metamorphism associated with Late Silurian-Early Devonian deformation is based on K-Ar ages of 403±17 m.y. and 389±21 m.y. in the Cape Columbia region (Frisch, 1974). According to Trettin *et al.* (1972) these dates may represent

heating associated with an intrusive episode rather than regional metamorphism. Regional metamorphism associated with the Ellesmerian Orogeny was related to intrusion, and was tentatively interpreted as Middle Devonian (Trettin *et al.*, 1972).

According to Trettin *et al.* (1972), the metamorphic grade on northern Ellesmere Island generally increases to the northwest, away from the axis of the Franklinian Geosyncline and towards the core of the Pearya Geanticline (Fig. 2). Metamorphic processes may have been more or less continuous in the core of the Pearya Geanticline, and the metamorphic phases known on land may represent intervals during which the geanticline expanded into northern Ellesmere Island.

3. Igneous Intrusion

Devonian and older intrusions of the Innuitian Mobile Belt are restricted to northern Axel Heiberg and Ellesmere Islands, and were described by Trettin (1971b) and Trettin *et al.* (1972). They are restricted to the Pearya Geanticline (Figs. 3 and 6), and were emplaced during its magmatic stage (Fig. 4). A mafic-ultramafic body near M'Clintock Inlet (Fig. 1) was dated as 376±16 m.y., suggesting a late Early Devonian event. Lower Paleozoic mafic dikes and sills in northwestern Ellesmere Island are post-Silurian, pre-Carboniferous, and probably were emplaced during an episode of crustal extension following the Ellesmerian Orogeny. Mafic sills and/or dikes on extreme northwestern Ellesmere Island may be related to Lower Devonian basic volcanism of northern Axel Heiberg Island.

Granitic plutons in northern Axel Heiberg and northern Ellesmere Islands (Trettin *et al.*, 1972) apparently are related to early and mid- Paleozoic orogenic activity. Some have been dated at 390±20 m.y., or Early Devonian (Frisch, 1974). A number of plutons were considered by Trettin *et al.* (1972) to be a genetic group of three assemblages. Others yield K-Ar age determinations of 345±15 m.y. and 325±14 m.y., and lie near the Devonian-Mississippian boundary (Trettin, 1971b). These are minimum values, and so suggest intrusion in Devonian time, presumably during the folding and regional metamorphism of the Ellesmerian Orogeny (Fig. 4). Granitic intrusions on northern Axel Heiberg Island are post-kinematic (Trettin, 1969a, 1971b) and are younger than Lower or Middle Silurian strata which they intrude (Trettin *et al.*, 1972). Radiogenic ages vary from 325±25 m.y. to 360± 25 m.y., and probably indicate contemporaneity with the Ellesmerian Orogeny.

4. Volcanism

Lower Paleozoic volcanic rocks occur in a narrow belt along the present Arctic continental margin on Ellesmere and Axel Heiberg Islands, and there are three main ages (Trettin *et al.*, 1972; Trettin, 1976). Middle Ordovician volcanics are most widespread, and occur as part of a volcanic island arc. Silurian volcanics are partly pyroclastic and brecciated, and they appear to be related to deformations that produced the Rens Fiord Uplift (Fig. 6). Lower Devonian (and/or younger) volcanism occurred on northern Axel Heiberg Island, and probably was related to crustal extension that affected the Rens Fiord Uplift and produced the Svartevaeg Trough to the east of it.

Volcanic activity originated at different times and localities, and probably is mainly hidden beneath the continental shelf. Most volcanics are intermediate to somewhat silicic, whereas basic rocks are subordinate, and rhyolite is scarce or absent.

E. Sequence 3: Simple Evolution

Sequence 3 of the Innuitian Mobile Belt (Fig. 3) was dominated by the Pearya Geanticline, a major tectonic high that trends northeastward along the Arctic continental margin (Fig. 6). The sequence was also influenced by several smaller internal basement uplifts.

A cross-section through Ellesmere Island (Fig. 7) shows the general or simple evolution of the mobile belt during deposition of Sequence 3. This is beyond the influence of any cross-trending internal basement uplifts that obstructed the history of the sequence (Fig. 6) and therefore, is the simplest history of Sequence 3 that can be depicted. During the span of Sequence 3 the Franklinian Geosyncline formed, evolved, and terminated, with five phases being recognized in its history (Fig. 5).

1. Phase I: Late Proterozoic (Hadrynian ?) to Mid-Late Cambrian

Phase I of the Franklinian Geosyncline (Figs. 5 and 7) shows a progressive pattern of sedimentary development (Kerr, 1967a). A thick column of Upper Proterozoic (Hadrynian?) mainly clastic sediments is present in the axis of the Franklinian Geosyncline of Ellesmere Island. These rocks are confined to the axis, suggesting that the geosyncline began as a narrow downwarp in Proterozoic time. Thickness patterns suggest that the Franklinian Geosyncline had taken on a linear form by late Proterozoic time, but was still very narrow. The geosyncline, however, had its characteristic broad but linear form by early Early Cambrian time (Fig. 7), when great subsidence occurred and thick sediments accumulated. Thinner equivalent sedimentation occurred on the Arctic Platform to the southeast. The Pearya Geanticline may have been exposed in parts of Cambrian time, when it apparently was a sedimentary source, but was followed by encroachment of the Grantland Formation northwest into metamorphic basement (Trettin, 1976).

The Franklinian Geosyncline is interpreted to have formed initially as a downwarp on a continental crust

(Fig. 3). A hinge-line or flexure separated the geosyncline from the Arctic Platform, and is located where the overall sedimentary thickness increases most abruptly. The hinge-line was abrupt and rather narrowly defined during Phase I. It had a different location and nature during other phases of Sequence 3.

During Phase I the geosyncline had a general developmental pattern (Fig. 5) It may have begun initially in late Proterozoic time as a narrow downwarp of subaerially exposed continental crust. Then during the rest of Phase I it had a history of progressive widening by outward encroachment. By late Early Cambrian time, rocks encroached outward onto the Arctic Platform at Bache Peninsula (Kerr, 1967a), and southeastern Devon Island (Kurtz *et al.*, 1952). Clastic sediments were derived from both margins, the Shield in the southeast and the exposed Pearya Geanticline in the northwest. Through time marine influence increased, as shown by the increasing volume of carbonate rock. By the end of Phase I in Late Cambrian time, carbonate sedimentation dominated in the miogeocline and Arctic Platform, and may have been occurring throughout much or all of the southern Canadian Arctic. During Phase I in general, thick rocks accumulated in the geosyncline and lesser thicknesses on the Pearya Geanticline and Arctic Platform.

In Early Cambrian time central parts of the geosyncline were unstable tectonically as subsidence began and there was a great influx of clastic rocks from the southeast and the northwest. By Late Cambrian time there was broader subsidence, with encroachment out of the geosyncline onto the former source areas (Fig. 7). Cambrian carbonates were deposited throughout the miogeocline and were spread as a sheet far to the south on the Arctic Platform including Boothia Peninsula (Miall and Kerr, 1977, and in press), northwestern Baffin Island (Trettin, 1975), and Victoria Island (Thorsteinsson and Tozer, 1962). Fine grained clastic rocks that appear to be partly Upper Cambrian encroached unconformably onto metamorphic basement of the Pearya Geanticline (Trettin, 1976).

It was earlier thought that there was marine withdrawal and a disconformity at the end of Phase I (Kerr, 1967a). This unconformity is not present (Barnes, pers. com., 1975; Stuart Smith and Wenekers, 1977), so Phase I merged into Phase II during continuous sedimentation.

Phase I had a pattern of deepening and widening of the geosyncline, and it ended with a widespread marine incursion in which the hinge-lines became less distinct. The progressive pattern of encroachment during Phase I is broadly similar to the pattern occurring in the Appalachian and Cordilleran miogeoclinal belts of North America.

2. Phase II: Mid-Late Cambrian to Mid.-Late Ordovician

Phase II (Kerr, 1968a) was a long period of very widespread, continuous, and steady marine sedimentary accumulation in the Canadian Arctic (Figs. 5 and 7). A carbonate body that spans this phase occurs in the Franklinian Miogeocline and adjacent Arctic Platform. It extends from Victoria Island (Thorsteinsson and Tozer, 1962) to Somerset Island (Dixon, 1973; Miall and Kerr, 1977, and in press), northern Baffin Island (Trettin, 1975), eastern Ellesmere Island (Kerr, 1968a), and to northwest Greenland (Dawes, 1976). Sedimentation appears to have been continuous from Phase I to Phase II in the carbonate and clastic belt of the Franklinian Geosyncline. It may have been continuous also on the Pearya Geanticline, where Trettin (1976) reported that fine grained clastic rocks of the Cambro-Ordovician Grant Land Formation.

In Phase II the Franklinian Geosyncline reached its broadest extent, and marine deposition was most widespread. The hinge-line was more poorly defined than at any other time (Fig. 7), although it had about the same location as earlier. The entire Canadian Arctic received marine sedimentation, at times, but subsidence and water depth were greatest in the geosyncline. The geosyncline was rather symmetrical during Phase II. The clastic belt of the geosyncline contained a starved axial basin, the Hazen Trough, with deep water and euxinic conditions, where black graptolitic shale and cherty limestone were deposited (Trettin *et al.*, 1972; Trettin 1973a, 1976, and in press). In parts of Ordovician time, sedimentation was about 2.5 to 4 times as slow in the Hazen Trough as it was in miogeoclinal parts of the geosyncline (Trettin, 1973a). Shallower water existed on both sides of the trough. On the southeast was a miogeocline, where a thick wedge of sediments was deposited (Kerr, 1968a). It is mainly carbonate, with three intervals of gypsum-anhydrite, one of which may be a sabhka deposit (Mossop, 1973, and in press). It contains only minor clastic rock.

The northwestern part of the clastic belt and the adjacent Pearya Geanticline contain a great variety of rock types and rapid facies changes (Trettin *et al.*, 1972; Trettin, 1973a, b), including clastic rocks, volcanic rocks, and carbonates. This is due to the intermittent activity of the Pearya Geanticline. At times during Phase II the geanticline and nearby parts of the clastic belt were stable and formed a shelf where carbonate rocks were deposited. At other times the geanticline was an active volcanic arc depositing thick volcanic rocks in the nearby clastic belt (Fig. 7).

3. <u>Phase III: Late Ordovician to early Early Devonian</u>

Phase III (Figs. 5 and 7) is characterized by a large variety of rock types that show great variations in thickness and lithology. This phase was marked by a southeastward expansion of the Hazen Trough at the expense of the miogeocline. The carbonate belt was narrow, with thick rocks being deposited (Kerr, 1976a; Morrow and Kerr, 1978). It appears that the flexure between the Arctic Platform and the geosyncline on Ellesmere Island lay within the carbonate belt and was rather broad, with a gradual thickness change across it.

The Hazen Trough was the region of deepest subsidence during Phase III. In early parts of the phase, when water was very deep and sediment supply was low, graptolitic and shaly rocks were deposited there, the Cape Phillips Formation (Kerr, 1976a), and the partly equivalent upper member (cherty shale) of the Hazen Formation farther northwestward (Trettin, 1971a, 1973a, 1976). When sediment supply increased because of uplift of the Pearya Geanticline, those rocks were succeeded by the Imina Formation, a flysch deposit that was dumped rapidly into the pre-existing deep trough, and overlapped southeastward (Fig. 7).

In most places the shales of the Hazen Trough overlapped the carbonates of the miogeocline, indicating general and gradual expansion of the trough. In rare places the carbonate to shale facies change is very abrupt, and occupied the same geographic location from Late Ordovician to Early Devonian time (Thorsteinsson, 1958; Thorsteinsson and Tozer, 1957; Thorsteinsson and Kerr, 1968; Kerr 1976a).

A narrow reef occurred within the Hazen Trough, partly or completely isolated from the main carbonate shelf and enclosed by the Cape Phillips Shale. This reef extended at least intermittently and perhaps continuously from northern Melville Island (Tozer and Thorsteinsson, 1964) to Ellesmere Island (Kerr, 1976a) and to northern Greenland (Mayr, 1976a), a distance of about 1700 km. The Imina Formation (Trettin, 1971a, 1973a, 1976) is a flysch-like deposit of turbidites that is as much as 2700 m thick (Fig. 7). It overlaps gradationally southeastward from the coastal plain and shelf to the Hazen Trough, so that its base is youngest in the miogeocline, where it grades in a complex way into equivalent formations. The flysch deposit largely filled up the trough that earlier had been a starved basin (Fig. 7). This flysch deposit was the precursor of the Ellesmerian Orogeny (Fig. 4)

The Hazen Trough, which is the axial zone of deep water facies, had its maximum southeastward extent at the expense of the miogeocline during this phase, when shale deposition was widespread in the miogeocline. The carbonate belt of the geosyncline was very narrow.

Through most of Phase III the Pearya Geanticline was not strongly positive and a section at M'Clintock Inlet (Trettin, 1969b) shows the sequence. During an early part of this phase the geanticline was a carbonate shelf, in middle parts it received flysch deposits, and in latter parts it again received carbonates. Lateral shifts of the northwestern margin of the Hazen Trough resulted from expansion and contraction of the Pearya Geanticline. The Imina flysch was deposited on the geanticline when it contracted and carbonates were deposited there when it expanded. The Hazen Trough was deep during Phase III, with flysch deposits mainly in the northwest, and black shale farther southeast (Fig. 7). Subsidence probably exceeded sedimentation during an early part of Phase III, but later this trend reversed as uplift of the margins of the geosyncline began.

The closing part of Phase III was a narrowing of the geosyncline, an infilling, and a general shallowing that was influenced by the forthcoming Ellesmerian Orogeny (Fig. 7). The geosyncline was narrowed from the northwest by the gradually rising Pearya Geanticline, evidence of the raising being seen in the shallower facies in the late part of this phase in the clastic belt (Trettin *et al.*, 1972; Trettin, 1976). Narrowing of the geosyncline from the southeast was due to an expansion and uplift of the Arctic Platform (Kerr, 1968b, 1977a). The end of Phase III is delineated by the unconformity produced by that uplifting event as the hinge-line moved northwestward.

4. <u>Phase IV: early Early Devonian to Middle Devonian</u>

During Phase IV there was a very different pattern of facies in the rocks of Sequence 3. This resulted from the developing Ellesmerian Orogeny, which gradually destroyed the geosyncline and terminated deposition of the sequence (Kerr, 1976a).

Deposition gradually became very restricted during this time of progressively more widespread emergence (Figs. 5 and 7), as the Franklinian Geosyncline was gradually filled in and the water-covered region was progressively reduced in size. This resulted largely from the Ellesmerian Orogeny, and associated increasing tectonic activity of the Pearya Geanticline (Trettin *et al.*, 1972; Trettin, 1978). Phase IV embraces an early part of the orogeny. At the same time that the Pearya Geanticline served as a source of sediments from the northwest, the Arctic Platform and Shield were also raised (Kerr, 1968b, 1977a), and became a source of sediments from the southeast.

A typical cross-section of Phase IV from the Arctic Platform to the Hazen Trough (Fig. 7) involves rocks located far enough from linear cross-trending basement uplifts so as not to be affected by them.

Phase IV began when rapid uplift occurred at approximately the same time in the Pearya Geanticline and the Arctic Platform (Fig. 7), both of which contributed to narrowing the Franklinian Geosyncline.

At the beginning of Phase IV, in Early Devonian time, broad gentle uplift of the Arctic Platform caused it to expand generally. The zone of flexure between platform and geosyncline shifted basinward and remained there through Phase IV. As a result of this shift southern and eastern parts of the miogeocline were converted into Arctic Platform. This general expansion has been documented most clearly on central Ellesmere Island (Fig. 7), but probably affected the entire southeastern margin of the geosyncline (Kerr, 1968b). The base of the sequence in the southeast is an angular unconformity produced by uplift of the Arctic Platform, and that unconformity disappears northwestward within the geosyncline. The unconformity is succeeded on the platform and nearby parts of the miogeocline by synorogenic to post orogenic formations. These were deposited while uplift of the Arctic Platform occurred, and encroached onto the platform when uplift ceased (Fig. 7). These clastic rocks were succeeded by a narrow belt of thick carbonates in the miogeocline and equivalent thinner carbonates on the platform (Kerr, 1976a). The true starved basin of earlier times (Trettin *et al.*, 1972) had largely disappeared by uplift and infilling, and the geosyncline in this phase contains no true shales.

Phase IV included marine and nonmarine rocks. It ended in the upper Middle Devonian with the beginning of deposition of a clastic wedge of largely nonmarine rocks including the Okse Bay and equivalent Formations (Kerr, 1974, 1976a, Embry and Klovan, 1976).

In Phase IV there was a marked increase in the uplift of the Pearya Geanticline, as it also broadened and became a much more active sedimentary source than in earlier phases. Throughout Phase IV the Ellesmerian Orogeny progressed and accelerated. This orogeny originated in and emanated from the Pearya Geanticline. In this phase the geanticline

continued to be a source of sediments to the geosyncline adjacent to it. In earlier parts of this phase a narrow Hazen Trough existed and received flysch deposits (Fig. 7). In later parts of Phase IV, as shallowing continued, those rocks were succeeded by molasse type rocks. Northwestern parts of the clastic belt which earlier had been filled by flysch deposits may at this time have been raised and eroded, to be redeposited in the narrower basin to the southeast.

Plutonic intrusives were emplaced in the Pearya Geanticline and nearby clastic belt in Early Devonian time (Trettin et al., 1972, Frisch, 1974), and apparently produced the uplifting. There seems to have been an acceleration of uplift of the geanticline during this phase. It culminated with broadening which included uplift and erosion of nearby parts of the clastic belt of the geosyncline (Fig. 7).

A fragmentation of the geanticline and clastic belt apparently occurred during Phase IV, as a result of the intrusion and uplift. At least one deeply subsiding basin developed, the Svartavaeg Trough (Trettin, 1969a), in which very thick Lower to Middle Devonian clastic and volcanic rocks were deposited.

The Grantland Uplift, which occurs within the clastic belt and trends along it, had more than 3 km of Ordovician and Silurian strata removed from it before being overlain by the Sverdrup Basin succession (Trettin *et al.*, 1972). The uplifting of that structure probably was contemporaneous with and a part of the Ellesmerian Orogeny, and may have been initiated in Early Devonian time (Fig.7). Because of its northeastward trend, the Grantland Uplift did not interfere with the general deformation of the geosyncline that was produced by Ellesmerian Orogeny, but instead appears to have been a smoothly integrated part of that deformation. There may have been deposition in basins associated with the Grantland Uplift (Fig. 7), and later eroded.

5. Phase V: late Middle and Late Devonian, and Possibly Early Mississippian

Phase V, of Sequence 3, includes the last events in the history of the Franklinian Geosyncline (Fig.7). It was intimately controlled by and coeval with the culmination of the Ellesmerian Orogeny.

The marine and partly nonmarine rocks of the earlier phase were succeeded by an upper Middle and Upper Devonian clastic wedge (Kerr 1974, 1976a; Embry and Klovan, 1976), mainly nonmarine clastic rocks. These rocks were deposited in a narrow basin occupying the southeastern side of the miogeocline and nearby parts of the Arctic Platform. They were deposited during a time of high relief, and engulfed and covered the earlier marine rocks.

Beyond the influence of cross-trending linear basement uplifts (Fig. 7), Phase V has a simple and progressive history. The Pearya Geanticline, which had begun its rapid rise during Phase IV, continued rising during Phase V, and broadened farther. Fine grained clastic rocks that were derived from it were spread progressively farther southward and eastward. Because of the general uplift they were largely nonmarine. They are now preserved widely in the region of the former carbonate belt. Deposition of these sediments may have been partly contemporaneous with deformation of that belt. The huge clastic wedge of Middle to Upper Devonian rocks that covered the Franklinian Geosyncline and adjacent parts of the Arctic Platform constitutes Phase V. This writer considers that the source was largely the Pearya Geanticline, and the exposed and eroded northwestern parts of the clastic belt of the geosyncline itself. Embry and Klovan (1976) suggest that the source was mainly uplifted regions of northeastern and eastern Greenland.

It appears that those parts of the Franklinian Geosyncline that were not near internal basement uplifts, had a very similar history during Phase V. This applied along the full length of the geosyncline, from Ellesmere Island (Fig. 7), to Cornwallis Island (Thorsteinsson and Kerr, 1968; Thorsteinsson, 1973), to Bathurst Island (Kerr, 1974), and on to Banks Island (Miall, 1976a, b).

F. Pearya Geanticline

The Pearya Geanticline (Figs. 2, 6, 7) is a linear tectonic high that trends northeastward along the Canadian Arctic Continental Margin. It is exposed today only on northwest Ellesmere Island and northern Axel Heiberg Island. The term Pearya Geanticline was introduced originally by Schuchert (1923), for an exposed borderland of crystalline rocks that he presumed to be of Precambrian age. The term fell into disuse when Thorsteinsson and Tozer (1960) applied the term eugeosyncline to a belt that included the crystalline rocks as well as the nearby lower Paleozoic clastics and volcanics. At that time the crystalline rocks were interpreted as lower Paleozoic intrusions. The term Pearya Geanticline was reintroduced by Trettin *et al.*, (1972), who recognized that the tectonic belt that included northwestern Ellesmere Island and Axel Heiberg Island had been mainly positive, although it had been the site of sedimentation at times. It is now known that there is an exposed core of gneissic Precambrian rocks on northern Ellesmere Island that were metamorphosed about 800 to 1000 m.y. ago, and therefore are of about Grenvillian age (Sinha and Frisch, 1975, 1976). This supports the use of the term geanticline.

The term Pearya Geanticline now applies to a broad mainly positive belt that includes a Precambrian crystalline core, the mainly clastic lower Paleozoic rocks that were deposited upon it, and the lower Paleozoic plutonic rocks that were intruded into it. The Precambrian age of the core suggests that the geanticline and the Shield are connected beneath the Franklinian Geosyncline (Fig. 3).

The Pearya Geanticline and the Franklinian Geosyncline are difficult to delineate precisely. Their histories were so long and complex that their mutual boundary varied in location with time. The geanticline (Fig. 6) was mainly a positive feature and source area, but at times it received thin deposition and at those times was part of the clastic belt of the geosyncline (Fig.2). Northwestern parts of the geosyncline also behaved from time to time as a source area. Thus the boundary between the geanticline and geosyncline is a broad and transitional zone between a source area and a deep basin (Fig.7).

The Pearya Geanticline had a magmatic stage that spanned Sequence 3 of the Innuitian Mobile Belt (Fig. 4). Its main period of tectonic activity was the Ellesmerian Orogeny, which spanned Phases IV and V of Sequence 3 (Figs. 5 and 7). The formation of the geanticline may have been largely the result of plutonic intrusives. The geanticline was the main source of the clastic sediments that filled the Franklinian Geosyncline and spread onto the adjacent Arctic Platform. During periods of gentle expansion or uplift of the Pearya Geanticline shallow water rocks of the clastic belt of the Franklinian Geosyncline were deposited on it. In the periods of more extreme uplift the geanticline was eroded, and probably metamorphosed or intruded. In periods of contraction or subsidence increasingly deeper water sediments encroached onto the geanticline.

The Pearya Geanticline was a moderate source of clastic and volcanic sediments during Phases I and II of the Franklinian Geosyncline (Sequence 3), and had carbonate, clastic and volcanic rocks deposited on it from time to time (Figs. 5 and 7). In Phase II it was positive and volcanic while the adjacent Hazen Trough was deeply negative. A metamorphic event in the Pearya Geanticline is recorded by radiogenic dates between 465 ± 19 m.y. and 389 ± 21 m.y., falling in an Early Ordovician to Early Devonian range (Frisch, 1974). Stratigraphic evidence suggests that this metamorphism was before the Middle Ordovician, because unmetamorphosed Middle Ordovician rocks are unconformable on the metamorphosed Cape Columbia complex (Trettin *et al.*, 1972). This metamorphic event therefore preceded the Ellesmerian Orogeny.

In Phase III the Pearya Geanticline became a substantial source of flysch sediments that began to rapidly fill the Hazen Trough. Periodic uplift was prevalent during this phase, but there were also times when it received sediments. The youngest widespread sedimentary unit known in the Pearya Geanticline in the region of the typical cross-section (Fig. 7) is the Silurian Marvin Formation. Deposition of this formation occurred in the upper part of Phase III, and preceded the Ellesmerian Orogeny.

In Early Devonian time, at the beginning of Phase IV, the Pearya Geanticline began to be uplifted more rapidly, and was intruded by plutonic rocks (Fig. 7). This began the Ellesmerian Orogeny, which originated in and emanated from the Pearya Geanticline.

The oldest plutonic rocks that were intruded into the Pearya Geanticline or adjacent parts of the clastic belt of the geosyncline and that are attributed to the Ellesmerian Orogeny have yielded dates of 390 ± 20 m.y., and 390 ± 18 m.y. They are of Early Devonian age and were emplaced during Phase IV. They occur in two places, and were associated with forceful emplacement (Frisch, 1974). The youngest plutonic rocks attributed to the Ellesmerian Orogeny may have been emplaced in the Pearya Geanticline in an early part of Phase V, in about late Middle Devonian time (Fig. 7). This includes a small high level granitic pluton that yielded an age of 360 ± 25 m.y. (late Middle Devonian), with confidence limits ranging from late Early Devonian to Early Mississippian. It was considered to have cooled quickly, and the age is approximately the age of emplacement. Plutons with younger radiogenic ages are known, but they were considered to reflect slow cooling or post-intrusive events (Trettin *et al.*, 1972.) From Early through Late Devonian time, spanning Phases IV and V (Fig. 7), the Pearya Geanticline was affected by the Ellesmerian Orogeny. It was uplifted, and most of it probably was exposed and subject to erosion. The geanticline provided an influx of clastic sediments to the geosyncline. Fragmentation of the geanticline and clastic belt also occurred during the orogeny. This fragmentation produced the large northwest trending Svartevaeg Trough in northern Axel Heiberg Island, where thick Lower Devonian clastic and volcanic rocks were deposited (Trettin, 1969a). Folding and uplift of the geanticline probably continued through Middle Devonian time, but had been largely completed by Late Devonian time, probably by Frasnian time. A unit in the Yelverton area of northern Ellesmere Island that is only slightly deformed (Mayr, 1976b) has yielded fossils of Middle or Late Devonian age (late Givetian or Frasnian; D.C. McGregor, pers. com., 1978).

The metamorphic grade of the exposed gneissic core of the Pearya Geanticline indicates that between 8 km and 30 km of uplift and erosion occurred there prior to deposition of upper Paleozoic strata of the Sverdrup Basin (Trettin *et al.*, 1972). There probably was more than one phase of uplift of the geanticline. An important uplift occurred in Early Devonian time in an early part of the Ellesmerian Orogeny, coinciding with Phase IV of the geosyncline, when plutonic intrusion and fragmentation occurred. Another important uplift probably occurred in late Middle and Late Devonian time, coinciding with the later part of the Ellesmerian Orogeny, during Phase V of the geosyncline when widespread clastic sedimentation occurred (Fig. 7).

G. Ellesmerian Orogeny, Unobstructed Pattern

The Ellesmerian Orogeny (Fig. 4) was a tectonic event that intruded and deformed the rocks of Sequence 3 of the Innuitian Mobile Belt. It terminated and destroyed the Franklinian Geosyncline, and brought the deposition of Sequence 3 to a close. The Ellesmerian Orogeny overlapped in time with Phases IV and V of Sequence 3. It originated and spread from the Pearya Geanticline. Its ultimate cause was the plutonic intrusion and uplift of the geanticline. The Ellesmerian Orogeny produced relatively simple effects where it was not obstructed by cross-trending basement uplifts (Fig. 7).

There were precursors of the Ellesmerian Orogeny, which were the development of a foredeep in the geosyncline during Phase II (Fig. 7), and the subsequent partial filling of the foredeep by the flysch deposits during Phase III. The Ellesmerian Orogeny *sensu stricto*, and as here considered, began in Early Devonian time, when substantial uplift, intrusion, and fragmentation began in the geanticline.

In an early stage of the Ellesmerian Orogeny in Early Devonian time, during Phase IV of Sequence 3, the Pearya Geanticline was raised substantially and began to be fragmented and intruded by igneous rocks. By late Middle to Late Devonian time the geanticline had been raised much higher, and parts of the clastic belt of the geosyncline were raised with it. The marine basin that formerly occupied the carbonate belt or miogeocline was eliminated by this broad uplift. The slope produced was sufficient so that parts of the geosynclinal column were transported away from the geanticline. Folds and thrust faults developed within the moving mass, presumably driven by gravitational forces. This resulted in southeasterly overriding in structures within the Franklinian Geosyncline on Ellesmere Island (Thorsteinsson and Kerr, 1972a,b; Kerr, 1967b, 1973b,c), and southerly overriding on Bathurst Island (Kerr, 1974). These involved decollement within the Proterozoic rocks of Phase I, as well as within evaporite units of Phase II.

The Ellesmerian Orogeny involved great uplift and fragmentation in the Pearya Geanticline during Sequence 3 (Fig. 6). At this time the Rens Fiord Uplift also was raised, and Lower Devonian rocks were deposited east of it in the Svartevaeg Trough of northern Axel Heiberg Island (Trettin, 1969a). The great folding event of the Ellesmerian Orogeny occurred somewhat later. Folding was largely completed in the Pearya Geanticline and nearby parts of the clastic belt of Ellesmere Island by Late Devonian, probably by Frasnian time. Lesser deformation continued on northern Ellesmere Island, but ended before deposition of the mid-Mississippian (Viséan) Emma Fiord Formation (Thorsteinsson and Tozer, 1970; Trettin, 1973b).

The Ellesmerian Orogeny was the final event or culmination that ended the Franklinian Geosyncline. It produced overriding that affected the entire geosyncline, but died out rapidly southeastward within the Arctic Platform (Fig. 7).

It appears that deformation by the Ellesmerian Orogeny did not begin in the miogeocline until later than in the geanticline and the clastic belt. There may have been some folding contemporaneous with the Middle-Upper Devonian clastic wedge of the miogeocline. The intense folding of the miogeocline however involved the clastic wedge in the deformation (Kerr, 1974, 1976a; Embry and Klovan, 1976), and rocks in that wedge are as young as latest Devonian (Famennian). Deformation of the miogeocline had been completed prior to deposition of the mid-Mississippian Emma Fiord Formation.

The Ellesmerian Orogeny deformed the entire Franklinian Geosyncline and Pearya Geanticline, and its general effects are shown in a simple cross-section (Fig. 7). The Ellesmere-Greenland Fold Belt was formed at this time, and extended from eastern Grinnell Peninsula through Ellesmere Island to northern Greenland. Folding was most intense in the northwest in and near the geanticline, and diminished in intensity to the southeast. The age and extent of the Ellesmerian Orogeny are depicted symbolically in a tectonic time diagram (Fig. 4).

At the end of the Ellesmerian Orogeny in latest Devonian to early Mississippian time the entire Canadian Arctic Archipelago region which now includes islands and channels (Fig. 6) may have been exposed and undergoing erosion. The known extent of the geosyncline and geanticline are shown (Fig. 2). The region now covered by the Sverdrup Basin probably also was part of those deformed features (Trettin *et al.*, 1972, Meneley *et al.*, 1975; Balkwill, 1978). The Franklinian Geosyncline continued at least as far southwest as Banks Island (Miall, 1976b). Those parts of the Pearya Geanticline that now lie under water northwest of the Arctic Archipelago presumably were exposed after the Ellesmerian Orogeny as well. It is probable also that the region that now forms the Canada Basin of the Arctic Ocean was a land area at that time. If so then the Franklinian Geosyncline and Pearya Geanticline made up a linear mobile belt lying within a large continental area. The Sverdrup Basin and presumably the Canada Basin subsequently developed on this eroded landscape, on a widespread angular unconformity.

H. Basement Uplifts Affecting Sequence 3

1. Foreword

Sequence 3 had a rather simple and straightforward evolutionary development in parts not affected by cross-trending internal basement uplifts (Fig. 7). The complete development of Sequence 3, however, was much more complicated than that, because of basement uplifts that extended across the basin (Fig. 6), and interfered with its development. Internal basement uplifts that strike across the northeastward depositional trend of the Pearya Geanticline and Franklinian Geosyncline caused Sequence 3 to have a much more complicated history.

Basement uplifts complicated the history of Sequence 3 in two ways. (1) They had a history of periodic uplifts that locally interrupted deposition of Sequence 3 both in the geosyncline and platform, and caused unconformities as well as affecting facies and thickness patterns. (2) The uplifts were large basement cored bodies or plugs that had been forced upward or intruded into Sequence 3. The basement cores were more rigid than the surrounding sedimentary rocks and created a local anisotropy. When the rocks of Sequence 3 were subsequently deformed by the Ellesmerian Orogeny, the structures that formed within that sequence in the region of the basement uplifts were drastically different from the structures elsewhere, because of that anisotropy. Thus the structure of the uplifts interfered with and partly controlled the deformation of nearby parts of the geosyncline during the Ellesmerian Orogeny. The nature of the basement uplifts will be described below (Figs. 6 and 8), and subsequently their effects on the evolution of Sequence 3 will be discussed (Figs. 9 and 10).

There are four major basement uplifts in the Canadian Arctic Islands in addition to the Pearya Geanticline (Fig. 6). Three of these affected the Franklinian Geosyncline. A fourth, the Minto Arch, affected the Arctic Platform, and may also have affected the geosyncline at its western end.

2. Boothia Uplift

The Boothia Uplift (Figs. 6, 8, and 9) is a north trending, north plunging, structure at least 650 km long that had a number of discrete pulses of vertical uplift (Kerr, 1977a; Miall and Gibling, 1978). A cross-section of the Boothia Uplift

(Fig. 8) shows that it formed by the vertical upward movement, by basement faulting, of a block of the Precambrian Shield called the Boothia Horst. The raising of this crystalline horst carried the overlying sediments with it, and deformed them to form the Cornwallis Fold Belt, a broad anticlinorium controlled by the basement faulting. The horst rose by means of numerous faults that trend northward and are concentrated near the margins of the uplift. In the deepest structural levels that can be observed, these faults are vertical. At higher elevations, or higher structural levels, the faults splay outward to become high angle reverse faults, or steep thrusts. At still higher levels they die out to become overturned folds, which in turn merge into asymmetric folds. Thus the amount of vertical displacement on each zone diminished and died out upward.

The Boothia Uplift is mushroom-shaped in cross-section (Figs. 8 and 9), because the faults and folds splay out at higher elevations. This presumably was due to rock expansion, that resulted from unloading by erosion and the consequent reduction of geostatic pressure. Each level within the raised uplift was brought to a higher elevation, and its confining pressure was reduced by the erosion that occurred above it, hence the uplift expanded.

The Boothia Uplift had one or more pulses of uplift in Late Proterozoic time, when parts of Sequence 2 were removed and remnants of that sequence were left on either side. It was mildly positive in Cambrian to Silurian time during Phases I to III of Sequence 3 (Fig. 9). The main uplift however, was the Cornwallis Disturbance in Silurian and Devonian time (Fig. 4), which included four pulses of uplift and erosion totalling at least 5300 m (Kerr, 1977a). In each pulse of the Cornwallis Disturbance the Boothia Uplift was raised and eroded, while deposition continued on either side.

The Boothia Uplift trends northward (Fig. 6), and this probably was largely controlled by the original northward trends of the gneisses in the Shield

that forms its crystalline core. It also plunges northward so that crystalline core is exposed in the south, and the folded sedimentary cover is exposed in the north (Fig. 8). The Boothia Uplift extends across the Arctic Platform and the exposed parts of the Franklinian Geosyncline. It probably extended completely across the geosyncline to connect with the Pearya Geanticline; however, this suggested northern extension is now covered by the Sverdrup Basin (Fig. 6).

The Cornwallis Disturbance occurred before the Ellesmerian Orogeny, and was a deformation of much lesser intensity. The situation that existed at the end of the Cornwallis Disturbance is shown in a simplified way (Fig. 8), and is called the basic structure of the Boothia Uplift (Kerr, 1977a). This structure formed mainly in Devonian time during Phase IV of Sequence 3 (Fig. 5), while the Ellesmerian Orogeny was affecting the clastic belt, but before it affected the carbonate belt. The basic structure of the Boothia Uplift (Fig. 8) existed before the geosyncline was deformed by the Ellesmerian Orogeny (Fig. 4), so that it was modified by the orogeny.

3. Bache Peninsula Arch

The smallest and least active basement uplift in the Canadian Arctic Islands is the Bache Peninsula Arch of Ellesmere Island (Fig. 6). It extends westward from the Arctic Platform across the miogeocline, and farther west is covered by the Sverdrup Basin.

This arch was first active in Ordovician and Silurian time (Kerr, 1968a), when it was a mildly positive swelling or uparching that subsided less than flanking basins. Its main activity was in early Early Devonian time, when it was raised and eroded to become the source of red clastic sediments. At this time the part of the arch lying within the geosyncline was converted to part of the Arctic Platform. From then on the arch behaved as part of the Arctic Platform with only thin sediments accumulating there.

Raising of the Bache Peninsula Arch did not produce any structural deformation that can now be observed at the surface. The extent of the arch has been determined by thickness and facies variations, and by the extent of the angular unconformity it produced. The mechanism of formation may have been similar to that of the Boothia Uplift (Fig. 8), but much weaker. It may have a mildly upfaulted basement core, with deformation dying out upward in the sedimentary column. Bache Peninsula Arch did not establish structural trends that were strong enough to influence later structural deformation now visible at the surface; however, such deformation may have occurred at depth.

4. Rens Fiord Uplift

The Rens Fiord Uplift (Fig. 6) is a fault bounded structure exposed on northern Axel Heiberg Island (Trettin, 1969a; Trettin *et al.*, 1972). Its core is composed of Cambrian(?) and Ordovician rocks. The earliest known phase of uplift occurred in Late Silurian to Early Devonian time near the onset of Phase IV of Sequence 3 (Fig. 5). The Rens Fiord Uplift was then the source of coarse red bed delta complex that prograded eastward into the Hazen Trough to form the Lower Devonian Stallworthy Formation, about 3000 m thick. Small plutonic intrusives in the core of the uplift are late Middle to Upper Devonian 360 ± 25 m.y.), and apparently were intimately related to a pulse of uplift. This uplifting was a part of the Ellesmerian Orogeny and the intrusions are post-kinematic.

The main activity of the Rens Fiord Uplift was one or more strongly upward pulses when more than 3 km of rocks were eroded from it on northern Axel Heiberg Island. These were between Early Devonian time and mid-Mississippian time, and were during the span of the Ellesmerian Orogeny.

The Rens Fiord Uplift probably connected northward with or was a part of the Pearya Geanticline (Fig. 6). The activity of the two occurred at about the same time, as the rise of the Rens Fiord Uplift was approximately contemporaneous with the Ellesmerian Orogeny. The rise of both also was approximately contemporaneous

with that of the Boothia Uplift. The Rens Fiord Uplift clearly was an important structure in Devonian time, extending south-southeastward from the Pearya Geanticline, and reaching most of the way across the Franklinian Geosyncline. It apparently established a south-southeastward structural trend on Axel Heiberg Island at that time, and this trend has dominated later deformations there.

The Rens Fiord Uplift may have been produced by a mechanism somewhat similar to that of the Boothia Uplift (Fig. 8).

5. Minto Arch

The Minto Arch on Victoria Island (Thorsteinsson and Tozer, 1962, 1970) exposes a small area of the Precambrian Shield and a large area of Proterozoic sediments (Fig. 6). Structural trends within the arch are parallel to its outline. The arch is overlain unconformably by Cambrian and Ordovician sediments. In most places these dip away gently, but in places there is a faulted contact. The Minto Arch was active as a structural high at some time after deposition of the Ordovician sediments, but the time of this activity is unknown. It is very likely that the arch was raised in Devonian time, contemporaneous with the uplifting of similar basement highs elsewhere in the Canadian Arctic Islands, and its mechanism may also have been similar. The arch may be connected with the Pearya Geanticline but this is speculative.

6. Summary of Basement Uplifts

Phase IV of Sequence 3 was dominated by the activity of basement uplifts. The Pearya Geanticline is a huge basement uplift (Figs. 4, 6 and 7) that was active throughout Sequence 3. It is overwhelmingly larger and more important than the other uplifts in the region (Fig. 6), so therefore has been treated separately in this report. The remaining uplifts which are partly or entirely within the geosyncline, are much smaller, and are fairly similar to each other in size though they vary in amount of vertical uplift, particularly along strike. Four of these basement uplifts trend crosswise to the Franklinian Geosyncline, and a fifth, the Grantland Uplift, trends along the geosyncline.

The several basement uplifts that were partly or entirely within the geosyncline had their main period of activity at about the same time, beginning in Early Devonian. This coincided with the beginning of Phase IV of Sequence 3 (Fig. 5), with the raising of the Pearya Geanticline, the onset of the Ellesmerian Orogeny, *sensu stricto* , and with the general narrowing of the geosyncline by expansion of the Arctic Platform (Fig. 7). In Phase IV there was a great influx of clastic sediments and rapid infilling of the geosyncline. The sediments were derived partly from sources outside the geosyncline, the Pearya Geanticline and the Arctic Platform, and partly from uplifts within the geosyncline. The coincidence in time and the similarity of behaviour to that of the Pearya Geanticline suggests that the raising of these uplifts was associated with and perhaps was a side effect of the Ellesmerian Orogeny, which was dominated by the raising of the Pearya Geanticline. The uplifts also apparently were related to the general narrowing of the geosyncline as well.

The ancient structural trends of the basement uplifts relative to the Pearya Geanticline, whether parallel, oblique, or perpendicular, was a very important influence on later sedimentation and deformation in the Canadian Arctic. The type of arrangements are exemplified by the two extreme cases. The Boothia Uplift trends at right angles to the Franklinian Geosyncline and Pearya Geanticline, and as such represents one extreme. The Grantland Uplift extends along the geosyncline and is parallel to the geanticline, and as such represents the other extreme. Other basement uplifts seem to be either nearly perpendicular, or quite parallel to the geosyncline.

I. Sequence 3, Obstructed Evolution

Several basement uplifts that trend crosswise to the Franklinian Geosyncline (Fig. 6) obstructed the normal evolution of Sequence 3. They first influenced facies patterns as the sequence was deposited, and subsequently influenced structures as the sequence was terminated and deformed. Of these uplifts, the obstructing effects of the Boothia Uplift are best known (Figs. 9 and 10). This uplift affected the basin that formed Sequence 3, and caused local interference or obstructions in the general evolutionary pattern of that sequence. The other uplifts may have had similar but less marked obstructing influences.

The Boothia Uplift extends from the Arctic Platform across the Franklinian Geosyncline and probably connects with the Pearya Geanticline (Fig. 6). It influenced the evolutionary development of Sequence 3 in two ways.(1)Its influence on sedimentary thicknesses and facies of the sequence is shown by cross-section (Fig. 9).(2)Its influence on the structural deformation of the sequence by the younger Ellesmerian Orogeny is shown in plan view (Fig. 10).

In those parts of the line of cross-section far from the Boothia Uplift (Fig. 9) the overall development of Sequence 3 was much like that which occurred elsewhere remote from uplifts (Fig. 7). In both cross-sections then the five phases of Sequence 3 can be readily summarized (Fig. 5) and involve: Phase I, origin and general widening; Phase II, starved basin; Phase III, activity of the geanticline begins to fill the starved basin with flysch deposits; Phase IV, substantial uplift as Ellesmerian Orogeny begins; Phase V, increasing uplift provides a clastic wedge, followed by southward overriding. The history was radically different, only in the vicinity of the Boothia Uplift. Much of this cross-section (Fig. 9) is entirely covered by rocks of the Sverdrup Basin (Fig. 6), but those parts of it that are exposed suggest that its history there was similar to the history of the sequence on Ellesmere Island (Fig. 7).

The effects of the Boothia Uplift on rocks of Phase I are uncertain, because that part of the column is poorly exposed (Miall and Kerr, in press), or known only from the sporadic wells (Mayr, 1978). Thinning over the uplift is suspected. During Phase II the Franklinian Geosyncline appears to have developed regionally in its normal fashion. The Boothia uplift may have been mildly positive then as well (Fig. 9).

During Phase III in the development of Sequence 3, sedimentation was interrupted briefly by a minor uparching of the Boothia Uplift.

Pulse 1 of the Cornwallis Disturbance occurred in Early Silurian time (Figs. 4 and 5), when the Boothia Uplift was arched gently upward. About 450 m of sediments were eroded from its central part on western Grinnell Peninsula (Kerr, 1977a). An unconformity that developed on the southern part of the Boothia Uplift in Early Silurian time (Miall and Kerr, in press) probably is also due to Pulse 1. This was a minor interruption in the sedimentation of the miogeocline, and does not appear to have substantially altered sedimentary patterns there. Thus during Phase III of Sequence 3 the overall evolutionary pattern of the Franklinian Geosyncline continued, largely unaffected by the Boothia Uplift, which probably was a mild arch with an overall thinner sedimentary column.

The Boothia Uplift began to have a major influence on Sequence 3 during Phase IV. During Pulse 2 of the Cornwallis Disturbance (Fig. 4) the Boothia Uplift again was raised, and this time at least 4140 m of rocks were eroded (Fig. 9). From this time on the part of the geosyncline affected by the Boothia Uplift behaved instead as an enlarged part of the Arctic Platform and thereafter received only thin sediments. Thick sediments continued to be deposited on either side of the Boothia Uplift however, within the region that continued as geosyncline. Large amounts of clastic rock derived from the uplift were deposited in the parts of the geosyncline near the uplift.

Pulse 2 of the Cornwallis Disturbance occurred in early Early Devonian time (Fig. 4). It coincided with the beginning of Phase IV of the Franklinian Geosyncline and the onset of the Ellesmerian Orogeny in the Pearya Geanticline. It also coincided with the raising of the Rens Fiord Uplift (Trettin, 1969a), and the Bache Peninsula Arch (Kerr, 1968b, 1976a), and with the general northwestward expansion of the Arctic Platform by migration of the southeastern hinge-line (Fig. 7). These events doubtless were related, and were controlled by some broader connected subcrustal mechanism.

After Pulse 2, sedimentation resumed on the Boothia Uplift (Fig. 9), as part of a widespread regional subsidence, and sedimentation continued there through much of the rest of Devonian time (Kerr, 1977a). The Boothia Uplift had a subsequent pulse during Phase IV, which was Pulse 3 in late Early Devonian time. This was much weaker than Pulse 2, and a lesser thickness of rocks was eroded from the uplift.

The Boothia Uplift had a substantial effect on sedimentary patterns of Sequence 3, by disrupting the normal pattern of sedimentation in a north trending linear belt (Fig. 6). The effects on sedimentation during Phases I, II and III were minor but during Phase IV they were substantial (Fig. 9).

During Phase IV the Franklinian Geosyncline was evolving regionally in its normal fashion (Fig. 7). The Pearya Geanticline was being raised and eroded, and the Ellesmerian Orogeny was progressing toward its culmination. In the region of the Boothia Uplift, there was a markedly different pattern (Fig.9). During Phase IV several uplifts fragmented the geosyncline and contributed to its filling there in the same way and at the same time as did the Pearya Geanticline. By analogy with the Rens Fiord Uplift, it is suggested that the Boothia Uplift was connected on the north with the Pearya Geanticline.

In Phase V of Sequence 3 the Pearya Geanticline continued to rise, and folds and thrusts were beginning to form in the geosyncline because of the Ellesmerian Orogeny. The Boothia Uplift was subjected to the last upward pulse of the Cornwallis Disturbance (Figs. 4 and 5) in Late Devonian time (Kerr, 1977a). After that pulse the Boothia Uplift possessed what is referred to as its basic structure (Fig. 8), that of an anticlinorium cored by a crystalline horst (Fig. 9, Phase V; Fig. 10, left). In the line of cross-section (Fig. 9) mass transport during Phase V that resulted from the Ellesmerian Orogeny was directed southward away from the geanticline. In this part of the geosyncline , the Boothia Uplift had a remarkable effect on the deformation of the Ellesmerian Orogeny and interfered with some of that mass transport (Fig. 10).

In the region of the Boothia Uplift the developmental history of the Franklinian Geosyncline and Arctic Platform was affected in two ways. First the Boothia Uplift influenced or modified regional facies patterns (Fig. 9). Secondly it influenced or obstructed the otherwise rather simple deformation produced by the Ellesmerian Orogeny (Fig. 10).

The basic structure of the Cornwallis Fold Belt (Fig. 8, and Fig. 9, Phase IV) was in existence prior to the overriding of the Ellesmerian Orogeny. In plan view at this time the basic structure (Fig. 10, left) was an anticlinorium, with a crystalline horst as a core, trending northward across an otherwise undeformed geosyncline. After the Ellesmerian Orogeny a very different situation existed (Fig. 10, right). In regions on either side of the Boothia Uplift where there was no interference by a pre-existing north trending structure, the southward overriding of the Ellesmerian Orogeny deformed the Franklinian Geosyncline rather simply. This rather simple deformation formed the Parry Islands Fold Belt in the west, and the Ellesmere-Greenland Fold Belt in the east. These belts contained anticlines that were sub-parallel to the Pearya Geanticline from whence they were initiated, and sub-parallel to the depositional trend of the Franklinian Geosyncline. Both belts formed by the folding of an undeformed geosynclinal pile, and structures developed parallel to the trend of the geosyncline and overrode toward the Arctic Platform. The Ellesmerian Orogeny here involved a southward directed decollement within one or more evaporite layers that lay within the rocks of Phase III, and the folding that this produced was a relatively surficial phenomenon (Kerr, 1974).

In and near the Boothia Uplift the rocks of Sequence 3 developed into quite a different situation during the Ellesmerian Orogeny than they did elsewhere (Figs. 9 and 10).

The Boothia Uplift obstructed the rocks of the geosyncline that were being transported southward. A zone of interference developed within the Cornwallis Fold Belt, and that belt was deformed in an unusual way. The columns on either side were carried southward or southeastward by decollements. The strong north trending structural grain that existed within the Cornwallis Fold Belt prevented a similar decollement from developing there. Nevertheless the great force of the southward moving mass was being imposed on the Cornwallis Fold Belt, and deformed it in a more complicated way (Fig. 10). Vertical strike-slip faults developed within the Cornwallis Fold Belt, following approximately the axes of the pre-existing anticlines. It is inferred that at depth these faults followed fundamental near vertical normal faults in the crystalline basement that had controlled the anticlines of the Cornwallis Fold Belt (Fig. 8). The strike slip faults in the western part of the Cornwallis Fold Belt were left-lateral, while those in the eastern part of the fold belt were right-lateral. The actual southward movement of material within the Cornwallis Fold Belt diminished from the edges toward the center.

The Rens Fiord Uplift (Fig. 6) may have interfered with the Ellesmerian Orogeny in a way similar to that of the Boothia Uplift. This cannot be determined because much of the Rens Fiord Uplift is covered by younger rocks.

The Bache Peninsula Arch may have interfered with the Ellesmerian Orogeny to a small degree. There does not appear to have been an influence in the stratigraphic level that is presently exposed; however there may have been interference at greater depth. The relationship of the Minto Uplift to the effects of the Ellesmerian Orogeny is unknown.

The Franklinian Geosyncline, Pearya Geanticline, and Arctic Platform are large tectonic features with a tendency to develop regionally as shown in Figure 7. Most cross-sections drawn through those features between Banks Island (Miall, 1976a) and northern Greenland (Dawes, 1976) probably show a generally similar history. The basement uplifts (Fig. 6) are smaller tectonic features that produced local abberations in the three features (Fig. 9). All these features, of course, developed in partial contemporaniety.

J. Continent as a Substructure Prior to Ocean Development

In latest Devonian time, at the end of the Ellesmerian Orogeny, all of northern Canada including the entire region of the Canadian Arctic Archipelago probably may have been above sea level and exposed (Fig. 7, Phase 4). Trending northeastward were the folded Franklinian Geosyncline and Pearya Geanticline, apparently both with high relief. Farther south was the exhumed Arctic Platform, where a thin sedimentary column covered much of the present Canadian Shield. The Shield itself probably was only locally exposed, in highs such as the Boothia Uplift, where there had been substantial Devonian erosion.

Little is known about the history of the water covered region northwest of the Canadian Arctic Archipelago. The Pearya Geanticline probably was very broad and may have included much or all of the region that is now continental shelf and slope adjacent to the Canada Basin. The continent at this time also may have occupied adjacent parts of the Canada Basin. Recent investigations (Churkin, 1969; Vogt and Ostenso, 1970; Ostenso and Wold, 1973) have concluded that the Canada Basin is underlain by oceanic crust. It may however contain downfaulted continental rocks, because according to King *et al.* (1966), the rocks have continental characteristics.

The end of the Ellesmerian Orogeny in latest Devonian or early Mississippian time is an important turning point in the history of northern North America. It represents the end of a constructional phase in which stratigraphic Sequences 2 and 3 were deformed, and solidified in succession as part of a large continent. This primeval continent, largely above sea level, may have occupied the entire region shown by Figure 1, including both land and sea areas of the present day.

III. FRAGMENTATION PHASE: EVOLUTION OF THE OCEAN BASINS AND BRANCHES WITHIN THE CONTINENT

1. Foreword

The fragmentation phase of northeastern North America began shortly after the Ellesmerian Orogeny in latest Devonian or Early Mississippian time (Fig. 4), and continued

to the present day. This phase appears to have begun with a primeval continent, largely expose that included the entire region shown by Figure 2. The Sverdrup Basin and possibly the Canada Basin first developed as rifted and subsiding basins, unconformably upon part of this exposed continental landscape. Subsequently major rifting events produced the present configuration of lands and seas in the Canadian Arctic.

A. Boreal Rifting Episode

The Arctic and Atlantic oceans are connected today via the Canadian Arctic by means of structurally controlled channels between them. Those structures resulted from rifting that emanated cratonward from the two oceans. Rifting appears to have begun first in the Canada Basin, and occurred intermittently there during deposition of Sequences 4, 5, and 6 (Fig. 4). This rifting event, which may have originated in the Alpha Ridge, is here named the Boreal Rifting Episode. Side effects of this rifting reached some distance southeastward into the continent, where it apparently caused the Sverdrup Basin to form by crustal fracturing. The Boreal Rifting Episode was an unsuccessful attempt to connect the Arctic and Atlantic Oceans. It did not break fully through the continent to the southeast, and therefore was aborted.

B. Sequence 4: Mid-Mississippian (Viséan) to Early Cretaceous (Pre-Isachsen Formation)

1. Foreword

The rocks of Sequence 4 of the Innuitian Mobile Belt (Figs. 3 and 4) were deposited in several basins (Fig. 11), the Sverdrup Basin, the Banks Basin, and presumably an open Arctic Ocean that became the Canada Basin. They also were deposited at times on the Sverdrup Rim, an intermittently positive feature, that overlay the Pearya Geanticline during its nonmagmatic stage.

The tectonic framework of Sequence 4 apparently was dominated by the events that took place in the Arctic Ocean. This basin appears to have been forming as a major depocenter during the time span of Sequence 4, by a rifting mechanism, possibly centered on the Alpha Ridge (Fig. 11) and named here the Boreal Rifting Episode (Fig. 4). The Sverdrup Basin and Banks Basin are smaller depocenters, located on a primarily continental area. They formed in response to events in the Canada Basin, are marginal to it, and are separated from that major basin by the Sverdrup Rim and Storkerson Uplift (Fig. 11).

2. Sverdrup Basin

The evolution of the Sverdrup Basin (Fig. 11 has been summarized recently by Balkwill (1978), where much of the following data were presented. Stratigraphic nomenclature of the Sverdrup Basin has been discussed by Thorsteinsson and Tozer (1970), Plauchut (1971), Thorsteinsson (1974), and Henao-Londoño (1977), and subsidence rates by Sweeney (1977). The framework of the basin was discussed by Trettin *et al.* (1972), and its petroleum potential by Stuart Smith and Wennekers (1977).

The Sverdrup Basin contains mid-Mississippian (Viséan) and younger rocks up to 13 000 m thick, and formed as a pericratonic basin along the northwest margin of Arctic Canada after the Ellesmerian Orogeny. It lies unconformably on folded rocks of the Franklinian Geosyncline and Pearya Geanticline (Fig. 3). Sequence 4 is mainly concordant in the axial part of Sverdrup Basin, but there are unconformities along the margins.

Sequence 4 includes the lower part of the column of the Sverdrup Basin (Fig. 4). Three phases are recognized within Sequence 4 there, each phase represented by a stratigraphic assemblage (Fig. 12). These are similar to the first three phases of the Sverdrup Basin as outlined by Balkwill (1978), with slight modification here in Phase III. These phases are (I) late Paleozoic (Carboniferous and Permian), when evaporites and marine muds were deposited in the axial region, and carbonates and mature sands on the margins; (II) early Mesozoic (Lower to mid-Upper Triassic), when great thicknesses

of siltstone and shale accumulated in axial parts of the basin, with lesser thicknesses of sand on the margin; (III) middle Mesozoic (mid-Upper Triassic to Lower Cretaceous), when widespread terrigeneous clastic rocks accumulated slowly during transgressions and regressions.

The history of the three phases of Sequence 4 was one of continuous sedimentation with minor interruptions of deposition (Fig. 12). During Sequence 4 the Pearya Geanticline was in its nonmagmatic phase, and was mildly negative to positive (Fig. 4).

The southern and eastern margin of the Sverdrup Basin is at the major basal angular unconformity, with the basin lying on folded older rocks and dipping regionally northwestward (Figs. 11 and 12). This margin was also the approximate location of the shoreline during much of Phases I to III of the Sverdrup Basin. On its north side the Sverdrup Basin dips southeastward regionally. On Ellesmere and Axel Heiberg Islands the Pearya Geanticline was exposed through most of the Sverdrup Basin history, so the present northwestern boundary there was the approximate shoreline at most times. Farther southwestward the Sverdrup Basin was bordered by the Sverdrup Rim, which was exposed from time to time during Phases I to III (Meneley *et al.*, 1975; Balkwill, 1978). This rim is aligned with and presumably is underlain by the Pearya Geanticline. It contains rocks equivalent to much of the Sverdrup Basin, but they are thinner, presumably due to the positive tendency of the geanticline beneath. A central high region of thinner rocks extended across the Sverdrup Basin (Meneley *et al.*, 1975), making two depocenters. Balkwill (1978) designated the depocentre to the west the Barrow segment and the one to the east the Axel Heiberg segment.

PHASE I: Mid-Mississippian to Late Permian

The Sverdrup Basin began to take on its form as a subsiding basin shortly after the Ellesmerian Orogeny ended (Figs. 4 and 11). The oldest rocks of the basin are the middle Mississippian (Viséan) Emma Fiord Formation, which is sporadically exposed on its northern and southern edges (Thorsteinsson and Tozer, 1970; Trettin, 1973b; Kerr, 1976b). The Emma Fiord Formation occurs in small, downwarped and downfaulted structural lows that formed sedimentary embayments on the fractured margins of the Sverdrup Basin (Fig. 12). On the southern margin at northern Grinnell Peninsula (Kerr, 1976b), relief developed gradually during Emma Fiord deposition by downwarping and faulting.

The Emma Fiord Formation occurs on the northern margin of the Sverdrup Basin, but in the central part this level is deeply buried and unknown. There probably was a similar tectonic setting throughout the Sverdrup Basin, with nonmarine Mississippian rocks deposited sporadically throughout it in narrow basins.

Mid-Mississippian (Viséan) time appears to represent a very early stage of crustal fracturing with local small subsiding basins. This may have been the beginning of the Boreal Rifting Episode (Fig. 4), an extensional episode that followed the compression and exposure by the Ellesmerian Orogeny. Initial fracturing in mid-Mississippian time in the Sverdrup Basin was a transition that graded into the later widespread fracturing of younger parts of Phase I of the Sverdrup Basin (Fig. 12).

During later parts of Phase I, the crustal fracturing event increased and the Sverdrup Basin began to subside greatly. A widespread conglomerate unit was deposited throughout the basin in latest Mississippian to early Pennsylvanian time. This unit is called the Canyon Fiord Formation in the southeast and the Borup Fiord Formation in the northwest (Thorsteinsson, 1974). It was deposited during a time when there was high relief, and active faulting occurred throughout the Sverdrup Basin. The crustal fracturing or extension developed local highs and lows within the overall low that formed the Sverdrup Basin. The highs were horsts or anticlines that provided a source of clastic sediments. The lows were grabens or synclines that received those sediments.

The crustal fracturing event that began to form the Sverdrup Basin during Phase I is well displayed on northern Grinnell Peninsula (Kerr, 1976b), but probably affected the entire Sverdrup Basin in much the same way. In central parts of the Sverdrup Basin for example, renewed differential uplift of the Grantland Uplift occurred during Phase I (Nassichuk and Christie, 1969; Trettin *et al.*, 1972; Mayr, 1976b)

It appears that the great crustal fracturing and faulting that initially produced the Sverdrup Basin was gradually but generally replaced during Phase I by subsidence or sagging. Gradually in the later part of Phase I the basin took on the form of a broad sag between the craton and the Pearya Geanticline. Relief was reduced by erosion, the proportion of conglomerate in the column progressively diminished, and finally fine grained marine sedimentation prevailed (Fig. 12).

By Pennsylvanian and Permian time marine deposition was more widespread. Encroachment out of the southern part of the basin produced overstepping by Pennsylvanian rocks. The Melvillian Disturbance (Fig. 4) produced local disconformities between Pennsylvanian and Permian rocks of the southern margin of the Sverdrup Basin, according to Thorsteinsson and Tozer (1970). At this time the Sverdrup Rim was breached (Meneley *et al.*, 1975; Balkwill, 1978), and deeper water facies followed. The Melvillian Disturbance apparently represents stepped-up activity in a long lasting otherwise gradual crustal fracturing event referred to here as the Boreal Rifting Episode.

Continued broader subsidence during Phase I downwarped the Sverdrup Basin regionally, and thick sediments accumulated. The basal conglomeratic rocks of the basin were mainly nonmarine. They grade up to and were succeeded by fine grained marine rocks, mainly limestones along the margins, and evaporites and shales along the axis (Fig. 12). Thorsteinsson (1974) distinguished several concentric facies belts in the Upper Carboniferous to Lower Permian rocks. There were clastic rocks around the margin, carbonates within this, and deep water evaporites and shales in the middle. The axial facies during Phase I formed a thinner column than the marginal facies (Fig. 12). In the lower part of the axial facies there are thick evaporites. They consist of gypsum and anhydrite in outcrop (Thorsteinsson, 1974; Davies and Nassichuk, 1975; Wardlaw and Christie, 1975), but include large volumes of halite as well in the subsurface (Davies, 1975). The upper part of the axial facies during Phase I was mainly shale.

The Canada Basin apparently was in existence during Phase I of the Sverdrup Basin (Figs. 11 and 12. This suggestion is supported by upper Paleozoic faunas which have a circum-Arctic affinity (Nassichuk, 1975) and which appear to have entered the Sverdrup Basin from the northwest (Nassichuk and Davies, in press). The Canada Basin may have been initiated at the same time as the Sverdrup Basin, and their origins probably were related. The crustal fracturing process that developed the Sverdrup Basin during Phase I (Fig. 12) may have been a marginal effect of the more intense crustal fracturing or rifting, and plate separation that produced the Canada Basin. These deformations may have originated in an active spreading center on the site of the Alpha Ridge (Fig. 11).

PHASE II: Early Triassic to mid-Late Triassic (Norian)

Phase II of the Sverdrup Basin (Fig. 12) includes Lower Triassic to mid-Upper Triassic (Norian) rocks (Balkwill, 1978). The oldest Triassic rocks in the basin are as old as any Triassic rocks known (Tozer, 1967). They lie on a widespread regional disconformity that marks the beginning of Phase II. This is a basin-wide disconformity resulting from marine regression, and separates the Permian and Triassic Systems (Thorsteinsson, 1974).

Phase II is represented by two correlative facies, a marginal and an axial basin facies (Fig. 12). The marginal facies occurs in the east and south, and consists of sandstone and conglomeratic sandstone. It grades very abruptly basinward to a thick succession of marine siltstones and shales that were deposited as turbidites. Some conglomeratic sandstone is present along the

northwestern rim, but little detritus was contributed from that direction. The basinal succession in Phase II is as thick as 4500 m, which is almost as thick as the remainder of the Mesozoic succession. Phase II differed from Phase I in that the axial facies was considerably thicker than the adjacent marginal facies (Fig. 12), the opposite to conditions during Phase I. Phase II also differed from Phase I in having no significant carbonates or evaporites. Phase II differed from later phases in having deeper water.

Phase II ended in mid-Late Triassic (mid-Norian) time, with a shallowing, when an Upper Triassic sandstone began to prograde widely across the Sverdrup Basin. The main large diapirs of the Arctic Islands formed by halokinesis (migration due to loading), a long continuing process that began in this phase (Triassic time). By mid-Late Triassic time the combined thickness of Triassic and upper Paleozoic rocks that overlay the Carboniferous Otto Fiord evaporites in eastern parts of the Sverdrup Basin was about 6000 m. Halokinesis was initiated by this loading (Balkwill, 1978).

Diapirs in the Arctic Islands occur only in central parts of the Sverdrup Basin, because they originated in the Otto Fiord Formation, an axial facies. The natural gas pools in central parts of the Sverdrup Basin are related to these diapirs and are discussed later. There are numerous evaporitic piercement domes in the Sverdrup Basin (Gould and de Mille, 1964; Thorsteinsson, 1974). They originated in the Otto Fiord Formation, an evaporite unit within the rocks of Phase I (Fig. 12). Some developed halokinetically, beginning probably during Phase II in Late Triassic time (Balkwill, 1978). They had intermittent activity that continued during deposition of younger rocks (Fig. 12). Other domes were generated later, during Tertiary folding and faulting. Evaporite has reached the surface and is exposed today in many structures, and in others it has reached only part way to the surface.

Phase III - Mid-Late Triassic (Norian) to early-Early Cretaceous (Valanginian)

Phase III of Sequence 4 is similar to the third phase (middle Mesozoic) of the Sverdrup Basin as outlined by Balkwill (1978). A difference is that he included older parts of the Isachsen Formation and equivalent rocks at the top of the third phase, and the present paper places the entire Isachsen Formation in the lower part of Sequence 5 (Fig. 12).

The Upper Triassic, Jurassic, and lower Lower Cretaceous rocks that constitute Phase III of Sequence 4 are characterized by alternating sandstone and shale formations (Fig. 12). The shales are marine, while the sandstones are largely but not wholly nonmarine. Throughout nearly all of Phase III the craton southeast of the Sverdrup Basin was a sedimentary source made up of the older fold belts and the Arctic Platform. As a result the rocks are predominantly nonmarine sands along the southern, southeastern, and southwestern edges of the Sverdrup Basin. Moreover, along that margin there are several angular unconformities with very slight angularity, that disappear basinward into continuous successions. Phase III included several southerly derived blanket sands that were spread fully across the Sverdrup Basin, and only two are shown (Fig. 12). In certain axial parts of the Sverdrup Basin it appears that shales were deposited more or less continuously throughout Phase III. Some sand however was deposited on the Sverdrup Rim and was largely enclosed by these shales (Fig. 12).

Upper Mesozoic clastic sediments of the Sverdrup Basin (Phase III) were largely derived from the southeast, presumably the folded Franklinian Geosyncline, the Arctic Platform, and the Precambrian Shield. The lowest rocks of Phase III are the upper or nonmarine part of the Heiberg Formation, a widespread Upper Triassic to Lower Jurassic delta deposit that prograded across the Sverdrup Basin from the southeast and represented almost complete marine withdrawal from that basin. Fine grained marine equivalents of this great column of sand must exist still farther northwest beneath the Arctic Coastal Plain and Continental Shelf, as part of a thick and long lived continental terrace wedge. This implies a proto-Arctic Ocean northwest of the Sverdrup Basin on the site of the Canada Basin by Late Triassic time. A subsequent Jurassic marine transgression deposited widespread marine shales almost fully across the basin. From late Early Jurassic until latest Jurassic time (late Volgian) shallow conditions of low relief continued in the Sverdrup Basin, and thin marine and nonmarine shales and sands alternated. From latest Jurassic to early Early Cretaceous (late Valanginian) time, sedimentation rates increased as deltas prograded into Sverdrup Basin from the west, southeast and northeast. This culminated by early Early Cretaceous (late Valanginian) when there was a basin wide marine regression and widespread exposure. This regression resulted from the deltaic alluviation of a thick succession of sands, directed principally from the southeast (Roy, 1974; Balkwill, 1978). This delta progradation deposited the Isachsen Formation, which was partly contemporaneous with tectonism (Rahmani, 1977). This widespread Isachsen Formation of sands marked a fundamental change in sedimentary and tectonic patterns in the Canadian Arctic. The top of Phase III of Sequence 4 is drawn in this paper in Valanginian time at the base of the Isachsen Formation of the Sverdrup Basin, which marked the onset of a pulse of increased tectonism.

The lower part of the Sverdrup Basin, which makes up Phases I, II, and III of Sequence 4 of the Innuitian Mobile Belt, had a long history that began with crustal fracturing and subsequently merges into a broader subsidence. This interval formed the lower part of the Sverdrup Basin, and ended with the beginning of deposition of the Isachsen Formation.

During the period of its history that makes up Sequence 4, late Paleozoic to early Early Cretaceous (Valanginian), the Sverdrup Basin appears to have been a small basin, lying on the continent, adjacent to a larger developing oceanic basin, the Canada Basin (Fig.12). The events that affected the Sverdrup Basin in this interval probably were secondary events, marginal to larger scale rifting events that occurred in the adjacent Canada Basin. It happens that rocks of the Sverdrup Basin are well exposed and well described (Balkwill, 1978), so that events that took place in the Canada Basin can be inferred from it. The Canada Basin had a long history of sporadic deformation, with the Sverdrup Basin region reflecting side effects, specifically crustal fracturing or rifting. It is probable that the Canada Basin of the Arctic Ocean began to form in late Paleozoic time, by the Boreal Rifting Episode, and has been in existence ever since. There is strong evidence from faunas (Tozer, 1961) that it has been in existence continuously from Late Triassic time onwards.

Balkwill (1978) pointed out that the depth to which the Sverdrup Basin would have subsided if not loaded with sediments is about the same as the present depth of the Canada Basin, i.e. about 3500 m. The Sverdrup Basin appears to have formed by vertical tectonics, prior to and in an early stage of plate breakup, in a manner similar to that outlined by Falvey (1974) for basins on other rifted continental margins.

3. Banks Basin

The Banks Island area lies southwest of the Sverdrup Basin (Fig. 11), and was part of an unstable cratonic margin in Jurassic and later time (Miall, 1975). The Banks Basin (Figs. 11 and 13) is a narrow basin that at times was continuous with or merged into the Sverdrup Basin. It was bordered on the northwest by the Storkerson Uplift, which is the southwest continuation of the Sverdrup Rim.

The rocks of Sequence 4 of the Innuitian Mobile Belt that are known in the Banks Basin are of late Early Jurassic to early Early Cretaceous age (Fig. 13). They are equivalent to rocks of Phase III of the Sverdrup Basin (Fig. 12). They are thick marine sands and shales. At this time the Banks Basin was part of a narrow marine trough that continued northeastward toward Sverdrup Basin, and southwestward toward the Beaufort Sea. This trough was bordered on the northwest by the exposed and positive Sverdrup Rim, which separated the subsiding Banks Basin from the deeper Canada Basin. This was a period of low to moderate tectonic activity, with relative uplift and subsidence, but apparently not strong faulting.

C. Sequence 5: Early Early Cretaceous (Valanginian) to Late Cretaceous (Campanian/Maastrichtian)

Sequence 5 begins with the Lower Cretaceous Isachsen Formation and includes all rocks up to the base of the Eureka Sound Formation. The sequence represents a marked change in sedimentary patterns from the earlier sequence (Figs. 4 and 12). The Isachsen Formation and equivalent rocks are a widespread wedge of nonmarine fine grained sandstone that represents a marine regression and alluviation.

The Isachsen Formation is rather thick in the Sverdrup Basin (1400 m) and generally conformable. The formation is older in the basin, being early Early Cretaceous (Valanginian). It becomes younger where it encroached onto the cratonic shelf (Fig. 12), there being of mid-Early Cretaceous age (about Aptian). The Isachsen Formation also is thinner on the margins of Sverdrup Basin (150 m), and in the wide cratonic shelf area where it encroached or overstepped southeastward, eastward, and southwestward, far beyond the long established margin of the Sverdrup Basin. Prior to the Isachsen Formation when Sequence 4 was being deposited (Fig. 12), the Sverdrup Basin was a distinct basin with clearly definable limits. The regression and subsequent overstepping event that deposited the Isachsen Formation began Sequence 5, in which there was a much broader continental shelf basin system (Fig. 12). The Sverdrup Basin was the deepest subsiding element within that system. The region to the south and southeast, where Sequence 5 extended beyond the bounds of the Sverdrup Basin, was principally a cratonic shelf.

The Isachsen sands of the Sverdrup Basin were derived mainly from the craton southeast of the margin of the Sverdrup Basin (Roy, 1974). The younger Isachsen sands of the cratonic shelf probably were derived from locally high parts of that shelf. The Isachsen Formation also occurs widely southeast of the normal bounds of the Sverdrup Basin on Banks Island (Fig. 13), where it is on an unstable cratonic margin (Miall, 1975). The rocks of the cratonic overlap assigned to the Isachsen Formation lapped out of the Sverdrup Basin to Bathurst Island (Kerr, 1974), and central Ellesmere Island (Thorsteinsson and Tozer, 1970), and extended far to the southeast.

Basal sands of Sequence 5 are present in the northern Baffin Island region (Jackson and Davidson 1975; H.R. Balkwill, pers. com., 1978). They are younger there than the basal sands of farther northwest, and equivalent to the Hassel Formation (Fig. 12). Similar basal sands are also present on western Greenland (Henderson *et al.*, 1976). Basal sands of this sequence also diachronously overstepped onto the Anderson Plain of the northern Canadian mainland (Young *et al.*, 1976). The encroachment onto the craton by the Isachsen, Hassel, and related basal formations was probably as a broad shallow epicontinental sea, rather than in deeply rifted basins. This overstepping in the Canadian Arctic was part of and synchronous with a worldwide marine transgression onto cratonic regions (Kent, 1976). It took place over much of the North America Continent.

The Isachsen Formation of the Sverdrup Basin was deposited during an early Early Cretaceous faulting event that began in Valanginian time (Rahmani, 1977). This faulting presumably was an extensional event that fractured the crust, subsided the Sverdrup Basin, and allowed an abnormally great thickness of the Isachsen Formation to accumulate in the depressed parts of the basin. The faults that produced this fracturing apparently did not extend upward far enough to cut Mesozoic rocks (Fig. 12). The Hassel Formation of the cratonic shelf represents a younger widespread transgression southeastward in mid-early Cretaceous (about Aptian) time. While this transgression occurred onto the cratonic shelf, faulting and sedimentation continued within the Sverdrup Basin where deposition was mainly shale (Fig. 12). Fault bounded troughs apparently developed beneath the Sverdrup Basin, producing locally thick depocenters in the sags that formed above them. The pattern of these sags indicates that they deepened to the northwest, and they presumably connected with perhaps thicker rocks of the continental shelf and were tied to the Canada Basin. The troughs developed at the time of deposition of the Isachsen Formation, which is thickest in the troughs, but the formation later spread more widely through Sverdrup Basin. The pulse of faulting that occurred in the Sverdrup Basin and the Banks Island region during Isachsen deposition probably was a marginal effect of an enlarged pulse of the

Boreal Rifting Episode in the nearby Canada Basin of the Arctic Ocean.

The post-Isachsen rocks of Sequence 5 were alternating shales and sands in the Sverdrup Basin. The rocks assigned to Sequence 5 in this paper have a total thickness of about 3000 m in the depoaxis of the Axel Heiberg segment of Sverdrup Basin (Balkwill, 1978). Rifting appears to have continued after Isachsen deposition, because the Christopher Formation of the Sverdrup Basin represents very rapid accumulation and presumably further subsidence. Post-Isachsen rocks of Sequence 5 include in ascending order, the Christopher Formation (marine silty shale with basalt flows), the Hassel Formation (coaly sandstone), and Kanguk Formation (acidic black shale). These rocks reflect respectively, a subsidence and marine inundation, a marine withdrawal and alluviation, and a starved euxinic marine basin. A similar sequence of events occurred overlying the Isachsen Formation in the Sverdrup Basin (Plauchut, 1971), and Banks Basin (Miall, 1975; Plauchut and Jutard, 1976). The Early Cretaceous overlap reached western Greenland (Henderson *et al.*, 1976), where the succession is similar to that of Bylot Island. The pattern of rifting and cratonic overlap generally supports the view of Jeletzky (1978) that oscillations in Cretaceous sea levels in northern North America resulted from tectonic events there.

Sequence 5 ended with widespread marine withdrawal from Sverdrup Basin, Sverdrup Rim, the cratonic shelf, and the Banks Basin. This withdrawal coincided with the onset of a new phase of tectonism that deposited the nonmarine dominated rocks of the Eureka Sound Formation. This tectonism was dominated by the Eurekan Deformation (Kerr, 1977a). It began in Late Cretaceous (Campanian/Maastrichtian time), and drastically changed sedimentary patterns. The top of Sequence 5 is at the onset of deposition of the Eureka Sound Formation, a mainly nonmarine sandstone. In places this contact is gradational but in others it is unconformable.

D. Connecting of Oceans by the Canadian Arctic Rift System

The Atlantic and Arctic Oceans are connected by a major seaway system that extends through the North American Continent. This seaway makes up the famous Northwest Passage. It includes the Labrador Sea, Baffin Bay, Parry Channel, and other channels within and adjacent to the Canadian Arctic Islands (Fig. 1). This seaway was formed by geological events, and is still controlled by structures of the Canadian Arctic Rift System (Kerr, 1973a). The rift system is a northwest trending branch of the Mid-Atlantic Rift System, and formed by two major plate tectonic episodes.

The Canadian Arctic Rift System formed as a primeval continent broke up into plates and subplates. Plate activity took place in both the Atlantic and Arctic Oceans, and was propagated from these oceanic areas into the continent where the two rift deformations met. The Canadian Arctic Rift System formed as as result of fault propagation cratonward from both its ends at various times. The rifting episode that emanated southeastward from the Canada Basin was the Boreal Rifting Episode (Fig. 4), and began first. The episode that actively brought about the ultimate structural connection of the two oceans however, was mainly the Eurekan Rifting Episode (Kerr, 1977a). That deformation was propagated northwestward from the Atlantic Ocean, and occurred during deposition of Sequences 6 and 7 in the Canadian Arctic region.

E. Sequence 6: Late Cretaceous (Campanian/Maastricthtian) to Mid-Tertiary (Miocene)

Sequence 6 was deposited while two plate tectonic episodes were affecting the North American Arctic, the Eurekan Deformation (Kerr 1977a), and the Boreal Rifting Episode (Fig. 4). There were two main depocenters where thick sedimentary columns of Sequence 6 accumulated (Figs. 14 and 15). The largest depocenter was in the northwest. It includes a thick Cenozoic to modern continental terrace wide (Fig. 2) that extends along the margin of the Arctic Ocean. The history of that wedge was dominated by the history of the adjacent Canada Basin. The other includes the Baffin Bay Basin and the Labrador Sea (Fig. 16A). The Queen Elizabeth Islands Sub-plate (Fig. 17) is a triangular-shaped block between those oceanic areas.

The two main depocenters are connected by rifted branches within the channels of the Canadian Arctic Archipelago. These channels form smaller local basins with lesser thicknesses of sediments (Fig. 14). Lancaster Sound for example projects westward from Baffin Bay and is fault-controlled (Barrett, 1966; Jackson *et al.*, 1977). It contains Lancaster Aulacogen (Kerr, in press), which has a westward-thinning extension of the thick sedimentation of the Baffin Bay Basin (Fig. 14). There also were several largely isolated, moderate-sized depocenters within the archipelago that are not confined to rifted channels. The central part of the Canadian Arctic Islands, is the area least affected by rifting and forms a broad northeast-trending height of land (Figs. 14 and 15). Fault controlled branches from the two oceanic areas project varying distances into this height of land.

A diagrammatic cross-section through the Canadian Arctic Islands (Fig. 15). shows how that area may have existed at the end of Sequence 6, in early Tertiary time, in an intermediate stage of the Eurekan Deformation. It shows the main height of land, the Cornwall Arch, and the Sverdrup Rim.

The Canada Basin of the Arctic Ocean apparently had been in existence continuously since late Paleozoic or earlier time, and therefore presumably includes rocks older than Sequence 6. Thus the thick column along the Arctic margin northwest of Sverdrup Rim (Figs. 14 and 15) probably includes upper Paleozoic, Triassic, Jurassic, and Cretaceous rocks in its lower part. Sequence 6 then presumably lies above these rocks along the margin of the Canada Basin (Fig. 15), having accumulated contemporaneously with the Eurekan Deformation (Fig. 4).

Sequence 6 apparently was deposited as plate movements expanded the Canada Basin and the Baffin Bay Basin more or less simultaneously. These plate movements formed the basins within which Sequence 6 was deposited by downfaulting them, but also raised adjacent areas to serve as sedimentary sources.

The Atlantic and Arctic Oceans became partly connected via channels within the Queen Elizabeth Islands. That island area is partly land and partly sea.

It is therefore a structural and physiographic transition between a branch of the Atlantic Ocean (Baffin Bay), and the Arctic Ocean.

F. Plate Reconstructions

1. Foreword

Sequence 6 in the North American Arctic was closely controlled by the plate tectonic processes that formed lands and seas,because the sequence was deposited while those tectonic events occurred. However, there are differences in interpretation of tectonics, which focus on two major problems. These problems are (1) the origin of the Canada Basin and its relationship to Alaska and Canada, and (2) the relationship of Greenland to the rest of North America. These are discussed below.

2. Canada Basin

The Canada Basin (Fig. 1) is a large deep part of the Arctic Ocean lying northwest of the Queen Elizabeth Islands, and southwest of the Alpha Ridge. It has depths greater than 3500 m over a wide area. The edge of the continental shelf between the Canada Basin and the Queen Elizabeth Islands trends northeast, and is remarkably straight. It is a lineament controlled by the Kaltag Fault (Yorath and Norris, 1975).

Numerous contrasting views exist on the origin of the Canada Basin. They have been summarized by by Churkin (1973) and by Sweeney and Haines (1978). On the one hand Meyerhoff (1970, 1973) states that the Canada Basin and other features in the Arctic Ocean have been in existence since Late Proterozoic or earlier time. Standing in contrast are views that the Canada Basin is a relatively younger feature. These latter ideas can be grouped into theories that suggest it is formed by (1.) subsidence of continental crust,and (2.) by rifting and lateral plate movements. Shatskiy (1935), Puscharovskiy (1960, 1978), and Atlasov (1964) concluded that the Canada Basin formed when the continental crust subsided.

A rifting origin for the Canada Basin was first proposed by Carey (1958). A rifting origin has been supported by numerous authors, who differ on its age and mechanism. Churkin (1969), suggested that the Canada Basin is at least as old as Paleozoic, Lambert (1973) that it developed in Carboniferous time, Taillcur (1973) that it formed in post-Triassic time, Herron *et al.* (1974) that it formed in Jurassic time, Sweeney *et al.* (1978) that it began to form in Early Cretaceous time, and Vogt and Ostenso (1970) that it is no older than Late Cretaceous. Geophysical work indicates that the Canada Basin is a true ocean basin,with an abyssal plain floored by oceanic crust (Ostenso and Wold, 1973; Hall, 1973).

This paper concludes that the Canada Basin of the Arctic Ocean is a long persisting oceanic basin, formed by the Boreal Rifting Episode which was active intermittently between late Paleozoic (Mississippian) and Tertiary time (Fig. 4). It was suggested by Lambert (1973), Sweeney (1977), and Balkwill (1978), that extensional formation of the Canada Basin also may have produced extension in the Sverdrup Basin. That relationship is supported herein. The age of the Sverdrup Basin, which formed by rifting beginning in Carboniferous time, therefore is used here as a key to determining the age of the Canada Basin. The Canada Basin may have formed at the same time, and it is presumed, therefore, that the thick wedge of sediments northwest of the Arctic Islands along the margin of the Canada Basin (Sobczak and Weber, 1973), may include rocks of Sequence 4 as old as Carboniferous (Mississippian).

The Canada Basin seems to have been persistent and the lower part of the thick continental terrace wedge of sediments northwest of the Canadian Arctic Islands (Fig. 3) is tentatively assigned to Sequence 4. It is presumed to be equivalent to the lower part of the Sverdrup Basin. The Canadian Arctic region (Fig. 1) during the deposition of Sequence 4 (Figs. 4, 11, and 12), apparently had a history in which extensional structures were propagated southeastward from the Canada Basin, part way into the continental crust. This intermittent event, the Boreal Rifting Episode, apparently had a time span from late Paleozoic (Mississippian) to mid-Tertiary (Fig. 4). The active spreading center may have been the Alpha Ridge (Figs. 11 and 14). The boundary between great extension in the Canada Basin, and lesser extension within the continent appears to be the Kaltag Fault, which Norris (1974) showed as extending along the continental margin. Yorath and Norris (1975) suggest that it had right lateral displacement on the North American mainland. This author considers that it has been a transform fault, with intermittent pulses of both normal and strike slip displacement between late Paleozoic and Cretaceous-Tertiary time. Adjacent to the Queen Elizabeth Islands its strike slip displacement may have been left lateral (Fig. 11).

3. Relationship of Greenland to the Rest of North America

The history of Sequence 6 was particularly dependent on the relationship of Greenland to the rest of North America (Fig. 1). Theories to explain this relationship fall into three categories (Fig. 16). Most theories agree on the timing of tectonic events, but disagree on their mechanism.

(1.) At one extreme are fixist theories and related oceanization theories (Fig. 16A). These both maintain that continents have always had their present geographic locations relative to each other. Shatskiy (1935) first suggested that large blocks of the continent foundered and have been oceanized where ocean now exists. This theory was modified by Beloussov (1968, 1970), who also maintains that certain marine areas formed when continental crust was downfaulted and oceanized to form oceanic crust. They are opposed to lateral drift of continents. Another fixist theory, which maintains that both continents and oceans are permanent and fixed is held by Meyerhoff (1970, 1973).

(2.) At the other extreme are theories that maintain that continents have moved great distances apart, and that their present continental shelf edges formerly were adjacent. These suggest that Greenland drifted several hundred kilometers northward along a strike slip fault in Nares Strait (Fig. 16B).

This lateral drift theory was first suggested by Wegener (1924), whose concept of continent drift resulted in large part from observations on Greenland. Later proponents of continental drift (Carey, 1958; Wilson, 1963a, 1965b) made little modification to Wegener's view concerning drift of Greenland. Modern authors have built on and modified Wegener's theory, and call it plate tectonics. Wilson (1963b) first suggested a Mid-Labrador Sea Ridge, and the existence of such a ridge was later confirmed (Drake *et al.*, 1963). The Labrador Sea is also now known to have linear magnetic striping (Hood *et al.*, 1967; Hood and Bower, 1975; Srivastava, 1978). Nearly all present day workers who discuss tectonics of this area

(Wilson, 1965b; Pitman and Talwani, 1972; Keen, Barrett *et al.*, 1972; Ross *et al.*, 1973; Harland, 1973; Herron *et al.*, 1974; Keen, Keen *et al.*, 1974; Pelletier *et al.*, 1975; Hyndman *et al.*, 1973; Irving; 1977; Kristofferson and Talwani, 1977; Srivastava, 1978), suggest that the oceanic crust of Baffin Bay formed because Greenland drifted hundreds of kilometers laterally away from Baffin Island by a strike slip fault in Nares Strait. They essentially conclude that the edges of the present continental shelves were once adjacent, which is the Bullard fit (Bullard *et al.*, 1965).

The present writer refers to the group of reconstructions made or apparently favored by those papers, as the *conventional plate tectonic theory*. This theory maintains that the continental shelf edges of Greenland and Baffin Island were formerly juxtaposed (Fig. 16B), and that Greenland drifted to the northeast along a left lateral strike slip fault in Nares Strait. Wilson called this the Wegener Transform Fault (1963a, 1965a), and indicated (1965b) that it had about 350 km of right lateral motion. Recent workers who support this are Newman (1977), and Newman and Falconer (1978), who conclude that there was 250 km of strike slip displacement, the amount shown in figure 16B.

(3.) A third theory (Fig. 16C), intermediate between the extremes outlined above, was proposed over ten years ago by this writer (Kerr, 1967b, c). It suggests that there has been only a small to moderate amount of lateral movement of the lands flanking Baffin Bay and Labrador Sea. The oceanic areas between formed by a combination of: (a) movement apart of plates (or drift) by rotation, and (b) subsidence and oceanization of a very large intervening segment of continental crust. The writer *suggests that the fixist theory and the conventional plate tectonic theory both are partly right.* This new proposal contains elements of and in fact reconciles both earlier theories. The reconstruction of western Greenland by Le Pichon *et al.* (1977) is somewhat similar to that of the present author (Kerr, 1967c, and herein), except that it involves somewhat more closure of the Labrador Sea.

The author's theory (Fig. 16C), which will be enlarged upon in future papers, integrates contrasting theories. It agrees with published conclusions that Labrador Sea and Baffin Bay contain oceanic crust with linear magnetic striping. However, it disagrees with the conventional plate tectonic interpretation of how these submarine features formed. This new theory is a plate tectonic theory in that it considers that plates have moved apart. But it also is an oceanization theory, because it suggests that large parts of Baffin Bay and the Labrador Sea contain foundered and oceanized continental crust (Kerr 1967c). The author suggests that existing plate tectonic theories and fixist theories have both gone too far and both need moderating.

4. Baffin Bay

The geology of Baffin Bay has been described in numerous pertinent papers over the years by Oliver et al. (1955), Pelletier (1966), Kerr (1967c), Hood and Bower (1975), Murray *et al.* (1970), Clarke and Upton (1971), Keen, Barrett *et al.* (1972), McMillan (1972), Henderson (1973), Hyndman *et al.* (1973), Keen and Barrett (1973), Manchester and Clark (1973), Martin (1973), Ross (1973), Ross and Henderson (1973), Ross *et al.* (1973), Wallace (1973), Denham (1974), Wetmiller (1974), Beh (1975), Daae and Rutgers, (1975), Ross and Falconer (1975), Jackson *et al.* (1977), and Srivastava (1978).

McMillan (1972) summarized the depth to magnetic basement along the Labrador Shelf, Baffin Shelf, and area east of Devon Island. Along most of this coastline Precambrian Shield rocks crop out, generally in high sea cliffs. The depths to magnetic basement in the submarine areas indicate that there is a sedimentary wedge about 6000 m thick along the Baffin Island shelf. Thus Baffin Bay, whatever else it may be, is a deep depositional basin with a thick column of sediments. This column probably is largely assignable to Sequence 6 (Fig. 15). The crust beneath central Baffin Bay was discussed by Keen, Barrett *et al.* (1972), who showed that the Mohorovic Discontinuity in most places is at a depth of about 10 km. They considered that the deep central region of Baffin Bay is underlain by oceanic crust. They found no conclusive evidence for a buried ridge beneath the bay, nor were magnetic lineations or fracture zones delineated at that time.

Pelletier *et al.* (1975) showed that there is a broad oceanic area in central Baffin Bay and suggested that the bay was formed by sea floor spreading. They designated an ocean-continent boundary on the basis of gravity and seismic refraction data, and superimposed this boundary on the 1000 m isobath of Baffin Bay. Their suggested boundary is located close to that obtained from aeromagnetic profiles (Hood and Bower, 1975), and they diverged no more than 50 km. Pelletier *et al.* (1975) suggested that there is a 40 km gap between oceanic and continental zones which they suggest is occupied by nonmagnetic material. They reported thick sediments in Baffin Bay

Srivastava (1978) concluded from magnetic anomalies that seafloor spreading commenced in Baffin Bay in early Paleocene/ time (anomaly 24), and ceased during the early Oligocene (pre-anomaly 13). In Davis Strait there is a great column of volcanic rocks (Park *et al.* 1971), which also crops out on adjacent West Greenland (Munck and Noe-Nygaard, 1957; Rosenkrantz, 1970; Noe-Nygaard, 1974; Clarke and Pedersen 1976); and on eastern Baffin Island (Clarke and Upton, 1971). Pelletier *et al.* (1975) reported that the main oceanic crustal layer in Baffin Bay is only 4 km thick, and omitting the sedimentary column, is significantly thinner than the normal thickness of 7 km for the crustal layer of most mature ocean basins. They followed Keen and Barrett (1972), who suggested that Baffin Bay has an oceanic crust overlain by a thick sequence of sediments.

Pelletier *et al.*, (1975) suggested that geometrically the closure of Baffin Bay and Labrador Sea is effected in a reasonably satisfactory manner by two constructions that include rotation about a pole in Lancaster Sound, and translation along Nares Strait. Newman (1977) and Newman and Falconer (1978) support the suggestion that Greenland and Baffin Island have moved apart a great distance (Fig. 16B), because of their conclusion that there is 250 km of left lateral displacement in Nares Strait. This follows essentially Wilson's (1965b) reconstruction, and the conventional theory of plate tectonics.

5. Labrador Sea

The geology of the Labrador Sea has been described in numerous pertinent papers by Kranke (1947), H. Holtedahl (1958), O. Holtedahl (1970), Drake *et al.* (1963), Wilson (1963b) Kerr (1967c), Fenwick *et al.* (1968), Grant and Manchester (1970), Mayhew *et al.* (1970), Le Pichon *et al.* (1971), Grant (1972, 1975), McMillan (1972), Austin (1973), Hyndman (1973), Hood and Bower (1975), Athavale and Sharma (1975), Beh (1975), McWhae and Michel (1975), Van der Linden (1975a, b), Van der Linden and Srivastava (1975), Kirstofferson and Talwani (1977), Wade *et al.* (1977), and Srivastava (1978).

The depth to magnetic basement was summarized by McMillan (1972), who showed that there is a thick wedge of sediments, along the western margin of Labrador Sea, as much as 13 000 m thick. He suggested that the western continental shelf is underlain at depth by fault blocks that formed by growth faults over which the sedimentary column is draped. The part of this column that was deposited as active tectonism occurred is assignable to Sequence 6.

Srivastava (1978) concluded from magnetic anomalies that active sea floor spreading commenced in the southern Labrador Sea during the Campanian (anomaly 32) and in the northern Labrador Sea during the Maastrichtian (anomaly 28). The spreading ceased in the Labrador Sea as well as in Baffin Bay during the early Oligocene (pre-anomaly 13).

6. Nares Strait

A most crucial area to the interpretation of the origin of Baffin Bay and Labrador Sea is Nares Strait (Figs. 1 and 16), which separates Greenland from Ellesmere Island. There are two opposed views on the amount of displacement in Nares Strait.

(1) The conventional plate tectonic theory requires several hundred kilometers of left lateral strike slip faulting in Nares Strait. (Wilson, 1965b; Pitman and Talwani, 1972; Keen, Barrett *et al.* 1972; Ross *et al.* 1973; Herron *et al.* 1974; Keen, Keen *et al.* 1974; Pelletier *et al.* 1975; Hyndman *et al.* 1973; Kristofferson and Talwani, 1977; Srivastava, 1978; Newman, 1977; Newman and Falconer, 1978). The most recent work (Newman, 1977; Newman and Falconer, 1978), uses a figure of 250 km, and this displacement was used in the conventional plate tectonic reconstruction (Fig. 16B).

(2) The integrated theory of plate tectonics (Fig. 16C) of Kerr (1967b, c) interprets the strait as a submarine rift valley, that behaved partly as a transform fault system, but concludes that there was very little strike slip movement. Evidence for

and against strike slip movement in Nares Strait is presented below

Recent suggestions that there has been great strike slip movement in Nares Strait have resulted primarily from marine geophysics, and nearly all of this has been done outside of Nares Strait itself. The reasoning seems to be: (1) Baffin Bay contains oceanic crust; (2) oceanic crust forms by spreading apart of continental blocks; (3) the only zone by which Greenland can have spread laterally from Baffin Island is along Nares Strait; (4) therefore Nares Strait contains a left lateral strike slip fault.

Early workers (Koch, 1929; Troelson, 1950) showed geological maps that included both sides of Nares Strait. From the evidence in those maps it was reasonable to suggest moderate displacement, but not hundreds of kilometers. Later work at Nares Strait (Kerr, 1967b), suggested that there was no more than several kilometers of strike slip offset. Subsequently Dawes (1973) concluded that the present state of geological knowledge could not rule out displacement up to 250 km. Thus Keen, Barrett *et al.* (1972) learned by personal communication with Dawes that a displacement of 150 to 200 km could not be disproved by the geology. They considered this to support their view (Keen, Barrett *et al.*, 1972) that a reconstruction as in Figure 16B is the most reasonable. After considerably more field work along the entire Greenland coast bordering Nares Strait, Dawes presently considers (pers. com., 1978) that the geology suggests essentially no lateral displacement along Nares Strait, but that minor displacement in a left lateral sense cannot be discounted. The present author holds to his original conclusion (Kerr, 1967b), that Nares Strait had only minor rotational opening from the south, and minor strike slip displacement. This strike slip displacement, which is left lateral, could be as much as 15 kilometers. An amount between 0 and 15 kilometers probably cannot be resolved by comparison of facies belts.

The conventional plate tectonic reconstructions (Fig. 16B) recognize that the amount of displacement is a matter of debate, but they then forget that the geology can put restraints on it. No reasonable interpretation of the geology of the two sides of Nares Strait can permit the left lateral displacement required

in reconstructions made by conventional plate tectonic theories (Fig. 16B), i.e. about 250 km. The geology places an upper limit of about 15 km on the displacement That distance is the amount used in the present reconstruction (Fig. 16C).

7. Summary of Greenland Relationships

The three theories outlined above were applied to the Labrador Sea - Baffin Bay region where Sequence 6 is present, and the three maps (Fig. 16) represent reconstructions using the three methods. The theories of Beloussov (1970) and Meyeroff (1973) would maintain that the bordering lands have had fixed positions similar to those of the present (Fig. 16A). Alternatively, most field workers or those who have speculated on the origin of the Labrador Sea and Baffin Bay (Wilson, 1965b; Pitman and Talwani, 1972; Keen, Barrett *et al.*, 1972; Pelletier *et al.*, 1975; Kristofferson and Talwani, 1977; Srivastava, 1978), favor the conventional plate tectonic theory (Fig. 16B). It indeed appears to be true that Labrador Sea and Baffin Bay formed by rifting and represent a branch of the rifted North Atlantic Ocean. The existence of rifting, along with evidence of marine geophysics, including magnetic striping, led to the further conclusion that Labrador Sea and Baffin Bay formed by great lateral movement apart of Greenland from the rest of North America. The lateral movement normally suggested in Labrador Sea is about 650 km if the edges of the present continental shelves were formerly joined and the displacement on Nares Strait is about 250 km (Fig. 16B).

The author's integrated theory of plate tectonics (Fig. 16C) suggests that Labrador Sea and Baffin Bay formed by a combination of lateral separation of plates by rotation, with foundering and oceanizing of a large remnant of continent between them (Kerr, 1967c). The restoration to an original pre-drift configuration (Fig. 16C) follows this reasoning and suggested that the lands bordering Baffin Bay and Labrador Sea moved apart by an amount that increases from northwestern Baffin Bay to southwestern Labrador Sea. The amount of restoration is measured by the movement back of the present 1000 m isobaths, which are approximately at the shelf edges. Those lines are not brought back completely together as was done by Bullard *et al.* (1965), and in other conventional reconstructions.

This integrated theory does not disagree with any of the geophysical observations or evidence from Labrador Sea or Baffin Bay, or with the geophysical interpretation of those observations. It disagrees only with the geological interpretation of those geophysical observations. This theory involves the lateral separation of plates, and in a reconstruction brings them back to their suggested original positions. However it implies that the plates did not travel as far as conventional plate tectonic theory suggests.

The origin of the Baffin Bay - Labrador Sea area is intimately related to the origin of the Canadian Arctic Islands. The history of each area puts certain restrictions on the interpretation that can be allowed in the other area.

G. Eurekan Deformation

The Eurekan Deformation (Kerr, 1977a) was a tectonic event that formed much of the southeastern part of the Canadian Arctic Rift System between Late Cretaceous (Campanian/Maastrichtian) and mid-Tertiary (Miocene or possibly later) time (Fig. 4). The nature of this deformation is compatible with the reconstruction of Greenland made earlier (Kerr, 1967c), and outlined above (Fig. 16C).

The Eurekan Deformation includes two related and complementary structural phenomena. These are the Eurekan Rifting Episode, and the Eurekan Orogeny, which are respectively the extensional and the compressional phases of a plate tectonic episode. The Eurekan Rifting Episode (Kerr 1977a) originated in the southeast, and advanced northwestward by propagating extension faults or rifts, that resulted as Greenland and the rest of North America moved apart. The Eurekan Orogeny (Thorsteinsson and Tozer, 1970; Balkwill, 1978) was a

in the southeast being subjected to extension was separated from the region to the northwest being subjected to compression, with an area of transition lying between them.

H. Eurekan Rifting Episode

The Eurekan Rifting Episode (Kerr, 1977a) was the rifting phase of the Eurekan Deformation. The rifting was solely responsible for producing most of the faults and fault controlled channels in the southeastern part of the Canadian Arctic Rift System. It was responsible for intensifying the faults in the northwest part of the rift system that had been initiated by the Boreal Rifting Episode.

The Eurekan Rifting Episode began in the southeast in the area of the Labrador Sea and later affected Baffin Bay. It apparently formed those waterways by plate separation and foundering (Kerr 1967c). By Late Cretaceous (?Campanian/Maastrichtian time, as Greenland and the rest of North America rotated apart, the rifts or faults had been propagated northwestward into the Canadian Arctic Islands. Rifting in the Eurekan Episode was greatest at sea, but it also caused extension within the islands themselves.

Lancaster Aulacogen (Figs. 14 and 17) is a complex asymmetrical linear graben, and is the type of structure that was formed by extension during the Eurekan Rifting Episode. On the line of cross-section (Fig. 17) the vertical displacement on the Parry Channel Fault during the rifting episode was about 8200 m, and the aulacogen contains 6000 m of unconsolidated to semi-consolidated Cretaceous to modern sediments. Displacement and thickness both increase eastward from that line toward Baffin Bay, and decrease westward. The aulacogen may contain rocks assignable to Sequences 5, 6, 7, and 8. Before the aulacogen developed the region had a rather flat topography, with platform facies lower Paleozoic sediments (Sequence 3) exposed, presumably horizontal to gently northwest dipping. The Canadian Shield was the foundation of the continent, with remnants of a Proterozoic basin (Sequence 2) lying above it.

The aulacogen was dominated by the Parry Channel Fault, which extends along its north side and developed in the Eurekan Rifting Episode. It was first suggested by Wegener (1924) in his early work on continental drift, that Parry Channel was fault-controlled. The Parry Channel Fault and other faults in the Lancaster Aulacogen are supported by various types of geophysical work (Gregory *et al.*, 1961; Barrett, 1966; Daae and Rutgers, 1975).

Sequence 5 may be present beneath Lancaster Sound in the lower part of the wedge-shaped sedimentary body of Cretaceous and younger sediments there (Fig. 17). It presumably is a remnant of a sheet like deposit of the cratonic shelf (Fig. 12). Most of the semi-consolidated column in the aulacogen, however, may be part of Sequence 6, which was deposited as the aulacogen formed during the Eurekan Deformation. Faulted sediments overlain by undeformed sediments have been reported in Lancaster Sound (Jackson *et al.*, 1977), where the uppermost 2 km of the sedimentary section are not cut by major faults. The faults in Lancaster Aulacogen therefore originated at depth and were propagated upward. Some of these faults die out upward, so there are folds rather than faults near surface, and the upper part of the column is not folded. Lancaster Aulacogen (Figs. 14 and 17) is an indentation from the Baffin Bay Basin into the Canadian Arctic Islands. It is the type of structure formed in the major channels within the Canadian Arctic Archipelago by the Eurekan Rifting Episode.

I. Eurekan Orogeny

The Eurekan Orogeny (Thorsteinsson and Tozer, 1970; Balkwill, 1978) was a secondary phenomenon of the Eurekan Deformation. It developed in response to the Eurekan Rifting Episode, and was the complementary compressional phenomenon. Regionally the area of orogeny was apically exposed to the rifted area, with the two separated by a large area that constituted several transform pivots, by means of which extensional deformation was transformed into compressional deformation.

At the onset of the Eurekan Orogeny sedimentary and tectonic patterns changed rapidly in the Sverdrup Basin. That basin had undergone almost continuous subsidence and sediment accumulation for about 260 m.y., from late

Paleozoic onward. In Late Cretaceous (Campanian/Maastrichtian) time the Eurekan Orogeny began to fragment and fold this basin, and Sequence 6 began to be deposited.

(Balkwill, 1978) recognized three phases of tectonism that he assigned to the Eurekan Orogeny:

(1) regional uplift and erosion of the Sverdrup Rim and broad intra basin arches in latest Cretaceous and early Tertiary;

(2) compressive folding and faulting of the eastern part of Sverdrup Basin in the interval between middle Eocene and early Miocene; and

(3) rejuventated uplift of some arches in Miocene, and possibly Pliocene and later time. The first of these is herein assigned to a late phase of the Boreal Rifting Episode. The second is considered to be the Eurekan Orogeny, and the third is assigned to the final phase of the Eurekan Rifting Episode.

In the Sverdrup Basin the Eureka Sound Formation is a syntectonic deposit that was deposited during the first two events outlined above. The oldest part of the formation, Upper Cretaceous to lower Tertiary (Paleocene), was deposited adjacent to uplifted intrabasin arches (Fig. 15). A higher part (Paleocene) was more widespread and encroached onto those arches after there was a general collapse within the Sverdrup Basin. Still higher parts of the formation (Eocene and possibly Oligocene), were deposited contemporaneous with the compressive folding and faulting of the Eurekan Orogeny.

The Eureka Sound Formation (Fig. 15) is syntectonic, deposited during the first two of the above three phases. The height of land (Figs. 14 and 15) appears to have been a sedimentary source during most of Eureka Sound time, with detritus being shed northwestward and southeastward from it by rivers that fed into the interarch basins and graben controlled channels.

J. Sequence 6: Penetration of the Continent by Rifting

The Canadian Arctic Rift System is a major feature that includes structures in Labrador Sea, Baffin Bay, the Queen Elizabeth Islands, and that part of the Kaltag Fault lying northwest of the Queen Elizabeth Islands (Fig. 18). The Kaltag Fault is a boundary that separates the Canadian Arctic Rift System from other rifted structures farther northwest.

The Canadian Arctic Rift System severed the North American Continent and connected the Atlantic and Arctic Oceans via the Canadian Arctic Islands. The connecting process involved a number of stages and is summarized in four sequential figures (Figs. 18-21). Two plate tectonic events that emanated from different oceans contributed to forming the Canadian Arctic Rift System. The initial penetration of the continent by the rift system was by southeastward propagation of extensional faults of the Boreal Rifting Episode during deposition of Sequences 4 and 5. This extension reached only into the central Queen Elizabeth Islands and was aborted there, apparently because of interfering structural trends. Most of the Canadian Arctic Rift System was formed by the Eurekan Rifting Episode, whose main activity occurred during deposition of Sequence 6.

Sequence 6 of the Innuitian Mobile Belt (Figs. 3 and 4) was deposited during a time when penetration of the North American Continent by the Canadian Arctic Rift System increased greatly. Increased crustal fracturing occurred in the Sverdrup Basin, presumably due to an enlarged pulse of the Boreal Rifting Episode. The major event during deposition of Sequence 6, however, was the initiation and rapid development of the Eurekan Rifting Episode, which was propagated northwest from the Atlantic Ocean through Labrador Sea and Baffin Bay into the Queen Elizabeth Islands. The rift system advanced relatively quickly there and drastically altered depositional patterns (cf. Figs. 11 and 14); however it was not able to break fully through the continent until somewhat after.

In Early Cretaceous time, prior to the tectonic events that controlled deposition of Sequence 6 (Fig. 4), the Canadian Arctic was part of a rather stable continent (Fig. 18). Much of the present island and continent area was covered by very shallow seas, where thin sediments of Sequence 5 were accumulating (Fig. 12). The

water probably was shallow on the Sverdrup Rim, and to the southeast or the cratonic shelf, and was somewhat deeper in the Sverdrup Basin. The Canada Basin of the Arctic Ocean already existed as an oceanic basin at this time, lying northwest of the continent. It may have formed adjacent to a spreading center occupying the Alpha Ridge. Rifted branches that emanated from the Canada Basin presumably projected some distance southeastward into the Queen Elizabeth Islands. These had been formed earlier as a result of the Boreal Rifting Episode that was active during Sequences 4 and 5, and which continued to develop at this time (Fig. 18).

Prior to the Eurekan Deformation (Fig. 18) the present Arctic Islands and the continent to the south presumably were part of a continental crust of normal thickness (Fig. 3), that was made up of thick crystalline crust in its lower part (Sequence 1), and sedimentary basins in its upper part (Sequences 2, 3, 4, and 5). This crust contained within it at that time gross regional structural trends that had been actively forming in Paleozoic time, but probably were mainly inherited from an underlying Precambrian crystalline basement. Some of the structural trends that are known to have existed at the onset of rifting (Fig. 18) are those of the sedimentary cover or basement uplifts. It is inferred that there was an even stronger structural grain existing in the crystalline basement beneath those uplifts, with the same trends. The structural trends of the crystalline basement controlled the deformation that was produced by the Boreal Rifting Episode, as well as by the younger Eurekan Rifting Episode and the Eurekan Orogeny.

In Late Cretaceous time, the Canadian Arctic Islands region was subjected to two rifting events (Fig. 19). The Boreal Rifting Episode emanating from the Canada Basin had enlarged activity that caused the Pearya Geanticline and its continuation southwest as the Sverdrup Rim to be actively raised (Fig. 19, early). According to Meneley *et al.* (1975) and Balkwill (1978), a few thousand meters of erosion occurred on the Sverdrup Rim in Late Cretaceous time, prior to Eureka Sound deposition. Marine withdrawal from Sverdrup Basin proceeded southward at this time, prompted partly by the uplift and erosion along Sverdrup Rim. Two arches within the Sverdrup Basin were also raised and several thousand meters of erosion occurred on them (Fig. 15). Included were the Cornwall Arch (Balkwill, 1973), which may overlie a northward extension of the Boothia Uplift (Fig. 18), and the southward plunging ancestral Princess Margaret Arch (Balkwill *et al.*, 1975), which overlies the Rens Fiord Uplift. That part of the Canadian Arctic Islands that included the Sverdrup Basin was being deformed into linear highs and lows. They were expressed at the surface as anticlines, but probably were fault blocks at depth, bounded by extension faults that were being propagated southeastward into the Canadian Arctic Islands from the Canada Basin. As this rifting deformation progressed and the structural highs within the Queen Elizabeth Islands were raised (Fig. 19, early), there was syntectonic deposition in the lows of thick Upper Cretaceous to lower Tertiary sandstone of the Eureka Sound Formation. By late Paleocene time, the ancestral Princess Margaret Arch also was an active sedimentary source (Balkwill, 1978), that formed a major element of the height of land separating the Atlantic and Arctic watersheds.

The northeast trending Pearya Geanticline, which is a Precambrian cored, lower Paleozoic structure, may have controlled the location of the younger Kaltag Fault, and thereby also the location of the margin of the Canada Basin (Fig. 19). Faults of the Boreal Rifting Episode were able to breach the Pearya Geanticline, but the distance they penetrated southeastward into the continent also appears to have been controlled by older structural trends farther southeast (Fig. 19).

Simultaneous with the Boreal Rifting Episode (Fig. 19, early) incipient rifts also were being propagated northwestward into the continental crust of the Canadian Arctic from a pull apart margin in the southeast. This was the Eurekan Rifting Episode. This rifting probably began to affect the Baffin Bay area by about latest Cretaceous to early Tertiary time, when tectonism also caused deposition of continental rocks of the Eureka Sound Formation in the Eclipse Trough of northern Baffin Island (Jackson and Davidson, 1975; H.R. Balkwill, pers. com. 1978). In this stage, in Late Cretaceous time, there apparently was rifting in the southeast and northwest, with little or none in the region between.

Incipient rifting may have begun to form the future Baffin Bay Basin by Campanian/Maastrichtian time (Fig. 19, early). At approximately this time, separation apparently was further advanced in the Labrador Sea. The Labrador sea had now reached the stage where magnetic lineations were forming in an oceanic crust. Srivastava, (1978) reports that magnetic lineations formed in the southern Labrador Sea by Campanian time, and in the northern Labrador Sea slightly later, by Maastrichtian time.

By early Tertiary (Paleocene) time, faulting and spreading in Baffin Bay were sufficient so that magnetic striping began to form (Srivastava, 1978). A pivotal area presumably existed at this time in the southeastern part of the Queen Elizabeth Islands, as a first order transform pivot (Fig. 19, late). separation within Baffin Bay created rotation, which occurred about the transform pivot area in the northwest. A four armed junction of rifted zones, or a quadruple junction, may have begun to form at this time in northwestern Baffin Bay, and is still expressed in today's topography (Fig. 1). Increased separation in the southeast by Paleocene time may have caused the reduced rifting of the Boreal Rifting Episode in the northwest, and the widespread collapse of highs in the Sverdrup Basin. Upper parts of the sandy Eureka Sound Formation overstepped onto the flanks of the formerly high structures, onto the Sverdrup Rim in Late Cretaceous to early Tertiary time, (Balkwill, 1978), and onto the Cornwall Arch (Balkwill, 1974) in early Tertiary (Paleocene—Eocene) time (Fig. 15, and Fig. 19, late).

The Banks Island area was also fragmented by local uplifts in Late Cretaceous time (Miall, 1975). This fragmentation may have extended eastward into Parry Channel as far as the Bathurst Island Fault Zone (Fig. 19), which formed in Late Cretaceous (Maastrichtian) time (Kerr, 1974).

Balkwill (1973, 1978) attributes the Late Cretaceous-early Tertiary fragmentation of the Sverdrup Basin and elevation of intrabasin arches to a mechanism of crustal fracturing. This fracturing was most prevalent in structures on the northwest side of the study area near the northwest margin of the Arctic Islands, and diminished southeastward (Fig. 19).

The present writer infers that these tectonic events stemmed from a larger event in the Canada Basin, perhaps a pulse of renewed rifting. Contemporaneous fracturing took place in the southeast and diminished northwestward. Thus, in Late Cretaceous to early Tertiary time, rifting apparently took place simultaneously both northwest and southeast of the major height of land (Figs. 14 and 19).

It appears then that, in Late Cretaceous time (Maastrichtian and possibly Campanian, there were two episodes of extensional or rifting deformation being propagated toward the Queen Elizabeth Islands, one from the northwest, and the other from the southeast (Fig. 19).

In the northwest fragmentation by rifting in Late Cretaceous time (Fig. 19, early), initiated marked uplift and erosion of Pearya Geanticline, Sverdrup Rim, and structures within the Sverdrup Basin (Balkwill, 1978), and raised the Storkerson Uplift (Miall, 1975). By latest Cretaceous to early Tertiary (Paleocene) time, rifting had become more advanced in the southeast, and discontinued in a large region in the northwest where there was then general collapse (Fig. 19, late). This led to widespread early Tertiary sedimentary overlap of some of the formerly uplifted structures by the Eureka Sound Formation (Fig. 15). The Pearya Geanticline remained high on Ellesmere Island, for the lower Tertiary strata on Ellesmere Island become progressively finer and more marine southward, suggesting a northerly source prograding southeastward (Bustin, 1977). This high may have been related to the proximity of a spreading center nearby to the north, occupying the Alpha Ridge.

The first order transform pivot developed southeast of a prominent pre-existing structural trend within the Queen Elizabeth Islands (Fig. 19, late). It appears that those trends impeded the northwestward advance of extensional faults of the Eurekan Rifting Episode, thereby controlling the location of the pivot. Subsequent rotation about that pivot produced the compression of the Eurekan Orogeny (Fig. 20).

The Eurekan Deformation reached its climactic phase (Fig. 20) in mid-Tertiary time (middle Eocene to early Miocene). In this stage, spread continued in Baffin Bay, and the rifts were propagated farther northwest. The advance of the rifts northwestward into the Queen Elizabeth Islands was impeded by the pre-existing structural trends (Fig. 18), some of which were transverse to their paths and deflected their propagation (Fig. 20). One rift zone was propagated westward into Lancaster Sound to begin the formation of Lancaster Aulacogen (Figs. 14 and 17). It was then further deflected southwestward by the Boothia Uplift. Another rift zone was propagated northward into Nares Strait but its advance also was impeded (Kerr, 1967b). Nares Strait appears to have had only minor strike slip displacement at this time, perhaps 15 kilometers (Fig. 20, enlargement).

At this time Baffin Bay continued to develop as a rifted basin, with the blocks on its sides that were in future to form Greenland and Baffin Island, rotating apart. A region in the northwest was subjected to compressional deformation, because the blocks bordering Baffin Bay continued to rotate apart yet, the rifts could not propagate to the northwest, northeast, or west. This situation produced the Eurekan Orogeny (Fig. 20). As the two plates that now contain Greenland and Baffin Island moved apart slightly, a broad region of continental crust between them apparently foundered and was oceanized, according to Kerr (1967c).

The climactic phase of the Eurekan Deformation (Fig. 20) involved compressive folding and faulting in eastern parts of the Sverdrup Basin and in the underlying rocks, in middle Eocene to early Miocene time, while extension continued in the Baffin Bay Basin. Thus for a time extension in the southeast (Eurekan Rifting Episode) apparently was coeval with compression in the northwest (Eurekan Orogeny). There was an area of transition between. This writer suggests that the extension in the southeast was the cause of the compressional deformation in the northwest. The effects of compressional deformation are most obvious in the Sverdrup Basin (Balkwill, 1978), because the existing rocks there were largely undeformed when compression began. Nevertheless compressional deformation of the Eurekan Orogeny also affected older rocks of the Franklinian Geosyncline and Arctic Platform that had been deformed previously (Kerr, 1967b; Kerr and Thorsteinsson, 1972). In those older rocks the effects of the Eurekan Orogeny are difficult to separate from earlier deformation. Eurekan deformation in the Sverdrup Basin (Fig. 20) was controlled in part by the regional stresses that were applied at the time, but also to a very large degree by the pre-existing structural trends within and beneath that basin (cf. Figs. 18 and 20). The rifts propagating northwestward were obstructed on their advancing ends by ancient structural trends (Fig. 20), yet opening continued at their older ends. This combination of events, of which the obstruction was a vital part, produced the compressional deformation of the Eurekan Orogeny (Kerr, 1967b, 1967c).

The extensional deformation that had been propagating southeastward from the Canada Basin in Late Cretaceous time (Fig. 19, early), apparently was later neutralized throughout a large region in latest Cretaceous to early Tertiary time by compressional effects that advanced northwest from the Eurekan Rifting Episode (Fig. 19, late). Subsequently (Fig. 20), this very widespread neutral state was soon replaced in a large part of the eastern Arctic by compressional deformation, and the neutral area became smaller and shifted westward.

The writer infers that the compressional deformation of the Eurekan Orogeny (Fig. 20) produced a general squeezing rather than a unidirectional overriding. It resulted from compression in the northwest that was apically opposed to the rotational opening of Baffin Bay and Labrador Sea in the southeast.

The youngest compressive deformation by the Eurekan Orogeny known on land is bracketed between middle Eocene and lower Miocene rocks on Axel Heiberg Island (Balkwill and Bustin, 1975). These are respectively Sequences 6 and 7 (Fig. 4).

Apparently the extensional deformation propagating to the northwest from Baffin Bay was stronger than that propagating to the southeast from the Canada Basin. The former compensated for and surpassed the latter in the broad area where compressional deformation occurred. The net effect in eastern parts of the Canadian Arctic Archipelago was that the compression resulting from the Eurekan Rifting Episode took over in the Sverdrup Basin area as the Eurekan Orogeny (Fig. 20). The central part of the Arctic Archipelago appears not to have been affected by the compressional deformation, which apparently diminished and died out toward the northwest.

All rocks younger than Maastrichtian (latest Cretaceous) are essentially unfolded along the Arctic Coastal Plain northwest of Ellef Ringnes Island (Meneley *et al.*, 1975), and there are have been continuous sedimentation there through the time span of the Eurekan Orogeny. This therefore lay in the neutral area (Fig. 20). Parts of the Canada Basin northwest of the area affected by compressional deformation may also have been neutral (i.e. no deformation) through the time span of the Eurekan Orogeny. In certain parts of the western Arctic, however, extension apparently existed without interruption throughout the Eurekan Deformation (Fig. 20). In the Banks Island area there was differential uplift and faulting from late Maastrichtian to Eocene time, resulting in deltaic sedimentation, without any interruptions by compressional deformation (Miall, 1975). The writer suggests that this faulting resulted from side effects of extensional processes that continued to operate in the Canada Basin through the time span of the Eurekan Orogeny. The Eurekan Orogeny was unable to interrupt and dominate the extensional faulting effects on Banks Island emanating from the Canada Basin (Fig. 20).

K. Sequence 7 (Miocene) Severing of the Continent by Rifting

Sequence 7 comprises the Miocene Beaufort Formation and equivalent rocks (Figs. 3 and 4). The sequence is exposed in two regions where the setting is very different. Deposition was partly contemporaneous with the final severing of the continent that was the last phase of the Eurekan Deformation.

In the Arctic Coastal Plain (Fig. 2) Miocene sands of the Beaufort Formation are undeformed. The formation lies unconformably upon folded older rocks and dips northwestward (Thorsteinsson and Tozer, 1970; Hills et al., 1974; Matthews, 1976), and apparently was not affected by tectonism of the Eurekan Deformation. It is part of a continental terrace wedge that thickens northwestward on northwest Ellef Ringnes Island. The base of this wedge is older than the Beaufort Formation, and farther northwest in the subsurface of the Arctic Coastal Plain it includes unfolded Upper Cretaceous rocks at the base (Meneley *et al.*, 1975). Thus the Eurekan Orogeny did not affect that area with compressive deformation, and there may have been essentially continuous deposition in the offshore area from Late Cretaceous (Eureka Sound) to Miocene (Beaufort) time as the Canada Basin subsided and grew.

Miocene rocks also occur on Axel Heilberg Island (Balkwill and Bustin, 1975) and northern Ellesmere Island (D.G. Wilson, 1976). In both places they were considered equivalent to the Beaufort Formation, and in both places these Miocene rocks are faulted conglomerates. These faulting events, which appear to be contemporaneous, are the youngest dated events of extensional faulting in the Arctic Islands.

Uplift, erosion and faulting of the Eurekan Deformation continued in the eastern Arctic in Miocene and possibly later time, because the Beaufort Formation on northern Ellesmere Island and on the Princess Margaret Arch was uplifted and faulted. These late faults may have been part of the final pulse of extensional faulting that finally severed the continental crust (Fig. 21), or adjustments to that pulse.

The Beaufort Formation of the Arctic Coastal Plain is not markedly affected by the post-Miocene faulting, but the Beaufort rocks of Axel Heilberg and Ellesmere Islands are. This is further evidence that the Eurekan Orogeny affected the eastern region but left the area to the west neutral (Fig. 20). This also suggests that the final breaking through of rifts to reach the Arctic Ocean was by means of the eastern region, including Nares Submarine Rift Valley more than by Parry Channel. In the western Arctic final extension may have been confined to the major fault zones of Parry Channel, which are not exposed.

Because lower Miocene rocks are faulted in the eastern Queen Elizabeth Islands, this indicates that the Canadian Arctic Rift System was still active. The pulse of activity in early Miocene or later time may have been its last. Because the rifting emanated from the southeast, it is probable that Baffin Bay, Nares Strait, and Parry Channel were also affected by that faulting event. It is likely that in those three waterways there will be a Miocene succession (Sequence 7) equivalent in age to the Beaufort Formation. If so it may be bounded by stratigraphic breaks. The lower break that separates it from Sequence 6 may result from the climactic event of the Eurekan Orogeny (Fig. 20). The upper break that separates it from Sequence 8 may result from the event that finally severed the continent (Fig. 20). The Eurekan Deformation ceased activity in mid-Tertiary (early Miocene or later) time, after extensional deformation breached the continent and reached the Arctic Ocean (Fig. There apparently has been no strong tectonism since then in the study region (Fig. 1).

The final event in the Eurekan Deformation was widespread extensional faulting that broke through the Queen Elizabeth Islands (Fig. 21). The Nares Submarine Rift Valley (Kerr, 1967b), whose advance had been impeded previously, continued to develop at this time and broke the rest of the way northward through the continent to connect with the Arctic Ocean. A westward projection of Lancaster Aulacogen (Kerr, in press) similarly broke through the Boothia Uplift and continued the rest of the way through the continent to form Parry Submarine Rift Valley and reach the Arctic Ocean.

The Miocene rocks of Sequence 7 making up the Arctic Coastal Plain (Fig. 1) have not been faulted (Thorsteinsson and Tozer, 1970), so the final event of the Eurekan Deformation may not have affected the Arctic Coastal Plain. It appears that faults in Parry Submarine Rift Valley broke westward through the Boothia Uplift at this time. In doing this they may have simply connected with existing older faults that had early formed the west end of that rift valley, but without substantially increasing the displacement on those western faults (Fig. 21).

With this final faulting event of the Eurekan Deformation a large triangular region became severed from the rest of the continent to become the completely fault bounded Queen Elizabeth Islands Sub-plate (Fig. 15). This breaking through by extension faults (Fig. 21) produced the final structural connection between the Atlantic and Arctic Oceans. It may have occurred contemporaneously with the rejuvenated uplift and faulting of the Princess Margaret Arch on Axel Heiberg Island, which cuts Sequence 7, early Miocene rocks (Balkwill and Bustin, 1975; Balkwill, 1978) That is the youngest extensional event documented in the Queen Elizabeth Islands, and is interpreted as a side effect of the main extensional pulse that produced the final breakthrough of rifted structures to reach the Arctic Ocean. Thus the structural connection of the Atlantic and Arctic Oceans apparently was achieved in early Miocene time or slightly later. The Alpha Ridge is not an active spreading center today. Its activity may have ceased when the continent was breached by the Eurekan Rifting Episode.

The severing of the Canadian Arctic by faults created the basic structural and physiographic configuration of the Arctic region (Figs. 1 and 21). Geological structure and tectonics controlled physiography on both very large and intermediate scales. The Canada Basin of the Arctic Ocean and the seaway forming Baffin Bay and the Labrador Sea were downfaulted as sedimentary basins, and are also deep physiographic basins. The channels within the Canadian Arctic Archipelago that connect the two main oceanic areas were downfaulted a lesser amount, and are now much shallower than the oceanic basins. The major channels bordering the Queen Elizabeth Islands contain rift valleys and are sub-plate boundaries. These rifts valleys are structural projections of the oceanic areas into the continent.

The Queen Elizabeth Islands Sub-plate also is made up of a number of lower order sub-plates that are separated by faults (Fig. 21). The faults controlled the shapes of the sub-plates, which in turn determined the shapes of the present day islands (Fig. 1).

The present study concludes that the Canadian Arctic Rift System formed the Labrador Sea, Baffin Bay, and the Queen Elizabeth Islands between Late Cretaceous (Campanian/Maastrichtian) and mid-Tertiary (early Miocene or later) time. The main rifting was part of the Eurekan Deformation, which was propagated northwestward from the Atlantic Ocean. It met with, overpowered, and aborted the Boreal Rifting Episode, which was of lesser strength and that simultaneously was being propagated southeastward from the Canada Basin of the Arctic Ocean. The separation of Greenland from the rest of North America apparently occurred by a combination of rotation apart of sub-plates and foundering of a continental blocks between. This occurred because of propagating faults that advanced northwestward. The nature and timing of the events in the separation and rotation of Greenland ~~events~~ are similar to those suggested in earlier papers (Kerr, 1967b, c). The timing conforms with that determined later from marine work in Labrador Sea and Baffin Bay (Pitman and Talwani 1972; Srivastava, 1978). There appears to have been no closing of Labrador Sea or Baffin Bay. Once they began to open they either continued to open or stopped, but did not close again.

The Haughton Astrobleme (Frisch and Thorsteinsson, 1978) is an impact crater of presumed meteoritic origin. Lake sediments of Miocene or possibly Pliocene age that lie in the crater. They were deposited soon after impact, which apparently occurred in about Miocene time, perhaps contemporaneous with the Beaufort Formation and Sequence 7. By the time of impact that part of Devon Island was above sea level and the present peneplain of exposed lower Paleozoic rocks had developed.

L. Sequence 8: Mid-Tertiary (post-Miocene) to Present Day Development

The Eurekan Deformation may have ceased rather suddenly after the early Miocene or later faulting event, (Fig. 21), for it appears that the Canadian Arctic Rift System has been dormant since then. There is very little evidence of faulting since that event. Several long lineaments occur in the Beaufort Formation of the Arctic Coastal Plain on Prince Patrick Island. These must be very recent faults because they create scarps in the modern soft weathering landscape (Tozer and Thorsteinsson, 1964; Thorsteinsson and Tozer, 1970). These are presumed to be faults, with a few feet of displacement which may be minor readjustments on older structures.

Fault activity after the Eurekan Deformation may have been more prevalent in the marine area, where the major zones of weakness of the Canadian Arctic Rift System exist. This system could have been reactivated there by adjustments to more global plate motions. The region of the Queen Elizabeth Islands has been very stable tectonically or partly in extension since the final faulting of the Eurekan Deformation, that is from about mid-Miocene onward, or after about 20 m.y. ago. During this interval the present marine areas have probably changed shape very little. They probably also have been water-covered continuously, except for near shore areas that are subjected to the vagaries of submergence and emergence due to glacial loading, isostasy, or eustatic sea level changes. There probably is no major post-Miocene stratigraphic break in the sedimentary sequence in the main channels. Whether deposition or erosion occurred in various parts of these basins since the end on the Eurekan Deformation probably depended on currents, sediment supply, and other factors, but probably

not on active tectonics. It seems likely therefore that the sedimentary column extending from the end of the Eurekan Deformation (post-early Miocene) will not have been interrupted by major tectonic events, and therefore should make-up a single sequence at sea (Sequence 8, Figs. 3 and 4). This sequence also will include deposits related to Pleistocene glacial events.

It may be impossible to separate Sequences 7 and 8 in much of the western Arctic, beneath and northwest of the present Arctic Coastal Plain (Figs. 2 and 4), because the Beaufort Formation there apparently was not deformed by the Eurekan Deformation (Fig. 21). That area may be like Banks Island (Miall, 1975), where the middle to upper Miocene Beaufort Formation is part of a major deltaic assemblage that continued to accumulate to the present day (Fig. 13).

The Canadian Arctic Rift System is now a nearly dormant structure within the North American Plate, as there is little seismic activity on it (Basham et al., 1977). There are numerous earthquakes, but their patterns indicate that tectonic forces characteristic of plate margins are not acting directly within the Canadian Arctic today. The present seismic activity may be mainly an expression of readjustment of existing structures, mainly the Canadian Arctic Rift System, to a regional stress field. Little plate tectonic activity is taking place now in northern Canada (Fig.1), because its internal rift system is dormant and the entire region is travelling as a part of the North American Plate. The overall rotation of that plate may be localizing any seismic activity within the study region along dormant sub-plate boundaries.

From the end of the Eurekan Deformation in mid-Tertiary (probably Miocene) time to the present, the Canadian Arctic (Fig. 1) has been subjected to erosion in some places and deposition in others, but not to strong tectonism. Sequence 8 was deposited in this time span and is still accumulating. The present day physiography was mainly achieved by the end of the Eurekan Deformation. Since then it has only been modified further by the sculpturing and infilling of erosional processes. The present physiography therefore resulted largely from the interplay of two things. It was dominated by the geological structures that had been produced by the Eurekan Deformation and by other similar structural events to the northwest. It was also modified by erosion in the period following the Eurekan Deformation. Structure exerted vastly greater control on regional physiography than did erosion.

Structural control of physiography applies mostly to larger features, e.g. the main linear margins. The Canada Basin is topographically low because it appears to have been largely downfaulted relative to the Arctic Islands. The Queen Elizabeth Islands Sub-plate (Fig. 21) forms a triangular island group (Fig. 1), because it is rift bounded. It is bordered on the south and east by submarine rift valleys. On the northwest the island group is bordered by the Kaltag Fault (Norris, 1974, Yorath and Norris, 1975; Norris and Yorath, 1979), which apparently controlled the continental margin there.

Baffin Bay and Labrador Sea apparently were largely downfaulted to become deep waterways. Simultaneous with and related to that downfaulting, bordering lands, Greenland, Baffin Island and Labrador, were raised, with high coastal mountain ranges developing along them. The main marine indentations into the Queen Elizabeth Islands group, such as Jones Sound, Hassel Sound, and others, appear to be grabens. The islands may have risen as horsts in conjunction with the formation of those grabens and as part of the same process.

The structural control of physiography applies also to structures of intermediate size, that are large enough to be shown on regional maps. Cornwallis Island for example is anticlinal (Thorsteinsson and Kerr, 1968), as are Cornwall and Amund Ringnes Islands (Balkwill, 1974: 1978). Ellef Ringnes Island may also be grossly anticlinal (Stott, 1968), with a structural low forming the channel to the east. Somerset Island (Fig. 21) is basically a horst, being almost completely surrounded by normal faults (Kerr and deVries, 1977).

A major height of land trending northeastward through the Canadian Arctic, which still exists, was produced largely by tectonic events during the Eurekan Deformation (Figs. 14 and 15). It is a mainly continental area lying between oceanic basins. The height of land was broken by narrow faulted channels during the Eurekan Deformation, when parts of it were deeply depressed to allow the oceans on either side to connect. The height of land has persisted since the Eurekan Deformation, and the narrow downdropped structures may have remained below sea level continuously since then. Since the Eurekan Deformation, the height of land has been modified by surficial processes operating both above and below sea level.

Fortier and Morley (1956) first suggested that the physiography of the Canadian Arctic Islands was produced by a combination of early Tertiary tectonic events and fluvial erosion, with the present marine channels being a structurally controlled former subaerial drainage system. That concept was enlarged by Pelletier (1966), Trettin *et al.* (1972), and Kerr (1977a). The writer considers that the original concept of Fortier and Morley (1956) remains largely correct. He suggests however, that the main channels may have been below sea level from an early stage of their formation, and that they were controlled only to a small degree by subaerial erosion. This applies particularly to the deeper or downstream parts of the channels, which were submerged in an early stage of tectonism, and may have remained below sea level ever since.

The Innuitian Ice Sheet (Blake, 1970, 1977) covered much or all of the Queen Elizabeth Islands region in Pleistocene time. This contributed to the sculpturing of the region, by preferentially scouring the valleys on land, and perhaps also the shallow submarine areas, including the fiords, and the shallower marine channels. Glaciation was followed by widespread post-glacial rebound. The amount of rebound appears to have been greater in the central Queen Elizabeth Islands than farther east, for the upper marine limit ascends westward on southern Ellesmere Island (Blake, 1970, 1975). It similarly increased westward on Devon Island (Kerr, 1977d).

With the retreat and melting of the glaciers, the Canadian Arctic Islands have emerged, such that there are marine raised beaches around many of the islands. The rate of post-glacial uplift in the Innuitian region was on the order of 1 to 2 mm per hundred years during the last six hundred years (Andrews, 1970), but this varies greatly from place to place. The rifting process that produced the Arctic Islands is now nearly dormant, the earthquake activity in the region is mild, and probably is limited to adjustments on existing structures (Basham *et al.*, 1977).

M. Mineral and Petroleum Deposits

1. Cornwallis Lead-Zinc District

The Cornwallis Lead-Zinc District is located within lower Paleozoic rocks in the northern part of the Cornwallis Fold Belt (Figs. 6 and 22). The lead-zinc deposits of that district all have a similar setting (Kerr, 1977b, c) and appear to have formed by a common mechanism. They were controlled by the tectonic history of the Boothia Uplift, having formed in the upper or sedimentary level of that uplift (Fig. 8).

The lead-zinc deposits characteristically occur in an Ordovician carbonate formation, in anticlines that were formed by the Cornwallis Disturbance. They are of Mississippi Valley-type, apparently having formed in caverns at low temperatures (as high as 105° C), by precipitation from formational fluids.

An early stage in the formation of the lead-zinc deposits was cavern formation (Fig. 22A). These caverns formed in the Ordovician Thumb Mountain Formation, and preferentially in the upper fossiliferous part of that formation. That formation normally is limestone, but is dolomitized in the vicinity of the lead-zinc deposits. The caverns were presumed by Kerr (1977b) to have formed in Early Devonian time, by karst solution during Pulse 3 of the Cornwallis Disturbance. Callahan (1978, 1979) suggested that they may have formed in an earlier erosional event, but this was considered to be unlikely by Kerr (1978, 1979). Whichever timing is correct, it appears clear that the deposits were controlled by the tectonic history of the basement uplift within the geosyncline (Figs. 8 and 9).

Mineralization occurred in the caverns, probably in Late Devonian time, apparently by precipitation when two appropriate fluids met (Fig. 22B). One of the fluids may have originated in a nearby black shale, obtaining one or more of the mineral components there, most probably the sulphur. The other essential components, metal ions, probably were carried in a second fluid that ascended from beneath. This second fluid may have originated in the deep crystalline basement of the Boothia Horst, which was being uplifted from time to time.

Various tectonic and erosional events that occurred subsequent to mineralization contributed to exposing the deposits. The Eurekan Rifting Episode was a major faulting event that formed the main channels in the Canadian Arctic, and occurred long after mineralization. It produced the Crozier Strait Fault Zone (Kerr and Ruffman, 1979), and was responsible for exposing the deposits on either side of Crozier Strait (Fig. 22C). This fault zone followed the structural trends of the Boothia Uplift that had been established during the Cornwallis Disturbance.

2. Oil and Gas Discoveries

Oil and gas Discoveries in the Arctic Islands have been summarized by Meneley (1976) and by Stuart Smith and Wennekers (1977). Source rock and maturation studies have been carried out by Snowdon and Roy (1975), Baker et al. (1975), and Powell (1978).

On Cameron Island (Figs. 1 and 6), live oil was discovered in two wells in Middle Devonian carbonate rocks that are part of a large isolated reef buildup. A subsequent follow up well has been abandoned (Stuart Smith and Wennekers, 1977), suggesting that this may not be an important discovery. The location of the reef buildup is shown in Figure 9, but its present setting is much more complicated because of subsequent deformation.

Exploration to 1977 had discovered 15 trillion cubic feet of gas in seven gas fields in the Western Sverdrup Basin, according to Meneley(1976). The fields are in two areas, and appear to be in approximately the same stratigraphic level in a Triassic-Jurassic sandstone body. One area is in the axial part of the Sverdrup Basin (Fig. 11) in the region of Ellef Ringnes Island and King Christian Island. These are related to diapiric structures (Figs. 12 and 23), and are in a sandstone unit that overlies and was raised by those evaporite diapirs. These gas bearing structures are ovate anticlines, with cores of evaporites which are not diapiric (Balkwill and Roy, 1977). The second area of gas fields is along the southern margin of the Sverdrup Basin on Melville Island (Fig. 11) and the nearby offshore region. These are the Sabine Peninsula fields (Fig. 23). They occur beneath a gentle regional angular unconformity (Meneley, 1976).

The size, shape, and mechanism of entrapment of gas were very different in the two regions (Meneley, 1976). The pools in the Ellef Ringnes Island region have on the order of 150 m of gas, and are about 8 km in average breadth. In the Sabine Peninsula area of Melville Island the pools are thinner, being on the order of 45 m thick, but are much wider, being on the order of 40 km across. About 80% of the gas reserves reported occur in the Sabine Peninsula fields (Fig. 23).

IV. SUMMARY AND DISCUSSION

The Canadian Arctic is a transition between two ocean basins, the Atlantic Ocean in the southeast and the Arctic Ocean in the northwest. The region of this transition that is most like a true continent geologically also is highest topographically. This is a northeast trending height of land that formed in Cretaceous-Tertiary time (Figs. 14 and 15), and persisted to the present, being expressed now by shallow bathymetry (Fig. 1). The height of land is the region that was least fragmented by the ocean forming plate tectonic process.

The Canadian Arctic is transitional first in a structural sense, with marginal parts of the Arctic being most oceanic in character and central parts of the Queen Elizabeth Islands Sub-plate (Fig. 15) most continental. Labrador Sea is rather like a deep oceanic basin, with a crustal structure that resembles the oceanic crust of a major ocean. It had moderate opening or separation of the two sides, also resembling a rifted ocean. Taking a step from there northwestward into Baffin Bay the crust is oceanic, but perhaps quasi-oceanic, and even less like the deep abyssal parts of a true ocean basin than is the crust of Labrador Sea. Magnetic lineations are poorly developed in Baffin Bay (Srivastava, 1978). It has very thick sediments (Pelletier *et al.*, 1975) and at least parts in the northwest have a crust that is intermediate between continental and oceanic crust (Wetmiller, 1974). In the region still farther northwestward, the Canadian Arctic Archipelago has a crust that is largely continental in character, with a thickness of about 37 km (Sobczak and Weber, 1973), but it has deep grabens and fissures in it as a result of crustal fracturing and extension.

A somewhat similar progression occurs from the Canada Basin of the Arctic Ocean southeastward to the central part of the Queen Elizabeth Islands Sub-plate. The Canada Basin is a deep oceanic basin with a large region deeper than 3500 m. The crust beneath it is of oceanic or quasi-oceanic character. It has certain geophysical attributes of continental crust (King *et al.*, 1966), but is far below sea level, approaching depths of an abyssal plain. Farther southeastward is a continental slope and shelf, with a thick wedge of sediments that forms a continental terrace wedge (Fig. 3). Still farther southeastward are the Queen Elizabeth Islands, still more continental, and with a crust of about 37 km thickness.

The Canadian Arctic is transitional between the Atlantic and Arctic Oceans in a physiographic sense also. A central height of land in the Arctic Archipelago trends northeastward, expressed in topography and bathymetry. This is a part of the larger Queen Elizabeth Islands, parts of which are above, and parts below sea level as relatively shallow channels between those islands, The physiographic transition exists also on either side. Southeast of the Queen Elizabeth Islands there are progressively deeper basins, first Baffin Bay, then Labrador Sea, and finally the Atlantic Ocean proper. The physiographic transition also occurs northwestward from the Queen Elizabeth Islands toward the Canada Basin of the Arctic Ocean. The physiographic transition was produced by the structural transition.

The Canadian Arctic Islands region (Fig. 1) lies between two immature oceans, and it apparently resulted from side effects of the processes that developed those two oceans. Both oceans resulted from breaks in a primeval continental crust. Both were controlled in important ways by the structural trends in the Precambrian crystalline shield of the continent within which and beside which they formed. The study region (Fig. 1) is that region where the two ocean forming events impinged on each other. Processes were different in nature and timing in the two oceanic areas, and this resulted in very different types of continental margins.

The Arctic Ocean margin of the Canadian Arctic Islands began its development very early. The trend of the margin can be traced back to the beginning of the Innuitian Continental Margin-type Mobile Belt (Fig. 3). That belt began to form along a northeast trend from the Banks Island region to northern Greenland, in Late Precambrian time. It may have followed structural trends already existing in the older crystalline rocks beneath that make up Sequence 1. The Innuitian Mobile Belt developed through several stages, each depositing a huge stratigraphic sequence. These sequences lie one above the other and are imbricated, the oldest being in the southeast and the youngest in the northwest. Sequence 2 of Proterozoic age, and Sequence 3, the Late Proterozoic to Late Devonian Franklinian Geosyncline, have northeast trends. They either established this trend during deposition, or more likely reinforced an older one. The northeast structural grain is now followed by the Arctic Ocean facing margin of the Canadian Arctic Islands. The history up to the end of Sequence 3 was the constructional phase of the continent (Fig. 4). It involved a linear geosyncline that may have lain within and upon a continental crust. It apparently did not involve the formation of a true ocean.

The fragmentation phase of the continent began in late Paleozoic time, probably Mississippian (Fig. 4), and has continued to the present. It formed the Canada Basin, Baffin Bay, and the other marine areas within the North American Continent (Fig. 1). The Canada Basin of the Arctic Ocean may have began to form during Sequence 4 by an unnamed rifting episode centered on the Alpha Ridge, that began shortly after the Ellesmerian Orogeny. The Arctic facing margin of the study area apparently was guided by the Pearya Geanticline. The Canada Basin appears to have formed by rifting, and perhaps also foundering of continental crust. Lesser rifting marginal to it extended into the continent and apparently produced the Sverdrup Basin. A major fault zone lies between. There may have had great down dropping on the northwest of this fault and lesser downdropping on the southeast. Therefore it must also have had a transform component, presumably left lateral. In Early Cretaceous time, as Sequence 5 was deposited, a new pulse of activity in the Canada Basin caused accelerated rifting which extended farther southeastward across the main fault zone and into the continent. This rifting event faulted the northwest marginal parts of the Canadian Arctic Islands, simultaneous with a great continent wide overlap of shallow seas southward and southeastward onto a cratonic shelf.

In Late Cretaceous time, the Canadian Arctic Rift System began to form more rapidly, mainly by activity in the southeast. The rift system is a branch that extended northwestward from the Mid-Atlantic Ridge through the Canadian Arctic, and separated Greenland from the rest of North America. Major rifting apparently was propagated northwestward by extension faults that produced rotation about a pivotal area in the northwest. For a time the two rift systems were both active, propagating toward each other. The extension that was being propagated to the northwest was the strongest. It first nullified the other rifting, subsequently caused compressional deformation, and later managed to break through the Canadian Arctic Islands to reach the Arctic Ocean. The active rifting that connected the Atlantic and Arctic Oceans stopped in mid-Tertiary time, about early Miocene or later, apparently as suddenly as it began. This suggests that the process of connecting the two oceans structurally ended the need for any further major rifting activity.

The Canadian Arctic Rift System is a major, continent-wide rift traversing a continent, that became dormant or frozen in a very early stage of its development. It contains a well exposed geologic record of the rocks from Precambrian to Recent time. These characters make it an excellent laboratory for the study of certain tectonic processes.

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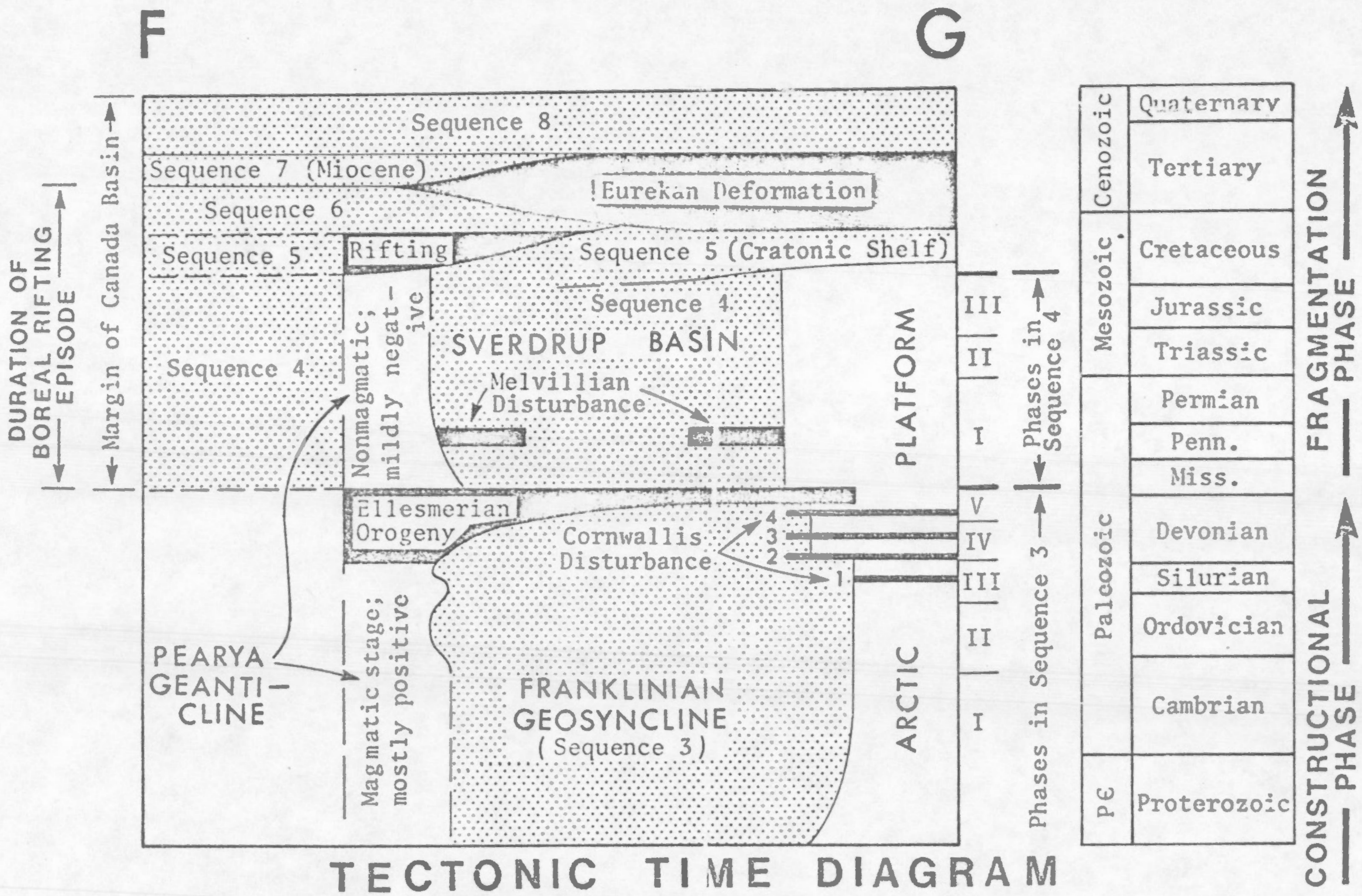

Figure 4. Tectonic time diagram for the Innuitian Mobile Belt. The stippled area shows a region where there was continuous or nearly continuous sedimentation. The solid black pattern indicates tectonic events. In the Ellesmerian Orogeny, Cornwallis Disturbance, and Melvilain Disturbance, erosion dominated, so these events mark wide unconformities. The Boreal Rifting Episode and the Eurekan Deformation were plate tectonic events, with formation of broad basins and wider sedimentation. This figure resembles a cross-section, with a location similar to that of Figure 3.

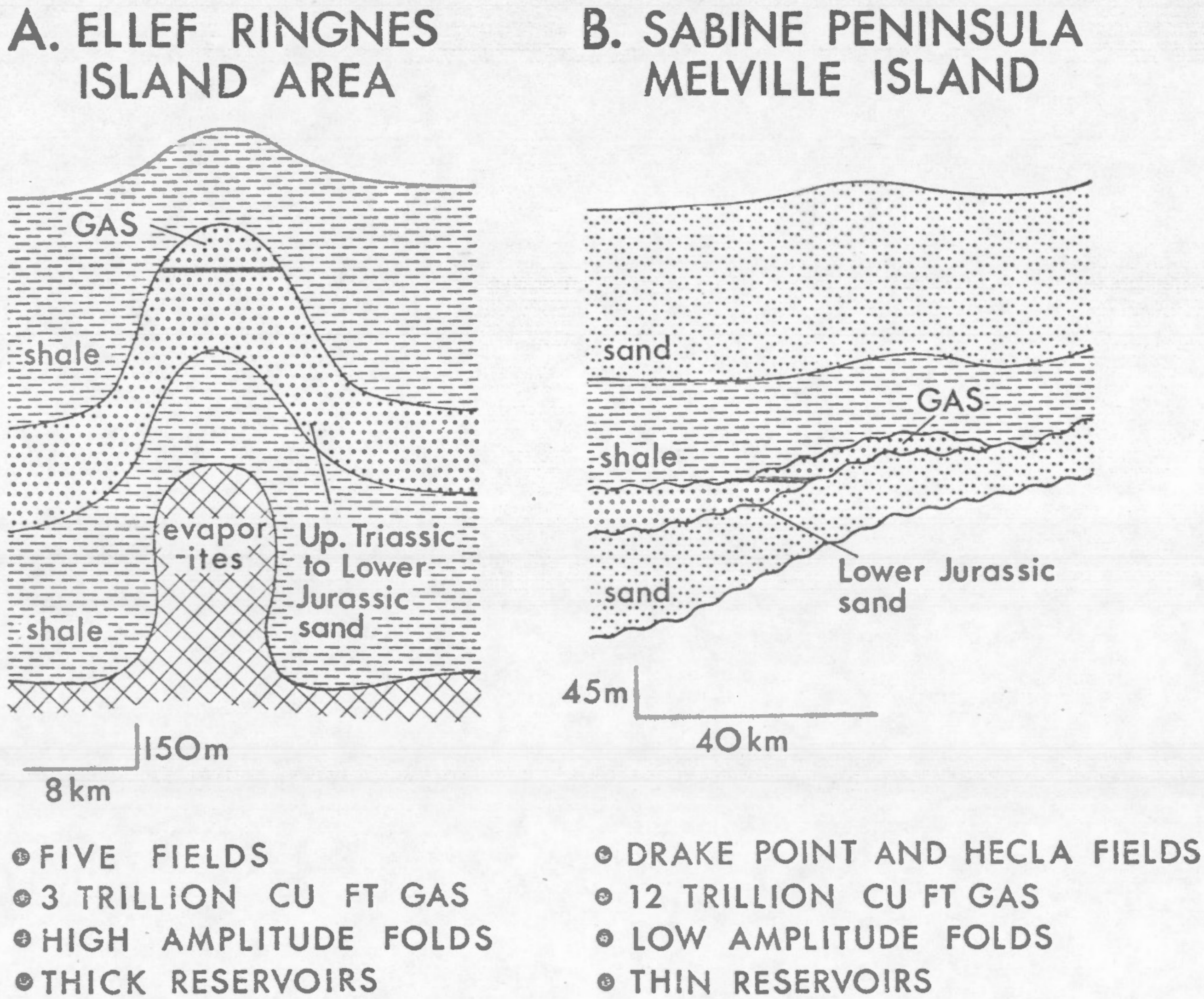

Figure 23. Generalized geology of gas discoveries in the Sverdrup Basin, summarized from Meneley (1976). Locations of these fields are shown in Figure 11.

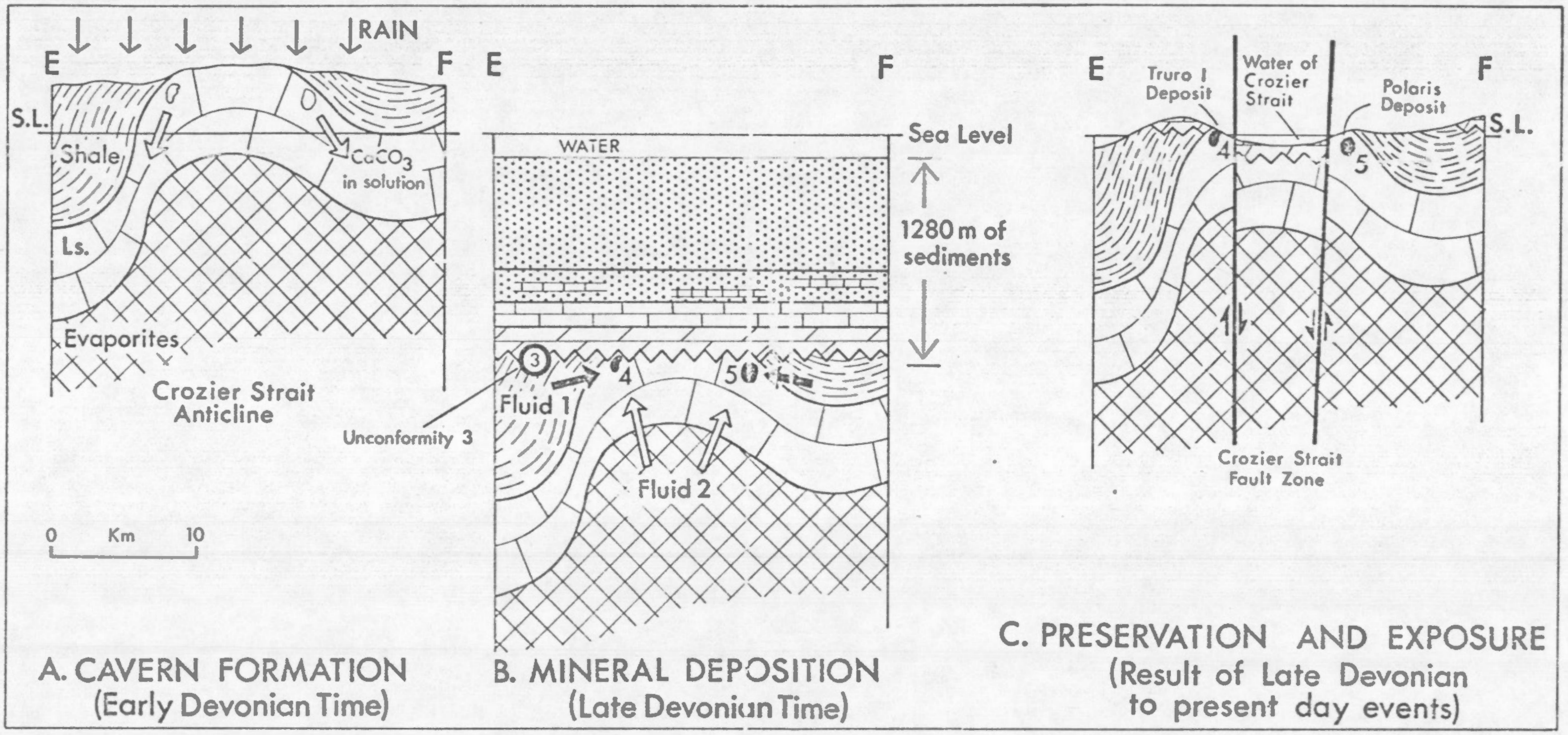

Figure 22. Sequence of events that formed typical deposits of the Cornwallis Lead-Zinc District. For location of these cross-section see Figures 6, 8 and 10. This is an enlargement of the area outlined in Figure 8. Cavern formation (A) took place by karstification in an erosional interval during the Cornwallis Disturbance. Mineral deposition (B) took place in the caverns by precipitation from formation fluids, apparently while the caverns were buried during the interval between Pulses 3 and 4 of the Cornwallis Disturbance. Preservation and exposure (C) resulted from Late Devonian and later deformation and erosion, particularly the Eurekan Rifting Episode, which formed the Crozier Strait Fault Zone.

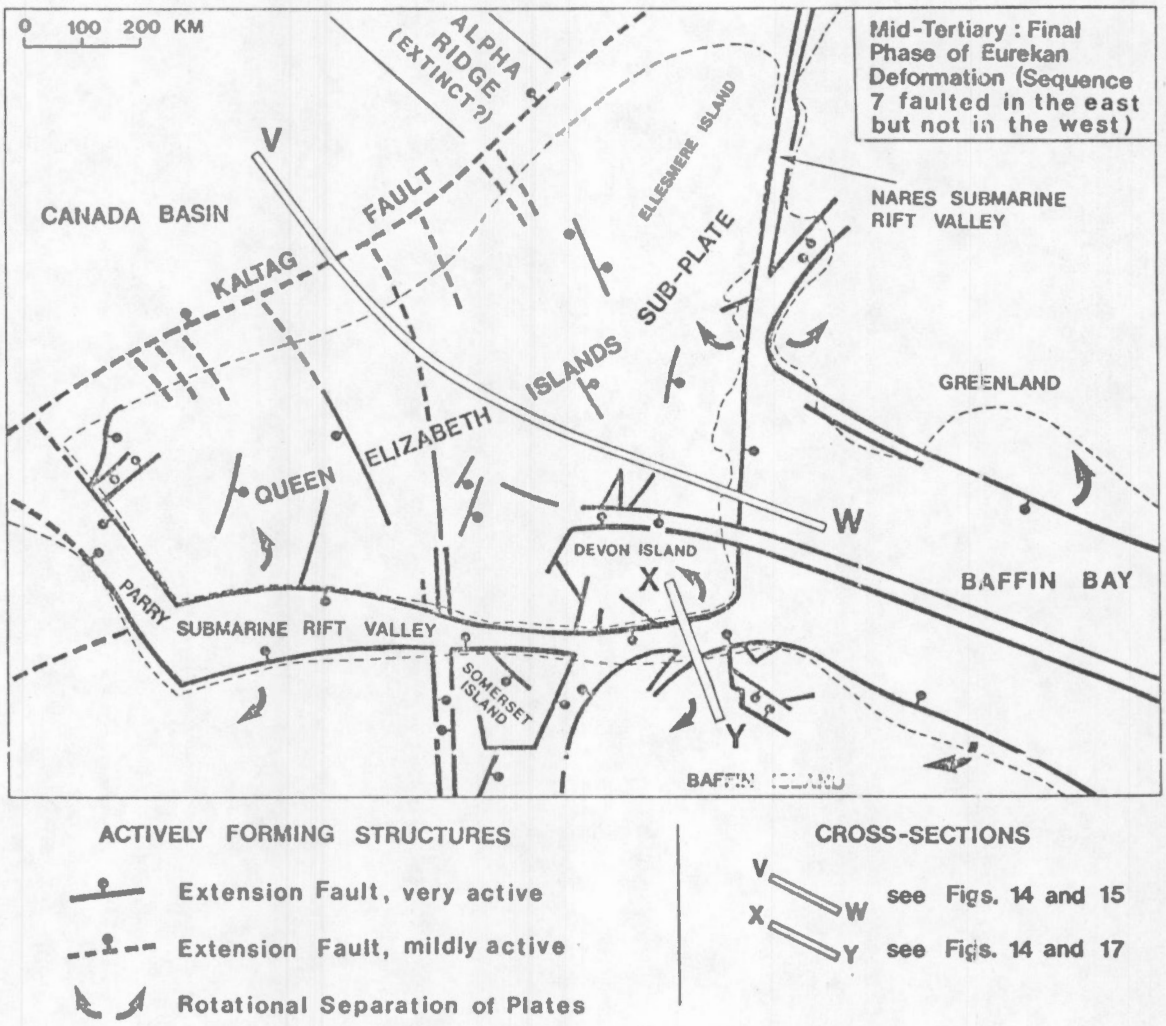

Figure 21. Final phase of the Eurekan Deformation (early Miocene or later), when the Queen Elizabeth Islands Sub-plate became completely surrounded by rifting zones. In this event only extension faulting occurred. Faults were able to break northward through the obstructions in the Nares Rift Valley and westward through obstructions in Parry Rift Valley. For the first time a structural connection from the Atlantic Ocean to the Arctic Ocean was made through the North American Continent. VW is location of cross-section in Figure 15; XY is location of cross-section in Figure 17.

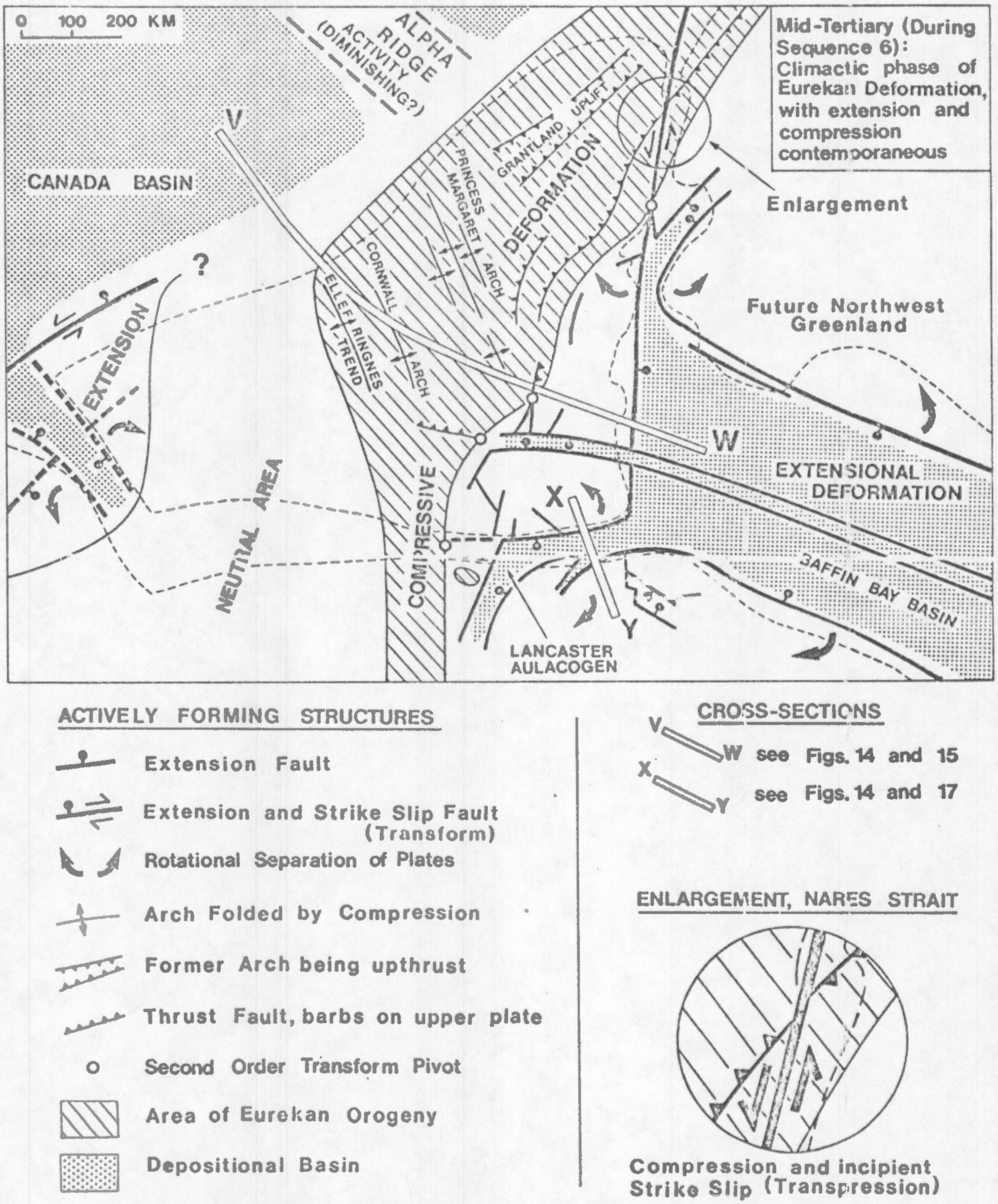

Figure 20. Climactic phase of the Eurekan Deformation, in mid-Tertiary time (between middle Eocene and early Miocene), when extension in the southeast (Eurekan Rifting Episode) caused compressional deformation farther northwest (Eurekan Orogeny). Farther west, in a large neutral area, extension that earlier emanated from the Canada Basin was neutralized by the Eurekan Deformation. In the extreme west, in the Banks Island area, extension of the Boreal Rifting Episode continued uninterrupted, presumably because this was too far west to be affected by the Eurekan Deformation. This followed shortly after the situation depicted in Figure 15.

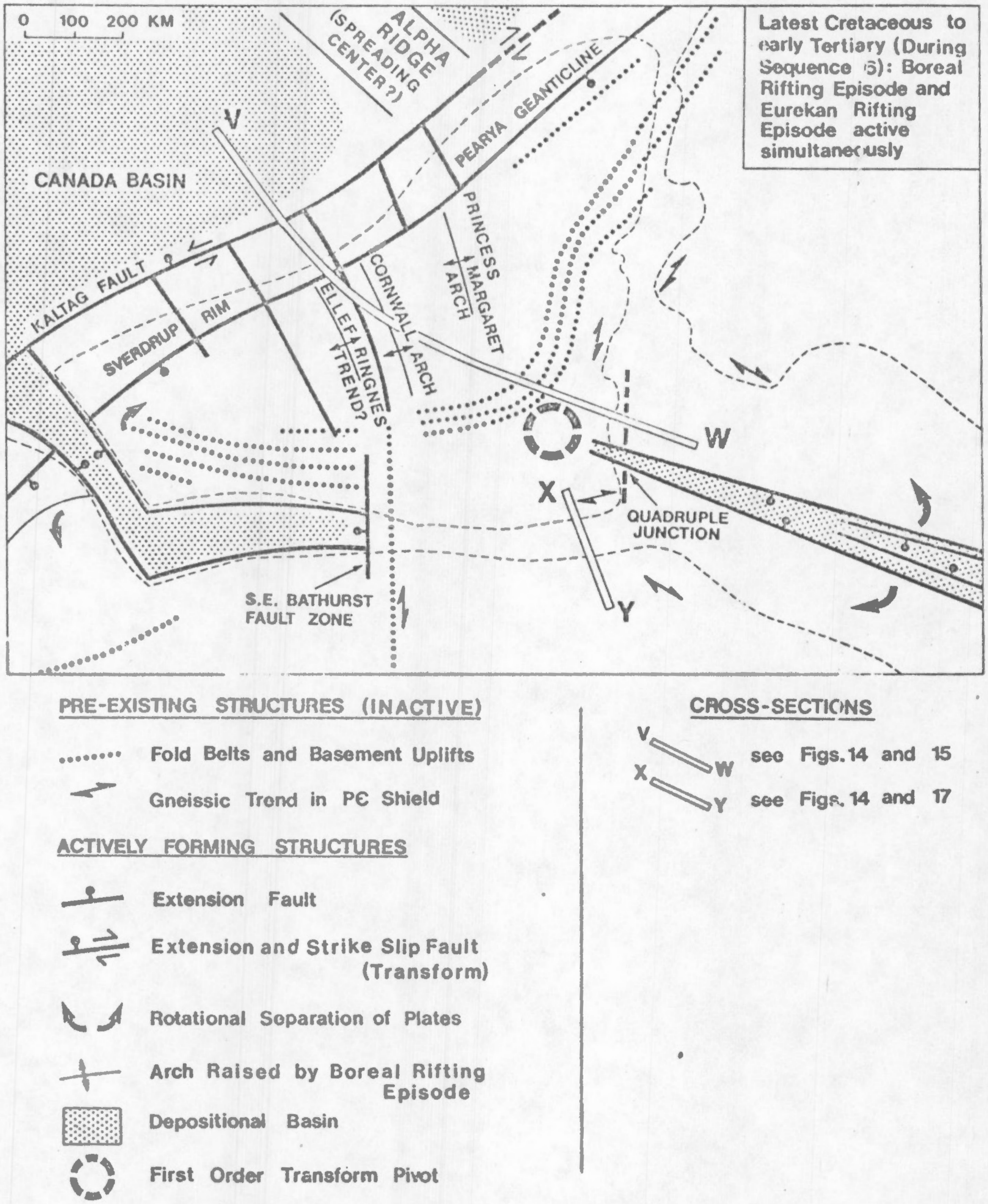

Figure 19. Simultaneous development of the Boreal Rifting Episode in the northwest and the Eurekan Rifting Episode in the southeast. Two time intervals are represented here. In Late Cretaceous time, crustal fracturing emanating from the northwest caused the Pearya Geanticline, Sverdrup Rim, Storkerson Uplift, and arches within the Sverdrup Basin to be uplifted and eroded. Parry Channel probably opened from the west, terminated at the Southeast Bathurst Fault Zone within the Boothia Uplift. Simultaneously crustal fracturing also was being propogated to the area from the southeast. Later in latest Cretaceous to early Tertiary time, further development of the Eurekan Rifting Episode began to neutralize the rifting effects of the Boreal Episode in the Sverdrup Basin. This resulted in widespread collapse of parts of the formerly uplifted arches, Sverdrup Rim, Storkerson Uplift, and the intrabasin arches, with sedimentary encroachment of Upper Paleocene to Eocene rocks onto each (Fig. 15).

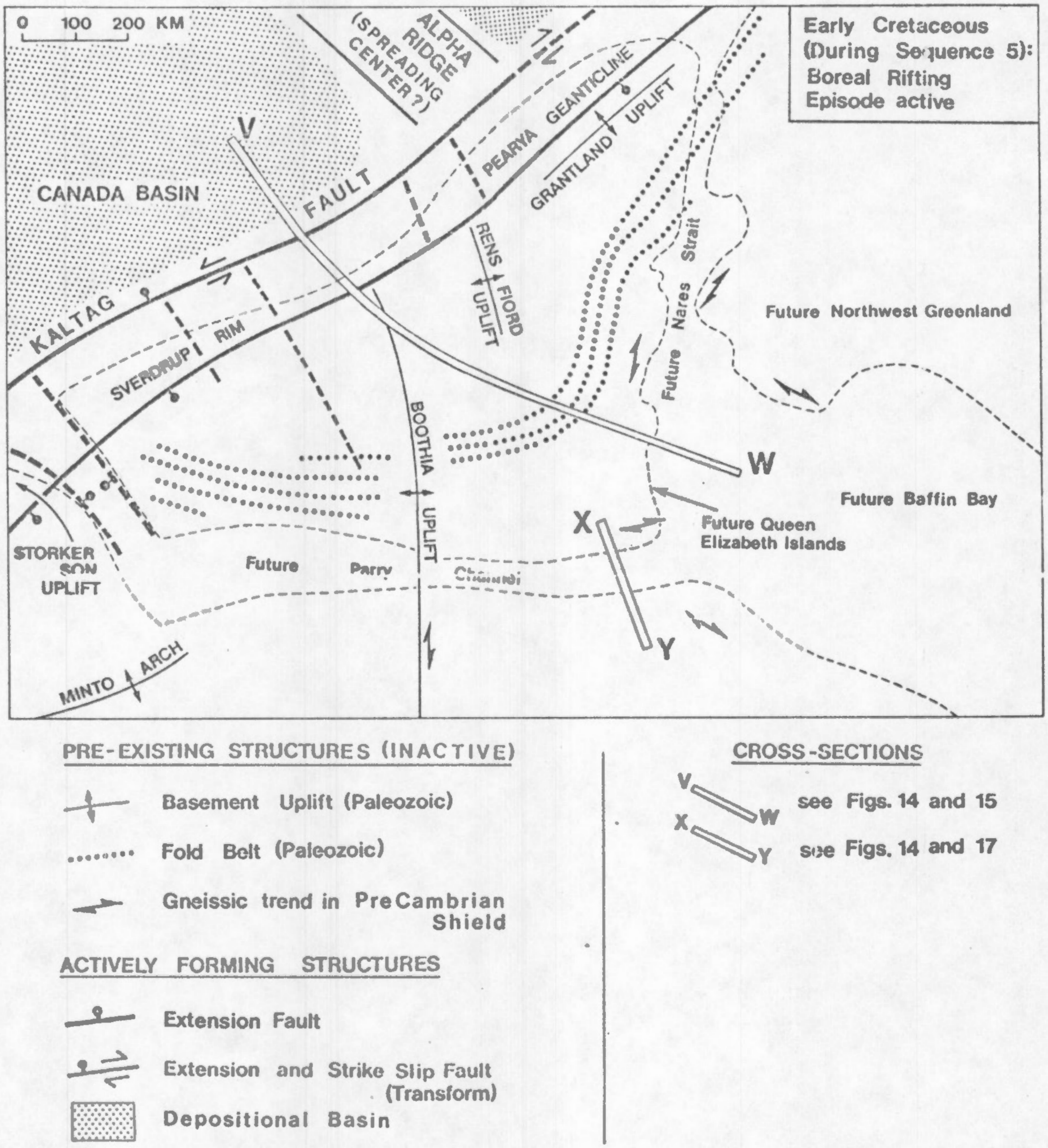

Figure 18. Early formation of the Queen Elizabeth Islands by the Canadian Arctic Rift System. Dashed lines show outline of Queen Elizabeth Islands and main channels. This represents the structural situation existing in Late Cretaceous time (pre-Eureka Sound Formation) that had been developed by the Boreal Rifting Episode (Fig. 4). The Canada Basin was being enlarged by the rifting episode. The continent was covered by an overlap of shallow seas that were depositing Sequence 5 (Fig. 12). The Pearya Geanticline and Sverdrup Rim were mildly positive. Trends of buried uplifts and gneisses influenced this and future rifting deformation.

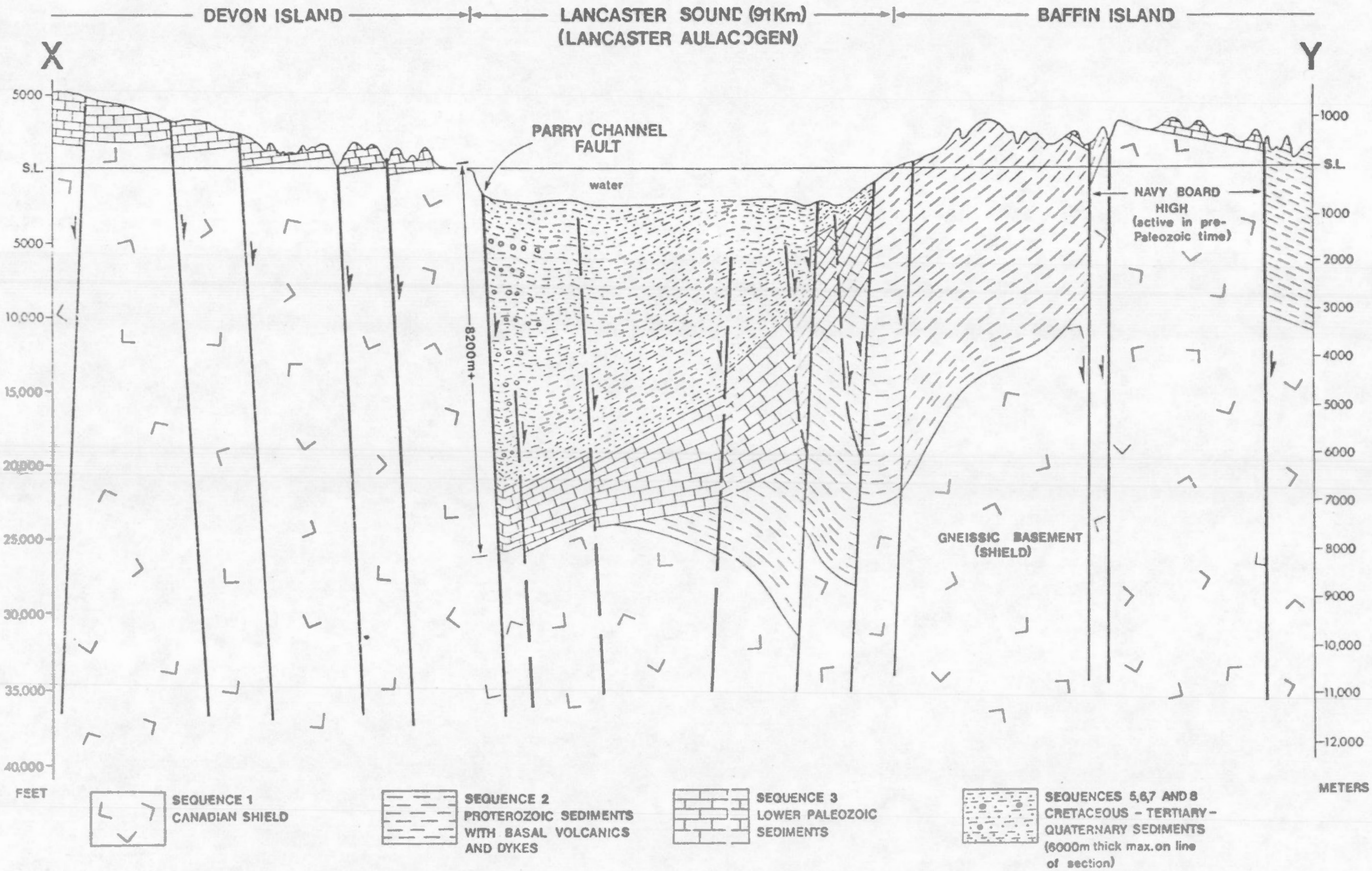

Figure 17. Cross-section through the Lancaster Aulacogen (after Kerr, in press). Location is shown on Figures 14, 20 and 21. The Cretaceous-Tertiary-Quaternary sedimentary column within the aulacogen probably constitutes Sequences 5, 6, 7, and 8, and the following history is presumed to apply. Sequence 5 may constitute the remnants of a sheet of rather uniform thickness, that was deposited on a cratonic shelf and covered the entire region of this cross-section before the aulacogen began to form (Fig. 12). This sequence later was preserved only in the aulacogen. Sequence 6 was deposited while the aulacogen was progressively forming (Fig. 20). Sequence 7 may have been deposited during the final phase of the Eurekan Deformation (Fig. 21). Sequence 8 was deposited after the period of active formation of the aulacogen.

THE CANADIAN ARCTIC RIFT SYSTEM

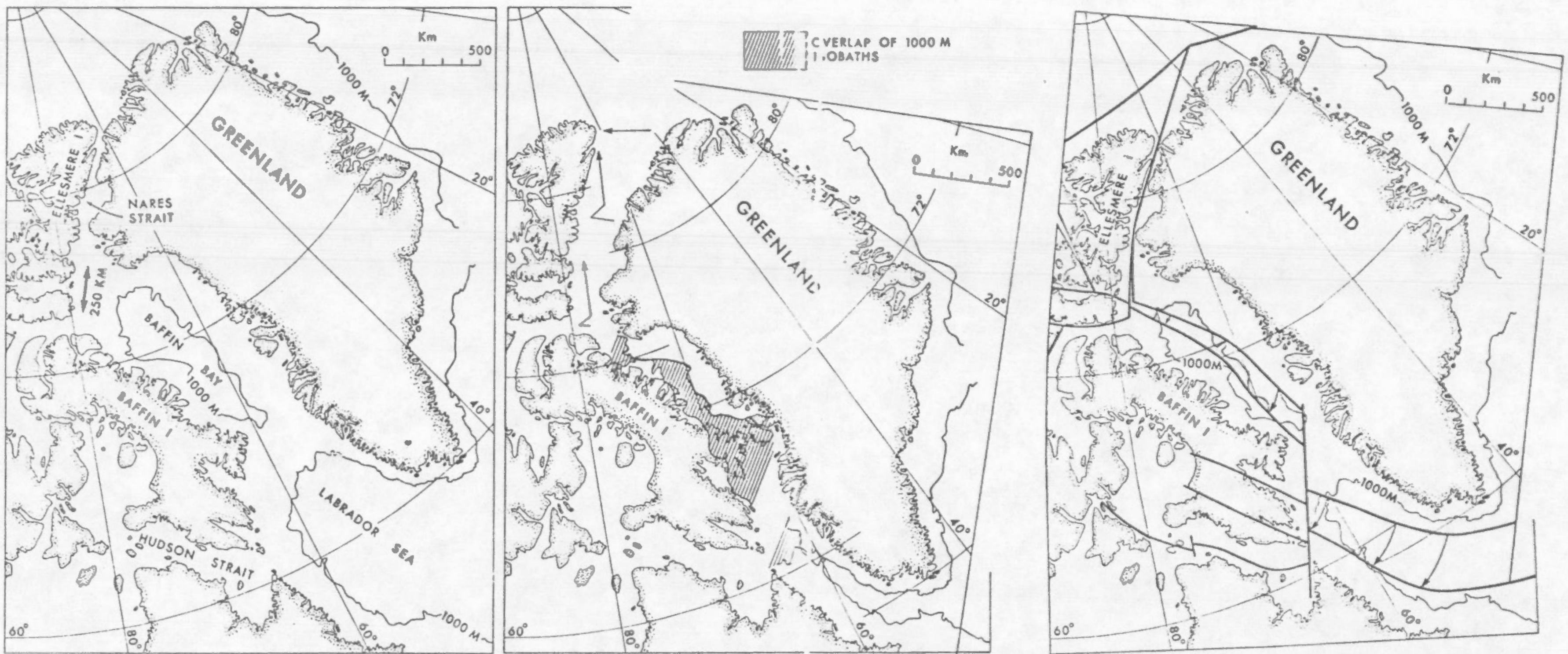

Figure 16. Current theories of tectonics that explain the relationship of Greenland to the rest of North America. (A) Fixist theories (after Beloussov, 1970, and Meyerhoff, 1973), in which there has been no lateral movement of Greenland. (B) Conventional plate tectonic theories (moving continents or moving plates, after Wegener, 1924; Carey, 1958; Wilson, 1965b; Dewey, 1972; Herron et al., 1974; Srivastava, 1978; Newman and Falconer, 1978; and most modern day workers), which consider that Greenland moved about 250 km along Nares Strait. The conventional reconstruction shown (after Srivastava, 1978) considers that Greenland first moved toward Ellesmere Island, closing up Nares Strait (horizontal arrows), and later had northward strike slip displacement (vertical arrows). (C) Restoration according to the integrated theory of plate tectonics (Kerr, 1967b, c, and herein). Greenland and Canada rotated apart, with the lateral movement increasing to the southeast, from Nares Strait to Baffin Bay to Labrador Sea. The amount of restoration made here is shown by curving arrows. Much of the great movement apart in Labrador Sea is accounted for by minor rotational opening in southern Nares Strait. The 1000 m isobaths in Baffin Bay and Labrador Sea cannot be brought back together. The gap between them that appears at first to be missing continental material can be accounted for by a foundered remnant of continental crust.

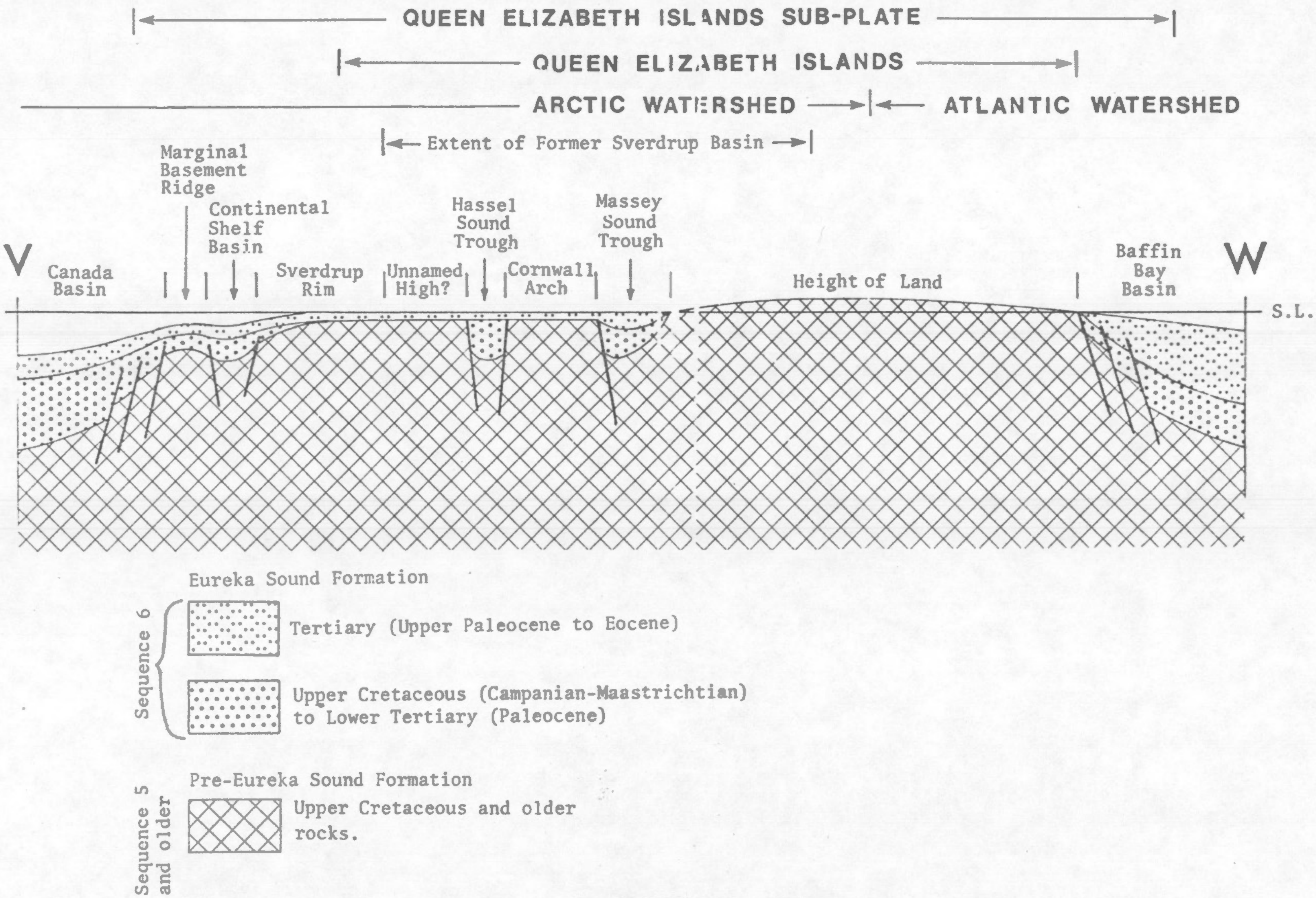

Figure 15. Generalized cross-section from the Canada Basin to the Baffin Bay Basin reconstructed to about Eocene time, showing Sequence 6 of the Innuitian Mobile Belt (see Figs. 3 and 4). The two basins had been actively forming as rifted oceanic basins, with the Queen Elizabeth Islands Sub-plate being a continental transition between. The lower part of Sequence 6 was deposited mainly in the northwest, as substantial activity of the Boreal Rifting Episode continued (Fig. 19, early). The upper part of Sequence 6 was deposited more widely, when the Boreal Rifting Episode had nearly expired, but the Eurekan Rifting Episode was in an early stage and accellerating (Fig. 19, late). The time of the reconstruction shown above is just prior to the climactic phase of the Eurekan Orogeny (Fig. 20), when Sequence 6 and older rocks were deformed by extension southeast of the height of land and by compression in the northwest.

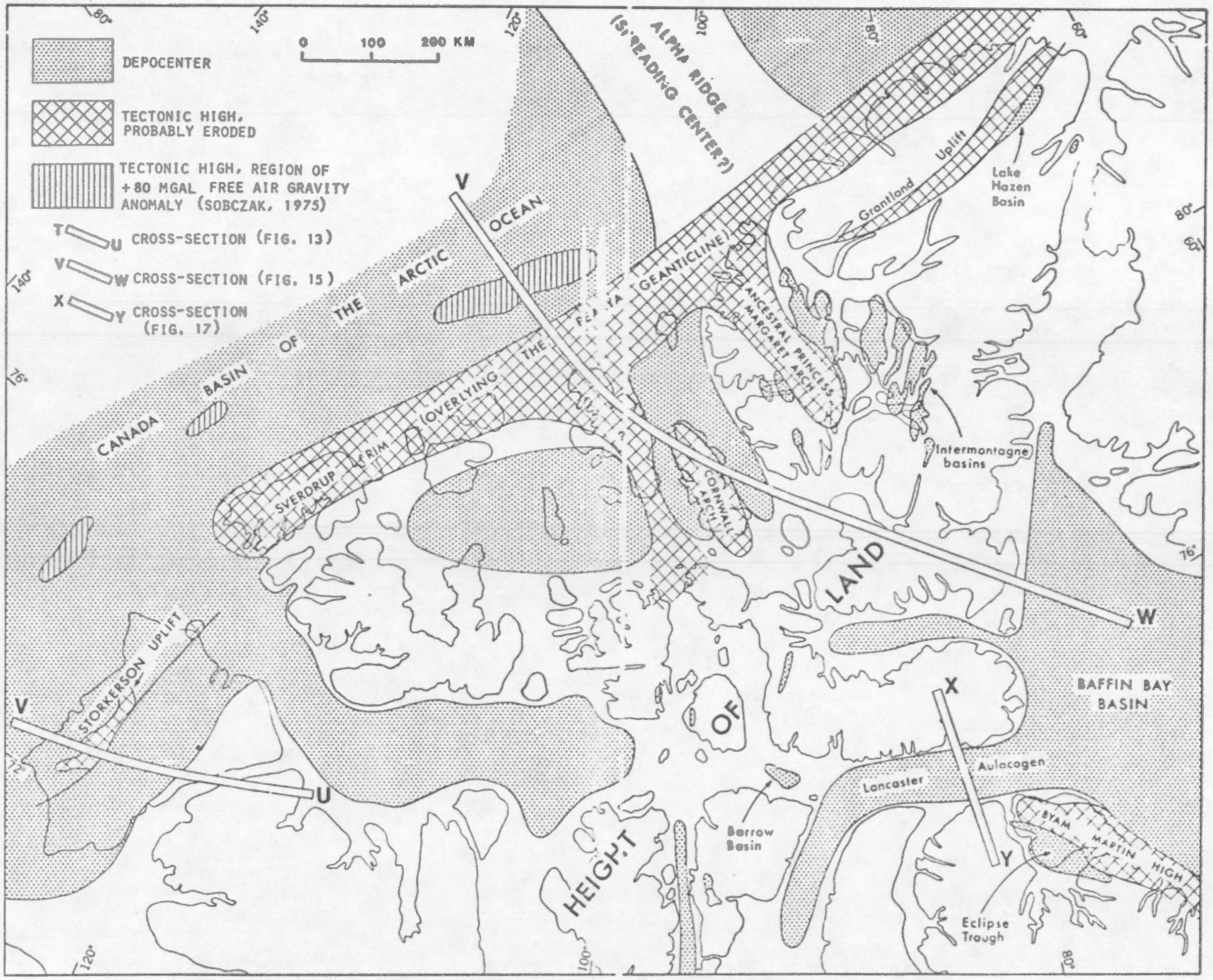

Figure 14. Paleography of Sequence 6, the Eureka Sound Formation and equivalent rocks, which extends from Late Cretaceous (Campanian-Maastrichtian) to early Tertiary (Paleocene-Eocene). This figure shows the transition that developed between the Atlantic and Arctic Oceans by means of the Canadian Arctic Rift System. The Canada Basin of the Arctic Ocean, and the Baffin Bay Basin were actively forming rifting basins, separated by a broad height of land. These two main depocenters became connected by the rifted channels that separate the islands of the Canadian Arctic Archipelago. Finger shaped projections of thinner rocks were deposited in parts o these channels.

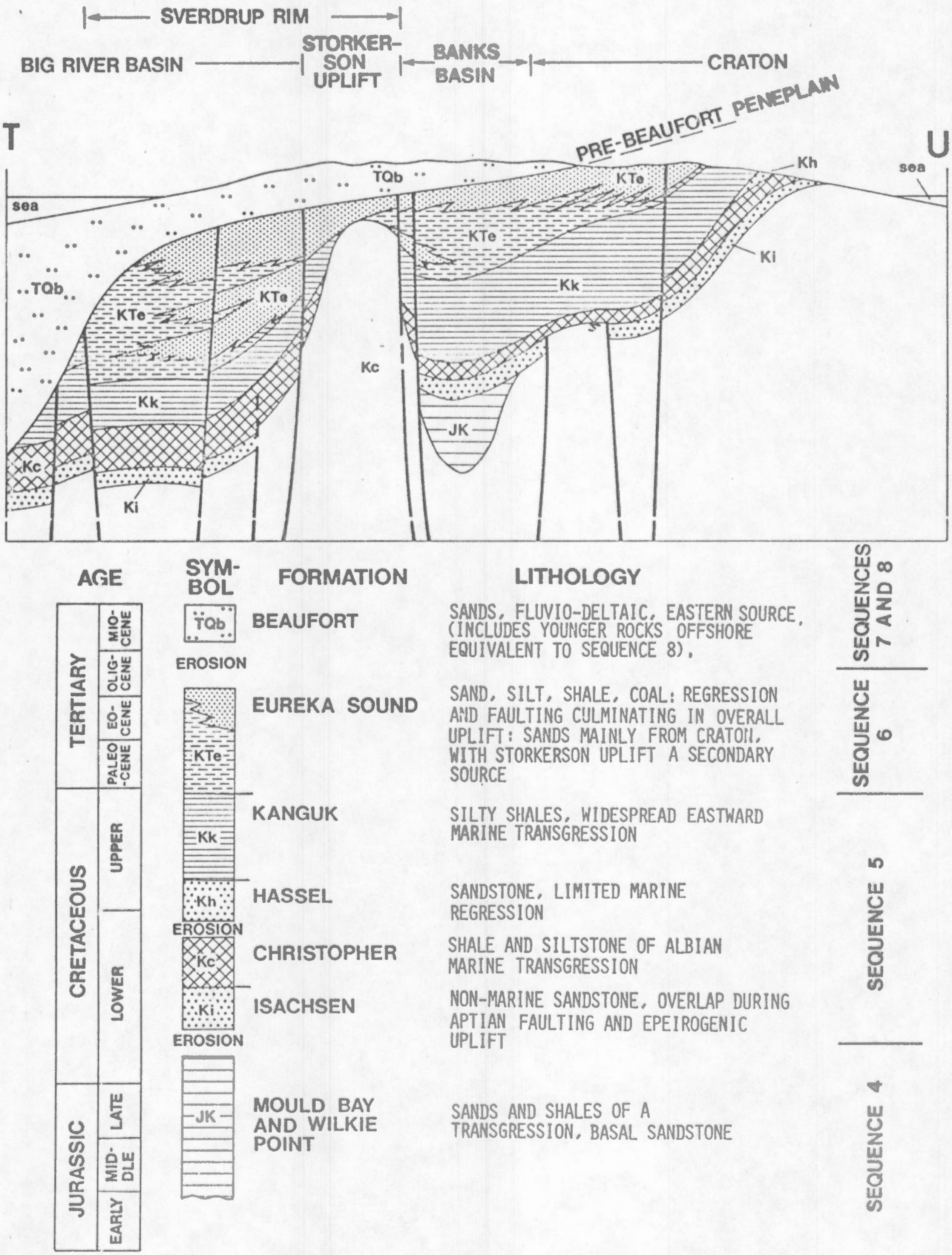

Figure 13. Cross-section showing the development of the margin of the Arctic Ocean Basin in the Banks Island area (after Miall, 1975). The approximate location is shown on Figure 11.

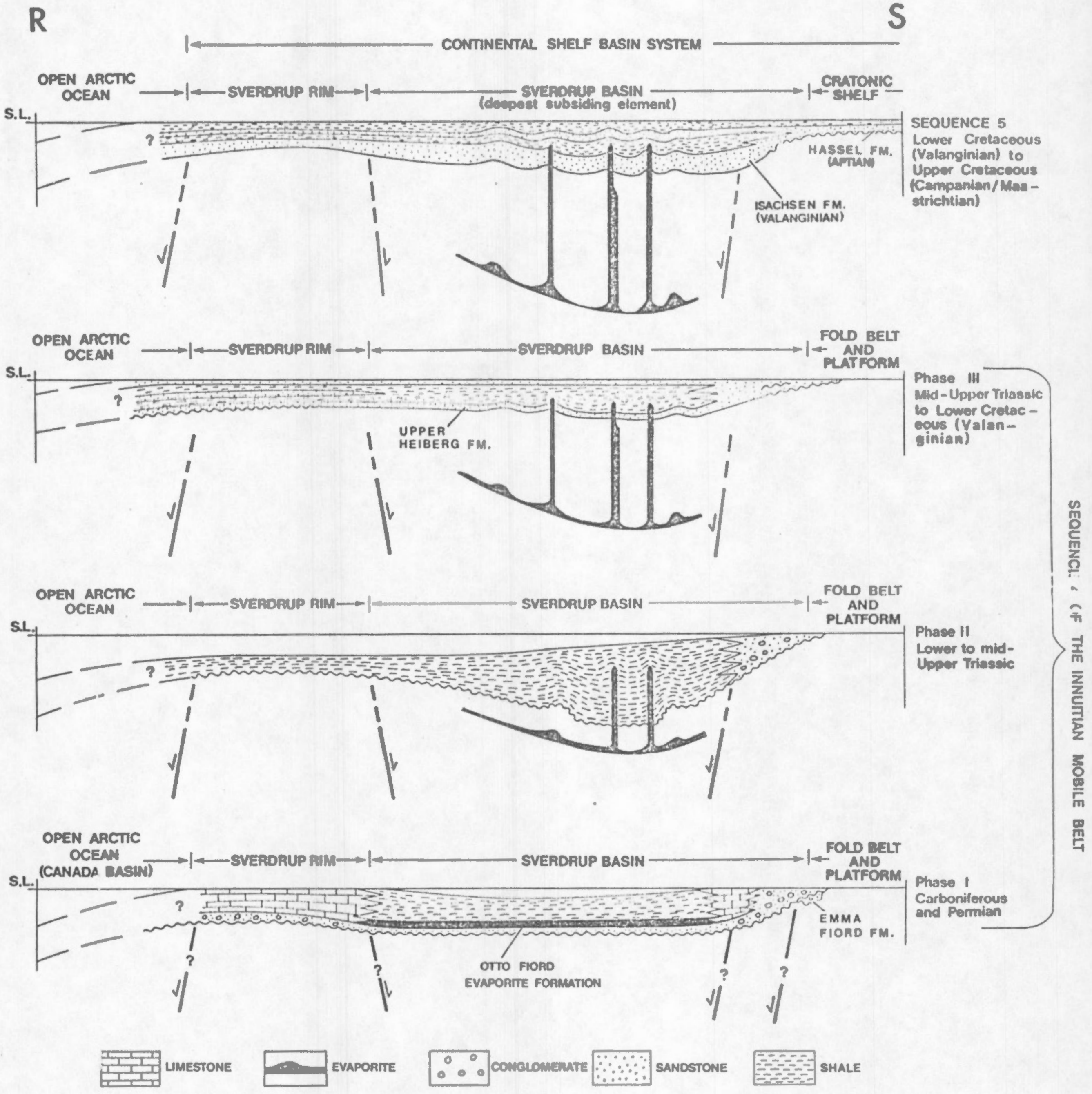

Figure 12. Cross-sections showing the history of Sequences 4 and 5 of the Innuitian Mobile Belt. The history of these two sequences was dominated here by the Boreal Rifting Episode (Fig. 4). Sections are diagrammatic and scale is not shown. The location of these lines of section is shown in Figure 11. Much of this diagram is adapted from Balkwill (1978).

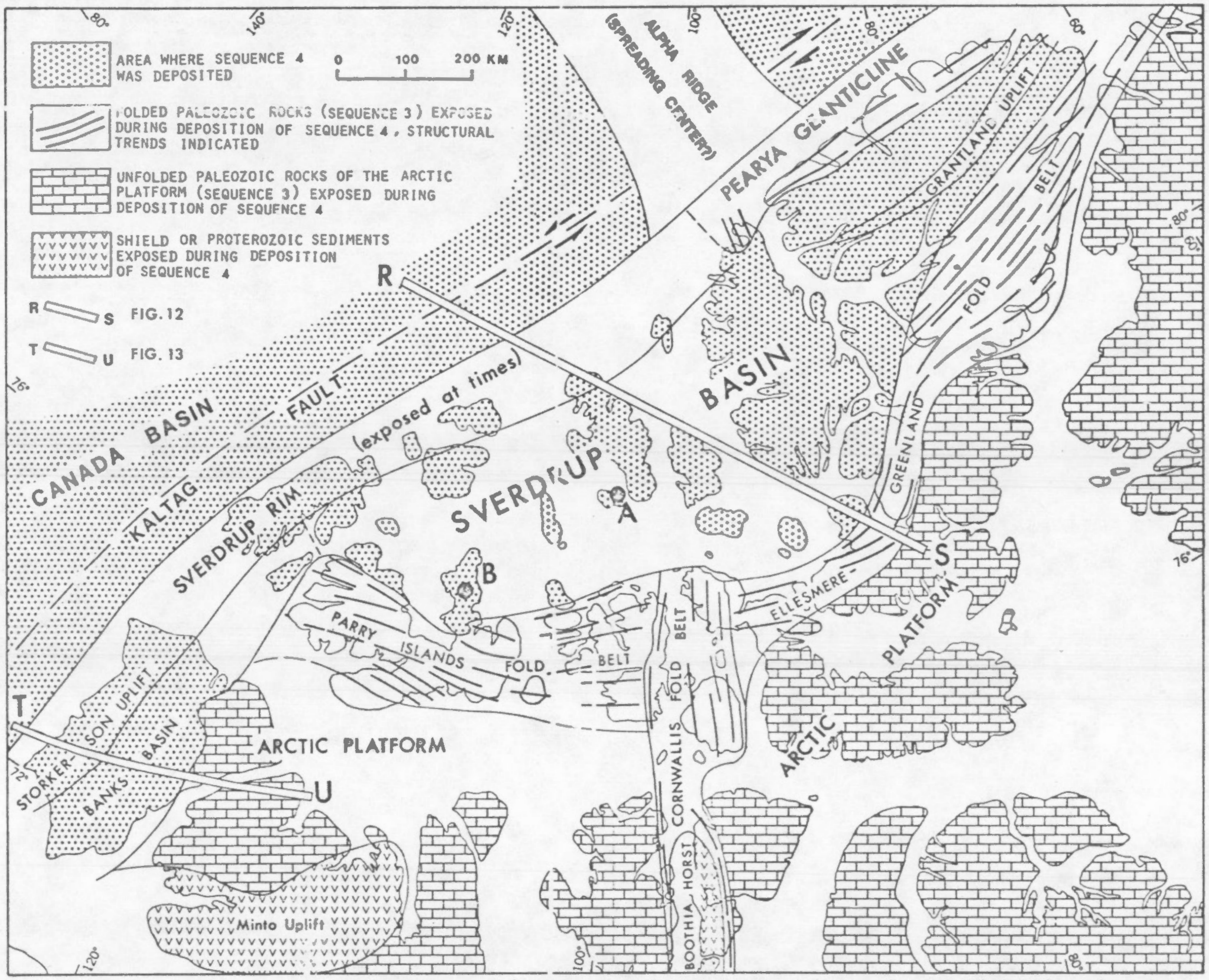

Figure 11. Paleogeography and tectonic setting of Sequence 4 (Figs. 3, 4, and 12), which includes rocks extending from Mississippian to lower Lower Basin, Banks Basin, Sverdrup Rim, and presumably rocks along the margin of the ancestral Canada Basin. Sequence 4 began the fragmentation phase of northern Canada (Fig. 4). A. Ellef Ringnes Gas Field Region (Figs. 12 and 23). B. Sabine Peninsula Gas Fields Region (Fig. 23).

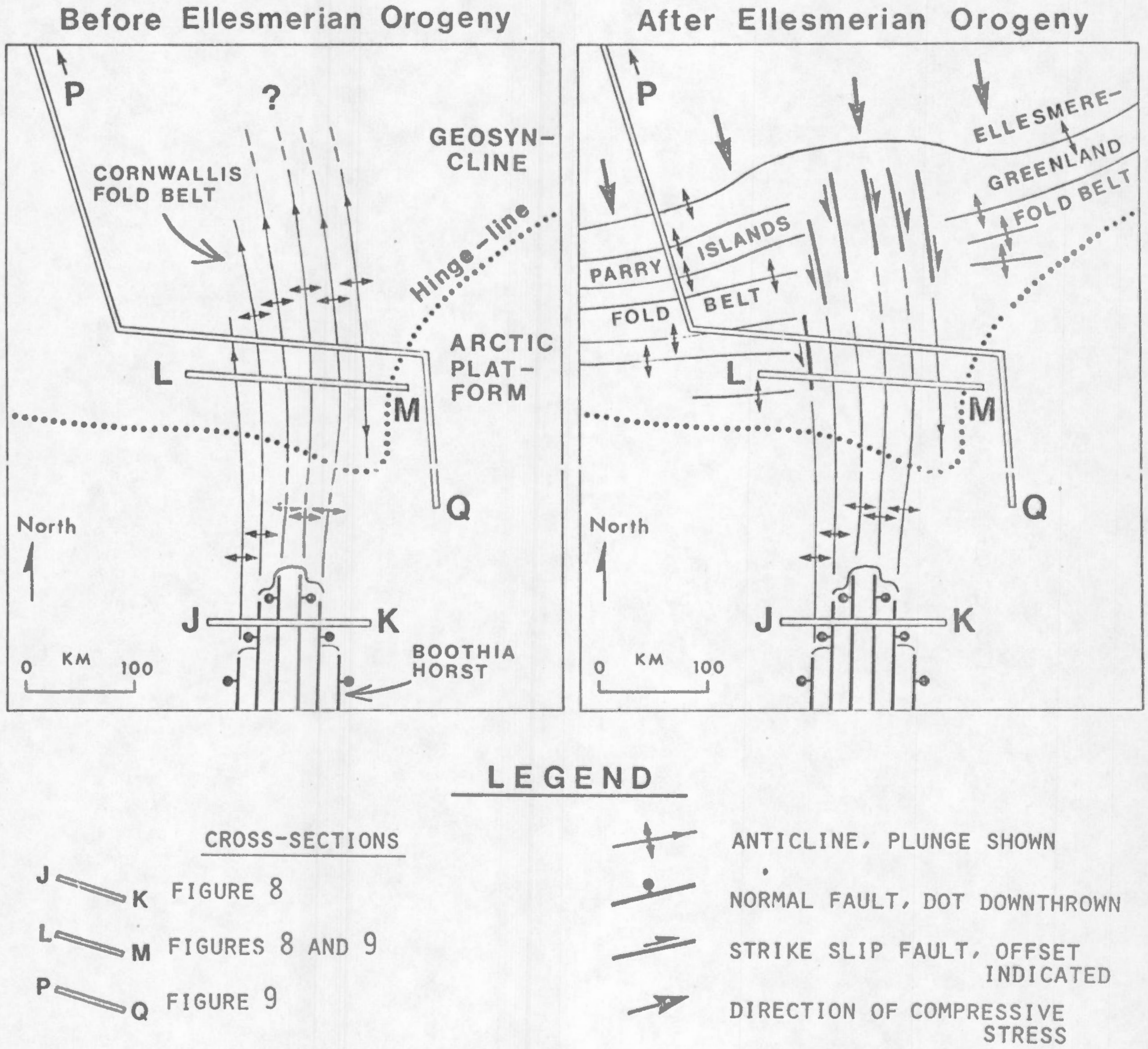

Figure 10. Maps showing how the Boothia Uplift interfered with the Ellesmerian Orogeny. Before: the basic structure of the Boothia Uplift was an anticlinorium (Cornwallis Fold Belt) draped over a crystalline core (Boothia Horst). After: southward overriding of the Ellesmerian Orogeny produced new fold belts in the previously unfolded parts of the geosyncline. Within the Cornwallis Fold Belt interference caused strike slip faults that are left lateral on the west side and right lateral on the east side.

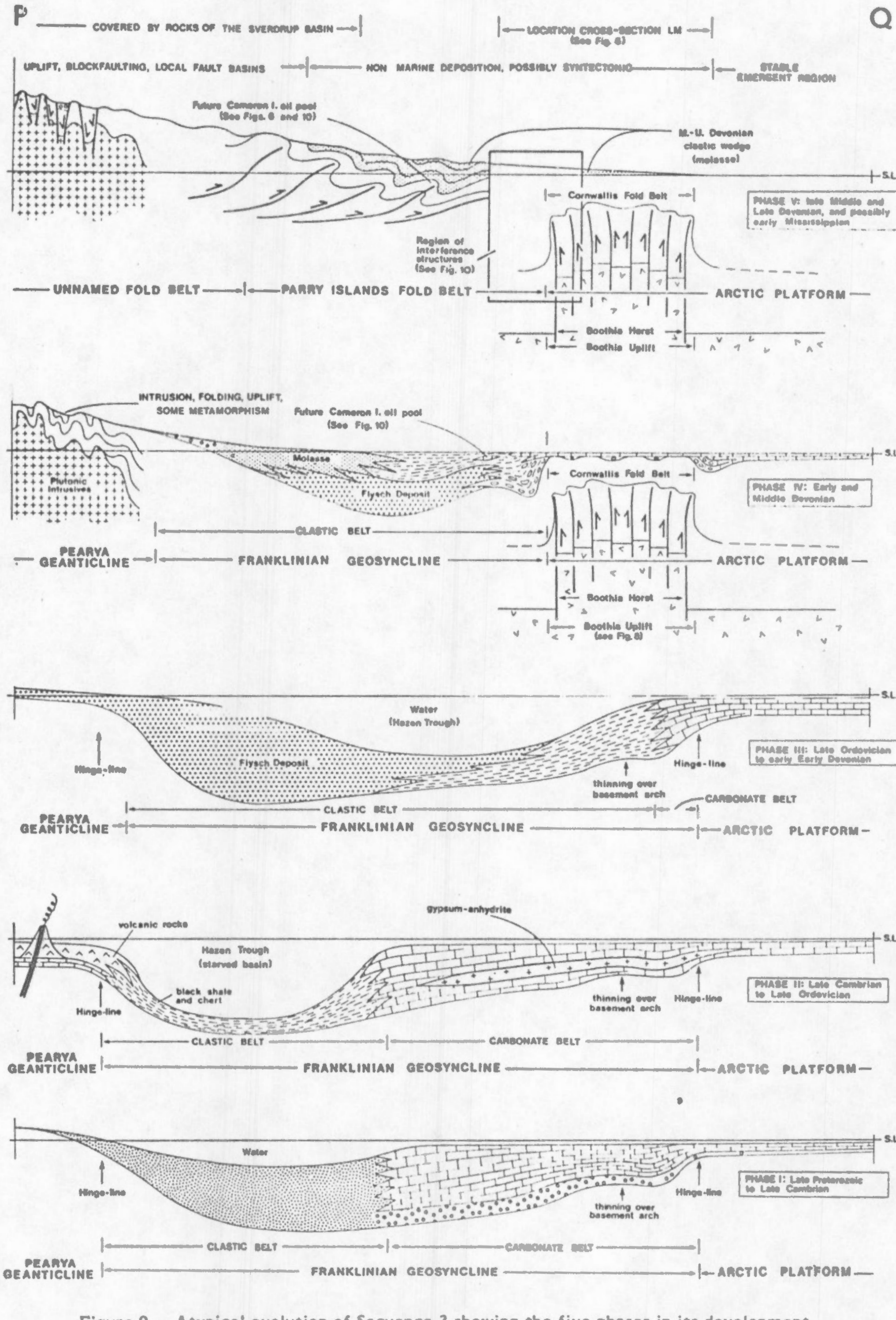

Figure 9. Atypical evolution of Sequence 3 showing the five phases in its development (cf. Fig. 7). This cross-section of the geosyncline was obstructed by a cross-trending basement uplift (see Fig. 6). In Phases I, II, III, and IV, the Boothia Uplift influenced facies patterns. In Phase V the existence of the uplift as a structure interfered with the deformation produced by the Ellesmerian Orogeny (see Fig. 10). Conventional lithologic symbols are used. The vertical scale is relative only, and varies from place to place. Where covered by the Sverdrup Basin the cross-sections are speculative. There they are similar to corresponding parts of Figure 7, and are compatible with available evidence.

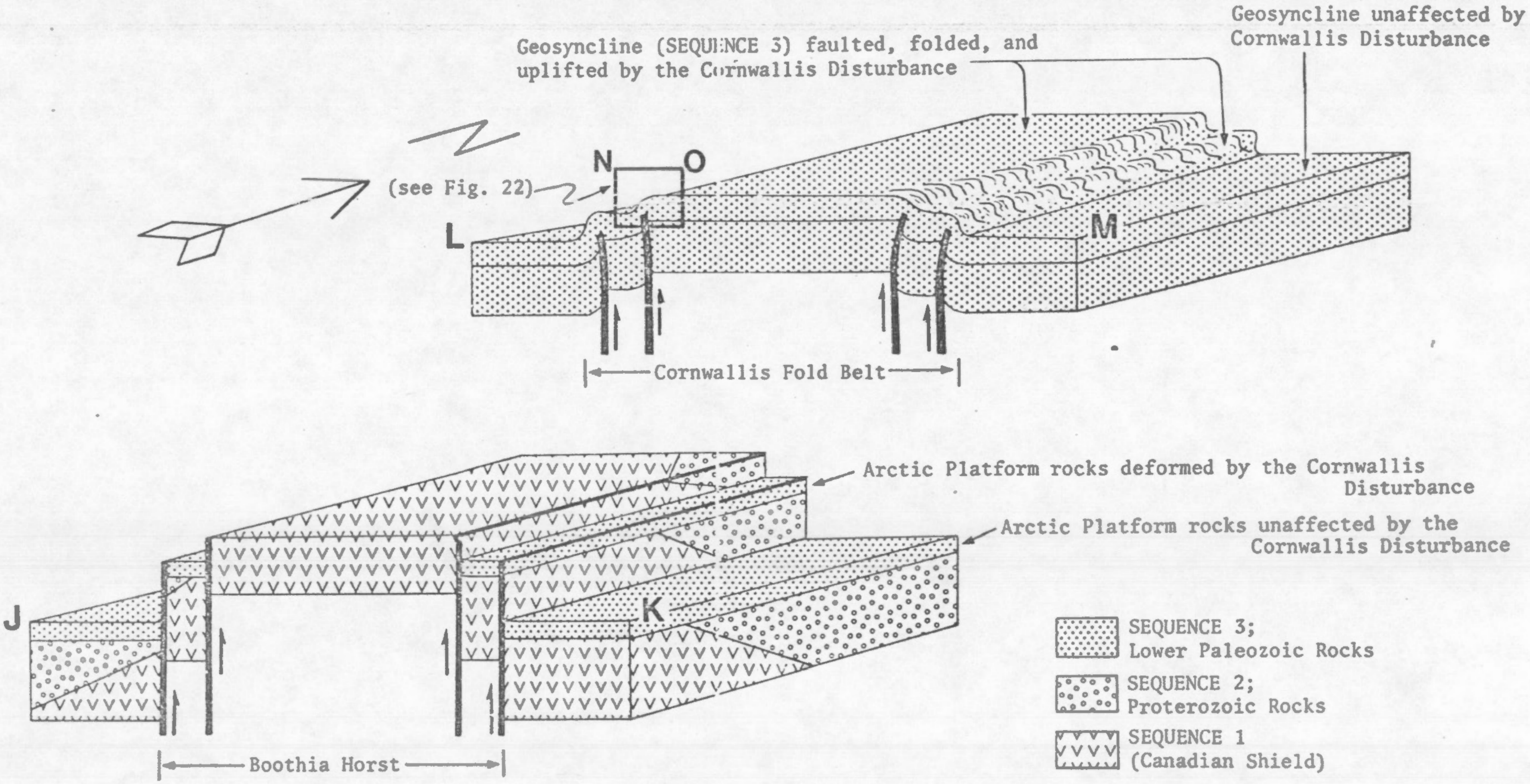

Figure 8. Formation of the Boothia Uplift by faulting and drape-folding of the Cornwallis Disturbance. This shows the basic structure of the Boothia Uplift, which was the structure existing at the end of its formation, and before its modification (Kerr, 1977a). For locations of cross-sections see Figures 6 and 10. The Boothia Uplift includes the Boothia Horst and the Cornwallis Fold Belt.

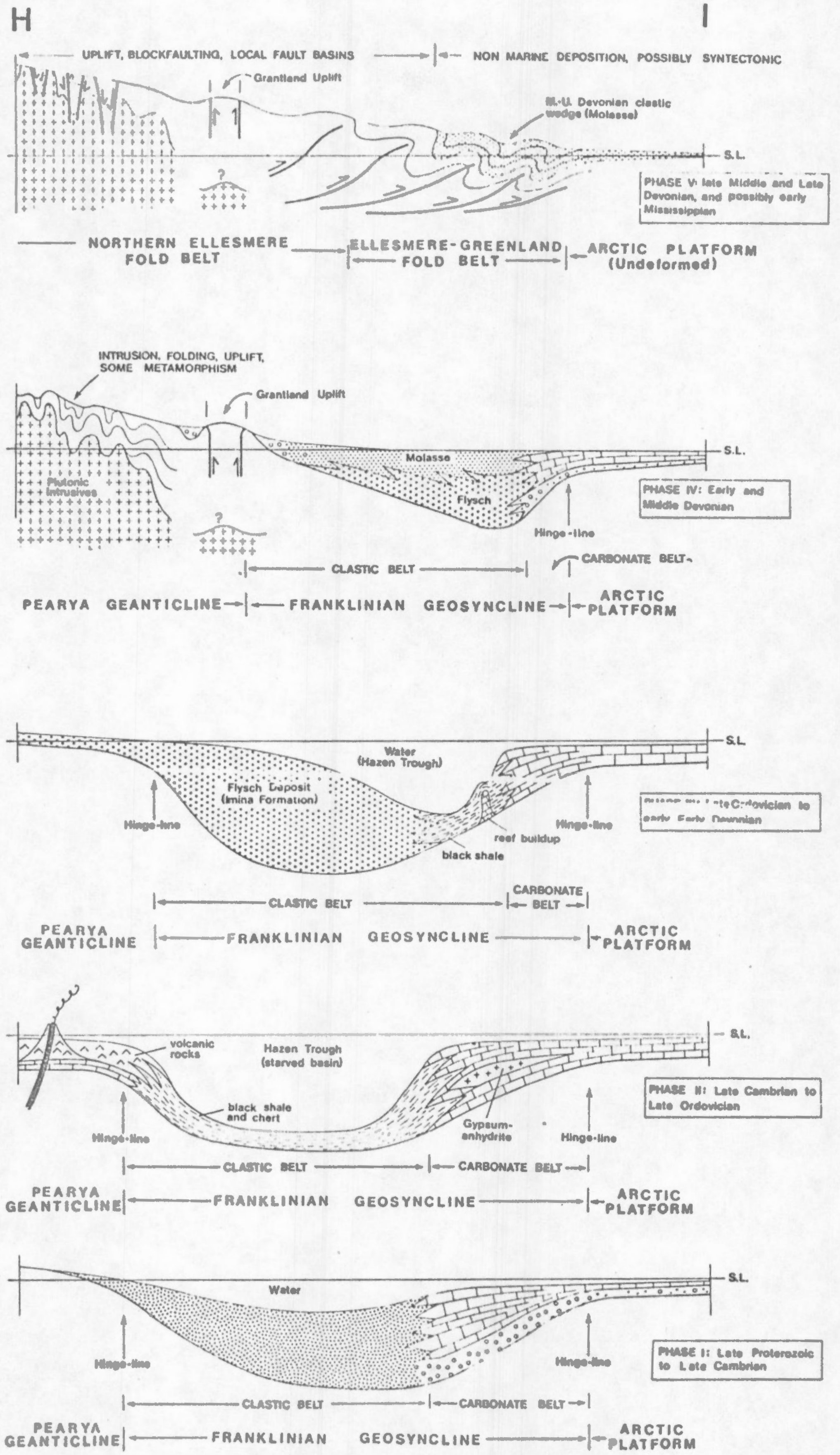

Figure 7. Typical evolution of Sequence 3, showing the five phases in its development that are summarized also in Figure 5. This is the simplest cross-section of the geosyncline, located where that basin was not affected by internal cross-trending basement uplifts. Sequence 3 was part of the constructional phase of northern Canada (Fig. 4). Coventional lithologic symbols are used. The vertical scale is relative only, and varies from place to place. For location see Figure 6.

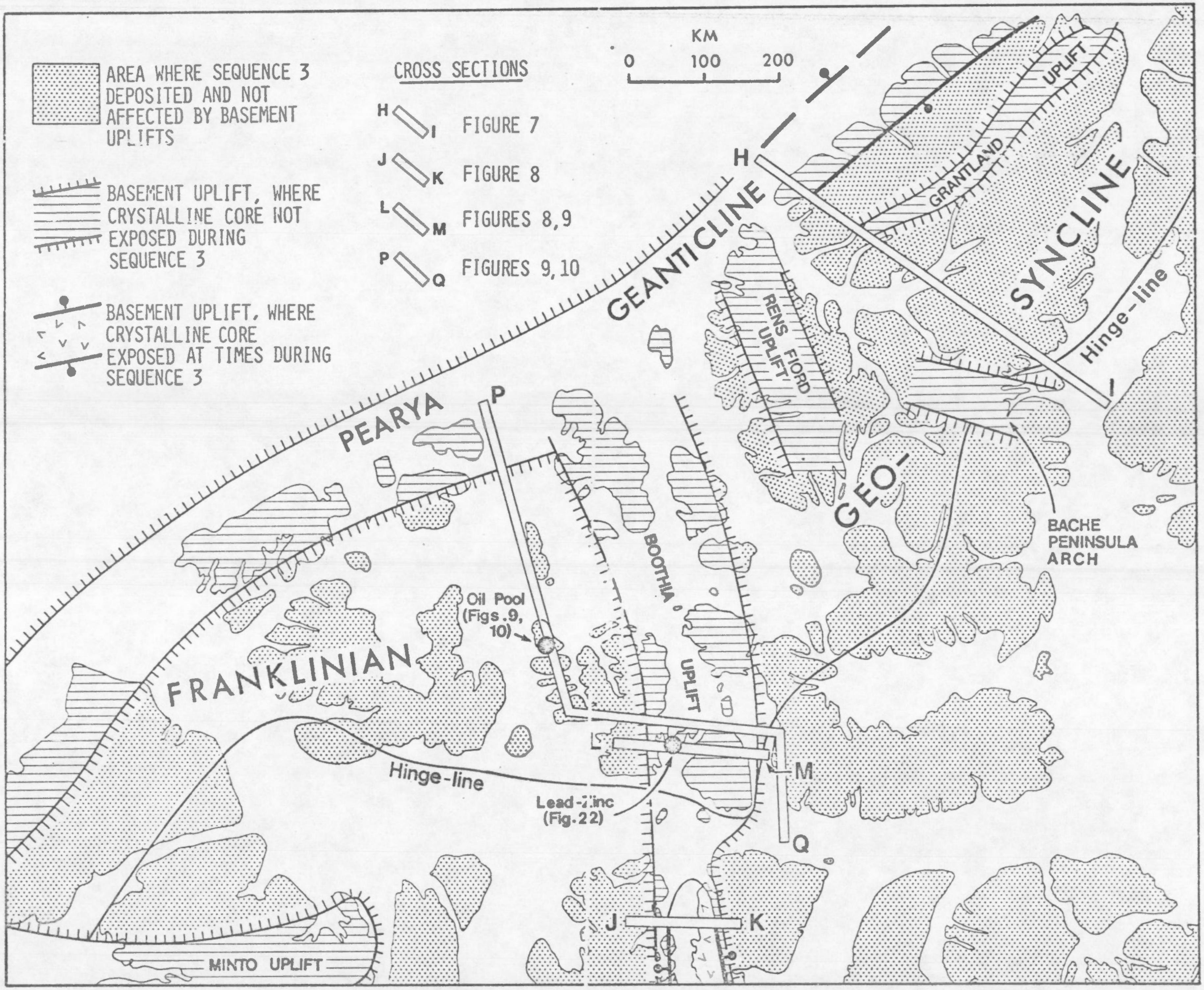

Figure 6. Lower to middle Paleozoic tectonic elements of the Canadian Arctic. The basement highs shown affected Sequence 3, which includes Upper Proterozoic to Upper Devonian rocks.

PERIOD	STAGE	PHASE	SEQUENCE 3 EVENTS
DEVONIAN	LATE	Phase V	Geanticline and much of geosyncline raised; southeastward overriding terminates geosyncline; (= late part of Ellesmerian Orogeny)
	MIDDLE		
	EARLY	Phase IV	Geosyncline narrowed from both sides and fragmented by basement uplifts; geanticline intruded, fragmented and raised to provide molasse; (= early part of Ellesmerian Orogeny)
SILURIAN	LATE	Phase III	Geosyncline narrowed slightly from NW with great infilling of Hazen Trough by flysch deposits; starved axial basin migrates southeast and becomes narrower; carbonates encroach widely onto Arctic Platform, with some onto geanticline.
	MIDDLE		
	EARLY		
ORDO-VICIAN	LATE	Phase II	Geosyncline broad; receives volcanics from geanticline; starved axial basin; carbonates spread onto platform with hinge-line indistinct.
	MIDDLE		
	EARLY		
CAMBRIAN	LATE	Phase I	Geosyncline began narrow; progressively widened by outward encroachment onto platform and geanticline.
	MIDDLE		
	EARLY		

DEFORMATIONS

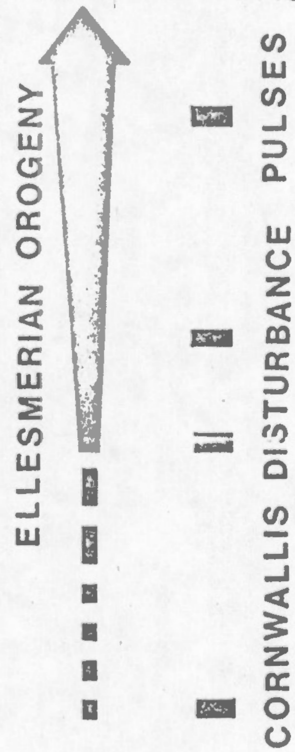

Figure 5. Phases in the history of Sequence 3 of the Innuitian Mobile Belt. Setting of this sequence is shown in Figures 3 and 4; its evolution is shown in Figures 7 and 9. Sequence 3 includes deposits of the Arctic Platform, Franklinian Geosyncline, and those on the Pearya Geanticline.

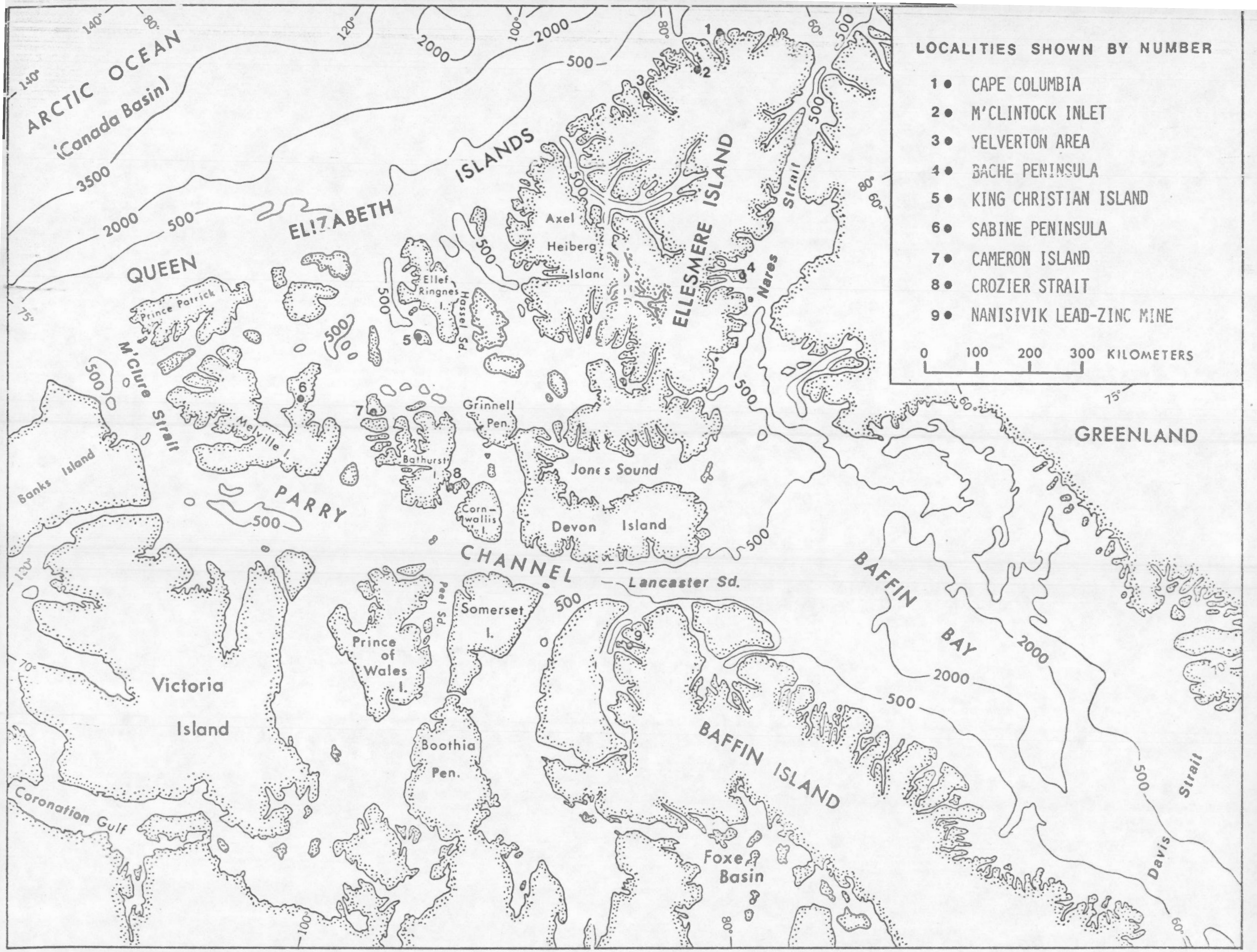

Figure 1. Index map of the Canadian Arctic Islands showing localities referred to in the text. Bathymetry is in meters.

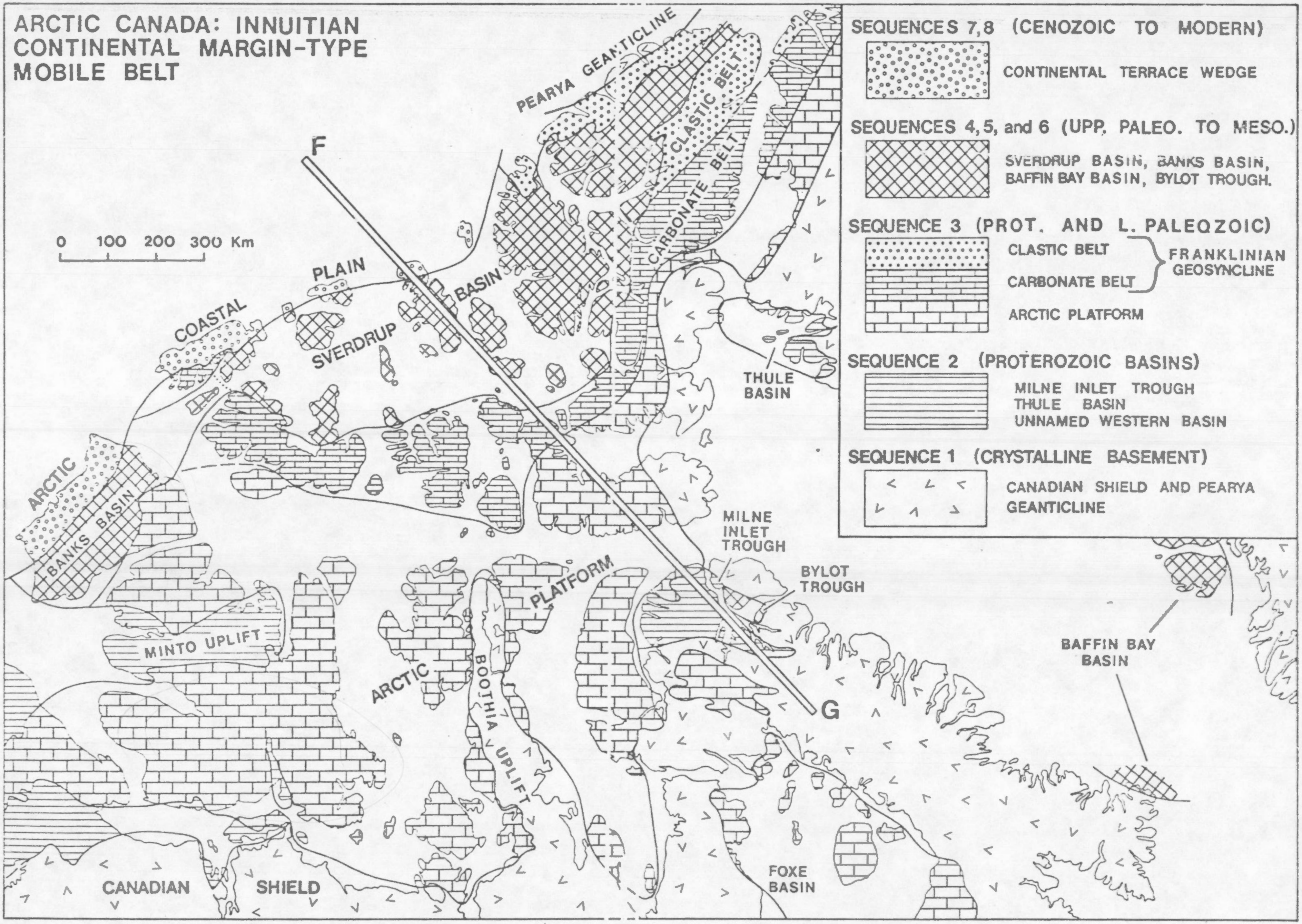

Figure 2. Stratigraphic sequences and geological provinces of the Canadian Arctic Islands. FG is location of cross-section (Fig. 3) and tectonic time diagram (Fig. 4). Sequence 8 is not shown; at sea it is water-covered and on land is thin alluvium.

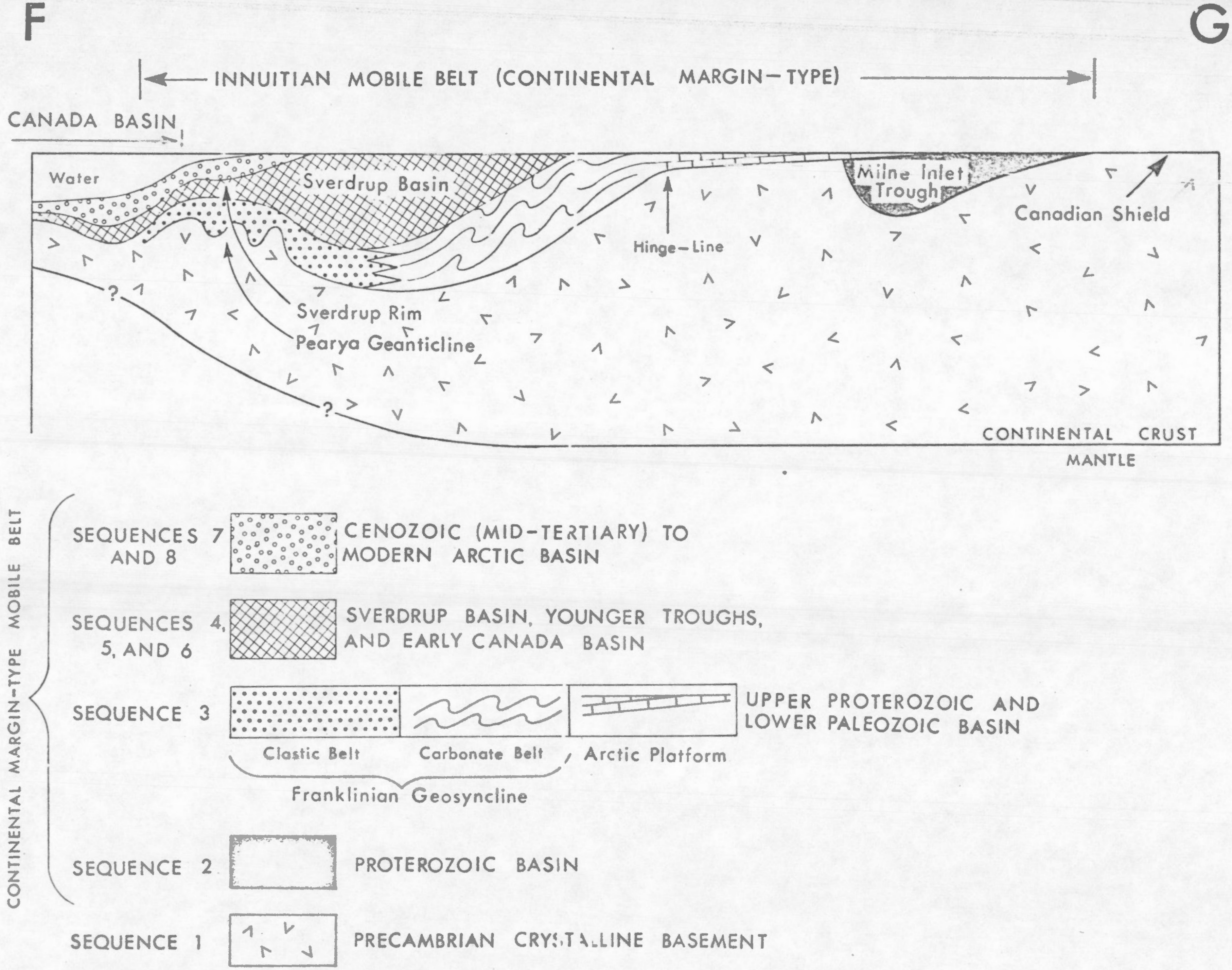

Figure 3. Cross-section of the Innuitian Mobile Belt in the Canadian Arctic Islands. Location is shown in Figure 2. Age relationships of sequences are shown in Figure 4.