Geological Survey of Canada
Commission Geologique du Canada

Open File 2045

TECTONIC FORMATION OF THE FORT NORMAN AREA, MACKENZIE CORRIDOR NORTHWEST TERRITORIES

G.K. Williams

Geological Survey of Canada (Calgary)
3303 - 33 Street NW
Calgary, Alberta T2L 2A7

CONTENTS

Illustrations

Figures

CONTENTS (cont.)

Figures

Tables

INTRODUCTION

This report is one of a series on the geology of the Mackenzie Corridor, an area north of latitude 60°N lying between the Precambrian Shield on the east and the Mackenzie and Richardson Mountains on the west (excluding the Mackenzie Delta). Most such reports deal with a specific time slice, e.g. Precambrian, Cambrian, Devonian . . . (Williams, 1985a, 1986a, 1987). Because of the unique tectonic history of the Fort Norman area, which includes the junction of the Keele Arch and Root Basin, a special report spanning Cambrian to Tertiary history, is warranted.

Tectonic evolution is illustrated mainly by a series of isopach maps. Such maps carry the implication that they record differential tectonic movements that prevailed during deposition. Although such is not always the case, isopach maps are virtually the only tool available; facies maps, surprisingly, are not particularly relevant.

Physiography and tectonic setting

Fort Norman, a small settlement on the north bank of the Mackenzie River, lies near the centre of the area of interest (Figs. 1, 2). The area is bisected by the northern end of the Mackenzie Plain which is a physiographic depression between the Mackenzie Mountains on the west and the Franklin Mountains on the east. Both mountain systems are products of Laramide (Late Cretaceous-Early Tertiary) compression. The Franklin Mountain ranges (Fig. 3) are large folds cut by steep reverse faults; most fault planes dip west. A few similar but generally smaller structures occur within the Mackenzie Plain but, for the most part, sediments of the Mackenzie Plain are only mildly deformed. Fort Norman lies at the dividing line between the southern and northern Franklin Mountains. The former are broad, linear, north by northwest trending ranges, the larger folds expose Proterozoic strata. The latter are narrower, arcuate structures which exhibit a variety of trends including a prominent east-west component, and expose strata no older than the Cambrian. The northern Franklin Mountains are considered to be genetically linked to the Colville Hills to the north (Cook and Aitken, 1973; Cook, 1983).

The map area contains parts of several tectonic elements (Fig. 2), of central concern in this report is the Keele Arch and its relation to the Root Basin. The term 'Root Basin' in conventional usage, applies to a depocentre that existed from Ordovician through Middle Devonian time (Gabrielse, 1967; Morrow and Cook, 1987). There is some evidence that this basin may have continued to be strongly negative into Late Paleozoic time (Williams, 1986b). Also, as will be brought out in this study, the Root Basin is only a segment of an earlier negative feature – the Mackenzie Trough – a system of north trending grabens or half grabens that extended from Root Basin in the south to Colville Hills in the north (Williams, 1987).

The Keele Arch (Cook, 1975) originated at a time near the Silurian/Devonian boundary. Significant uplift and erosion over the crest of the arch occurred in Early Devonian time (early phase of the Keele Arch), during the long post-Devonian to pre-Cretaceous interval, and near the Early/Late Cretaceous boundary (late phase of the arch). That part of the Keele Arch (Fig. 2) that lies south of the transverse fault (Fig. 3) became a local depocentre in Late Cretaceous-Paleocene time.

It was postulated previously (Williams, 1987) that a broad northerly trending zone of crustal weakness transects much of the Mackenzie Corridor linking the Mackenzie Plain and Franklin Mountains with the structures of the Colville Hills (Fig. 1). The rationale is that there is a genetic linkage between structures along this belt, including the Root Basin and Keele Arch, with deep seated faults, including those deduced to underlie the northern Franklins (Cook, 1983) and Colville Hills (Davis and Willott, 1978). This postulated belt of crustral weakness lies parallel to and west of the Fort Simpson (ca. 1.86 Ga) magmatic arc, defined in the subsurface by positive magnetic anomalies (Hoffman, 1987; Hildebrand et al., 1987).

This zone of weakness consists of several fault bounded basement blocks. Adjustments between the various blocks to crustal stresses (which varied from time to time and, at any given time, from north to south) influenced subsidence and structure, hence thickness and facies from Early Cambrian time to the present (Williams, 1987, p. 52). A summary of the main deep seated faults within the map area is given in Figure 4. Note that the dextral component of movement along the Brackett Fault is a post-Paleocene event, so this effect must be discounted on all earlier paleo-reconstructions.

Some exploration highlights

Reconnaissance scale (1:250,000 and 1:500,000) geological maps were produced by Aitken and Cook, 1974; Balkwill, 1971; Cook and Aitken, 1974, 1976a; Douglas and Norris, 1963; and Gabrielse, Blusson and Roddick, 1973.

Drilling in the area stems from 1920, with the spudding of a well, near an oil seep, that led to discovery of the Norman Wells oil field (Fig. 5). A spurt of drilling activity occurred between 1942 and 1944, this was part of the war-time Canol Project (Hume and Link, 1945; Hume, 1954). Drilling has been sporadic since then. Probably the most surprising result was from Shell Keele River L-04, drilled in 1965; some 2 km of Cambrian to Devonian sediments that, on regional considerations should have been present, were missing, and the Cretaceous lay on Cambrian shale. Aside from the Norman Wells oil field (in a Devonian reef) the only significant petroleum show to date is from Candel et al. East MacKay B-45, drilled in 1971 – a recovery of oil from Early Paleozoic carbonates; this oil has been traced to a Late Cretaceous source rock (AGAT, 1977, Feinstein et al., 1988a).

Our understanding of the stratigraphy has developed slowly and is still inadequate, a consequence of the rather complex depositional history which is, in turn, a consequence of a complex tectonic history. Some landmark papers, in chronological order are Hume, 1945; Bassett, 1961; Ziegler, 1967; Tassonyi, 1969; Aitken, Macqueen and Usher, 1973; Norford and Macqueen, 1975; Yorath and Cook, 1981.

Mountains within the map area exhibit some rather unusual geometries and have long intrigued structural geologists. A master control mechanism has been sought in crystalline basement (Goodman, 1954); in a decollement within Cambrian salt (Cook and Aitken, 1973), and in a sub-salt system of northerly trending dextral wrench faults (Cook, 1983). The existence of deep-seated transverse fault systems in the area has been postulated by Cecile, Cook and Snowdon, 1982; and Aitken and Pugh, 1984.

Sources of data

Most of the data used to construct the various maps and cross-sections are derived from a study of logs, cuttings and core of the wells drilled in the area (Fig. 5); the well names appear on Table 1. Locations of measured field sections are shown on Figure 6, with the sources listed in Table 2.

Extensive reflection seismic data over the map area is available through COGLA. Only a fraction of this material was consulted and then only in a cursory fashion. Geophysical data play virtually no part in this report.

PROTEROZOIC FOUNDATION

Depth to crystalline basement is unknown. The map area is underlain by several kilometres of unmetamorphosed Proterozoic sediments. Figure 7 shows some of the nomenclature, approximate thicknesses, dominant lithologies and calculated ages of Proterozoic strata within and adjacent to the map area. Correlation between eastern and western outcrop belts is after Young, 1977, and Aitken and Pugh, 1984. Figure 8 shows distribution of outcrops and, for those wells that penetrated Proterozoic strata, the drilled thicknesses and dominant lithologies.

The Hornby Bay and Dismal Lakes Groups (Kerans et al., 1981) represent two cycles of terrestrial siliciclastic to shallow marine carbonate strata deposited on a westward sloping stable crust. Above these sediments are 3 to 4 km of flood or plateau basalts of the Coppermine Group. These basalts, with the associated Mackenzie dyke swarm, are elements of an important and very widespread mafic igneous event dated about 1.2 Ga (Kerans, 1983).

Sediments of the Mackenzie Mountains supergroup and of the Shaler Group are strikingly similar and are probably products of the same Proterozoic seaway (Young, 1981). They represent a mature orthoquartzite-carbonate suite deposited on a tectonically stable, northwest to west sloping cratonic platform (Aitken, 1982).

The youngest Proterozoic, or Windermere equivalent strata, are known only from the southwest corner of the map area (belt A, Fig. 8) and farther west in the Selwyn Basin. These strata are dominanted by poorly sorted, somewhat immature clastics including glacigene sediments; both shallow and deep water facies occur. Other lithologies are mature sandstones, shallow water dolomite, and iron formation. Rapid facies and thickness changes are characteristic of most units. In contrast to older Proterozoic sediments, Windermere aged strata were deposited in an unstable tectonic environment marked by syndepositional faulting and elongate, northerly trending sub-basins. These extensional features have been attributed to the onset of rifting (Eisbacher, 1981) an event that . . . . "may have culminated in continental fragmentation in Phanerozoic time" (Young et al., 1979, p. 125).

At Cap Mountain (southeast corner of Fig. 8) some 1830 m of Proterozoic sediments are exposed (Douglas and Norris, 1963; Aitken et al., 1973). These are shallow water sediments dominated by red or purple mudstones. In spite of the confident assertion (Aitken and Pugh, 1984, p. 141) that these rocks . . . "can be related, on a unit-to-unit basis, with part of the Hornby Bay - Dismal Lakes succession . . ." (some 350 km distant) these rocks at Mt. Cap are undated. They may indeed be as old as the Hornby Bay Group, or they could, in part or in total, be Windermere correlatives, as suggested by Gabrielse et al., 1973 (p. 28).

On Figure 7, subsurface column, a suggested surface to subsurface correlation is indicated. With the exception of Sammons H-55, located within sight of outcropping Katherine quartzites, all subsurface identifications must be regarded with scepticism. The basalt (Nogha O-47 and Bele O-35) probably belongs to the Coppermine Group, however, basalts occur higher, e.g. in the upper part of the Little Dal Group (Norris and Aitken, 1982), in the Copper Cycle (Ruelle, 1982), and within Windermere equivalent strata (Gabrielse et al., 1973, p. 158). For the remaining wells, the identifications are even more tentative – which brings up the subject of the Fort Norman Structure (Fig. 8). Aitken and Pugh, 1984, on the basis of subsurface correlations, deduced the presence of this Proterozoic flexure, or fault, northwest-side-down. As will become evident later, there is substantial evidence for transverse faults, both ancient and modern, in the Fort Norman area. Aitken and Pugh's Fort Norman Structure may well be valid, but the evidence on which it was deduced is extremely tenuous.

The rifting event that preceeded and influenced Windermere sedimentation is known from the Selwyn Basin, west of the map area (Eisbacher, 1981; Young et al., 1979; Park and Aitken, 1986; Yeo, 1981). Extensional tectonics continued, or were reinstated in that area, as well as the Richardson Trough, through part of Paleozoic time (Cecile, 1982). An extentional tectonic regime also influenced the map area through much of Paleozoic time (This report; Williams, 1987; AGAT, 1977). By analogy, therefore, Late Proterozoic, rift-related structures might be anticipated within the map area. Perhaps the deep seated, northerly to northeasterly trending faults deduced by Davis and Willott, 1978 (Colville Hills), and by Cook, 1983 (northern Franklin Mountains) originated at this time. In this regard the Proterozoic history of the Mackenzie Arch is pertinent.

In the southwest corner of Figure 8 (Mackenzie Mountains) the paleogeology at the sub-Cambrian erosion surface is depicted by belts A (youngest), B and C (oldest Proterozoic strata). The belt of deepest sub-Cambrian erosion (C) marks the axis of the Mackenzie Arch, as redefined by Aitken et al., 1973, p. 37. In the opinion of these authors the arch was not positive at the time of deposition of the Mackenzie Mountains supergroup; these sediments (belts B and C, Fig. 8) appear to thicken uniformly across the arch. Through Windermere time the axis of the future arch acted as a hingeline, with a greatly increased rate of subsidence to the west (*ibid.*, p. 37). Available field evidence, while compatible with the above interpretation, is not conclusive. It is quite possible that the Mackenzie Arch was positive through Late Proterozoic time. Some seismic sections in the Fort Norman area indicate west to east thickening of Proterozoic strata (unpubl. data), this suggests that the arch was indeed positive. By implication, then, the north-south graben system underlying the Fort Norman area, evident from Cambrian and younger strata, may have been inherited from a Late Proterozoic tectonic regime.

PALEOZOIC SUBSIDENCE HISTORY

Lower and Middle Cambrian
(Saline River, Mount Cap and Mount Clark Formation)

Following a long period of uplift and erosion a late Early into middlde Cambrian marine transgression spread over most of the Mackenzie Corridor. Within the main depocentres the section consists of a basal sandstone (Mount Clark or Old Fort Island Formation) followed by shale with, in some areas, carbonate interbeds (Mount Cap Formation) capped by halite (Saline River Formation). Over the main arches the thin Cambrian section consists of mixed and interbedded fine clastics, anhydrite and dolomite; subdivision into the foregoing formations is difficult to impossible.

Throughout much of Early Paleozoic time an extensional tectonic regime prevailed within areas west of the map area (Cecile, 1982). A similar although less drammatic tensional regime prevailed throughout much of the Mackenzie Corridor (AGAT, 1977; Williams, 1987). A large part of the evidence for block-faulting is from the geometry of Cambrian sediments within the map area (Figs. 9, 10). The basic geometry is believed to be tilted blocks bounded by deep-seated down-to-the-west listric faults, as depicted by the cross-section. The term 'Mackenzie Trough' (Douglas, 1970) is used in preference to 'Mackenzie Plain Depocentre' (Williams, 1987). The term 'McConnell Arch' was introduced by Williams (1987). The term Redstone 'Arch' (Gabrielse, 1967; Gabrielse et al., 1973) is used in preference to 'Mackenzie Arch' which, in current usage (Aitken et al., 1973), is reserved for an earlier arch whose axis, although parallel, lay somewhat farther east (Fig. 8).

Figure 9 is modified from map 3 of Williams, 1987; since the latter was published two significant bits of information became available. In K'Alo B-62 the Cambrian section is over 1 km thick (T.D. is in the basal sandstone); this tends to confirm the concept of a boundary fault east of the Mackenzie Trough. The basal shale section in Tunago N-37 has been reinterpreted to be Cambrian, rather than Proterozoic (on the basis of fossil fragments); therefore the section here is over 1 km thick.

The thick (over 1 km) deposits within the Mackenzie Trough are flanked on the east by a much condensed section only a few tens of metres thick. As yet it cannot be documented that this radical west-to-east thinning is a consequence of syndepositional faulting, it remains however, a tenable hypothesis. To the west, the limit of these Cambrian strata appears to be a consequence of syndepositional tilting, intra Cambrian erosion as well as post-Cambrian erosion (Williams, 1987).

Uppper Cambrian – Middle Ordovician
(Franklin Mountain Formation)

Following the Cambrian salinity crisis normal marine circulation was restored and a thick blanket of carbonate was deposited over virtually the entire Mackenzie Corridor. This Late Cambrian transgression was more extensive than the Lower to Middle Cambrian event, covering basins and arches alike. Within basins the basal contact is conformable and gradational. Over arches not covered by the previous transgression the relationship appears to be onlap over eroded Cambrian or Precambrian terrain.

This extensive carbonate, mostly dolomite, is known as the Franklin Mountain Formation (Norford and Macqueen, 1975) over most of the Mackenzie Corridor, including most of the map area (Fig. 11). In the southwest, over and flanking the Redstone Arch, it is known as the Broken Skull Formation (Gabrielse et al., 1973). The Franklin Mountain Formation consists almost entirely of shallow water dolomite; the upper ± third is distinctive in that it contains many beds of white chert (Macqueen, 1970).

Following deposition of the Franklin Mountain Formation there was a widespread regression (Fig. 11). Almost all of the Mackenzie Corridor, including the map area, stood above sea level through most or all of Middle Ordovician time; this is the sub-Mount Kindle erosion interval (Norford and Macqueen, 1975). Franklin Mountain isopachs, therefore, reflect not only differential subsidence during deposition but differential uplift during pre-Mount Kindle erosion. Determining the syndepositional subsidence pattern is further complicated by the fact that, over the area occupied by the future Keele Arch, later episodes of erosion not only removed the Mount Kindle Formation but eroded deeply into, locally completely through, the Franklin Mountain Formation.

Figure 12 portrays basin-arch architecture that is the net result of Franklin Mountain subsidence plus pre-Mount Kindle erosion. To construct this interpretative map, it is necessary to ignore those thickness measurements that have suffered post-Mount Kindle erosion (future Keele Arch, these values will be considered later). In spite of sparse control it is apparent that the subsidence pattern resembles that of Early and Middle Cambrian time (Fig. 9).

Figure 13, a geological map of the pre-Mount Kindle erosion surface, confirms that the Mackenzie Trough remained relatively negative during the Middle Ordovician regression. The preserved extent of the upper cherty unit of the Franklin Mountain Formation follows, in a general way, the outline of the trough. On the flanking arches this unit had been eroded prior to Mount Kindle deposition (see Fig. 1 of Macqueen, 1970).

To the south the Mackenzie Trough merges with the Root Basin. In the deep part of the Root Basin, which lies south of the map area, marine conditions prevailed throughout Ordovician time and there is no widespread unconformity below the Mount

Kindle Formation. Throughout Middle Ordovician time, while most of the map area was above sea level, at least 500 m of carbonate with minor sandstone (Sunblood Formation) was accumulating within the Root Basin which was connected, south of the Redstone Arch, with the Selwyn Basin (see Fig. 10(3) of Ludvigsen, 1975).

In summary, the tensional tectonic regime that had prevailed throughout Early and Middle Cambrian time continued, without notable changes, through Late Cambrian into Middle Ordovician time. The cross-sections of Figure 14 display the geometry.

Upper Ordovician – Lower Silurian
(Mount Kindle Formation)

The Mount Kindle Formation (Norford and Macqueen, 1975) consists entirely of dolomite. Equivalent strata within the Root Basin and over the southern end of the Redstone Arch are within the Whittaker Formation (Douglas and Norris, 1961); the Whittaker Formation represents a longer time span including some Middle Ordovician time (Fig. 11). In the deep parts of the Root Basin the Whittaker Formation includes basinal limestone and shale (Ludvigsen, 1975).

Over most of the map area the upper surface of the Mount Kindle Formation is an erosional unconformity below either the Silurian-Devonian Delorme Group or the Lower Devonian Bear Rock Group. Only within the deeper parts of the Root Basin is the upper contact transitional (Gabrielse et al., 1978, p. 73). The iospachs of Figure 15, therefore, reflect erosion as well as syndepositional tectonics.

In the central part of the map area the Mount Kindle edge outlines the area of the future Keel Arch (Cook, 1975); presumably the formation originally covered this entire area with a thickness in the order of 100 m. This is deduced from the fact that the formation has the same lithology on all sides of the zero line. There are no indications from the facies to indicate the presence of a Keele Arch during Mount Kindle deposition.

From the isopachs one can make no firm deductions regarding tectonic behavior during Mount Kindle time over much of the map area because thinning over the highs is primarily a consequence of post-Mount Kindle erosion. Possibly the northern part of the Mackenzie Trough had ceased to exist as a negative feature. The map does confirm, however, that the Root Basin continued to be strongly negative. A transverse (easterly trending) deep seated, down on the south fault zone (Keele Fault Zone, Figure 4) probably was active through Mount Kindle time.

Upper Silurian – Lowest Devonian
(Delorme Group)

In this report 'Delorme' is used as a group term, embracing all strata between the Mount Kindle (or Whittaker) Formation and the Bear Rock Formation or Group. This is a complex group of lithologies including carbonates, evaporites and clastics, mostly of shallow water origin but including some basinal deeper water strata. In terms of older nomenclature (Douglas and Norris, 1960) the Delorme Group includes the Delorme (carbonates and shale), Camsell (carbonate breccia after evaporites) and the Sombre (dolomite) formations (Fig. 11).

West of the Redstone Arch where Delorme strata grade to a deeper water facies the Group is known to range from Upper Silurian into Lower Devonian (Perry and Lenz, 1978). Within the map area the strata are virtually unfossiliferous, hence many correlation problems exist. In many field sections, especially within the Root Basin, there is considerable doubt concerning the upper and lower contacts. Delorme thickness values of Figure 16 must, therefore, be regarded as approximations.

One source of possible error lies in subsurface identifications. In several wells flanking the Keele Arch strata included in the Delorme Group would be identified as Franklin Mountain Formation by some geologists. The suspect beds consists of very sandy dolomite or clastic-rich anhydrite, a lithology that better fits the Delorme paleogeography, as interpreted in this report. An additional uncertainty concerns the Tsetso Formation (Meijer Drees, in press), a thin clastic-rich suite of anhydrite, dolomite and redbeds within the Great Bear sub-basin southwest of Great Bear Lake. These undated beds differ from the overlying Bear Rock anhydrites mainly by their clastic content. The Tsetso Formation may be younger than any part of the Delorme Group and belong instead, with the Bear Rock Group.

Regardless of these many uncertainties isopachs of Figure 16 display the essential elements of uplift or subsidence through Late Silurian-Earliest Devonian time. The change from earlier maps is dramatic. A segment of the earlier Mackenzie Trough has been transformed into the Keele Arch.

Although no shore-line deposits are known the zero isopach probably approximates the northeastern extent of the Delorme sea. The sandy nature of the strata near the zero-edge, with sporadic chert pebbles, suggests proximity to an upland with considerable relief (Williams, 1971). Thus the northern half of the map area was occupied by a broad upland termed the Norman Wells High (Williams, 1975) the westernmost part of which is called the Carcajou Salient (Aitken et al., 1982). That part of the uplift from which Mount Kindle beds were completely removed by erosion is known as the Keele Arch (Cook, 1975).

In contrast, in the southern part of the map area the tectonic pattern displayed by previous maps remained essentially the same. The southern portion of the Mackenzie Trough, now known as the Root Basin, remained strongly negative, accumulating up to 2 km of sediments. As previously postulated (Williams, 1987), an extensional tectonic regime continued to prevail in the south; the north, in contrast may have undergone some compression. The Brackett Fault and, probably, the Keele Fault zone were active, down on the south, through Delorme time (Figure 4, panel 4).

Lower and Middle Devonian
(Bear Rock Group and Hume Formation)

The Hume and Bear Rock Group are lumped together because there is no consistent criterion, applicable over the entire map area, to map a time-significant contact between the two. The time represented by this package ranges from Emsian into

Eifelian. The Hume Formation consists of limestone and shaly limestone; equivalents in the Root Basin are the Headless shale and Nahanni limestone. Bear Rock is here used as a group term to embrace various more or less time-equivalent formations: the Bear Rock anhydrite, Arnica dolomite and Landry limestone. Relationships are shown on Figure 11. The upper boundary of the isopached package is an abrupt change from limestone to black shale. The lower boundary of the Bear Rock Group varies from transitional within the Root Basin to possibly unconformable where the Delorme Group is thin (e.g. Great Bear sub-basin) to definitely unconformable where the Delorme Group is missing. There is some evidence (not reviewed here) that the package progressively onlapped the Norman Wells High (Fig. 16), only the uppermost Bear Rock strata plus the Hume Formation covered the apex of the high or extended farther east. The original extent of the package probably includes the entire map area. The zero isopach (Fig. 17) is a consequence of post-Devonian – pre and/or intra-Cretaceous uplift and erosion.

For several reasons many of the thickness values of Figure 17 are of questionable value with respect to determining subsidence history: a) The upper contact is a drowned carbonate surface, therefore not necessarily a consistent stratigraphic horizon; b) Within the Great Bear sub-basin the upper part of the package is the pre-Cretaceous erosion surface; c) Within the Root Basin the position of the Bear Rock/Delorme contact is controversial; and d) There has been, undoubtedly, some solution of evaporites in the outcrop areas.

In spite of all the uncertainties Figure 17 indicates that the tectonic pattern of differential uplift and subsidence seen on the Delorme map continued into Middle Devonian time.

There are three interesting asides of uncertain significance. First, note that the belt of facies change, anhydrite (east) to dolomite (west) trends northwest across the Root Basin, seemingly indifferent to the isopach lines. Apparently factors other than the rate of subsidence influenced facies (climatic?). Second, there are some units within the southern part of the map area that can be correlated over wide areas but cannot be identified in the northwest. Examples are the Arnica Platform dolomite of the Root Basin and its approximate equivalent, the Ernestina Lake marker, farther east (Williams, 1975). Another example is the Ebbutt break of Law, 1971, which can be mapped north into the Great Bear sub-basin but cannot be identified northwest of the Norman Wells High. Third, the upper part of the Hume Formation in Tate J-65 consists of porous dolomite (a DST yielded a trace of oil). This is the only known occurrence of dolomite at this stratigraphic level anywhere north of the southern end of the Nahanni Range. A relationship (if any exists) between this anomaly and the Keele structure is not apparent.

Latest Middle Devonian

(Hare Indian and Kee Scarp formations)

Following the drowning of the widespread Hume limestone (probably induced by an eustatic event) the Mackenzie Corridor was invaded from the northeast by a clastic wedge. These clastics, shale, silt and fine sand with some limestone, are known as the Hare Indian Formation. The basal unit, the Bluefish Member, is a potential petroleum source rock (Snowdon et al., 1987).

The seaward or southwestern edge of this clastic wedge, in plan view, formed a sinuous pattern of lobes and reentrants; in vertical profile the edge was fairly steep, sloping from the intertidal zone down into water over 100 m deep within 50 km, or less. The tops of some of the banks were capped by a shallow water limestone – the Kee Scarp platform. Before the Hare Indian clastics could prograde entirely across the drowned Hume platform they were immoblizied by another deepening event. This second event, also probably a eustatic event, gave rise to the Kee Scarp reefs (Williams, 1985b, 1986a).

A segment of the seaward edge of the Hare Indian mud bank lies within the map area. Figure 18 shows the belts of outcrop or pre-Cretaceous subcrop of the Hare Indian and Kee Scarp formations. This map also shows the thickness of the Hare Indian Formation; note that, east of the subcrop belt, the isopachs are hypothetical restorations.

If there is any relationship between the geometry of the Hare Indian mud bank, or the development of the Kee Scarp reefs, to the tectonic setting of the area, that relationship is not obvious. The preserved portion of the seaward edge of the Hare Indian bank totals about 1000 km (see map of Middle Devonian facies belts, Williams, 1986a); the fact that a segment of the bank more or less coincides with the axis of the Keele Arch may be fortuitous.

Upper Paleozoic

No map is supplied for this interval. Only a thin eastward thinning wedge of the Upper Devonian Imperial Formation survived pre-Cretaceous erosion. This surviving wedge, up to about 800 m thick consists of marine shale, siltstone and fine sandstone. A thin limestone (Jean Marie-like limestone; Fig. 26) is present about 600 m above the base. By comparison with what is known about the Jean Marie and similar limestones farther south (Williams, 1986b) this limestone can be expected to be replaced seaward by a lime-rich clinoform decending to within a few metres of the base of the Imperial Formation. The argillaceous limestone that outcrops along the Norman Range known as the Jungle Ridge lentle (Tassonyi, 1969) may be such a marker.

How thick the Imperial Formation was, or how much other section may have once covered the map area prior to pre-Cretaceous erosion, is unknown. Several kilometres may well have been present. Spores and pollen of Carboniferous, Permian and Triassic ages are fairly common in Cretaceous rocks (Yorath and Cook, 1981, p. 31), indicating their former presence west of the map area.

Aside from the fact that the Imperial clastics originally covered the entire map area, and were subsequently removed by pre-Cretaceous or later erosion, the Late Paleozoic tectonic history remains unknown.

THE PRE-CRETACEOUS LACUNA

Within the map area there is no preserved sedimentary record from Late Frasnian to Albian, a time span of some 250 million years. For comparison, the Paleozoic history reviewed to this point spans some 200 million years. Many aspects of the tectonic history must, therefore, remain conjectural, at best, or unknown.

During the pre-Cretaceous lacuna the Mackenzie Corridor, including the map area, was uplifted, tilted westward, and deeply eroded. The erosion surface must have been a virtually featureless plain with few prominent topographic or structural anomalies. For the corridor as a whole a pre-Cretaceous geological map shows a westerly dipping homocline with Precambrian rocks in the east, Carboniferous or younger rocks in the west (see Fig. VIII-35, p. 446, Douglas, 1970). Discordance across the pre-Cretaceous unconformity is low, measurable in metres or fractions of metres per kilometre. There are two notable exceptions to this simple picture – the Bonnet Plume area (Williams, 1988) and over the Keele Arch within the map area.

Figure 19 reveals several anomalies. In the northwest corner the complexity is not a consequence of tectonism but of stratigraphy. Kee Scarp reefs in this area range from 0 to 200 m in thickness, and the Hare Indian Formation ranges from under 50 to over 200 m (Williams, 1985b). It is these thickness changes that produce the somewhat complex pattern. Actually the pre-Cretaceous structure here is a gently dipping homocline.

The ± 20 km offset of the subcrop belts across the Brackett Fault is a post-Paleocene phenomon. However, down-on-the-south movement during the post-Devonian erosional episode is indicated by the subcrop pattern, at least along the northeastern segment of this fault.

The remaining anomalies are related to the tectonic history of the Keele Arch; they will be discussed later (Second phase of the Keele Arch).

CRETACEOUS-TERTIARY HISTORY

Background, Mackenzie Corridor as a whole

Although present distribution is patchy, a consequence of erosion, Lower Cretaceous (Albian) strata may have once covered virtually all of the Mackenzie Corridor (see Figs. 9, 10, Jeletzky, 1971). In some areas the basal unit is a thin, mature, shield-derived, in part nonmarine sandstone of Aptian to Early Albian age (e.g. Gilmore Lake Member, Yorath and Cook, 1981). By Albian time the seaway was extensive, linking the Arctic with the rest of Western Canada. Into this seaway clastic detritus spread north and east, derived from the rising Cordillera orogen far to the southwest. Only the fine grained distal facies of this Albian package is preserved; the near shore facies were uplifted and cannibalized. Formation names for this package include the Fort St. John Group in the south (Stott, 1982), Arctic Red or Horton River formations in the north (Yorath and Cook, 1981).

Over most of the corridor there is evidence for profound regression near the Lower/Upper Cretaceous boundary (see Table 1, Dixon, 1986). In the south this is manifest by the seaward (eastward) spread of nonmarine conglomerates and sandstones of the Dunvegan Formation followed by a long hiatus (Stott, 1982). West of the map area the transition from Arctic Red shale to Trevor sandstone (both marine) represents a regression. North of the map area, near the Arctic coast, there is an erosional unconformity and long hiatus above the Horton River Formation (Yorath and Cook, 1981).

Upper Cretaceous strata are much less widespread, although Late Cretaceous seas probably covered most of the Mackenzie Corridor (see Figs. 11-20, Jeletzky, 1971). In the southern Yukon are thin remnants of marine to nonmarine clastics of the Kotaneelee and Wapiti formations (Stott, 1982). In the Bonnet Plume area fluviatile beds are preserved in a structural basin (Norris and Hopkins, 1977). Near the Arctic coast are marine shales of the Smoking Hills and Mason River formations, the former is renowned for a high content of organic material (Yorath and Cook, 1981). Within the map area nearly two kilometres of marine and nonmarine Late Cretaceous to Paleocene shale and molasse are preserved within a local downwarp (Brackett Basin).

Cretaceous – Tertiary of the map area

Yorath and Cook, 1981, provide a comprehensive description of the Cretaceous-Tertiary of the map area. More recent work has focused on the Brackett Basin, a review (Sweet et al.) is in press, an in depth report by the same authors is in preparation. Interpretations in this present report that differ in significant respects from Yorath and Cook, 1981 are discussed in the Appendices. The strata are illustrated by several cross-sections, for which Figure 20 is an index map.

Figure 21 shows the present distribution of Cretaceous strata; the patchy distribution, is a consequence of Tertiary to Recent erosion. This map also shows the main physiographic-tectonic elements: the Peel Trough on the northwest, the Great Bear Basin on the southeast, and, in between, the Keele Arch in its Cretaceous guise (Yorath and Cook, 1981, Fig. 3). The Brackett Fault divides the arch into two segments which have somewhat different tectonic histories. The North Keele Arch remains positive to the present whereas the South Keele Arch evolved into a Late Cretaceous-Tertiary depression known as the Brackett Basin (Sweet et al., in press). Figure 22 shows the nomenclature and correlation of the above areas . . . note the demotion of the Sans Sault from a formation to an informal member of the Arctic Red Formation (see Appendix 1).

Peel Trough

The bulk of the Peel Trough fill is Albian, the uppermost beds range into Turonian. No pre-Albian Cretaceous beds are definitely known, although some may be present (see Tassonyi, 1969, p. 134). The fill of the trough is almost entirely marine and fine grained, mostly shale and mudstone (Arctic Red facies) with, in the upper part of the section, interbeds of fine grained lithic sandstone (Trevor facies). At the base of the Arctic Red Formation is a ± 60 m thick sandy unit (Fig. 23). Although dominated by mudstone this basal unit contains several thin sandstone layers; abundant carbonaceous debris, traces of coal and scattered pebbles suggest a nonmarine influence. The transition from the Arctic Red (shaly) to Trevor (sandy) facies occurred within Albian time. There appears to be no widespread sedimentary break anywhere in the section, no unconformity at or near the Early/Late Cretaceous boundary. However, a local break has been seen in outcrops on Mountain River, west of the map area (68°28'N, 129°11'W) where conglomeratic sandstone with coal seams occurs within the Trevor facies (Yorath and Cook, 1981, p. 24); the age of these coaly beds is probably Cenomanian (*ibid.*, p. 66, C-10033).

Great Bear Basin

The Cretaceous sediments of this area (Fig. 24) are remarkably similar to those of the Peel Trough. The bulk of the sediments are Albian but they range into Late Cretaceous. They are dominated by shales and mudstones but fine grained lithic sandstones become increasingly common in the upper part. There is a basal sandy zone. Were it not for the intervening Keele Arch, where no Albian strata are present, Peel Trough nomenclature would be appropriate. There are three main differences: 1) The basal sandy unit is about twice as thick and includes, at its base, a significant amount of quartzose, coarse, pebbly sandstone (Unit A of Yorath and Cook, 1981). 2) The upper, or Trevor facies seems to contain a higher proportion of sand (quantitative data are not available). 3) A thin mature, quartzose sandstone occurs within the uppermost part of the section, this may indicate a significant erosional break within the section.

Keele Arch

As depicted on Figure 21 the Keele Arch is the broad, sinuous belt over which no definitely dated Albian strata are known to occur. North of the Brackett Fault there are only a few scattered outcrops of sandstone or pebbly sandstone, these are of questionable Albian or Cenomanian age (Yorath and Cook, 1981, p. 27; the Unnamed Basal Cretaceous Sandstone).

South of the Brackett Fault, in contrast, there are up to 2 km of Late Cretaceous to Paleocene strata. These sediments display an intertonguing transition from a coarse, nonmarine facies in the west to marine shale in the east. Relationships are displayed on cross-sections, Figures 25 to 31. The Slater River and East Fork formations are marine shales. The Little Bear Formation, mostly sandstone and mudstone, is mostly marine but contains coal seams in the upper part. The Summit Creek Formation, almost entirely nonmarine, contains thick conglomerate layers, tuff, and coal seams. It is interpreted to have been deposited as a northeastward prograding alluvial deposit (Sweet et al., 1988).

Tectonic implications

Albian

The Cretaceous Keele Arch (Fig. 21) is a broad sinuous belt that includes not only the ancestral Keele Arch (Fig. 15) but areas to the northeast and south. This Cretaceous pattern shows an obvious spatial coincidence with the Early Paleozoic Mackenzie Trough (Figs. 9, 12, 13). As will become evident later, it may be misleading to call this an arch, a term that suggests a simple broad uplift.

The term 'Peel Trough' may also be somewhat misleading in that it implies a linear depression that lay in juxtaposition with the source area. The Arctic Red and Trevor sediments are fine grained distal deposits of what must have been an extensive seaway. The non-silty Horton River shales some 500 km farther north represent even more distal deposits of the same seaway.

Yorath and Cook (1981), believed that the lower or oldest Albian strata of the Peel Trough lapped out from west to east against the rising Keele Arch. Aitken et al., 1982 believed that Albian sediments became more sandy in the same direction, providing more evidence for a rising Keele Arch. Both of these concepts are examined in Appendix 1 and found to be unsatisfactory. Cretaceous strata do thin rapidly from west to east but this is because of Tertiary-Recent erosion. There is insufficient data to document any depositional thinning or significant facies changes. If the north Keele Arch existed through Albian time, it need not have been as broad a feature as depicted on Figure 21.

The western limit of the Cretaceous fill of the Great Bear Basin coincides with the McConnell and St. Charles Ranges, and this limit is taken as the eastern margin of the Keele Arch. Yorath and Cook (1981), evisioned the Great Bear Basin as a true depositional basin with its western shore defined by the Keele Arch (*ibid.*, p. 41). This concept is examined in Appendix 3 and found to lack convincing evidence, although it cannot be ruled out.

Well log markers of the Great Bear Basin strata are shown on Log section, Figure 24. With the exception of the basal sandy unit (units A and B of Yorath and Cook, 1981), all markers diverge westward at rates of a few metres per kilometre. An example is shown on Figure 32. This geometry suggests deposition on a stable, westward tilting platform. From subsurface data there is no reason to believe that a Keele Arch lay to the west, at least through most of Albian time.

There is, however, some evidence for tectonic influence within the Great Bear Basin, Figure 33 shows thicknesses of the basal sandy zone (units A and B, Log section, Fig. 24) as well as the southern, depositional limit of the basal pebbly sandstone (Unit A). Although control points are few it could be argued that these earliest deposits of the Albian sea were locallized by a northeast trending topographic (tectonically controlled?) low. This concept leads to the inference that the Brackett Fault was active (southeast side down) in earliest Albian time.

The foregoing leads to an hypothesis: The Albian sea spread across the entire map area. A more or less uniform blanket of shale, mudstone and sandstone was deposited, thinning eastward but with little change in facies. At a time near the Early/Late Cretaceous boundary a northerly trending belt, which included the ancestral Keele Arch, was uplifted above sea level. Albian strata over the uplift were cannibalized and redeposited, spreading mainly eastward into the Great Bear Basin. As will be discussed later (Keele Arch, late phase) this was probably an eastward-directed compressive event, an episode of the Columbian-Laramide orogeny. Such a history is implied by diagrams of Dixon, 1986, see his Figures 5, 17 and 18.

Late Cretaceous-Tertiary (Brackett Basin)

The outline of the Brackett Basin is shown on Figure 34. The stippled area is the area in which the Latest Cretaceous-Paleocene Summit Creek Formation is known to be preserved plus an extension to the northeast that is covered by Quaternary lake deposits (Sweet, et al., in prep.). The areas of preserved Little Bear and Slater River Formation are somewhat greater.

This is a late Laramide (post-Paleocene) structural depression, a part of the Mackenzie Plain. Sediments of the Brackett Basin have been folded and faulted in harmony with the Laramide structures. However, in common with the Mackenzie Plain as a whole, Brackett Basin sediments are for the most part relatively undeformed (cross-sections, Figs. 24-31). Although the outline shown on the map, Figure 34 is partly a consequence of post-Paleocene downwarping, the Brackett Basin was also a local, syndepositional downwarp.

The oldest dated strata above the eroded Paleozoic floor of the basin are Cenomanian in age (see Appendix 2). Isopachs of the Slater River Formation are shown on Figure 35. West of the 200 m isopach the data are reliable, here the Little Bear/Slater River contact is a consistent marker (Log section, Fig. 36). Farther east the data are not a reliable guide to the basin's geometry because in this area the Little Bear Formation loses its identity (Log section, Fig. 37). The isopachs clearly show rapid west to east thinning and a high (thin) in the vicinity of Fort Norman. This phenomenon can be explained by onlap over a pre-existing topographic ridge. This condition is depicted by cross-sections Figures 27 and 28. Paleontological data permits, but does not confirm, this hypothesis. Other possibilities include syndepositional warping and/or intra Cretaceous erosion. Slater River shales are dark, only slightly silty and contain one to three thin radioactive beds rich in organic matter, including fish scales, one such marker occurs on Log section, Figure 37. The geometry of these radioactive markers could be interpreted as east to northeast dipping clinoforms (e.g. cross-sections, Figs. 26, 27). Thus the lithology, thickness and internal geometry of the Slater River Formation suggests deposition within a stagnant submarine depression, of which the Paleozoic floor had several hundred metres of topographic relief.

The character of the Little Bear Formation is illustrated on Log sections, Figures 36 and 37. In Figure 36 it is seen that several sandstone markers extend in a north-northwest direction for at least 75 km. This is strong evidence that depositional strike lay in this direction. Log section, Figure 37 shows that west to east thinning is in part a consequence of a shale-out of the upper sands (compare logs of I-70 and B-45). This section also demonstrates the rapid west to east decrease in grain size (regardless of the accuracy of the suggested correlation). Figure 38 shows isopachs of the Little Bear Formation; the depocentre appears to lie near to and parallel to the Brackett Fault.

For younger strata of the Brackett Basin there are insufficient data to construct isopach maps. Observations of facies changes, current direction indicators, and local thinning trends within the Summit Creek nonmarine strata tend to confirm that the Brackett Basin was a tectonically controlled, northeast trending depocentre (Sweet et al., in press).

Cross-sections, Figures 25 to 31 illustrate some of the structural complexities that lie below the Late Cretaceous cover. The patch of preserved Hare Indian Formation in the east (C-21, cross-section, Fig. 28) shows that the South Keele Arch is not a simple uplift but a composite feature, probably of several folds or fault blocks. This subject will be addressed later (Keele Arch, late phase) after a review of some Laramide structures.

ANTECEDENTS OF SOME FRANKLIN MOUNTAIN STRUCTURES

Several of the Laramide (Late Cretaceous-Early Teritary) structures of the Franklin Mountains coincide with more ancient tectonic features. This coincidence, sometimes obvious, sometimes subtle, bespeaks a genetic relationship.

McConnell Range

This is the easternmost major fold of the Franklin Mountains (Fig. 34), it is also the largest. Measured at the level where the Hume Formation outcrops, the structure measures 240 km in length by up to 35 km in width. Basically it is one large asymmetrical anticline with a broad, western limb dipping about 10° to the west and a narrow, steep eastern limb, near vertical in places. Over most of its length Cambrian strata are exposed in the core, patches of Proterozoic rocks are exposed in two culminations (Fig. 8). High angle thrust faults occur along most, but not all segments of the anticline crest; most of these dip west, but east dipping faults also occur. These core faults have little horizontal displacement. Some segments of the east limb display an unfaulted Cambrian to Cretaceous sequence; other segments, as mapped, show an east-directed thrust fault, with Paleozoic over-riding Cretaceous rocks. Nowhere is there evidence for any significant (i.e. kilometres) eastward displacement (see cross-sections AB and CE, Wrigley Map 1373A, Douglas and Norris, 1963). The amount of vertical uplift of the range is up to 3.5 km.

The belt now occupied by the McConnell Range coincides (within the limits of control) with a long-lasting down-on-the-west hinge belt. This hinge was most active through Cambrian to Middle Ordovician time (Figs. 9, 12) but activity continued at least through Middle Devonian time (Maps figs. 16, 17). The most dramatic thickness changes that have been documented occur in the southeast corner of Figure 12 where the Franklin Mountain Formation thins from over 783 m (S 61, Hill and Gully) to 257 m (E-11) within a distance of 15 km. Unfortunately there are few published measured sections along the length of the range; with more mapping other manifestations of this hinge belt may come to light.

It was speculated previously (Williams, 1987) that the fundamental tectonic control may have been a deep seated listric fault zone that, in response to an extensional regime, was active, down on the west, through most of Paleozoic time. In response to Laramide compressive stresses, up on the west movement created the McConnell Range.

St. Charles Range

Actually this range (Fig. 34) is unnamed. Mount St. Charles, which lies on the north bank of the Great Bear River, is one of its more conspicuous peaks.

This range lies north of and on trend with the McConnell Range. A structural saddle with at least 1.5 km of relief separates the two ranges, also the St. Charles Range seems to be offset a few kilometres to the east.

Like the McConnell Range, this is basically an asymmetrical anticline for most of its length with the steepest limb, vertical in places, on the east. Thrust faults, dipping either east or west, disrupt the crest; there is little horizontal displacement. The St. Charles structure differs from the McConnell Range in being narrower (only a few km) having only about 1.5 km in vertical relief, and in the fact that no strata older than the Franklin Mountain Formation are exposed. It is probable that the exposed Mt. St. Charles structure is detached from its Proterozoic roots at the level of Cambrian salt.

Through Cambrian to Middle Ordovician time the down on the west hinge belt that underlies the McConnell Range continued northward beneath the present St. Charles Range (Maps figs. 9, 12). From Silurian time on the vertical movement along the St. Charles segment of the hinge was reversed to up on the west, and this lineament coincides with the eastern limb of the early Keele Arch (Fig. 15).

Figures 17 and 18 show no anomalies associated with the St. Charles lineament, this may be due to lack of control. However, there is a pre-Cretaceous anomaly. Near the northern end of the range, on its western flank, a basal Cretaceous sandstone, tentatively dated Late Albian or Cenomanian, lies with an angular unconformity on the cherty unit of the Franklin Mountain Formation (Yorath and Cook, 1981, p. 27). The discordance at the unconformity is about 15°, the pre-Cretaceous dip was to the east. It was noted (*ibid.*, p. 28) that "This is the only locality observed in the northern Franklin Mountains wherein localized pre-Cretaceous deformation and erosional truncation can be demonstrated". Although one cannot draw any firm conclusions from one outcrop, this is an indication that the St. Charles structure of today (a post-Paleocene feature) is probably a rejuvenation of an intra- or pre-Cretaceous fold. By analogy, other Franklin Mountain Ranges may be rejuvenated structures for which no evidence has yet been found.

MacKay Range

The Mackay Range is a narrow, sinuous, northwesterly trending, faulted, asymmetrical (east limb steepest) anticline. Saline River redbeds outcrop in the core. The west flank contains a normal (for the region) succession of Franklin Mountain to Hare Indian formations overlain unconformably (but as far as we know with structural concordance) by Cretaceous. The east flank, mainly covered, contains a few exposures of steeply east dipping Bear Rock breccia. No Paleozoic strata younger than the Hume Formation are known to occur on the east flank. Although there are no exposures of a thrust fault, the range is assumed to have been thrust eastward over Cretaceous strata. Paleocene strata east of the range are steeply tilted, confirming late Laramide movement of structure (Yorath and Cook, 1981, p. 36).

A clue to earlier instability in the vicinity of the MacKay Range is seen in the nature of the pre-Bear Rock unconformity. At the northern end of the range at least 130 m of Franklin Mountain dolomite are present; near the southern end, some 20 km distant, the Bear Rock lies directly on Saline River redbeds. This is a north to south down cutting rate in the order of 6 or 7 m per kilometre. MacKay B-45, drilled about 4 km east of the range, encountered 495 m of Franklin Mountain dolomite (base not reached). Although no Bear Rock beds are present in the well, the pre-Bear Rock east to west down cutting between the well and the range must have been in the order of 125 m per kilometre. Figure 39 shows some of the implications:

1. The MacKay Range area was uplifted and deeply eroded prior to Bear Rock deposition.
2. Devonian sediments, presumably including Bear Rock, Hume, Hare Indian and Imperial formations, covered the entire area.
3. The sense of movement reversed and, prior to Cretaceous deposition, the paleo MacKay structure, at the level of the Saline River Formation top, had been 'unbent'.
4. Laramide structural movement was focused by the ancient feature, be it a fault or zone of flexing.

Another clue to pre-Laramide structure can be seen on the pre-Cretaceous geological map, Figure 19. A pronounced down on the west flexure is evident from the narrow subcrop belt of the Bear Rock, Hume, Hare Indian and Canol formations. In an east-west direction across this belt the pre-Cretaceous unconformity shows a discordance in the neighbourhood of 5 degrees, some 2000 m of section is cut out over a distance of 25 km. Over most of the Mackenzie Corridor, in contrast, the discordance is measured by metres or fractions of metres per kilometre. The coincidence of this pre- or intra-Cretaceous flexure with the MacKay Range raises the possibility of a deep seated master fault zone.

Norman Range

The Norman Range is a northwesterly trending, gently west dipping slab some 100 km long by 10 to 15 km wide exposing Saline River to Imperial formations. To the southwest is a synclinorium floored by Cretaceous strata; to the northeast are uplands formed by folded and faulted Paleozoic carbonates. Two of several possible interpretations of the Norman Range structure were discussed by Cook and Aitken, 1976b. Their favoured interpretation, involving a structural high at the level of the Proterozoic indicates ". . . the possibility that an earlier structure related to the pre-Upper Cretaceous Keele Arch . . . has localized the Laramide structure (*ibid.*, p. 322). Their second interpretation involved a low angle east-directed thrust fault, originating in the Saline River salt, with movement in the order of 13 km.

Later a gravity profile was run across the range to test the above models, the thrust model was preferred (Lawton, 1982). One of the reasons for this preference was that to satisfy the faulted Precambrian model requires the presence of up to 2 km of some dense rock, such as dolomite, at depth, and this was considered to be unlikely. Given the paucity of our knowledge of Proterozoic history in this area such a surmise is unwarranted; there may well be 2 km of Little Dal dolomite, or even several kilometres of Coppermine basalts. At any rate, the gravity data do not rule out the existence of a deep-seated structure under the Norman Range.

The nature of the pre-Bear Rock unconformity along the Norman Range is similar to that described for the MacKay Range. Along the length of the range the preserved thickness of the Franklin Mountain dolomite decreases from 625 m in the north to only 145 m at Bear Rock, which is the south end of the range. As in the MacKay area, a nearly complete section of Franklin Mountain dolomite, including the cherty unit, is preserved in the mountains only a few kilometres northeast of Bear Rock (Mahony Lake sheet, Cook and Aitken, 1974). Also, as is the case of the MacKay Range, the Norman Range lies along the pre-Cretaceous hinge line (Fig. 19). Before being offset by the Brackett Fault this hinge line extended in a north-northwest direction for 150 km, from north of Norman Wells to the south end of the Keele Arch.

THE KEELE ARCH

Early Phase

Following Cook, 1975, the early Keele Arch can be defined as the oval area within the zero isopach of the Mount Kindle Formation (Fig. 15). In pre-Mount Kindle time this area had been part of the Mackenzie Trough (Fig. 14). The origin of the arch cannot be precisely dated but it must have been rising through much of Delorme time (Late Silurian-Early Devonian) and it was not entirely onlapped until late Bear Rock time (Emsian-early Eifelian). This is a span of some 30 to 40 million years. This arch was a local feature superimposed on a much larger positive element – the Norman Wells High (Fig. 16).

The cross-sections of Figure 40 show structural profiles across the arch using the pre-Bear Rock erosion surface as datum, this discounts any topographic relief and is, therefore, only an approximation. These cross-sections show that there was over one kilometre of uplift along the axis of the dome. It must be emphasized that we have no firm evidence regarding the Proterozoic strata. The sections, as drawn, show concordance, however, as a thick Cambrian salt layer is known to be present, and salt flow can be anticipated, the actual profile on the Proterozoic surface may have been quite different. In view of the relatively shallow burial of the salt (1-2 km) and the competent nature of the overlying rocks the concept of a salt anticline as the prime cause of the uplift seems to be out of the question. Proterozoic strata must have been involved.

Reconstructing the early history of the arch is rendered difficult by what happened later. As discussed earlier it can be assumed that the arch was onlapped by the Bear Rock-Hume package (which buried any remnant topography of the erosion surface) which was itself covered by a kilometre or more of post-Hume Paleozoic sediments. Then, as part of a craton-wide pre-Cretaceous uplift, these post-arch sediments were deeply eroded and, over a part of the early Keele Arch, the pre-Bear Rock erosion surface was exhumed and no doubt modified to some extent. Over that area wherein all Delorme and Bear Rock strata are missing (Fig. 19) we cannot know for certain how much erosion occurred during the early phase of uplift. In the following discussion it has been assumed that most of the erosion into the Franklin Mountain and older strata took place during the pre-Bear Rock event; the main supporting evidence is that pre-Bear Rock erosion down to the Saline River Formation can be documented on the west side of the early Keele Arch (Cook, 1975).

Some details concerning the nature of the early Keele Arch are revealed by Figures 41 and 42, the former shows the pre-Bear Rock/Delorme geology, the latter shows isopachs of the eroded Franklin Mountain Formation. The right lateral offset along the Brackett Fault is a late Laramide phenomenon, therefore to visualize the paleoconditions this offset must be mentally retracted. The uplift has the appearance of a broad northerly plunging dome except at its southern end where, it must be emphasized, we have the least control. The axis of the uplift lies, approximately, along the MacKay and Norman ranges.

Cross-section, Figure 43, drawn along the north-south axis is a diagrammatic depiction of the junction of the Keele Arch and Root Basin. In spite of sparse control in this area, it is virtually necessary to postulate a deep seated, transverse, down on the south fault zone. During the 30 to 40 million years that the Keele Arch was rising and being eroded the Root Basin was subsiding. Marine sediments in the order of 3 or 4 km thick in the deep part of the Root Basin represent the depositional vacuity plus the eroded section of the Keele Arch (Franklin Mountain, ± 1 km; Mount Kindle, ± 1 km; Delorme, ± 1.5 km). That dramatic thickness changes occur over short distances is documented by data from the Summit Anticline (Fig. 3). On the north end of the anticline Delorme sediments are either missing or too thin to be mapped separately (Aitken and Cook, 1974, p. 13). In Red Dog K-29, drilled some 30 km to the south on this anticline, Delorme strata are over 1300 m thick (base not reached). The precise position or orientation of the Keele Fault zone is unknown, no obvious clues to its existence are revealed by the surface geology.

Figures 41 and 42, can be interpreted in different ways, depending on what one accepts as the most likely datum. Assuming the erosion surface was fairly level, the maps depict a north plunging half dome, assuming high topographic relief, they depict the valley of a major river. These two concepts combined lead to the idea of an incised core of a rising half dome through which a southward flowing river delivered clastics and dissolved salts into the Root Basin. Delorme strata of the Root Basin are characterized by an abundance of siliciclastics, varying from floating quartz grains to very sandy or silty dolomite or anhydrite in the northern end of the basin to the siltstones of the basinal strata at the southern end (orange siltstone member of the Cadillac Formation). Morrow and Cook, 1987, have demonstrated that these clastics had a northern source, *ibid.*, p. 52 and Fig. 22). The Franklin Mountain Formation of the Keele Arch and adjacent areas is a logical source for some of these clastics, it is silty, sandy, and contains thick beds of chert. The occurrence of angular chert pebbles within as well as erratic thickness variations of the Delorme dolomite seen on the southeastern flank of the Keele Arch (Williams, 1971) attest the presence of considerable topographic relief during deposition (see Keele Passage, Williams, 1975).

Near the southern end of the Keele Arch salt solution and collapse structures would have contributed to the development of the axial valley. What effects would the dissolved salt have had on the water chemistry of the Root Basin? If the salt layer extends far to the south within the Root Basin (Fig. 9) it would have been deeply buried early in Paleozoic time. For example, by the end of Hume time it would have lain at a depth of about 5 km (Isopach maps, Figs. 12, 15, 16, 17). What effect would this thermal conduit have had on heat flow under the Keele Arch? Perhaps the thermal anomalies found by Feinstein et al., 1988b, over the southern end of the Keele Arch are in some way related to this question (reflectance values are plotted on Figures 25 to 31).

Keele Arch, Late Phase

Consideration of material discussed in the chapters on antecedents of some of the Franklin Mountain structures and on the Cretaceous-Tertiary history of the map area has led to the following hypothesis:

> Compressive stresses that created the Franklin Mountains occurred in two discreet pulses. The final pulse, which gave the ranges their present form, occurred in post-Paleocene time – a late Laramide event. The previous pulse occurred in Late Albian time. It is a moot point whether this first pulse should be ascribed to the latter part of the Columbian orogeny or to an early phase of the Laramide orogeny. To avoid that issue it is herein called the St. Charles event, so named because it is on this range that evidence for this event can be seen in outcrop (Yorath and Cook, 1981, p. 27).
>
> The degree or intensity of folding produced by this event is not clear, but within a ± 100 km wide belt, now generally known as the Keele Arch, uplift amounted to at least 1 km. Albian sediments, which had previously been continuous across this belt, were entirely stripped away exposing and modifying the original pre-Cretaceous erosion surface. Possibly all of the major Franklin Mountain structures originated at this time.
>
> The youngest sediments affected by the folding are dated Late Albian. The oldest beds above the unconformity are dated questionably Late Albian to Cenomanian. It follows that this must have been a brief event, possibly only one or two million years duration. (Germs of this hypothesis are found in Cook, 1975 (his Fig. 41.3) and Yorath and Cook, 1981, p. 27, 28.)

On Figure 44 are two different methods of depicting the late phase of the Keele Arch. The small inner area, after Cook, 1975, is the area from which all Devonian were removed prior to Cretaceous deposition; this is actually only the core of a broader uplift. The larger area, after Yorath and Cook, 1981, is the area over which no Lower Cretaceous strata occur. The pattern of this larger area is remarkably similar in outline, but of reversed polarity, to the Early Paleozoic Mackenzie Trough (Figs. 9, 12, 13). This coincidence is surely significant. The same deep seated faults that reacted under tension to create the Mackenzie Trough reacted under compression to create the Keele Arch, late phase. According to the foregoing hypothesis the late phase Keele Arch is not, strictly speaking, an arch, but an uplift belt made up of a series of linear folds and faulted folds created during the St. Charles compressive event. South of the Brackett Fault three such folds are known or suspected: the St. Charles Range, the Police Island structure (cross-sections, Figs. 27, 28) and the ancestral MacKay Range. North of the Brackett Fault, aside from the ancestral Norman Range, the histories of the St. Charles and late Laramide events cannot be differentiated.

During the St. Charles event the old, pre-Bear Rock erosion surface of the South Keele Arch was exhumed, folded, faulted, and further eroded. The lesser fold axes depicted on Figure 41, those lying east of the main axis, are diagrammatic representations of the resulting subcrop pattern. Erosion and tectonics combined to create several hundred metres of topographic relief on the erosion surface. Probably some structures continued to rise during the Late Cretaceous inundation. Late Cretaceous strata onlap these structures, e.g., Police Island structure. (Subsurface data permit but do not confirm these concepts.)

Down on the south motion across the Brackett Fault during and after Late Cretaceous-Paleocene deposition created the Brackett Basin which preserved the historical record of the St. Charles event over the South Keele Arch. North of the Brackett Fault post-Paleocene erosion has destroyed virtually all of that record.

TRANSVERSE STRUCTURES

The Brackett Fault

Many geologists who have visited Fort Norman and gazed at Bear Rock, the cliff-like southern end of the Norman Range, must have wondered how the range could terminate so abruptly. The Mackenzie River flows serenely by, no rapids . . . no evident constriction, and, south of the river there is only a featureless plain. The simplist explanation would be that the roots of the range lie under the plain and were, for some reason, completely beveled by Late and post-Tertiary erosion.

Cecile et al., 1982, postulated the necessity for "a steeply dipping northeasterly trending tear fault" in an area of the Mackenzie Mountains southwest of Fort Norman, near the southern end of the Plateau Thrust (Fig. 3). The projected trend includes the Gambill Fault (*ibid.*, Fig. 15.3).

Once the concept of transverse faulting is introduced, the solution to the question – what happened to the southern continuation of the Norman Range? – becomes obvious: it is the MacKay Range, which lies some 20 to 25 km to the southwest of Bear Rock. The sense of drag motion on the once abutting ends of the two ranges agrees with dextral offset. Other topographic anomalies that support a transverse fault zone (but are neutral with respect to strike-slip movement) are: 1) the juxtaposition of the plateau-like southern end of the northern Franklin Mountains with the swamp covered northeastern end of the Brackett Basin; this lineament, followed by the Brackett River, marks an elevation change of about 300 m; 2) the northeastern segment of the Gambill Fault, mapped as a thrust fault, northwest side up.

The best evidence for significant dextral offset is provided by subsurface data, specifically Figures 41 and 42. The time of this movement must have been after or during the compression that created the Norman and MacKay ranges, i.e. post-Paleocene.

Dextral movement is only the most recent expression of the Brackett Fault. Activity along one or more segments of this lineament are discernable at several times, going back to the Cambrian:

- Cambrian to Early Ordovician (northeastern segment) down-on-the-north motion, marks the abrupt northern end of the McConnell Arch (Figs. 9, 12).
- Late Silurian-Early Devonian, down-on-the-south motion, coincident with the northwestern end of the Root and Great Bear basins (Fig. 16).
- Late Albian or St. Charles event (northeastern segment) down-on-the-south movement, preserving the Albian sediment of the Great Bear Basin.
- Late Cretaceous-Paleocene (central segment), down-on-the-south movement, creating the Brackett Basin.
- Post Paleocene (central segment), aside from ± 20 km of dextral movement, down-on-the-south movement, preserving the Brackett Basin fill from erosion.

On the basis of the above anomalies plus the various topographic and/or structural features a northeastern trending lineament extends from the central Mackenzie Mountains to Smith Arm of Great Bear Lake (as plotted on the various maps). Evidence for dextral movement is only obvious over the central segment of the lineament. If this is a tear fault (Cecile et al., 1982) over its western portion, it need not be so over its entire length. If the detachment level lies within the Cambrian salt, or deeper, the strike-slip movement could end in a series of thrust faults north of the lineament, such as the Norman and other Northern Franklin Mountain Ranges.

Keele Fault Zone

Whereas the transverse Brackett Fault is manifest by several obvious surface features, the Keele Fault Zone is not, this feature is deduced entirely from isopach maps. However, there may be a connection between this feature and the structurally complex north end of the McConnell Range (see Fort Norman map, Cook and Aitken, 1976a, especially the transverse fault south of Smith Creek, 64°32'N, 124°30'W).

The virtual necessity for down-on-the-south movement is illustrated by isopach maps, Figures 15 to 17 and the cross-section through the junction of the Keele Arch and Root Basin, Figure 43. The isopach patterns are similar on all of the above maps. It seems probable that down-on-the-south movement on both the Brackett and Keele Fault zones occurred from Upper Ordovician to, at least, Middle Devonian time. There is no evidence to document any later movement along the Keele Fault zone, however down-on-the-north movement may have influenced the development and preservation of the Late Cretaceous-Paleocene Brackett Basin.

SUMMARY

A broad, sinuous, northerly trending zone of crustal weakness extends nearly the full length of the Mackenzie Corridor linking the Mackenzie Plain, Franklin Mountains and Colville Hills. Extensional tectonics, beginning in Cambrian time, or earlier, resulted in a network of half-grabens along this belt through Cambrian into Middle Ordovician time. This negative feature is known as the Mackenzie Trough.

Beginning about Late Silurian time the northern part of the Mackenzie Trough was no longer under tension, some parts were under compression. Out of a central segment of the Mackenzie Trough the Keele Arch (early Phase) developed and, by early Middle Devonian time formed a broad, north plunging deeply eroded half-dome. Over the southern, structurally highest end of the half-dome erosion was down to the level of the Cambrian salt horizon. The southern part of the Mackenzie Trough remained strongly negative – this part is known as the Root Basin. The junction between the Keele Arch and Root Basin is a zone of easterly trending, south side down, deep seated faulting – the Keele Fault zone. Throughout much of Late Silurian to Middle Devonian time a south flowing river traversed the breached axis of the Keele Arch and debouched into the Root Basin, delivering the clastic component of the Delorme sediments.

The early Keele Arch, in concert with adjacent previously exposed areas, subsided and was plastered over with Middle Devonian and later Paleozoic sediments. Most of this sedimentary record, possibly several kilometres thick, was eroded away during the pre-Cretaceous lacuna. Except for the fact that there was Corridor wide uplift, westward tilting, and erosion, the detailed tectonic history of this 250 million year time span can never be known.

The map area, in common with the entire Mackenzie Corridor, plus at least the outer belt of the Mackenzie Mountains, was part of an extensive Cretaceous foreland basin. The source area for Cretaceous clastics, the rising Columbian-Laramide orogen, lay far to the southwest, deep within the Cordillera. Within the map area, above a thin and discontinuous basal sandstone, the initial Cretaceous deposits are of Albian age. These spread eastward and northward as a fairly uniform blanket of fine grained distal clastics. A Late Albian compressive event (the St. Charles event) created a broad upland over a belt that coincides, almost exactly, with the ancient Mackenzie Trough. This uplifted belt, which includes the ancient Keele Arch, was eroded and stripped of its Cretaceous cover. The old pre-Middle Devonian erosion surface was exhumed and compressed into a series of northerly trending folds, these are precursors of the modern Franklin Mountains.

The erosion surface left by the brief St. Charles event had a structurally controlled topographic relief of several hundred metres. Late Cretaceous-Paleocene clastics onlapped and buried this uneven surface. The Brackett Fault is a transverse fault of ancient lineage that divides the Keele Arch into north and south segments. Because of down-on-the-south movement during and after Late Cretaceous-Paleocene deposition, up to two kilometres have been preserved over the south part of the former Keele Arch, this depression is known as the Brackett Basin.

REFERENCES

Aitken, J.D.
1982: Precambrian of the Mackenzie Fold Belt - a stratigraphic and tectonic overview; in Precambrian Sulphide Deposits, R.W. Hutchinson, C.D. Spence and J.M. Franklin (eds.); Geological Association of Canada, Special Paper 25, p. 150-161.

Aitken, J.D. and Cook, D.G.
1974: Carcajou Canyon map area (96 D), District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 74-13.

Aitken, J.D., Cook, D.G., and Yorath, C.J.
1982: Upper Ramparts River and Sans Sault Rapids map areas (106 G, H), District of Mackenzie; Geological Survey of Canada, Memoir 388, with maps 1452A and 1453A.

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

Aitken, J.D. and Pugh, D.C.
1984: The Fort Norman and Leith Ridge structures: Major buried Precambrian features underlying Franklin Mountains and Great Bear and Mackenzie Plains; Bulletin of Canadian Petroleum Geology, v. 32, no. 2, p. 139-146.

Applied Geoscience and Technology (AGAT) Consultants Ltd.
1977: Lower Mackenzie energy corridor study, geological component; COGLA no. 51-3-6-2.

Balkwill, H.R.
1971: Reconnaissance geology, southern Great Bear Plain, District of Mackenzie; Geological Survey of Canada, Paper 71-11, 47 p.

Bassett, H.G.
1961: Devonian stratigraphy, central Mackenzie River region, Northwest Territories, Canada; in Geology of the Arctic, v. 1, G.O. Raasch (ed.); Alberta Society of Petroleum Geologists and University of Toronto Press, p. 481-495.

Cecile, M.P.
1982: The Lower Paleozoic Misty Creek Embayment, Selwyn Basin, Yukon and Northwest Territories; Geological Survey of Canada, Bulletin 335.

Cecile, M.P., Cook, D.G., and Snowdon, L.R.
1982: Plateau overthrust and its hydrocarbon potential, Mackenzie Mountains, Northwest Territories; in Current Research, Part A, Geological Survey of Canada, Paper 82-1A, p. 89-94.

Cook, D.G.
1975: The Keele Arch - a pre-Devonian and pre-Late Cretaceous paleo-upland in the northern Franklin Mountains and Colville Hills; in Report of Activities, Part C, Geological Survey of Canada, Paper 75-1, p. 243-246.

Cook, D.G.

1983: The northern Franklin Mountains, Northwest Territories, Canada - a scale model of the Wyoming Province; in Rocky Mountain foreland basins and uplifts, J.D. Lowell (ed.); Rocky Mountain Association of Geologists, Denver, Colorado, p. 315-338.

Cook, D.G. and Aitken, J.D.

1969: Geology Fort Good Hope, District of Mackenzie; Geological Survey of Canada, Map 4-1969.

1973: Tectonics of northern Franklin Mountains and Colville Hills, District of Mackenzie, Canada; in Arctic Geology, M.G. Pitcher (ed.); American Association of Petroleum Geologists, Memoir 19, p. 13-32.

1974: Geology of Norman Wells (96 E) and Mahony Lake (96 F) map areas, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 304.

1976a: Geological maps of Blackwater Lake (96 B) and Fort Norman (96 C), District of Mackenzie; Geological Survey of Canada, Open File 402.

1976b: Two cross-sections across selected Franklin Mountain structures and their implications for hydrocarbon exploration; in Current Research, Part B, Geological Survey of Canada, Paper 76-1B, p. 315-322.

Davis, J.W. and Willot, R.

1978: Structural geology of the Colville Hills; Bulletin of Canadian Petroleum Geology, v. 26, p. 105-122.

Dixon, J.

1986: Cretaceous to Pleistocene stratigraphy and paleogeography, northern Yukon and northern District of Mackenzie; Bulletin of Canadian Petroleum Geology, v. 34, no. 1, p. 49-70.

Douglas, R.J.W. (editor)

1970: Geology and economic minerals Canada; Geological Survey of Canada, Economic Geology Report no. 1.

Douglas, R.J.W. and Norris, D.K.

1960: Virginia Falls and Sibbeston Lake map areas, Northwest Territories; Geological Survey of Canada, Paper 60-19, Maps 1377A and 1378A.

1961: Camsell Bend and Root River map areas, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 61-13, Maps 1375A and 1376A.

1963: Dahadinni and Wrigley map areas, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Paper 62-33, Maps 1373A and 1374A.

Eisbacher, G.H.

1981: Sedimentary tectonics and glacial record in the Windermere supergroup, Mackenzie Mountains, northwest Canada; Geological Survey of Canada, Paper 80-27.

Feinstein, S., Brooks, P.W., Fowler, M.G., Snowdon, L.R., and Williams, G.K.
1988a: Families of oils and source rocks in the central Mackenzie Corridor: A geochemical oil-oil and oil-source rock correlation, in Sequences, Stratigraphy, Sedimentology: Surface and Subsurface, D.P. James and D.A. Leckie (eds.); Canadian Society of Petroleum Geologists, Memoir 15, p. 543-552.

Feinstein, S., Brooks, P.W., Gentzis, T., Snowdon, L.R., and Williams, G.K.
1988b: Thermal maturity in the Mackenzie Corridor, Northwest and Yukon Territories, Canada; Geological Survey of Canada, Open File Report no. 1944.

Gabrielse, H.
1967: Tectonic evolution of the northern Canadian Cordillera; Canadian Journal of Earth Sciences, v. 4, no. 2, p. 271-298.

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

Goodman, A.J.
1954: Tectonics of east side of Cordillera in western Canada; in Western Canada Sedimentary Basin, L.M. Clark (ed.); American Association of Petroleum Geologists, p. 341-354.

Hildebrand, R.S., Hoffman, P.F., and Bowring, S.A.
1987: Tecto-magmatic evolution of the 1.9 Ga Great Bear magmatic zone, Wopmay Orogen, northwestern Canada; Journal of Vulcanology and Geothermal Research, v. 32, p. 99-118.

Hoffman, P.F.
1987: Continental transform tectonics: Great Slave Lake shear zone (ca. 1.9 Ga), northwest Canada; Geology, v. 15, p. 785-788.

Hume, G.S.
1954: Lower Mackenzie River area, Northwest Territories and Yukon; Geological Survey of Canada, Memoir 273.

Hume, G.S. and Link, T.A.
1945: Canol investigations in the Mackenzie River area, Northwest Territories and Yukon; Geological Survey of Canada, Paper 45-16.

Jeletzky, J.A.
1971: Marine Cretaceous biotic provinces and paleogeography of western and Arctic Canada: illustrated by a detailed study of ammonites; Geological Survey of Canada, Paper 70-22.

Kerans, C.
1983: Timing of emplacement of the Muskox intrusion: contraints from Coppermine cover strata; Canadian Journal of Earth Sciences, v. 20, p. 673-698.

Kerans, C., Ross, G.M., Donaldson, J.A., and Geldsetzer, H.J.
1981: Tectonism and depositional history of the Helikian Hornby Bay and Dismal Lakes Groups, District of Mackenzie; in Proterozoic Basin of Canada, F.H.A. Campbell (ed.); Geological Survey of Canada, Paper 81-10, p. 157-182.

Law, J.
1971: Regional Devonian geology and oil and gas possibilities upper Mackenzie River area; Bulletin of Canadian Petroleum Geology, v. 19, no. 2, p. 437-486.

Lawton, D.D.
1982: Report of gravity survey across the Norman Range, Northwest Territories; Geological Survey of Canada, Open File Report no. 855.

Ludvigsen, R.
1975: Ordovician formations and faunas, southern Mackenzie Mountains; Canadian Journal of Earth Sciences, v. 12, p. 663-697.

Macqueen, R.W.
1970: Lower Paleozoic stratigraphy and sedimentology; eastern Mackenzie Mountains, northern Franklin Mountains; in Report of Activities, Part A, Geological Survey of Canada, Paper 70-1A, p. 225-230.

Meijer Drees, N.C.
in press: The Devonian succession in the subsurface of the Great Slave and Great Bear Plains, Northwest Territories.

Morris, W.A. and Aitken, J.D.
1982: Paleomagnetism of the Little Dal lavas, Mackenzie Mountains, Northwest Territories, Canada; Canadian Journal of Earth Sciences, v. 19, p. 2020-2027.

Morrow, D.W. and Cook, D.G.
1987: The Prairie Creek Embayment and Lower Paleozoic strata of the southern Mackenzie Mountains; Geological Survey of Canada, Memoir 412.

Norford, B.S. and Macqueen, R.W.
1975: Lower Paleozoic Franklin Mountain and Mount Kindle formations, District of Mackenzie: Their type sections and regional development; Geological Survey of Canada, Paper 74-34.

Norris, D.K. and Hopkins, W.S., Jr.
1977: The geology of the Bonnet Plume Basin, Yukon Territory; Geological Survey of Canada, Paper 76-8.

Park, J. and Aitken, J.D.
1986: Paleomagnetism of the Katherine Group in the Mackenzie Mountains: Implications for post-Grenville (Hadrynian) apparent polar wander; Canadian Journal of Earth Sciences, v. 23, p. 308-323.

Perry, D.G. and Lenz, A.C.
1978: Emsian paleogeography and shelly fauna biostratigraphy of Arctic Canada; in Western and Arctic Canadian biostratigraphy, C.R. Stelck and B.D.E. Chatterton (eds.); Geological Association of Canada, Special Paper No. 18, p. 133-160.

Ruelle, J.C.L.
1982: Depositional environments and genesis of stratiform copper deposits of the Redstone Copper Belt, Mackenzie Mountains, Northwest Territories, Canada; in Precambrian Sulphide Deposits (Robinson Volume), Geological Association of Canada, Special Paper 25, p. 701-738.

Snowdon, L.R., Brooks, P.W., Williams, G.K., and Goodarzi, F.

1987: Correlation of the Canol Formation source rock with oil from Norman Wells; Organic Geochemistry, v. 11, p. 529-548.

Stott, D.F.

1982: Lower Cretaceous Fort St. John Group and Upper Cretaceous Dunvegan Formation of the Foothills and Plains of Alberta, British Columbia, District of Mackenzie and Yukon Territory; Geological Survey of Canada, Bulletin 328.

Sweet, A.R., Ricketts, B.D., Cameron, A.R., and Norris, D.K.

in press: An analysis of the Brackett coal basin, Northwest Territories, Canada; in Report of Activities, Part , Geological Survey of Canada.

Tassonyi, E.J.

1969: Subsurface geology, lower Mackenzie River and Anderson River area, District of Mackenzie; Geological Survey of Canada, Paper 68-25.

Williams, G.K.

1971: A geological review of part of the Mackenzie Plain, for Western Decalta Petroleum Ltd., COGLA no. 19-1-5-50.

1975: Arnica Platform Dolomite (NTS 85, 95, 96, 105, 106), District of Mackenzie; Geological Survey of Canada, Paper 75-1C, p. 31-35.

1985a: Proterozoic geology of the Mackenzie Corridor; Geological Survey of Canada, Open File Report 1273.

1985b: Kee Scarp Play, Norman Wells area, Northwest Territories; Geological Survey of Canada, Open File Report 1228.

1986a: Middle Devonian Facies belt, Mackenzie Corridor; Geological Survey of Canada, Open File Report 1353.

1986b: Map of Upper Devonian and younger Paleozoic rocks, Mackenzie Corridor, District of Mackenzie; Geological Survey of Canada, Open File Report 1401.

1987: Cambrian geology of the Mackenzie Corridor, District of Mackenzie; Geological Survey of Canada, Open File Report 1429.

1988: A review of the Bonnet Plume area, east-central Yukon Territory (including Snake River, Solo Creek, Noisy Creek and Royal Creek areas); Geological Survey of Canada, Open File Report 1988.

Yeo, G.M.

1981: The late Proterozoic Rapitan Glaciation in the northern Cordillera; in Proterozoic Basin of Canada, F.H.A. Campbell (ed.); Geological Survey of Canada, Paper 81-10, p. 25-46.

Yorath, C.J. and Cook, D.G.

1981: The Cretaceous and Teritary stratigraphy and paleontology, northern Interior Plains, District of Mackenzie; Geological Survey of Canada, Memoir 398.

Young, G.M.

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

1981: The Amundsen Embayment, Northwest Territories; relevance to the upper Proterozoic evolution of North America; in Proterozoic Basin of Canada, F.H.A. Campbell (ed.); Geological Survey of Canada, Paper 81-10, p. 203-218.

Young, G.M., Jefferson, C.W., Delaney, G.D., and Yeo, G.M.

1979: Middle and late Proterozoic evolution of the northern Canadian Cordillera and Shield; Geology, v. 7, p. 125-128.

Ziegler, P.A.

1967: Guidebook for Canadian Cordillera, Field Trip, International Symposium on the Devonian System; Canadian Society of Petroleum Geologists.

APPENDIX 1

The Slater River Formation

The early history of this name is summed up in Tassonyi, 1969. The type area on Slater River, near 64°57'N, 126°15'W, consists of a few small outcrops of shale, there is no reliable thickness of the total section, neither upper nor lower contacts are exposed, and there are no diagnostic fossils. As used by Canol geologists the name was applied to Late Cretaceous shale outcrops near Fort Norman (Slater River Formation in the modern sense) as well as to Albian shales in the Norman Wells area and west to Mountain River (i.e. to all or part of what today would be called Arctic Red Formation).

Yorath and Cook (1981), applied the name to the shale section underlying the Little Bear Formation in the subsurface southwest of Fort Norman (see list, *ibid.*, p. 26). This area, which is the western side of the Brackett Basin, coincident with the southern end of the Keele Arch, is the only area in which the Slater River Formation is mappable. It is a layer of shale plus a thin discontinuous basal sandstone that lies below the Little Bear sandstone and above the Paleozoic erosion surface. The basal contact is an angular unconformity, the age of the underlying strata ranges from Middle Cambrian (Mt. Cap Formation) to Late Devonian (Imperial Formation). The upper contact appears to be conformable, however this point is uncertain. Slater River shale is soft, dark grey, bentonitic and generally non-silty or sandy except near its top and base. Accessory components include ironstone, *Inoceramus* prisms and fish remains; the latter are most abundant in thin radioactive zones. One to three radioactive zones (gamma log markers) occur within the unit; these markers are rich in organic material, including fish remains. The basal few metres, in some wells, contain traces of fine grained sanstone or, rarely, coarser sandstone.

As thus defined the Slater River Formation is over 400 m thick along the southwest edge of the Brackett Basin (a post-Cretaceous erosional limit). It thins rapidly to the north and east. The reason for this thinning is uncertain. It could be a result of onlap over pre-Slater River topography, a result of syndepositional warping, or a result of pre-Little Bear erosion.

The span of time represented by Slater River shale includes Turonian and possibly Cenomanian (D.J. McIntyre, pers. comm., 1988). The overlying Little Bear Formation is known to range from Santonian to Campanian, possibly younger (Yorath and Cook, 1981, p. 30). If the Little Bear/Slater River contact is conformable, as appears to be the case, Slater River shales may range into Santonian. To date no Coniacian aged rocks have been recognized in this area (*ibid.*, p. 44).

The Little Bear Formation can be mapped north from its type area on Little Bear River to about 65°N. It is probable that Slater River-aged shales are also present however, at this latitude, they are underlain by several hundred metres of Albian shale (see Dodo Canyon K-03, Fig. 4 of Yorath and Cook, 1981). In this area, there is no mappable break between Lower and upper Cretaceous shales. Farther northwest Slater River-aged strata fall within the Trevor Formation.

Within the Brackett Basin along, and east of, the Mackenzie River the Little Bear Formation is not mappable, either in outcrop or the subsurface. It is probable that the formation shales out from west to east, therefore shaly strata must be mapped as 'Slater River-East Fork undivided' (Yorath and Cook, 1981, p. 27). Slater River-aged shales are probably present farther east in Brackett Lake C-21 and K'Alo B-62 however no paleontological control is yet available for these wells. Another facies change that occurs within the eastern end of the Brackett Basin is an increase in the thickness and coarseness of the basal sandstone: 25 m thick in Brackett Lake C-21; 334 m thick, including a basal chert pebble conglomerate, in K'Alo B-62; a similar pebbly sandstone, 122 m thick, occurs in Old Fort Point E-30.

In summary the term 'Slater River' is useful only within the western end of the Brackett Basin where the Little Bear Formation is present but Albian shales are absent.

This unit is of Late Cretaceous age. The main unsolved problem is the relationship between these shales and Albian shales of the Arctic Red Formation along the margins of the Keele Arch. Is there an angular unconformity between these shales? Or, do Cretaceous shales onlap the arch, with only Upper Cretaceous shales covering the crest?

Slater River Formation (restricted), Age determinations

A **Outcrops:** (see Figure 45)

F1 North end of St. Charles Range
in Yorath and Cook, 1981, p. 64; C-16887, 1972 by W.W. Brideaux
Late Albian or Cenomanian

Note: this is from the basal Cretaceous sandstone. Sample from an overlying shale has been sampled but not yet processed (per Art Sweet, Sept. 88).

F2 Outcrops, northeast bank of Mackenzie River west of Bear Rock
in Yorath and Cook, 1981, p. 64, 65; C-30980, 1974, C-30981, 1974 by W.W. Brideaux
Late Albian-Turonian

New data in 6-DJM-1988, C-112273 and C-112274, C-112278 and C-112279;
Turonian

F3 North of Kelly Lake
in Yorath and Cook, 1981, p. 65; C-37209, 1975, by W.W. Brideaux
Probably Late Cretaceous

F5 South of Gambill Range
in Yorath and Cook, 1981, p. 58; C-5539, 1976, by W.W. Brideaux
?Albian

F6 Type area on Slater River
in Aitken and Cook, 1974 (Paper 74-13), p. 19; 5-TPC-1971
Near Early-Late Cretaceous boundary

F7 Grotto Creek
in Aitken and Cook, 1974, p. 19, by B.D.E. Chatterton (by pers. comm.)
Turonian

F8 Halfway Islands
in 6-DJM-1988; C-112285 and C-112286
Cenomanian

B **Subsurface**

Bluefish No. 1A
a) in Tassonyi, 1969, p. 109; Stelck
Turonian

Bluefish No. 1A

b) in 5-DJM-1988; 840 ft to 845 ft
Turonian to Santonian
1085 ft to 1245 ft
Turonian (base of Cretaceous at 1618 ft)

Redstone No. 1

in 5-DJM-1988
Turonian to Maastrichtian

Keele L-04

in Yorath and Cook, p. 27, citing A.P. Audretsch (pers. comm., 1979)
Cenomanian-Turonian

Tate J-65

in Yorath and Cook, 1981; p. 71, citing R.L. Cox; Campanian.

Fort Norman K-14

No new data. The Early Cretaceous age from strata above 61 m (W.W. Brideaux in Yorath and Cook, 1981, p. 75) may be spurious, as the Summit Creek Formation is mapped at the surface at this location. Upper Cretaceous pollen and dinoflagellates occur below the 61 m sample (ibid., p. 76).

APPENDIX 2

The Sans Sault 'Formation'

The term 'Sans Sault Formation' conjours up two concepts, both are, probably, misconceptions: 1) that Cretaceous strata of the Norman Wells-Fort Good Hope region are appreciably more sandy than equivalent strata farther west, and 2) that Lowermost Cretaceous strata lap out from west to east from the Peel Trough towards the Mackenzie River.

Sans Sault rapids on the Mackenzie River (65°42'N, 128°46'W) are caused by resistant ledges of Lower Cretaceous sandstone; outcrops on the east bank constitute the type section of the Sans Sault Formation. The term has been applied in several ways (see review in Tassonyi, 1969). The most recent consensus seems to be that the formation is a local sandy facies of Albian strata equivalent to the Arctic Red mudstones and shales (Aitken et al., 1982, Fig. 20). These authors state (*ibid.*, p. 31) that "The sandy facies (Sans Fault Formation) appears to be distributed in an arc about the northwestern termination of the northern Franklin Mountains". In Yorath and Cook (1981), the interpretation is similar except that the Sans Sault is considered to represent only the upper part of the Arctic Red Formation (*ibid.*, Table 2, p. 19), earlier Cretaceous strata are missing, presumably a result of west to east onlap. Both the above concepts lend support to a third concept: the Keele Arch was positive through Albian time (Yorath and Cook, 1981, Fig. 32 and text p. 10 and 43).

To critically examine the above postulates it is advisable to begin with subsurface data. Judging mainly from wire line logs (samples are inevitably badly contamined by cavings) there is little if any west to east increase in sand content within the Arctic Red Formation except, possibly, in the lowermost part of the section. In all wells drilled in the general Norman Wells-Good Hope region Cretaceous beds are dominantly shale and mudstone but there is a basal sandy zone, on average about 60 m thick. Although this basal unit is also dominantly mudstone it contains several well developed sand layers.

Most of the sand is very fine grained, coarse sand to pebbles occur near the base. This unit is probably marine (abundant glauconite) but carbonaceous material is abundant and, in some wells, there are traces of coal. There is, in other words, no good reason for any change in nomenclature of Albian strata between the Arctic Red River area (Arctic Red Formation) and the Mackenzie River area (Sans Sault Formation) except for the basal sandy zone. The term Sans Sault Member for the basal sandy unit would be appropriate. Tassonyi (1969, Fig. 11, wells 7, 11 and 16 to 19) applied the term 'Sans Sault Formation' to this basal sandy unit, unfortunately he also used the term to represent the entire Arctic Red equivalent (*ibid.*, well 9).

From the foregoing discussion it would seem that there might be some mix-up at the Sans Sault type section. The unit is described in detail in Yorath and Cook, 1981, p. 53. The section is faulted, but the displacement was considered to be negligable (*ibid.*, p. 22). The measured thickness is 422 m which (by coincidence?) is about that of the total preserved Cretaceous section in nearby wells. The section is dominantly mudstone or shale (at least 2/3 of the section is mudstone or covered). Most of the sandstone of the type section occurs in the upper 108 m, above the fault. This fact, in view of subsurface data, suggests that the fault in the outcrop section may be significant, with the basal sandy zone thrust over younger mudstones. Palynological data supports such a repeat. Macrofossils from above the fault were dated as "Early Lower Albian" by J.A. Jeletzky (in Yorath and Cook, 1981, C-84788 and C-84792; p. 60). Microfossils, mostly from below the fault, were described by W.S. Hopkins as a "scabby assemblage . . . far from completely age diagnostic" but probably no older than Middle Albian (*ibid.*, p. 59). Yorath and Cook (*ibid.*, p. 23) elected to believe that ". . . the discrepancy reflects differences in precision . . . between the two fossil groups . . ." More likely there is no real discrepancy; the section is faulted with well dated Early Albian basal sandy strata thrust over mudstones which, on "scabby" evidence, may be Middle Albian. Bearing on the age of oldest Cretaceous strata in this general area is the questionable identification of Aptian or Neocomian spore cases from a core from Loon Creek No. 1 (Tassonyi, 1969, p. 134; Yorath and Cook, 1981, p. 23).

In summary, there is little or no evidence that Lower Cretaceous strata become more sandy in a belt west of the Keele Arch and the concept of west to east onlap onto a rising (in Albian time) Keele Arch contradicts the best available paleontological evidence. The concept that the Keele Arch became emergent prior to or during Albian time may, or may not, be correct. Lithologic or paleontological evidence in favour of this concept (implicit in published discussion of the Sans Sault Formation) is equivocal (at best) or non-existent. The term 'Sans Sault' should be demoted to the basal sandy member of the Arctic Red Formation. This sandy member is analogous, in lithology, stratigraphic position, and age, with the Martin House Formation of the Snake River area, with the Langton Bay Formation of the Anderson-Horton Plains, and with units A and B of the unnamed Cretaceous strata east of Great Bear Lake (see Charts 1 and 2, Yorath and Cook, 1981).

APPENDIX 3

Western edge of the Great Bear Basin, Albian time

The Cretaceous fill of the Great Bear Basin is abruptly terminated, by post-Cretaceous erosion, along the eastern flanks of the St. Charles and McConnell Ranges. These ranges are essentially large asymmetrical anticlines with steep eastern limbs. In the vicinity of the map area the Hume Formation is present all along the eastern limbs, and is the youngest preserved Paleozoic unit; in other words, pre-Cretaceous erosion cut into but not through the Hume carbonates. Outcrops of Cretaceous strata east of the ranges are few and far between; no exposures of the Hume/Cretaceous contact have been seen. From the scant outcrop data it would appear that Cretaceous and Hume strata are structurally concordant along the east flank of the ranges.

The north-south line formed by the eastern limbs of the St. Charles-McConnell Ranges is taken as the eastern limb of the South Keele Arch by Yorath and Cook (1981), who believed that the Keele Arch was positive, and formed a low topographic barrier during part of Albian time, isolating the Great Bear Basin from the western Albian seaway. As evidence they cite east to west onlap as well as the condensed nature of the Cretaceous section in the belt east of the St. Charles Range (*ibid.*, p. 10, 41 and Fig. 32). The evidence is probably based on the fact that the basal sandstone (unit A) was not seen in outcrop and is known to be missing at one locality (Big Smith Creek, 64°41'N, 124°23'W, *ibid.*, p. 38). Thus the concept of east to west onlap of the basal pebbly sandstone is tenable, however this one example could be a local phenomenon. An outcrop of pebbly quartzose sandstone, about 150 m above an exposure of Hume limestone, occurs about 64°29'N, 124°15'W (Williams, 1971).

There is no other evidence for a condensed section along the eastern flank of the Great Bear Basin. On the Fort Norman and Mahony Lake maps (Cook and Aitken, 1974, 1976a) there is a covered belt about 3 km wide (unit K-1) between the east flanks of the

St. Charles and McConnell Ranges and the lowest Cretaceous outcrops. Given the high east dips characteristic of this belt there is ample room for the anticipated ± 1000 m of Cretaceous strata to be present. Well log markers of wells drilled within the Great Bear Basin show that all Cretaceous units, with the exception of the basal sandy unit (A and B) consistently diverge from east to west, and yield no hint that a Keele Arch existed in Albian time.

APPENDIX 4

Subsurface correlations, Brackett Basin

Subsurface markers used in cross-sections Figures 25 to 31 agree fairly closely with the formation picks of Yorath and Cook (1981), with the following exceptions:

1) The Summit Creek/East Fork contact has been lowered so that the "Sandstone Member of the East Fork Formation" of Yorath and Cook is within the Summit Creek Formation. This unit contains coal and pebbly beds, a lithology more akin to the Summit Creek than to the East Fork Formation. In several wells (J-65, B-30, L-66 and B-45) this revised contact marks the top of a soft (bentonitic?) shale that is particularly prone to caving, hence is a marker on caliper logs. This change may place some marine strata within the basal Summit Creek Formation, but it in no way alters the age relationship of the two formations.

2. Fort Norman K-14 spuds in the Summit Creek Formation, which outcrops near the well site. Underlying strata, between the Summit Creek beds and Paleozoic carbonate breccia do not outcrop; this section is mapped as ?East Fork Formation (Sweet et al., in press). Concerning material from the well samples (which he dated as Lower Cretaceous), W.W. Brideaux stated ". . . All of the samples yield Upper Cretaceous pollen and several contain assemblages of Upper Cretaceous dinoflagellates . . . One sample . . . contains only Upper Cretaceous pollen. These Upper Cretaceous assemblages are interpreted as being caved material" (in Yorath and Cook, 1981, p. 76).

   On cross-section Figure 28 the above interpretation has been ignored, and the strata are assumed to belong to the East Fork Formation.

3) Paleontological data that post-dates the Yorath and Cook memoir indicates that in Redstone No. 1 (cross-section Fig. 30) and Bluefish No. 1A (cross-section Fig. 29) the oldest Cretaceous are Late Cretaceous, probably Turonian but possible Cenomanian (D. McIntyre, pers. comm., 1988). This tends to confirm that the Slater River Formation of the Brackett Basin is entirely Late Cretaceous.

TABLE 1

List of (most) wells of report area listed by NTS area

95 N	63°N-64°N, 124°W-126°N	A-12	Johnson
		D-65	Dahadinni
		G-51	Silvan Plateau
		G-70	W. Wrigley
		I-46	Cloverleaf
		I-70	Dahadinni
		M-43A	Dahadinni
95 O	63°N-64°N, 122°W-124°W	E-11	Blackwater
		G-60	Fish Lake
		I-15	Ochre River
		I-54	Wrigley
96 B	64°N-65°N, 122°W-124°W	G-52	Blackwater Lake
		H-61	St. Charles Creek
		I-54A	Blackwater Lake
96 C	64°N-65°N, 124W°-126°W	A-28	Keele South
		A-49	Bluefish
		A-53	Windy Island
		B-30	Stewart
		B-45	East MacKay
		E-30	Old Fort Point
		I-01	Keele River
		I-70	Little Bear
		J-65	Tate
		K-14	Fort Norman
		K-29	Red Dog
		K-71	Bluefish
		L-04	Keele River
		L-21	Little Bear
		L-66	Police Island
		N-30	Great Bear River
		N-62	Keele
		P-78	Redstone
		No. 1A	Bluefish
		No. 1	Redstone
96 D	64°N-65°N, 126°W-128°W	A-37	Slater River
		K-53	Blueberry Creek
		O-33	Mirror

TABLE 1 (cont.)

96 E 65°N-66°N, 126°W-128°W	B-48	Norman Wells 36x
	F-27	Hoosier
	H-71	Oscar Creek
	K-03	Dodo Canyon
	N-22	Hoosier Ridge
	No. 1	Loon Creek
	No. 2	Loon Creek
	No. 1	Vermilion Ridge
96 F 65°N-66°N, 124°W-126°W	B-62	K'Alo
	C-21	Brackett Lake
	D-61	Wolverine Creek
	H-34	Whitefish River
	I-74	Mahony Lake
	K-76	Whitefish River
96 G 65°N-66°N, 122°W-124°W	F-62	Lost Hill Lake
	G-22	Losh Lake
	M-04	White
	M-07	Russel
	N-70	Grey Goose
97 K, L 66°N-67°N, 124°W-128°W	A-40	Good Hope
	N-37	Tunago
	O-35	Bele
	O-47	Nogha
106 H 65°N-66°N, 128°W-130°W	F-57	Maida Creek
	H-55	Sammons
	J-05	Hanna River
	K-68	Carcajou
106 I 66°N-67°N, 128°W-130°W	N-39	Ontadek

TABLE 2

Sources for cited outcrop sections

	Map Abbreviation	Reference
GSC Reports	62-33	Douglas and Norris, 1963
	70-1A	Macqueen, 1970
	73-9	Aitken et al., 1973
	74-13	Aitken and Cook, 1974
	74-34	Norford and Macqueen, 1975
	81-1A	Morrow and Meijer Drees, 1981
	DMW	Morrow, in press
Industry Reports	BW61	Brady and Wissner, 1961
	S61	Shell Oil Company, 1961
	S and A 59	Sproule and Associates Ltd., 1959
Other	MQ-3, Bear Rock	Macqueen, pers. comm.
	MN	W.S. Mackenzie, unpublished notes
	L. Jacques	Thickness calculated from Map 4-1969, Cook and Aitken, 1969

Figure 1

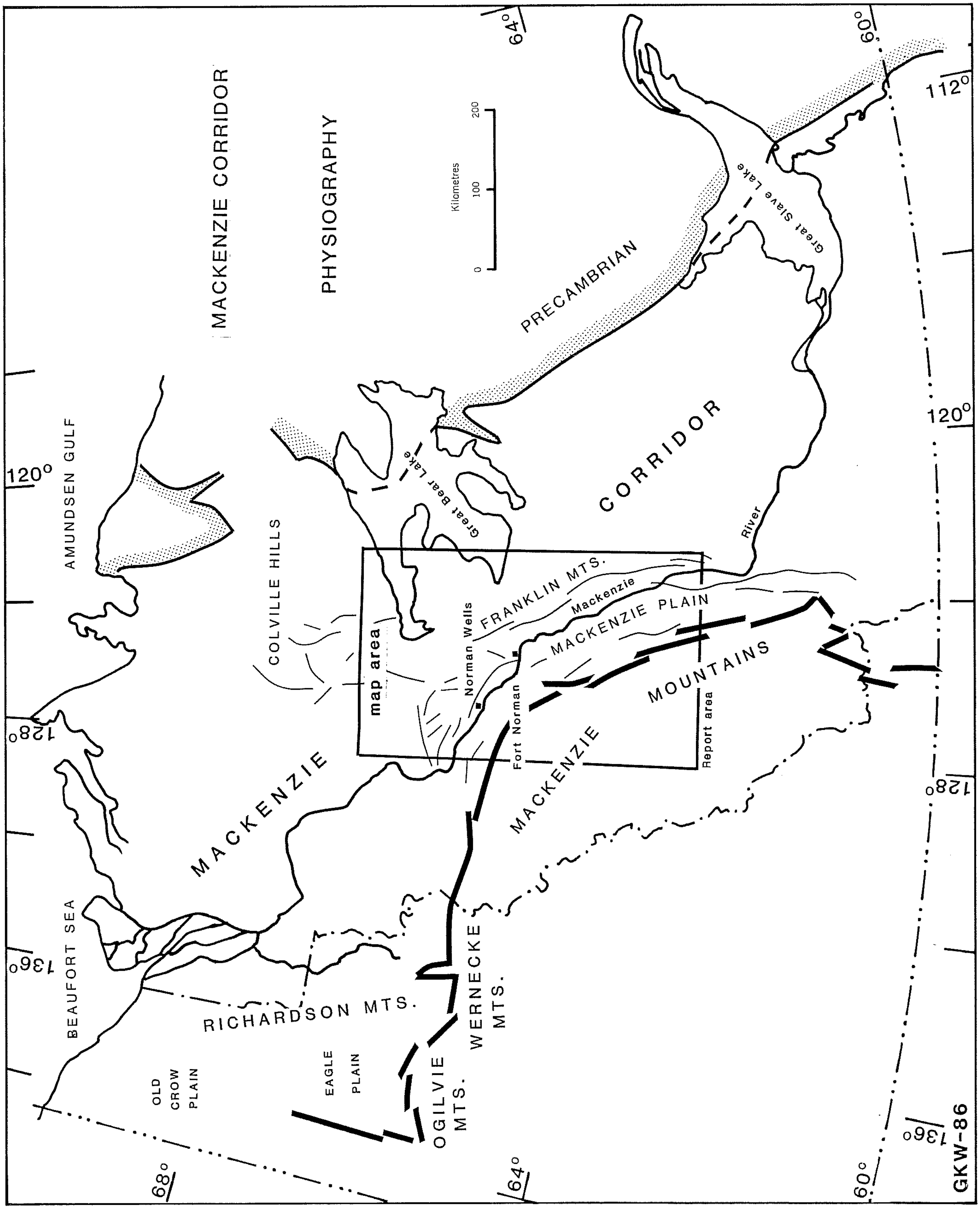

MACKENZIE CORRIDOR
PHYSIOGRAPHY
Kilometres
0
100
200
PRECAMBRIAN
Great Slave Lake
Great Bear Lake
CORRIDOR
River
Mackenzie
FRANKLIN MTS.
MACKENZIE PLAIN
Norman Wells
Fort Norman
map area
Report area
COLVILLE HILLS
AMUNDSEN GULF
MACKENZIE MOUNTAINS
MACKENZIE
BEAUFORT SEA
RICHARDSON MTS.
WERNECKE MTS.
OGILVIE MTS.
OLD CROW PLAIN
EAGLE PLAIN
GKW-86
64°
60°
112°
120°
128°
136°
68°
120°
128°
136°
64°
60°

Figure 2

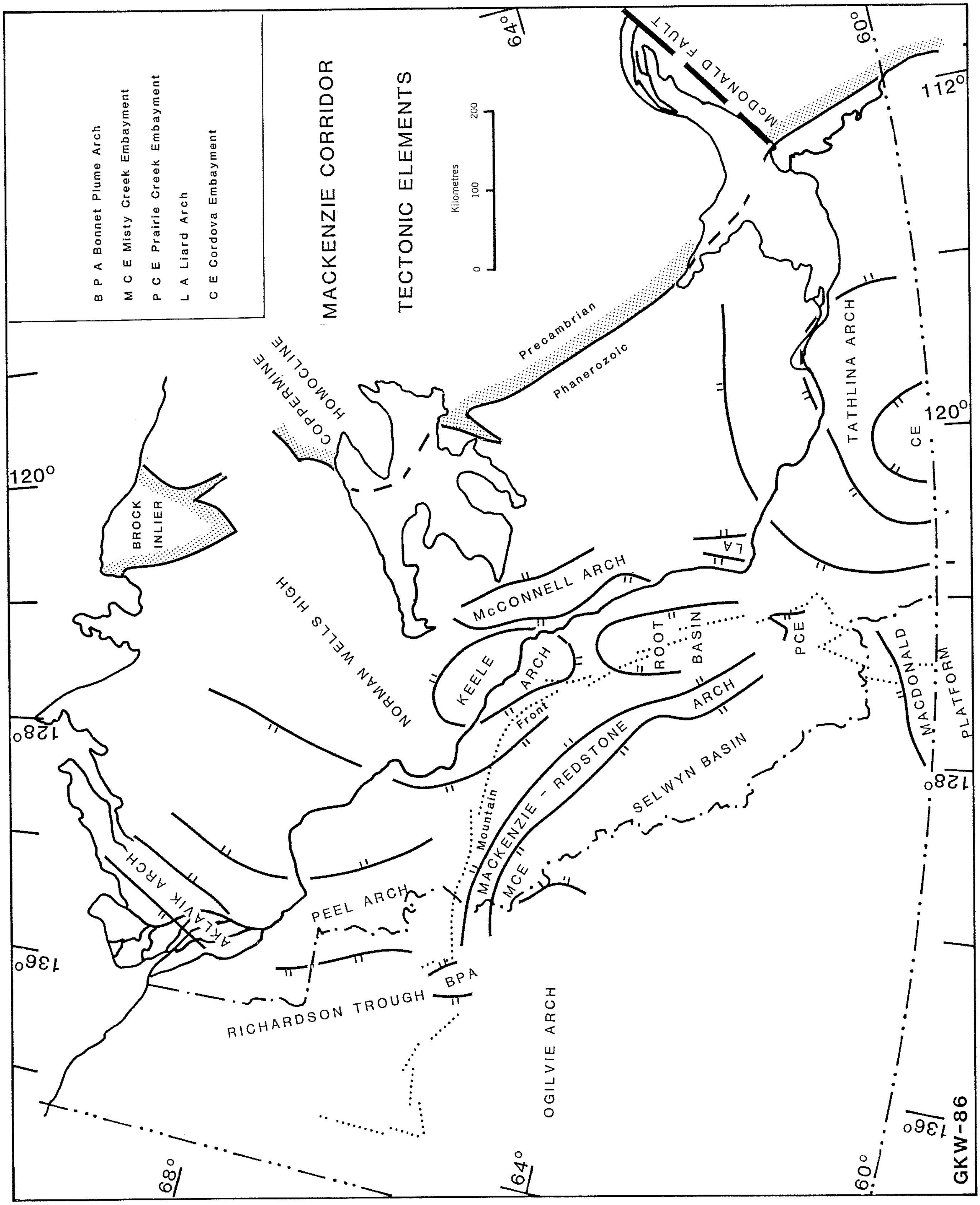
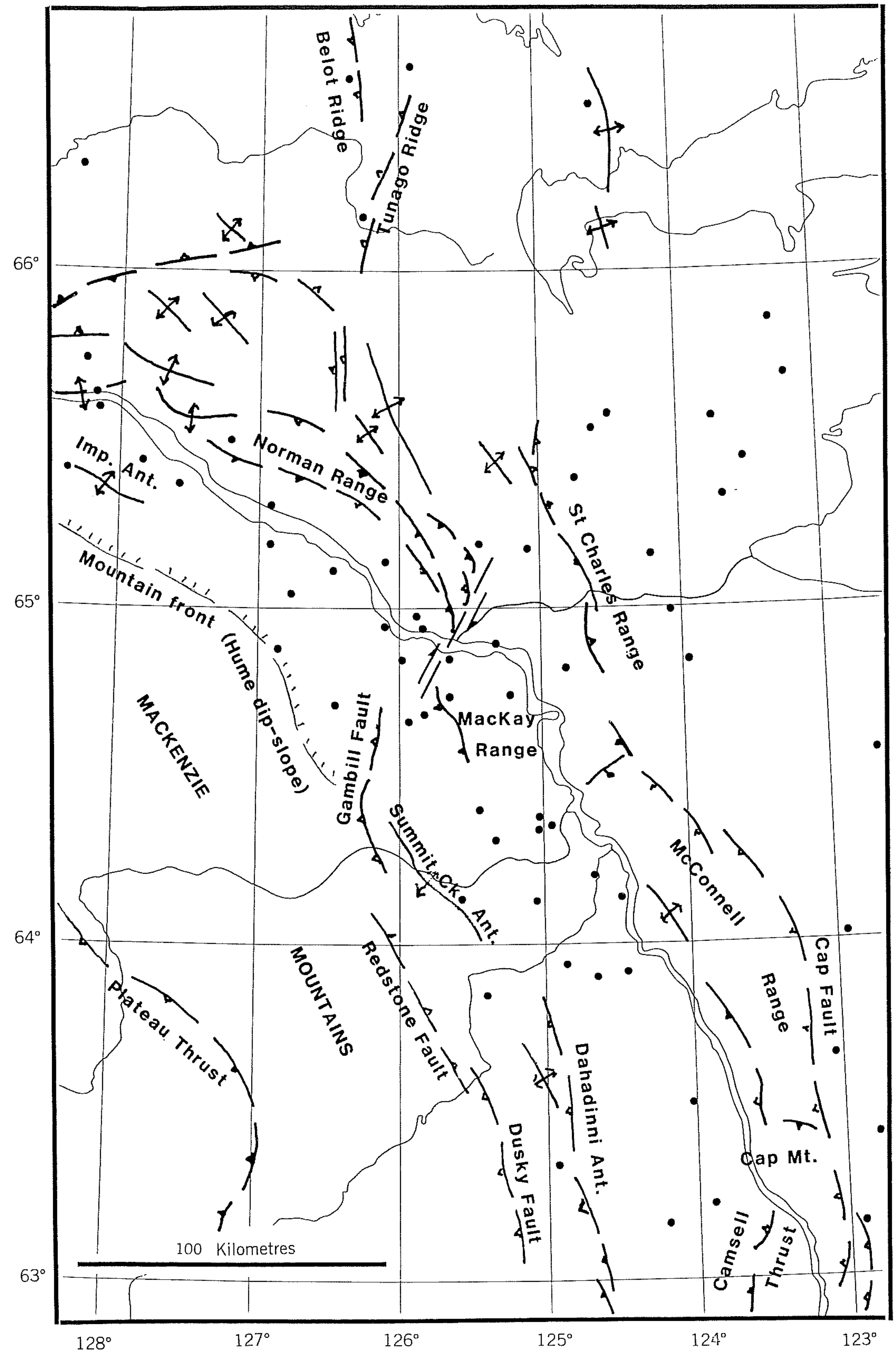
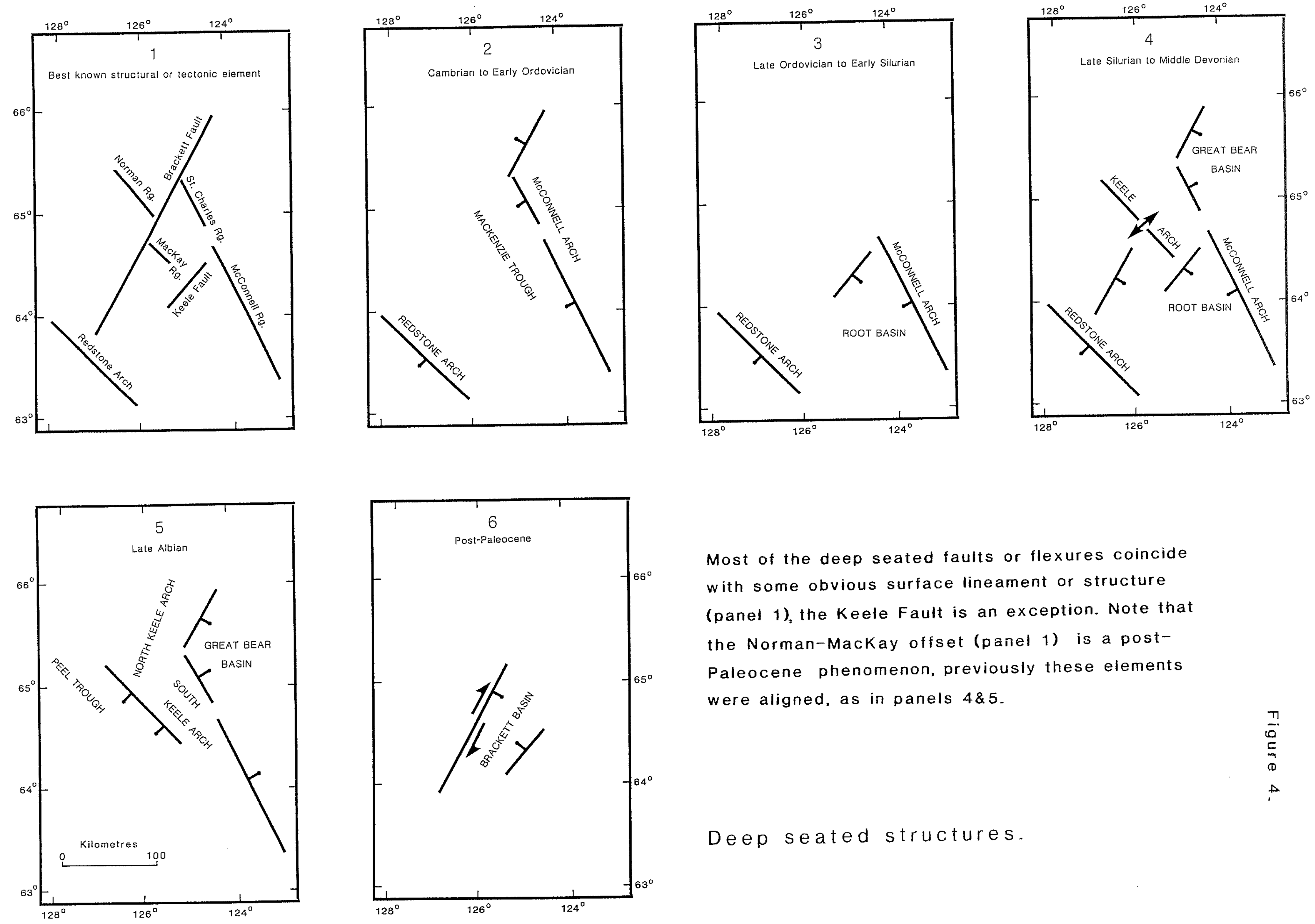

MACKENZIE CORRIDOR
TECTONIC ELEMENTS
B P A Bonnet Plume Arch
M C E Misty Creek Embayment
P C E Prairie Creek Embayment
L A Liard Arch
C E Cordova Embayment
Kilometres
0
100
200
McDONALD FAULT
Precambrian
Phanerozoic
COPPERMINE HOMOCLINE
BROCK INLIER
NORMAN WELLS HIGH
McCONNELL ARCH
KEELE ARCH
ROOT BASIN
LA
TATHLINA ARCH
CE
PCE
MACDONALD PLATFORM
MACKENZIE - REDSTONE ARCH
SELWYN BASIN
Mountain Front
MCE
PEEL ARCH
AKLAVIK ARCH
BPA
RICHARDSON TROUGH
OGILVIE ARCH
64°
60°
112°
120°
128°
136°
68°
GKW-86

Figure 3



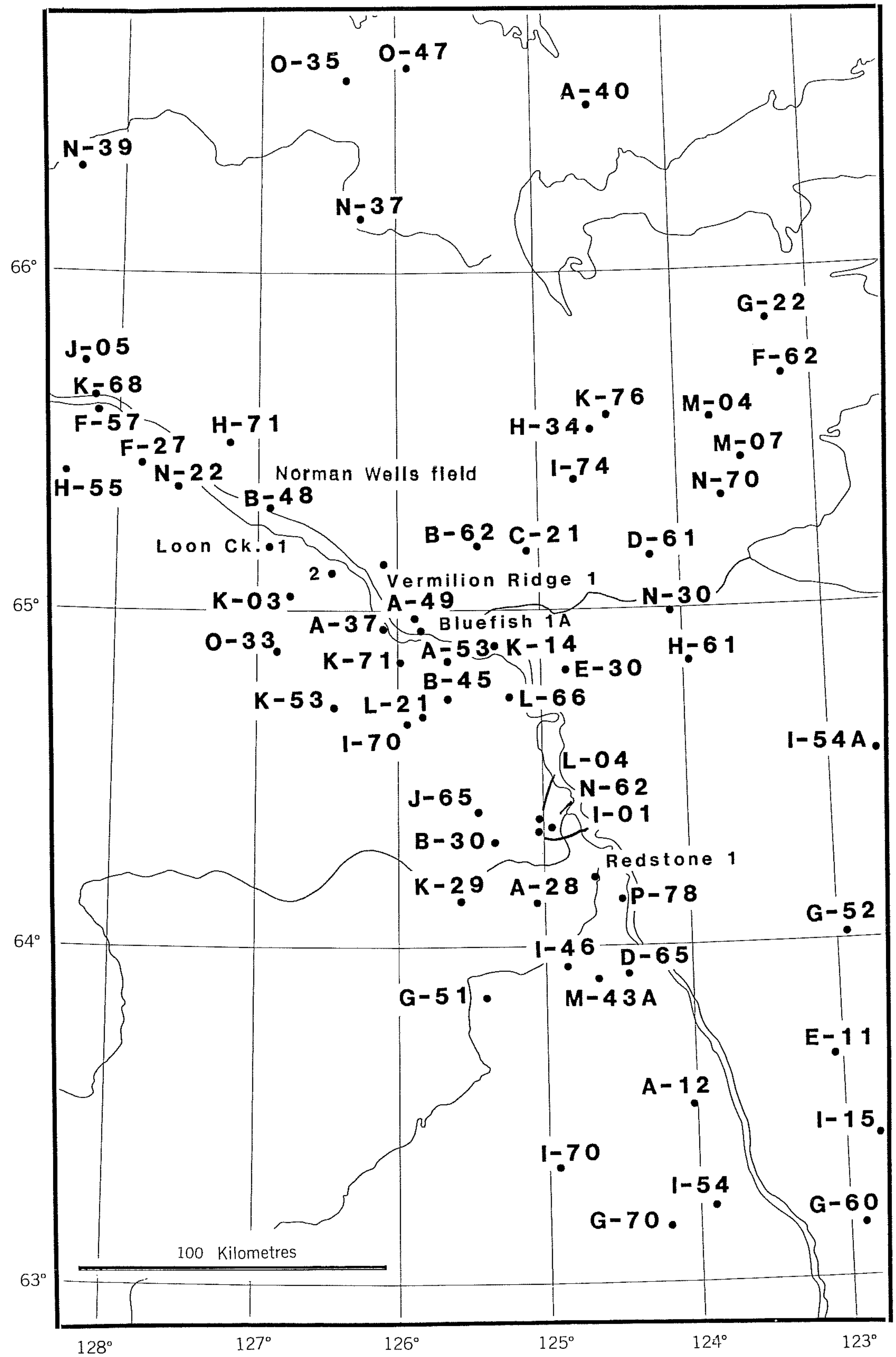
Laramide structures



Most of the deep seated faults or flexures coincide with some obvious surface lineament or structure (panel 1), the Keele Fault is an exception. Note that the Norman-MacKay offset (panel 1) is a post-Paleocene phenomenon, previously these elements were aligned, as in panels 4&5.

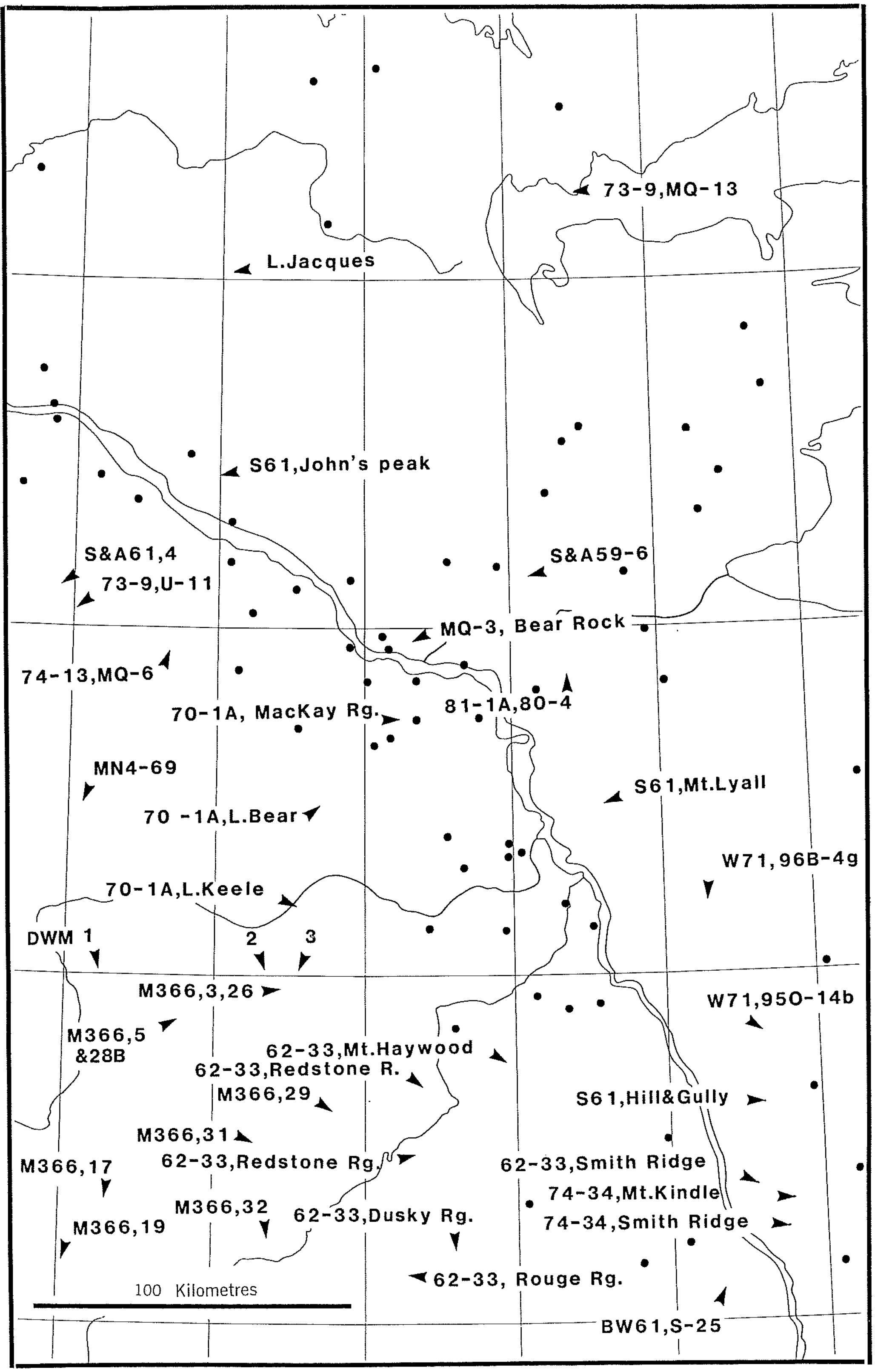
Figure 4.

Deep seated structures.



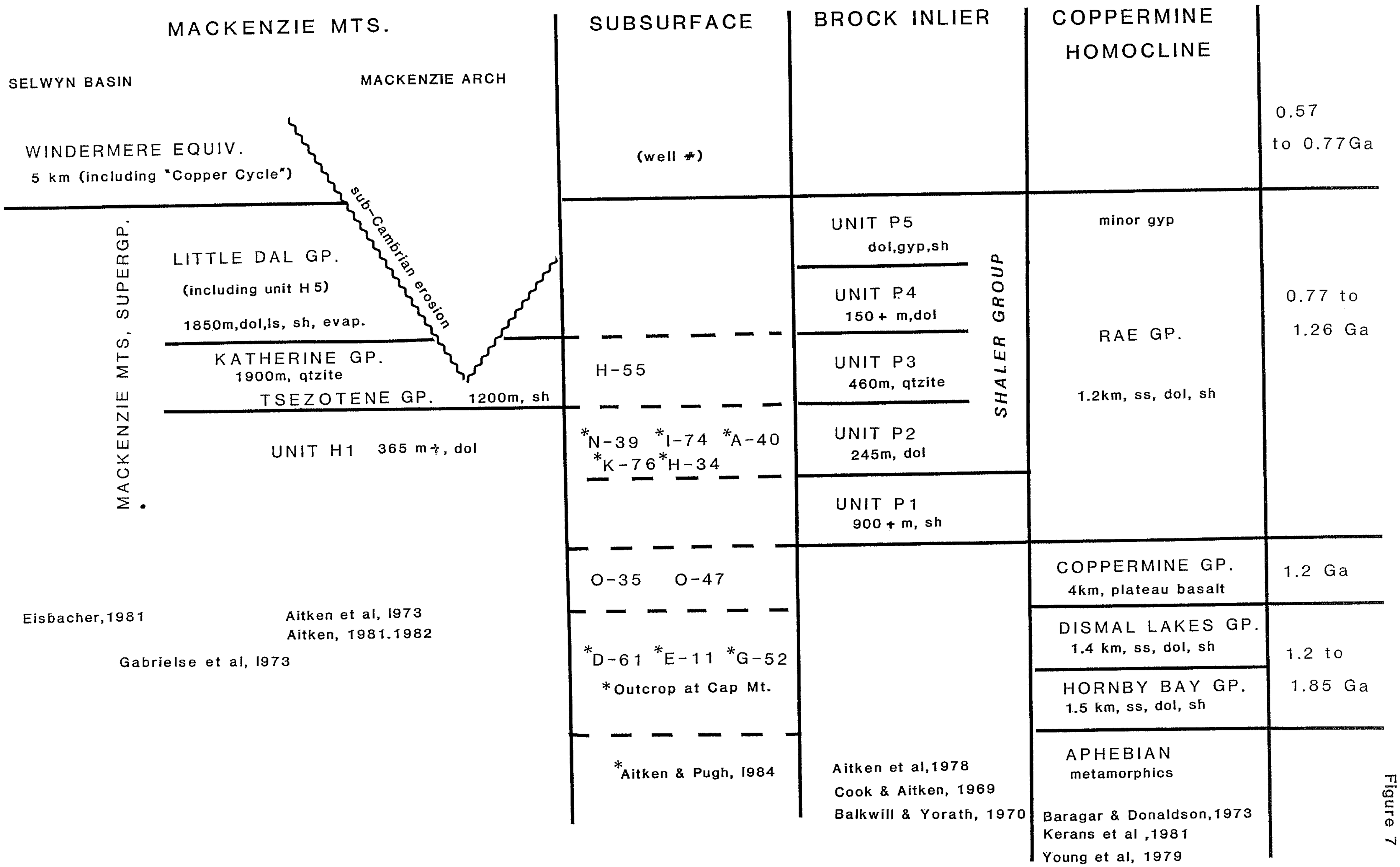
Well locations

Figure 6



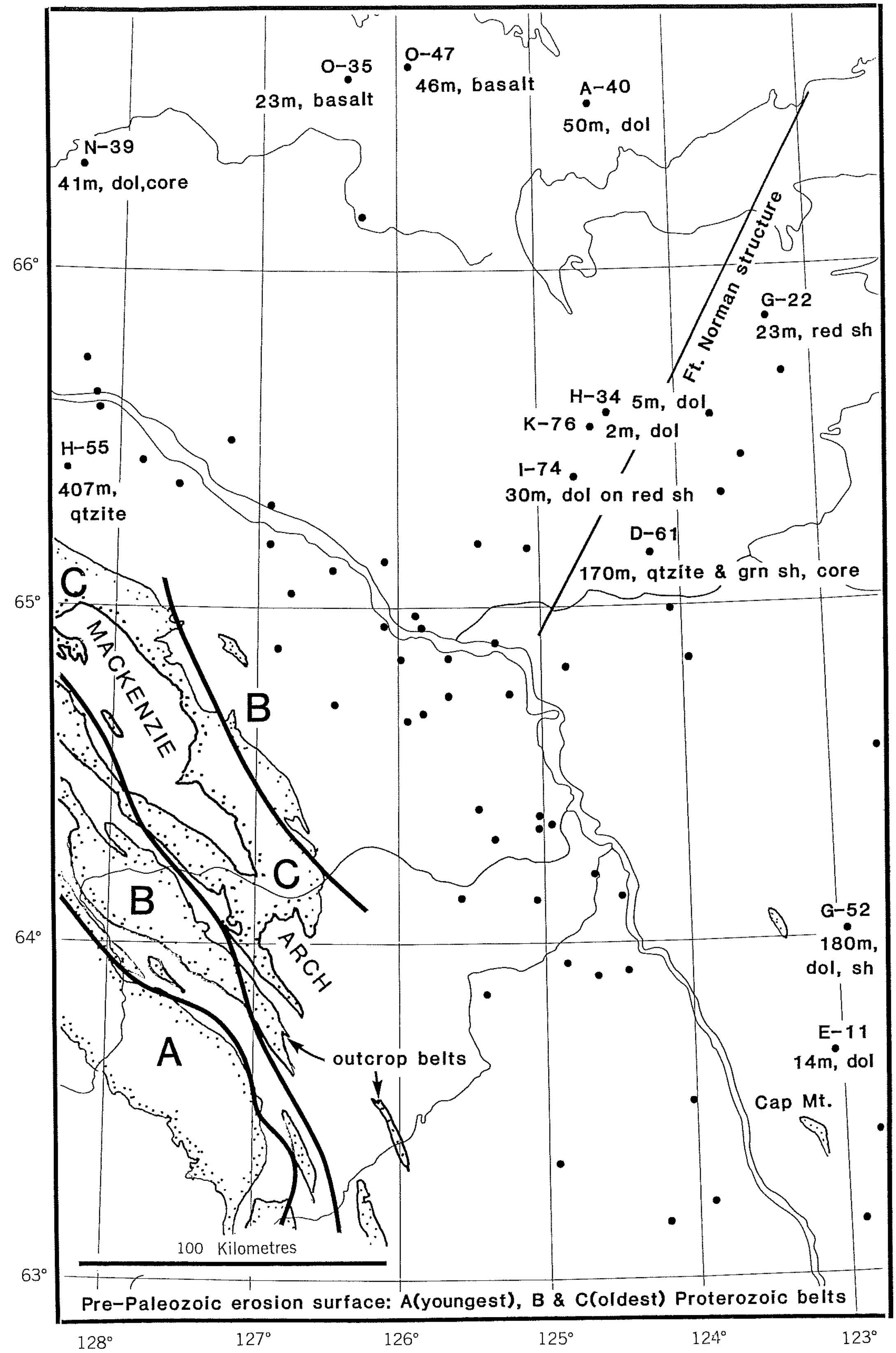
Field section locations

Figure 7



Precambrian correlation chart

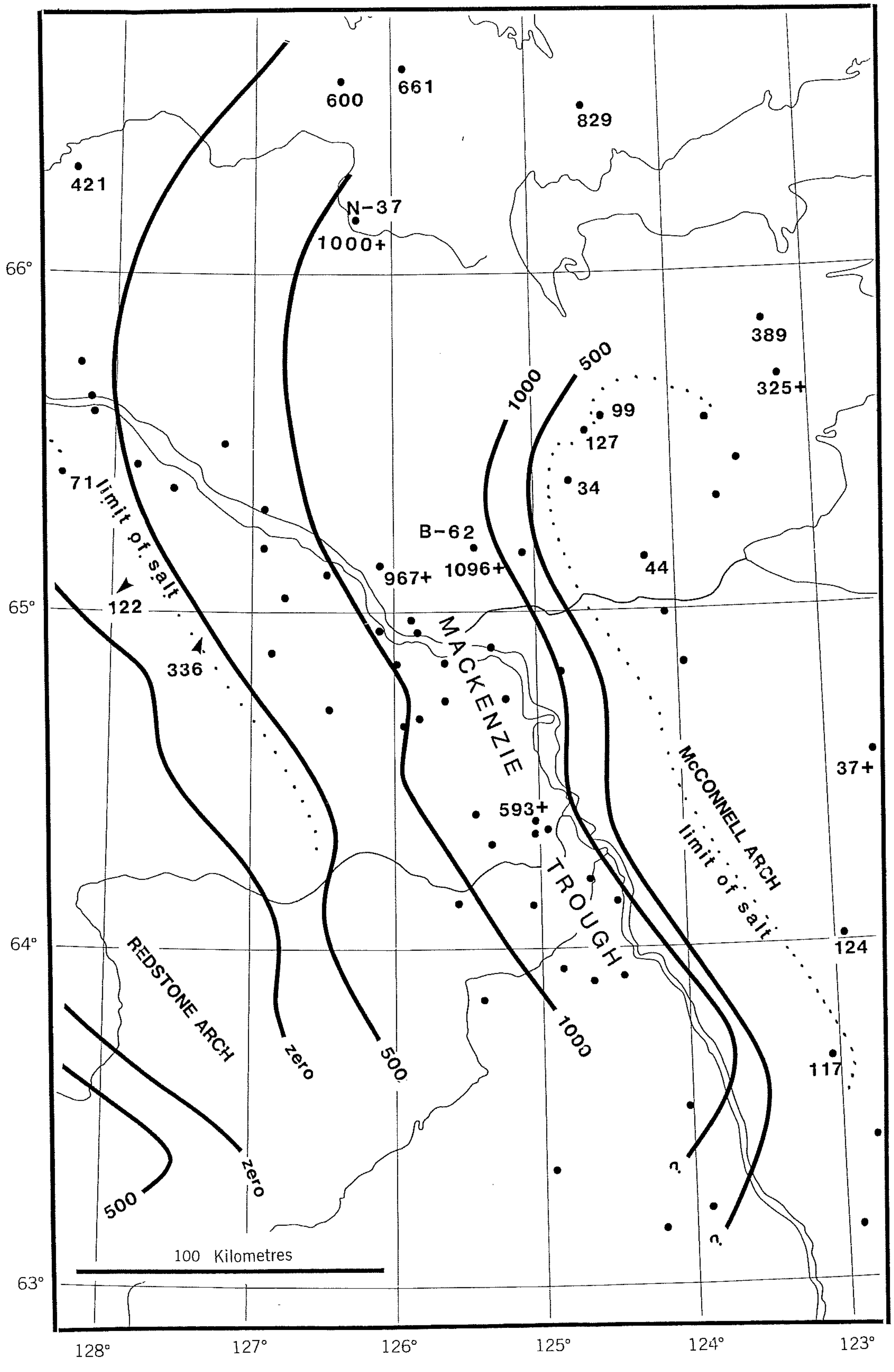
Figure 8



Pre-Paleozoic erosion surface: A(youngest), B & C(oldest) Proterozoic belts

Precambrian

Figure 9



Isopachs, Lower and Middle Cambrian

Figure 10

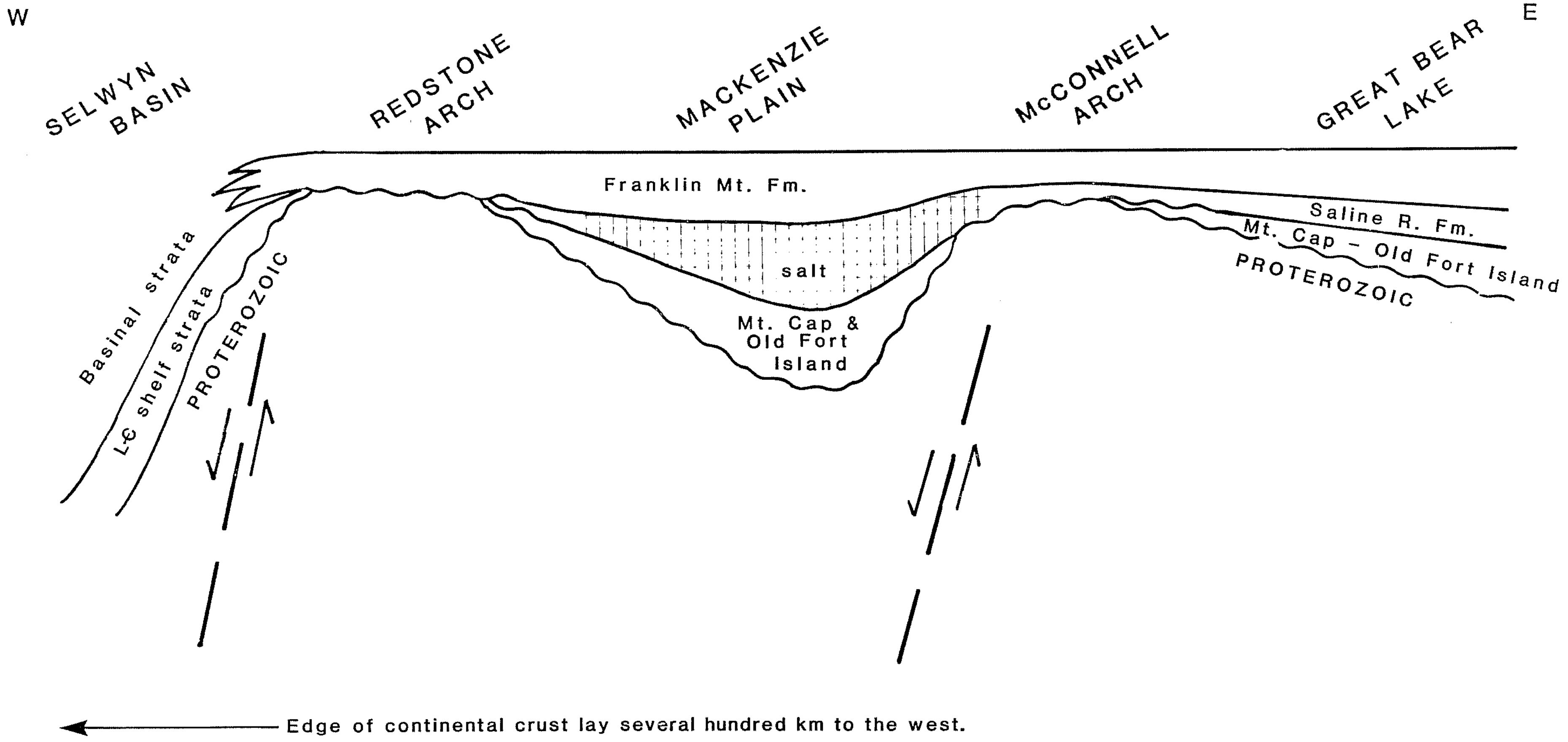
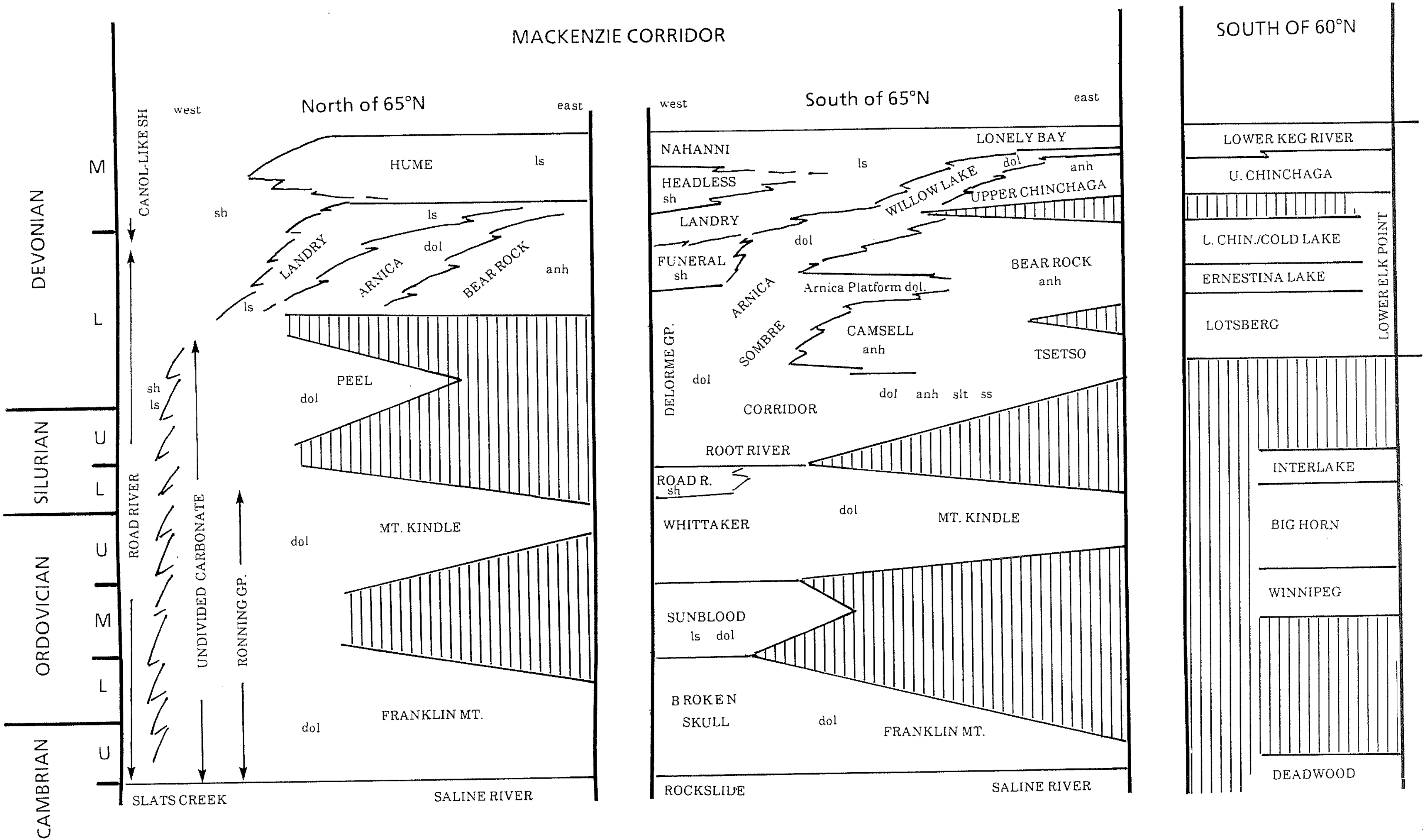


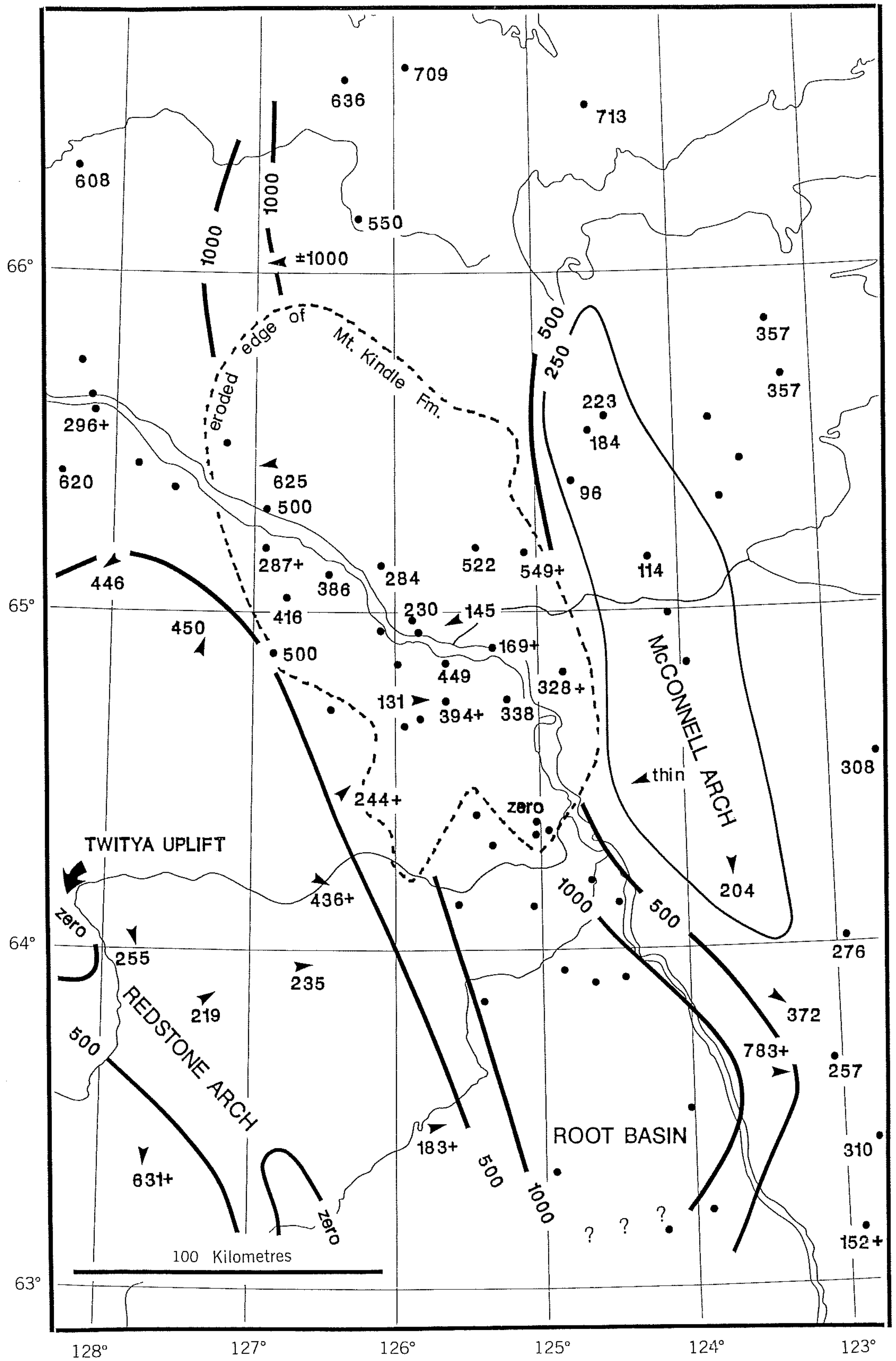
Fig. 10 Hypothetical early Paleozoic tensional tectonic regime.



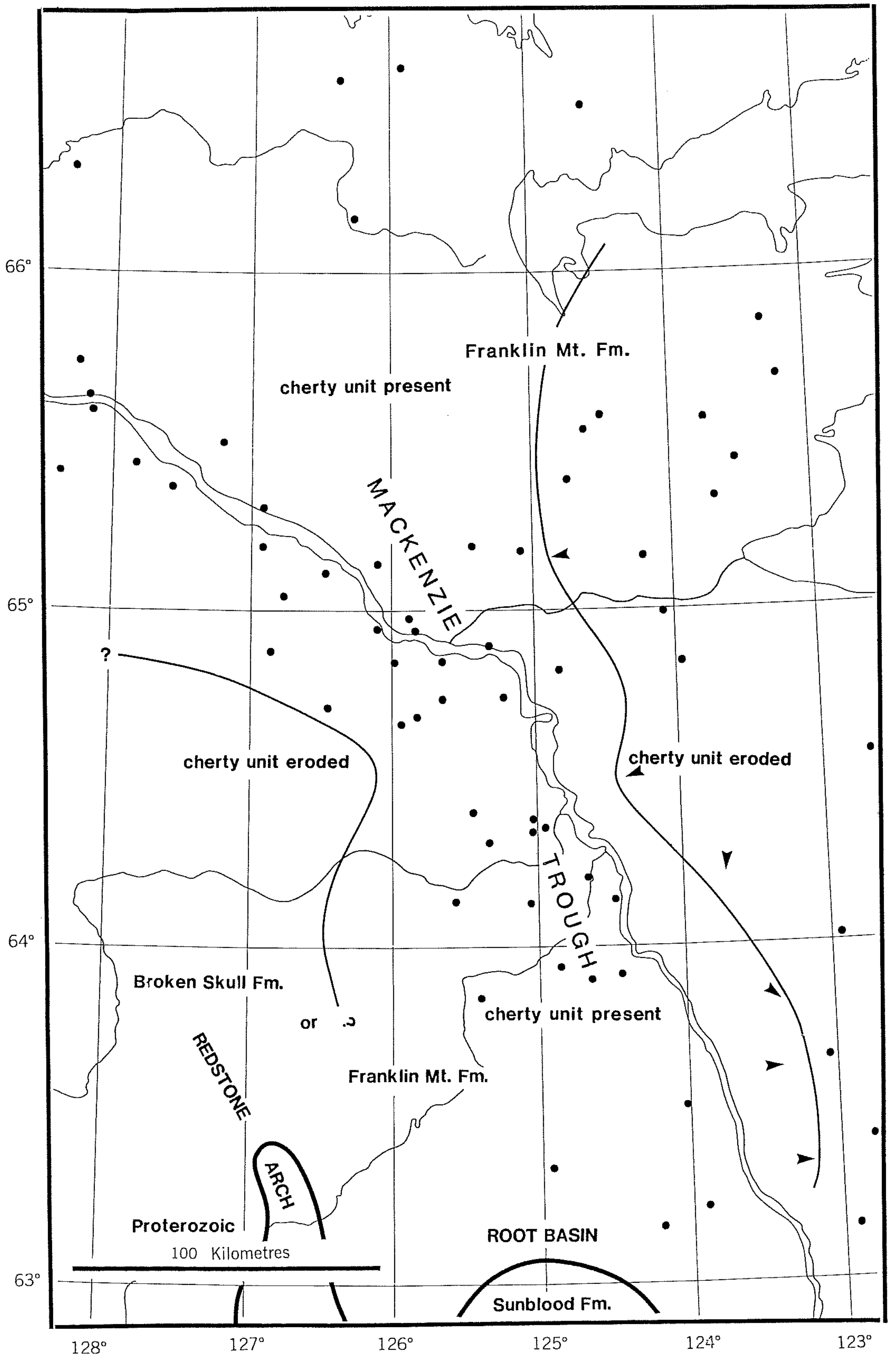
Upper Cambrian to Middle Devonian correlation chart

Figure 11

Figure 12

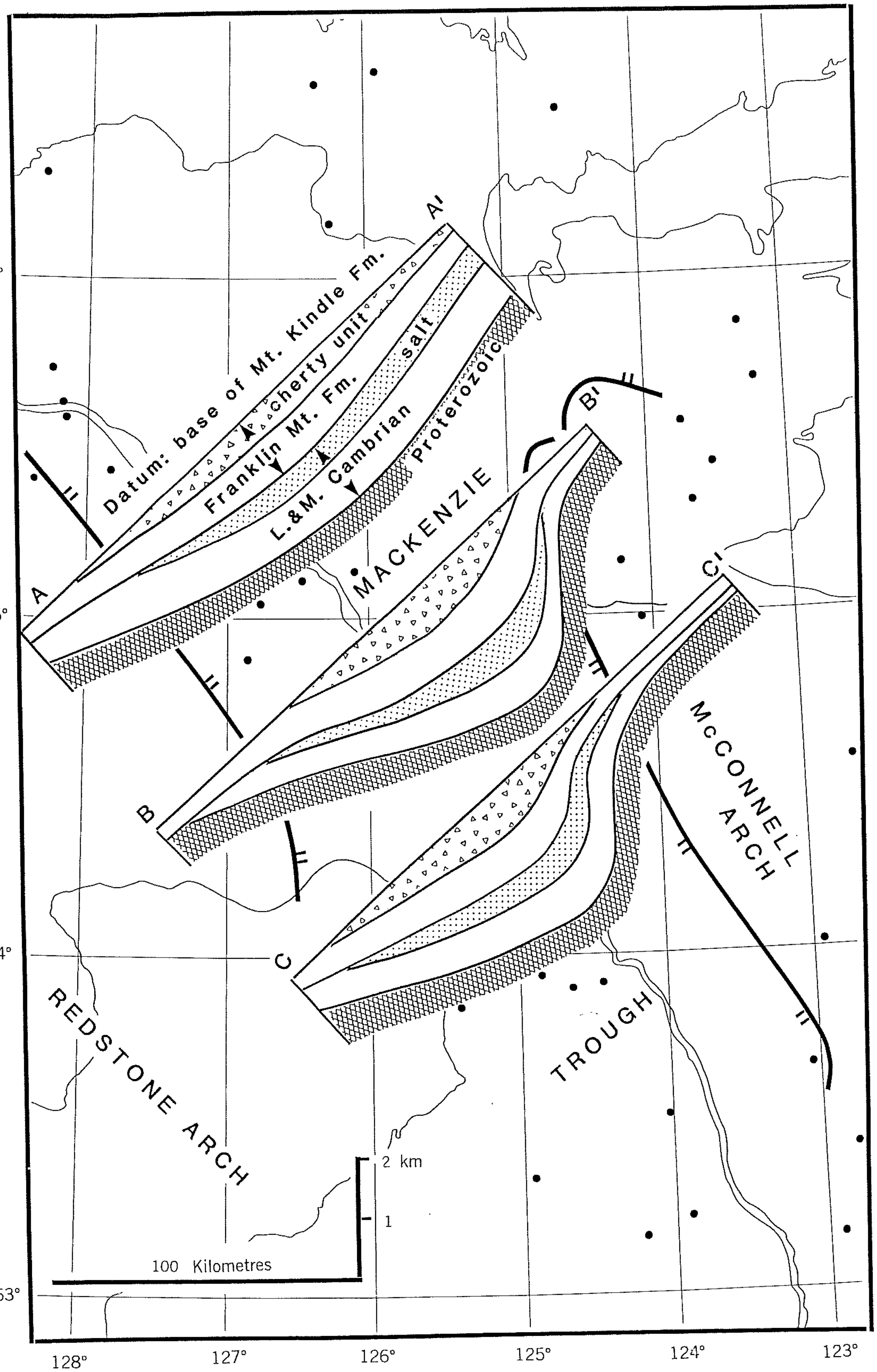


Isopachs, Franklin Mountain Formation.



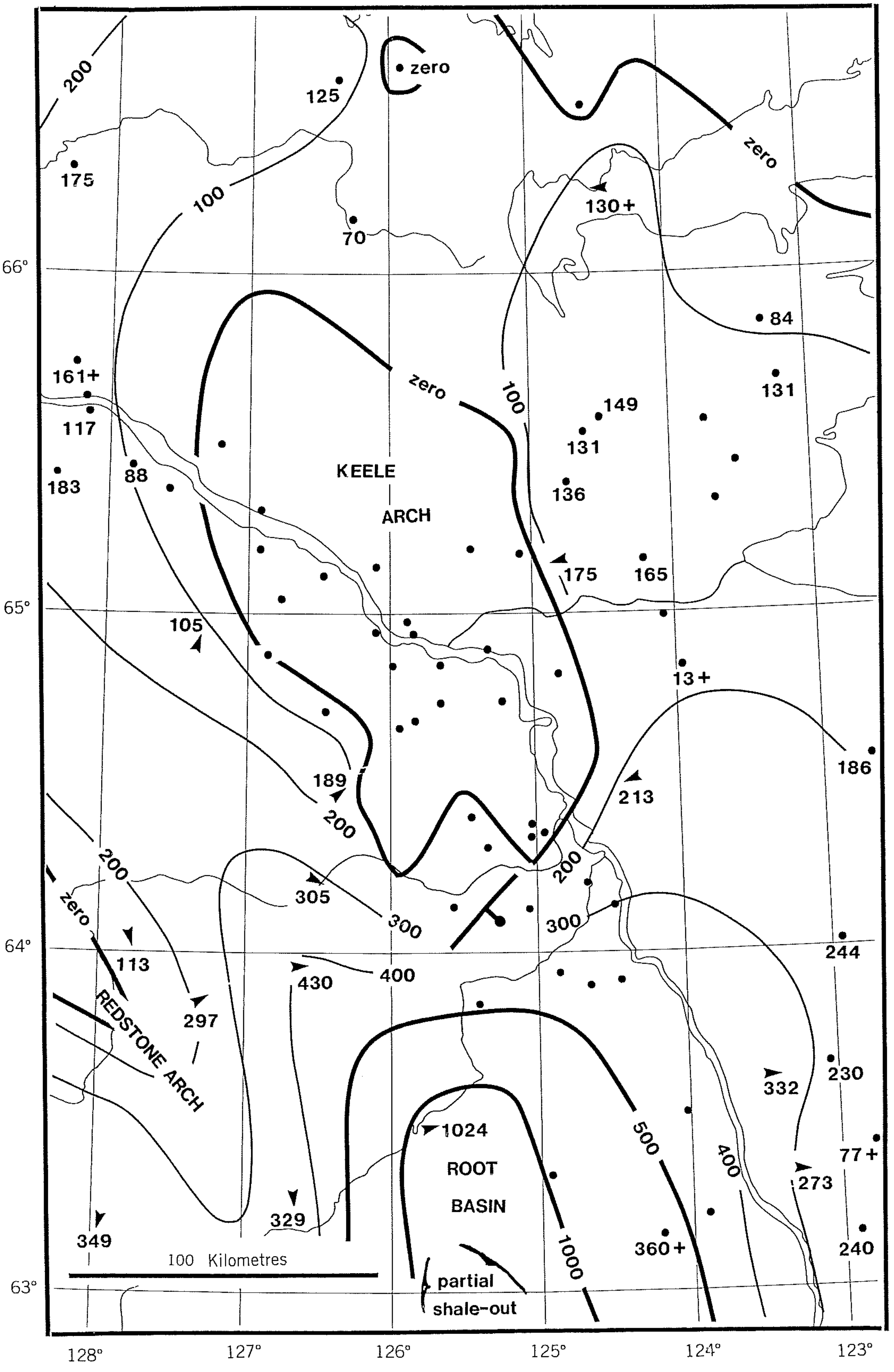
Pre-Mount Kindle geology

Figure 14



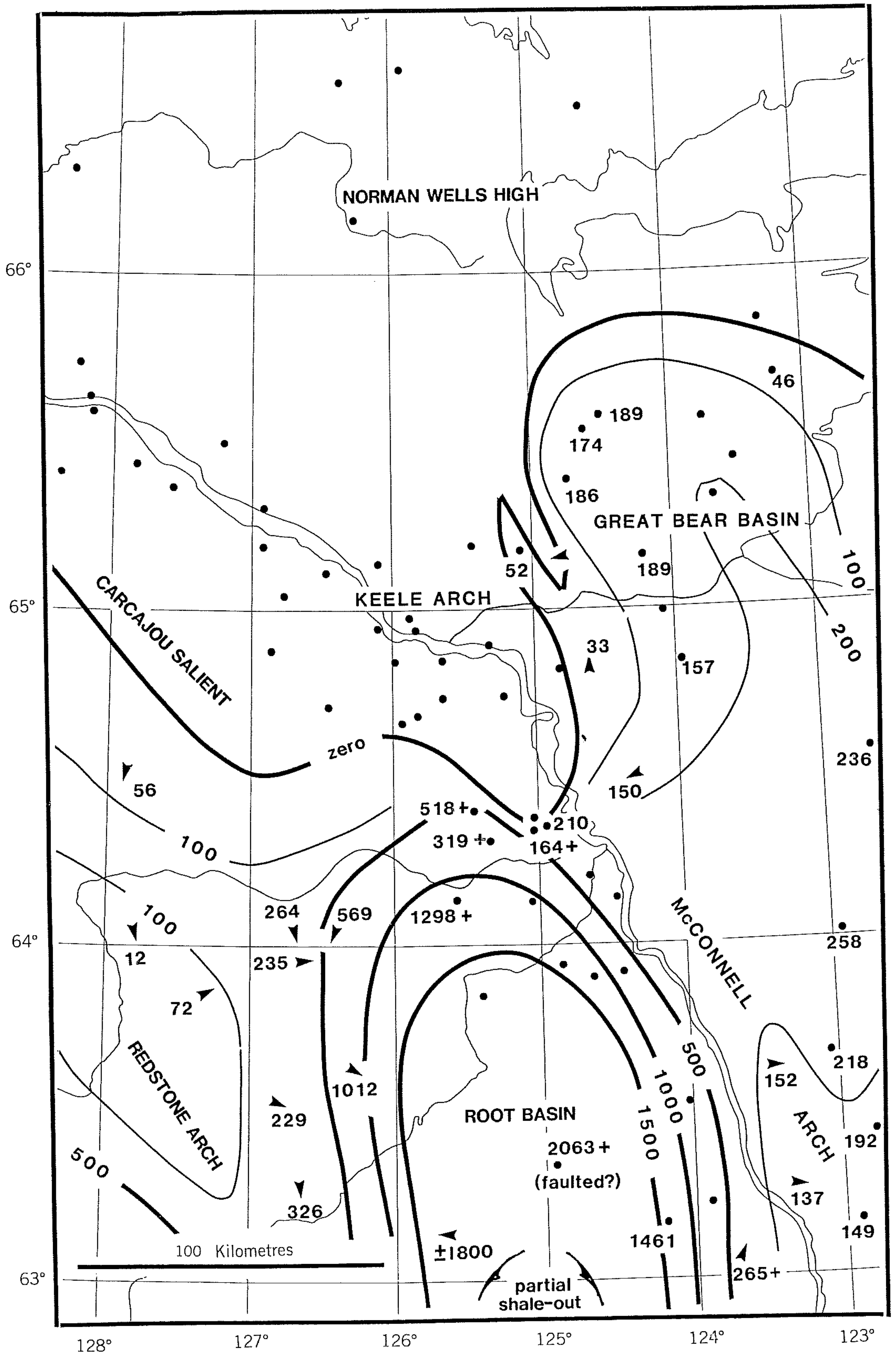
Cross-sections, Mackenzie Trough

Figure 15



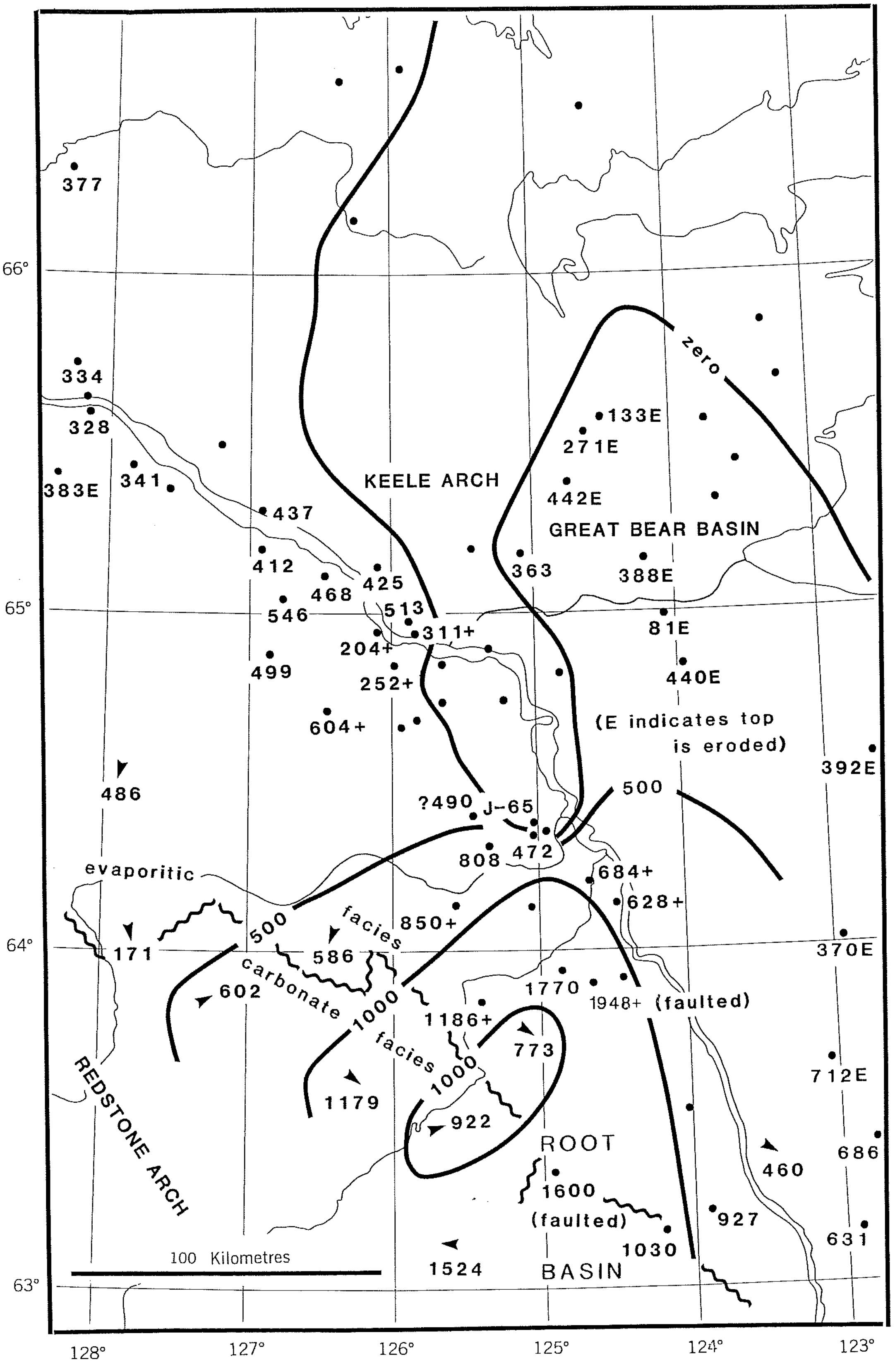
Isopachs, Mount Kindle Formation

Figure 16



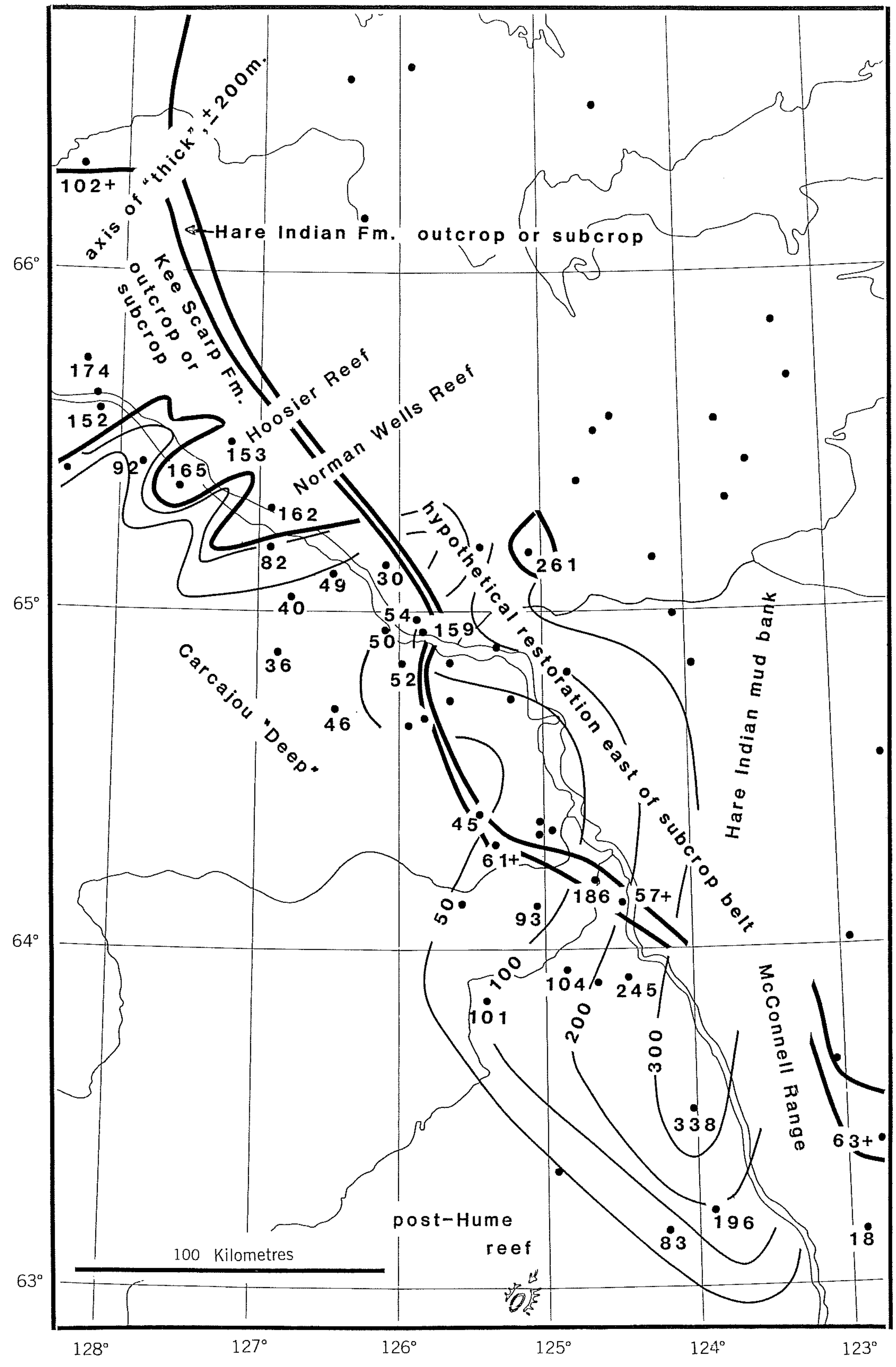
Isopachs, Delorme Group

Figure 17



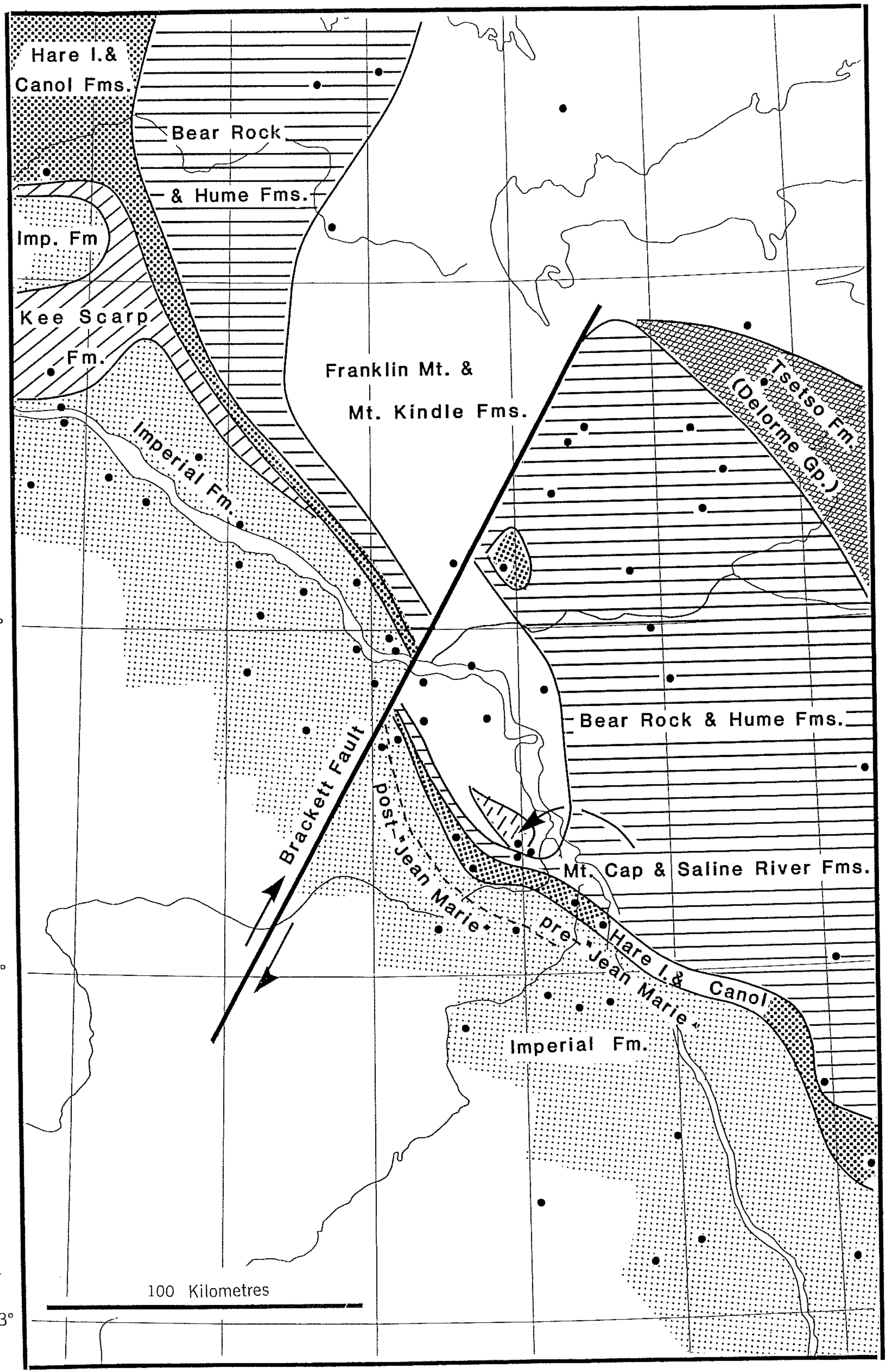
Isopachs, Bear rock and Hume Formations

Figure 18



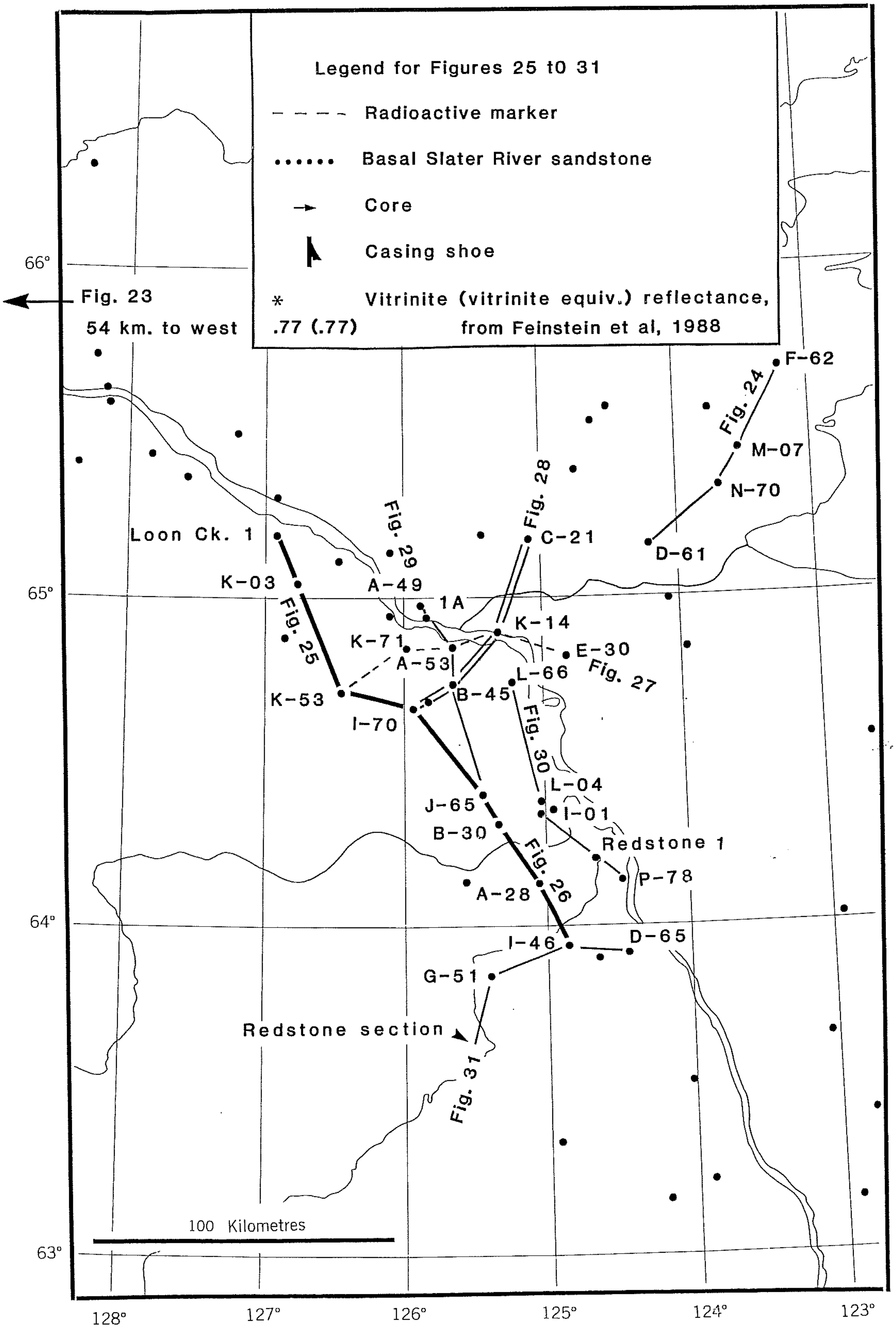
Isopachs, Hare Indian Formation

Figure 19

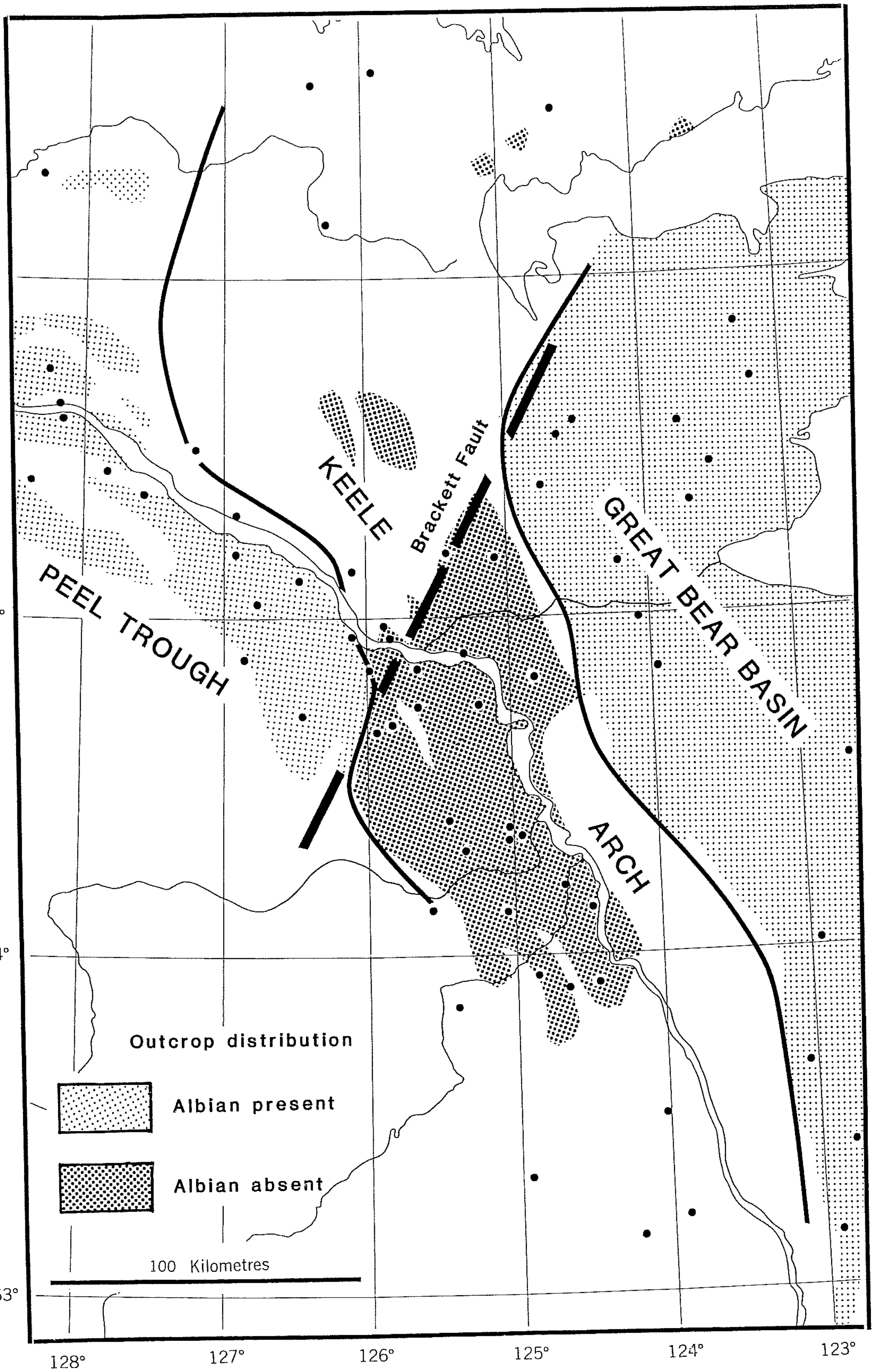


Pre-Cretaceous geological map

Figure 20

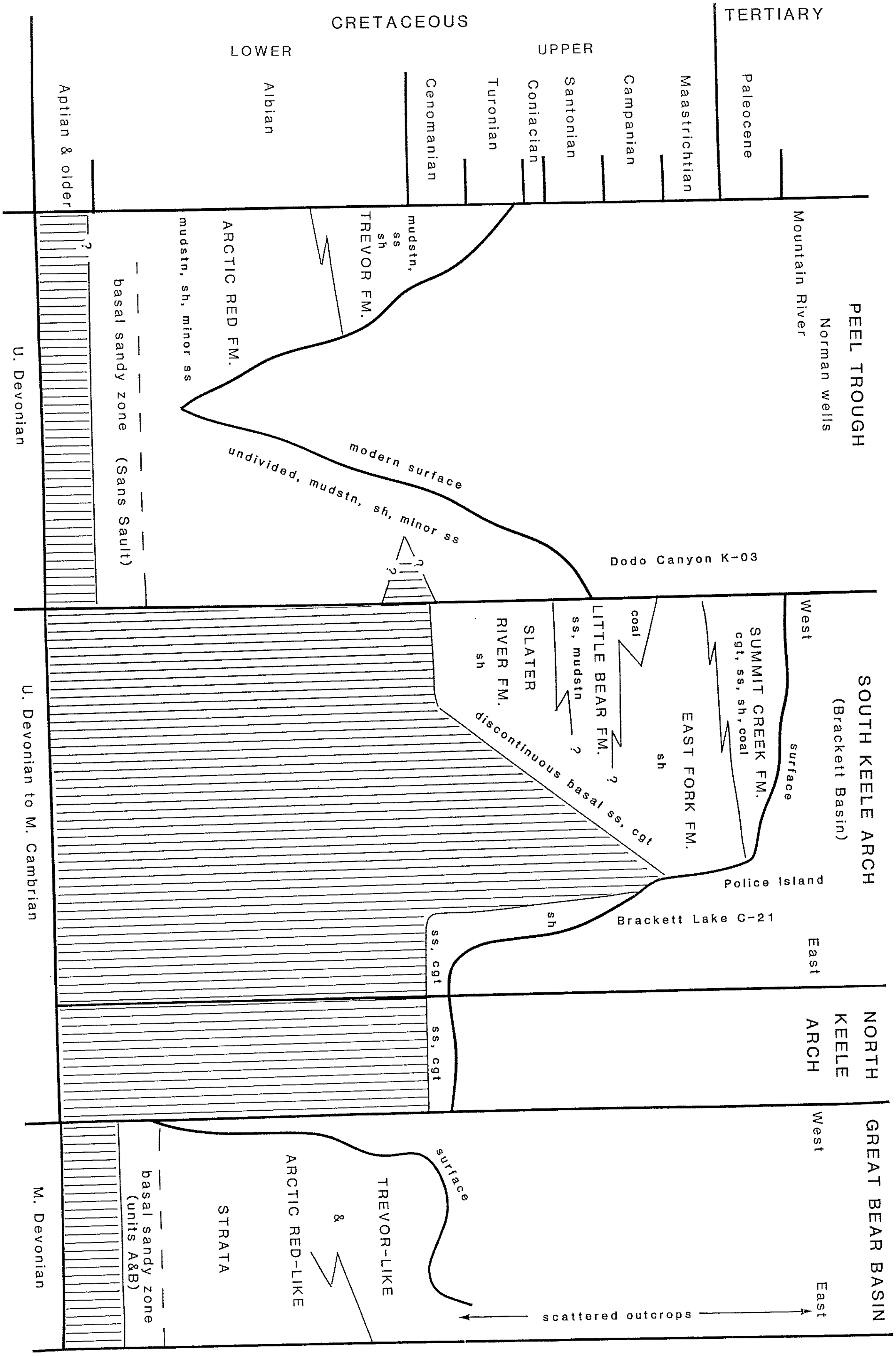


Index map for cross-sections that illustrate Cretaceous strata

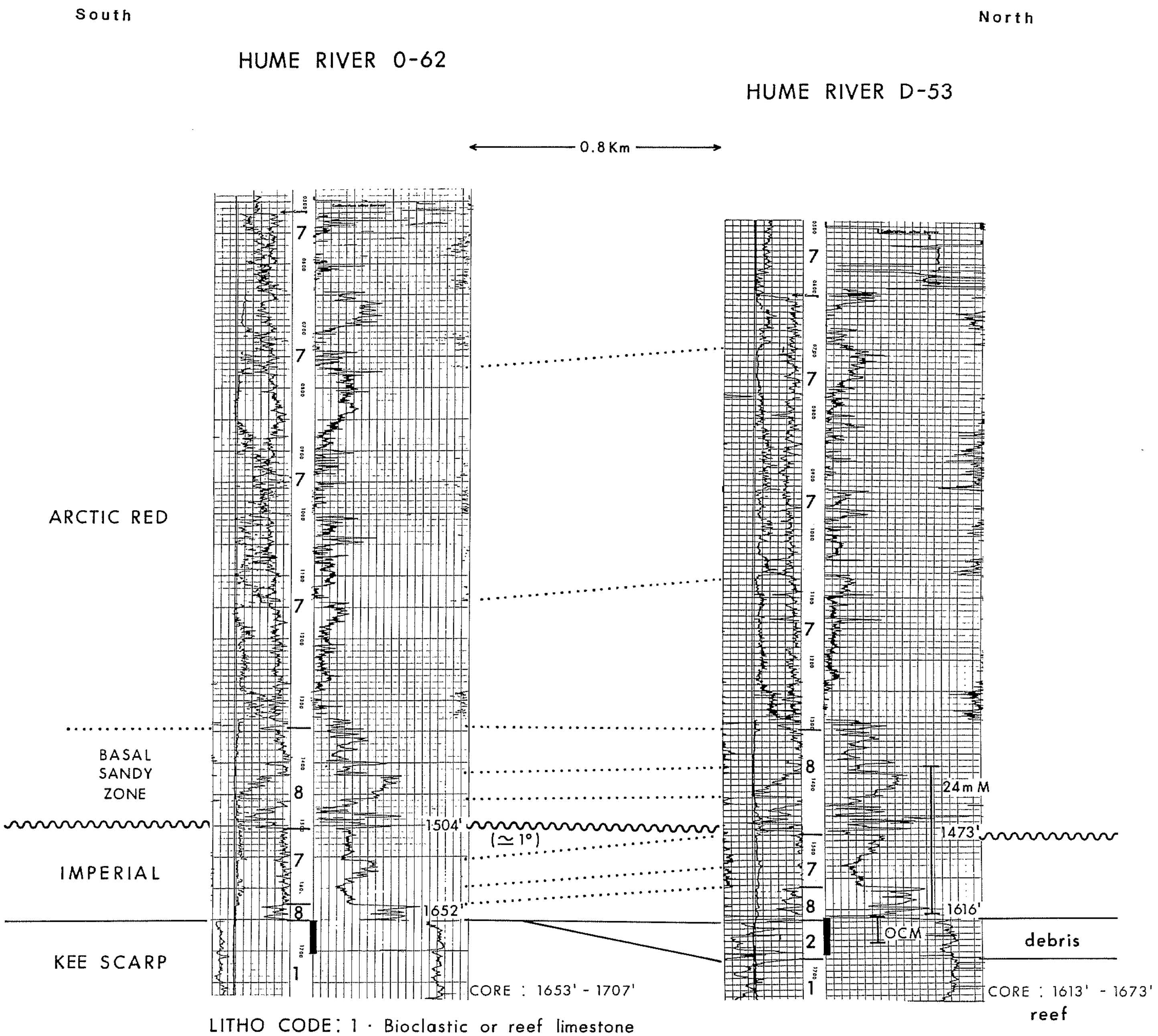


Cretaceous distribution, main tectonic elements

Figure 22



Cretaceous-Tertiary correlation chart



LITHO CODE: 1 · Bioclastic or reef limestone
2 · Reef debris
7 · Silty shale, siltstone
8 · Siltstone, sandstone, shale

Figure 23

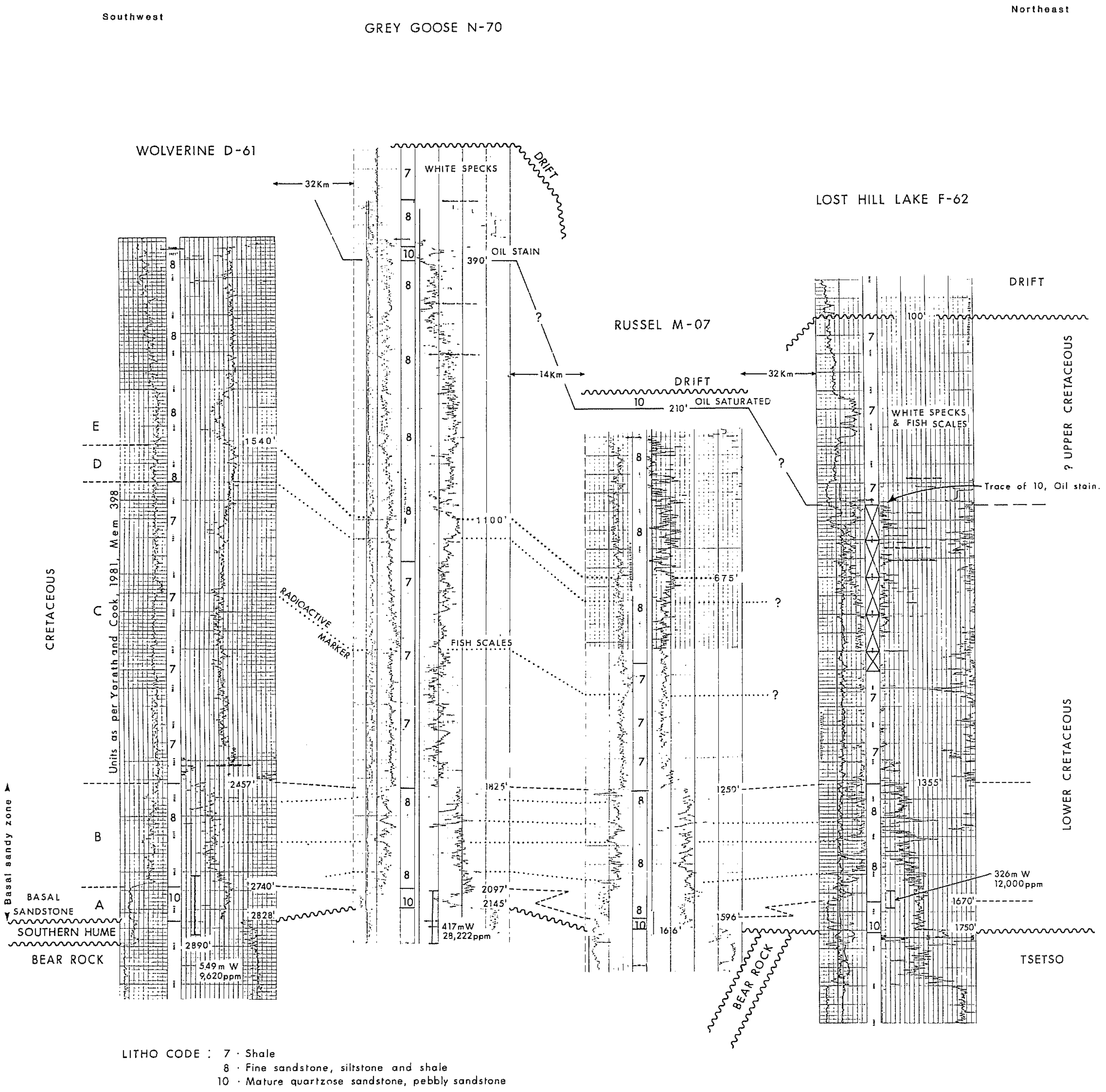

Southwest
Northeast
GREY GOOSE N-70
WOLVERINE D-61
LOST HILL LAKE F-62
RUSSEL M-07
32Km
14Km
32Km
DRIFT
WHITE SPECKS
OIL STAIN
390'
DRIFT
10
210'
OIL SATURATED
WHITE SPECKS & FISH SCALES
100'
? UPPER CRETACEOUS
Trace of 10, Oil stain.
1540'
1100'
675'
E
D
C
B
A
CRETACEOUS
Units as per Yorath and Cook, 1981, Mem 398
RADIOACTIVE MARKER
FISH SCALES
1825'
1250'
1355'
LOWER CRETACEOUS
2457'
2740'
2828'
2890'
549m W 9,620ppm
2097'
2145'
417mW 28,222ppm
1596'
1616'
326m W 12,000ppm
1670'
1750'
TSETSO
BEAR ROCK
BASAL SANDSTONE
SOUTHERN HUME
BEAR ROCK
Basal sandy zone
7
8
10
?
LITHO CODE : 7 · Shale
8 · Fine sandstone, siltstone and shale
10 · Mature quartzose sandstone, pebbly sandstone

Figure 24

Figure 25

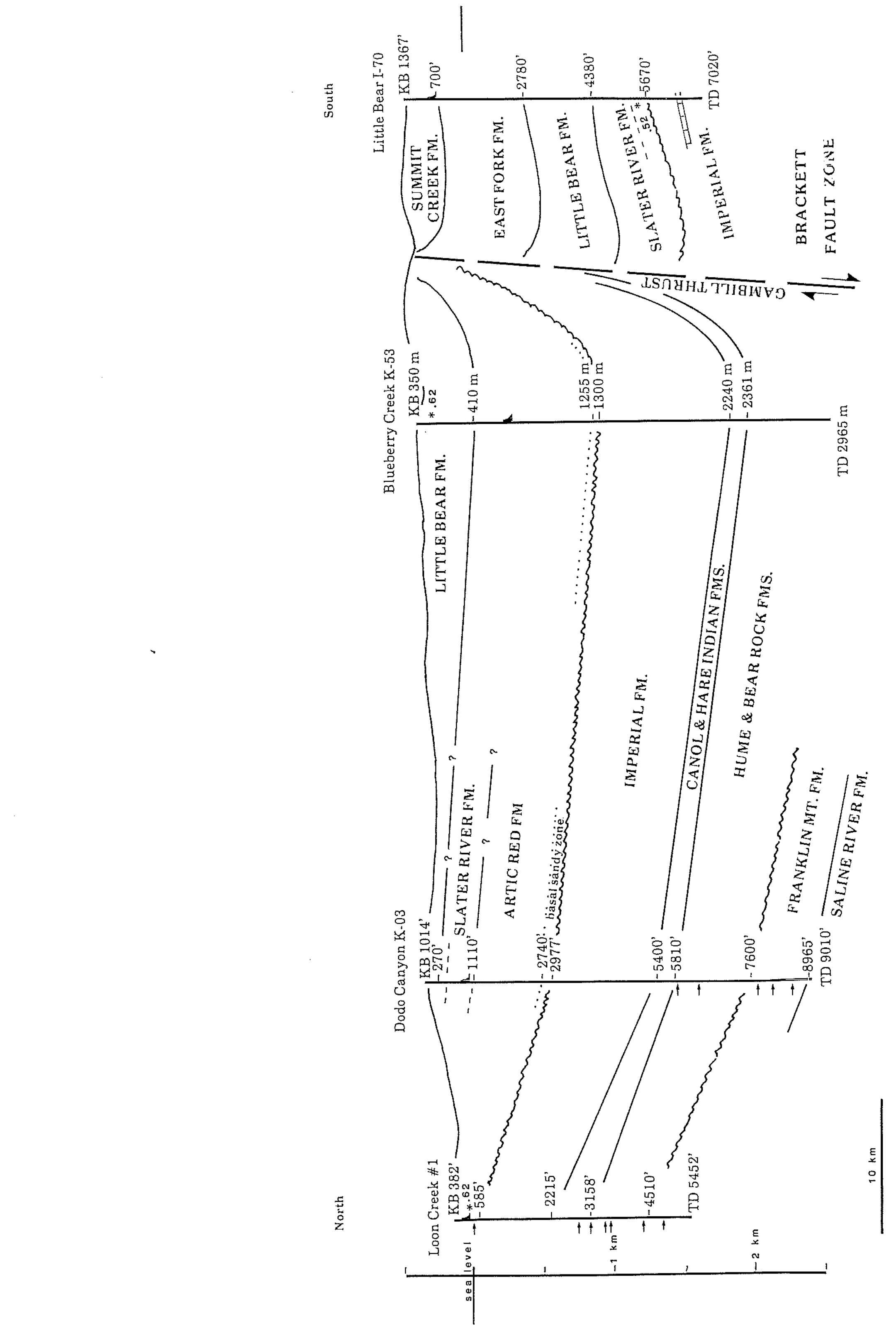

South
North
Little Bear I-70
KB 1367'
-700'
-2780'
-4380'
-5670'
*.52
TD 7020'
SUMMIT CREEK FM.
EAST FORK FM.
LITTLE BEAR FM.
SLATER RIVER FM.
IMPERIAL FM.
BRACKETT FAULT ZONE
GAMBILL THRUST
Blueberry Creek K-53
KB 350 m
*.62
-410 m
-1255 m
-1300 m
-2240 m
-2361 m
TD 2965 m
LITTLE BEAR FM.
IMPERIAL FM.
CANOL & HARE INDIAN FMS.
HUME & BEAR ROCK FMS.
Dodo Canyon K-03
KB 1014'
-270'
-1110'
-2740'
-2977'
-5400'
-5810'
-7600'
-8965'
TD 9010'
SLATER RIVER FM.
?
ARTIC RED FM
basal sandy zone
FRANKLIN MT. FM.
SALINE RIVER FM.
Loon Creek #1
KB 382'
*.62
-585'
-2215'
-3158'
-4510'
TD 5452'
10 km
sea level
-1 km
-2 km

Figure 26

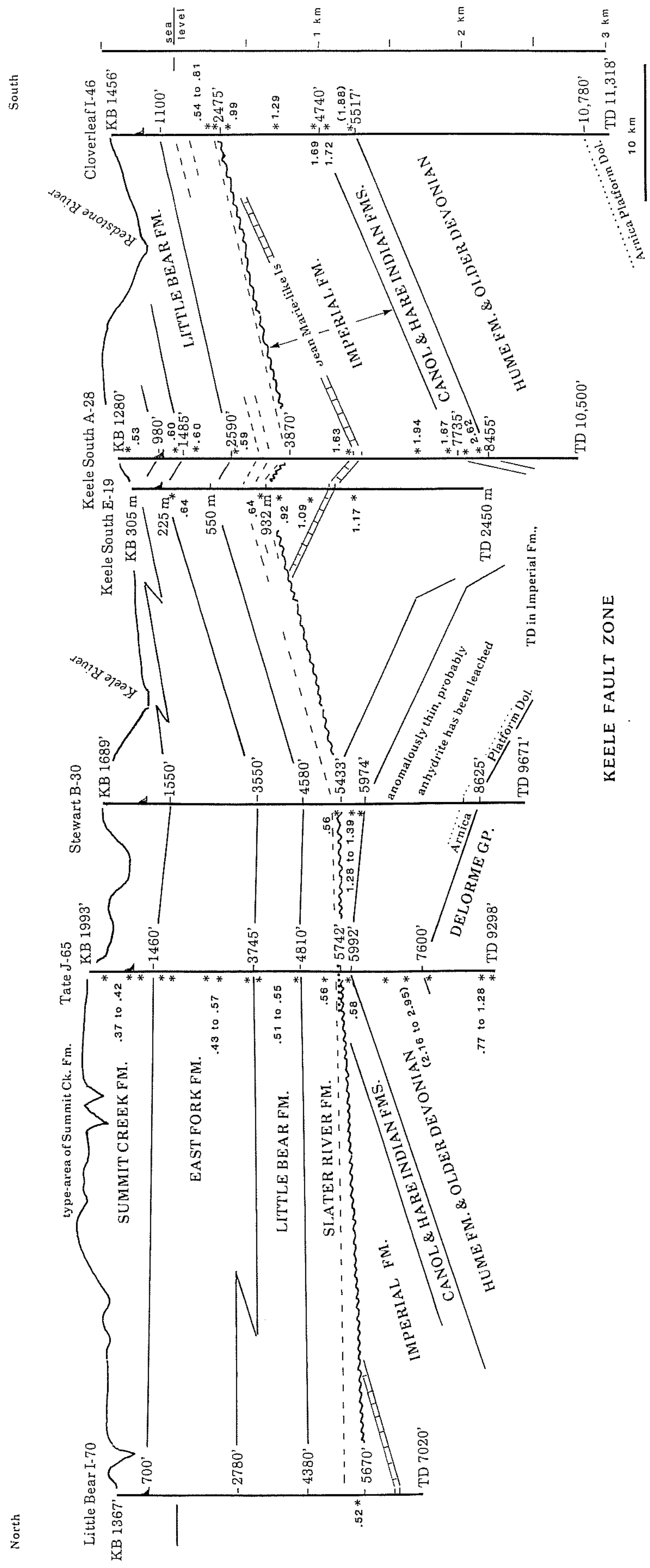
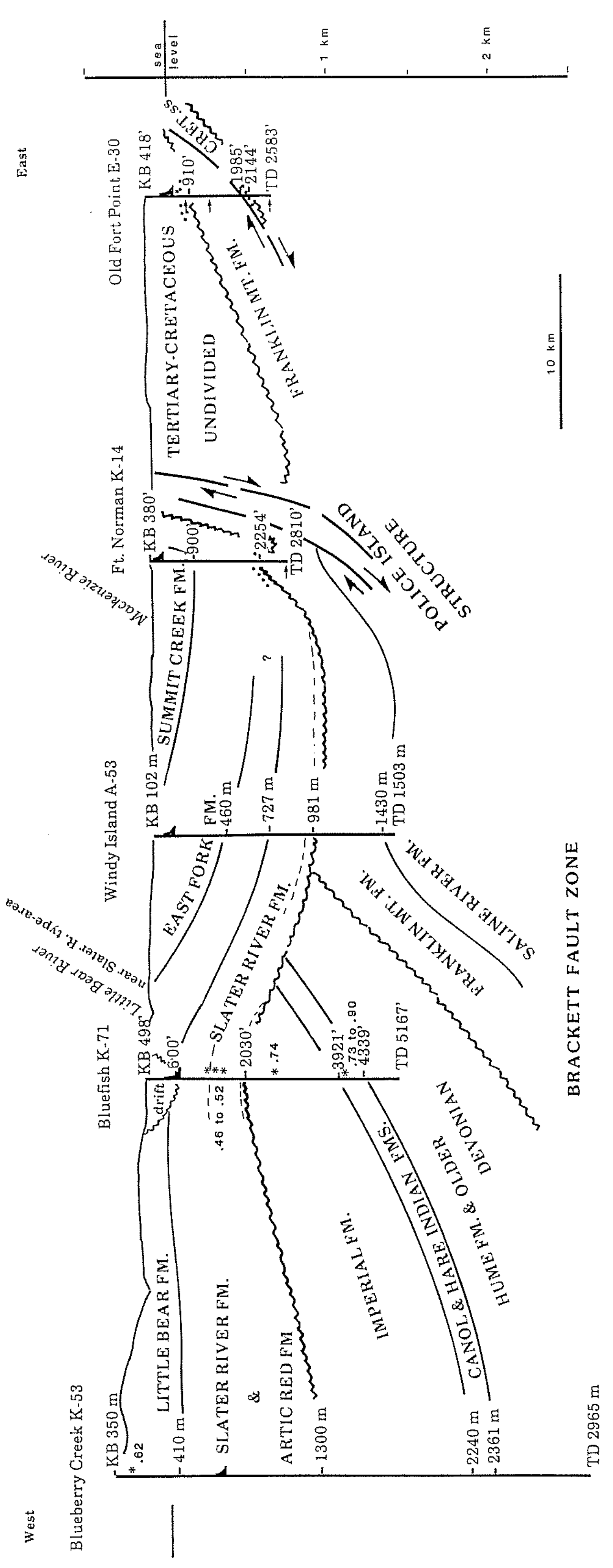

North
South
Little Bear I-70
KB 1367'
700'
2780'
4380'
5670'
TD 7020'
.52
type-area of Summit Ck. Fm.
Tate J-65
KB 1993'
1460'
3745'
4810'
5742'
5992'
7600'
TD 9298'
.37 to .42
.43 to .57
.51 to .55
.56
.58
(2.16 to 2.95)
.77 to 1.28
SUMMIT CREEK FM.
EAST FORK FM.
LITTLE BEAR FM.
SLATER RIVER FM.
IMPERIAL FM.
CANOL & HARE INDIAN FMS.
HUME FM. & OLDER DEVONIAN
DELORME GP.
Arnica
Stewart B-30
KB 1689'
1550'
3550'
4580'
5433'
5974'
8625'
TD 9671'
.56
1.28 to 1.39
Platform Dol.
anomalously thin, probably anhydrite has been leached
Keele River
Keele South E-19
KB 305 m
225 m
550 m
932 m
TD 2450 m
.64
.64
.92
1.09
1.17
TD in Imperial Fm.,
KEELE FAULT ZONE
Keele South A-28
KB 1280'
980'
1485'
2590'
3870'
7735'
8455'
TD 10,500'
.53
.60
.60
.59
1.63
1.94
1.67
2.62
Redstone River
Cloverleaf I-46
KB 1456'
1100'
2475'
4740'
5517'
10,780'
TD 11,318'
.54 to .81
.99
1.29
1.69
1.72
(1.88)
LITTLE BEAR FM.
Jean Marie-like ls
IMPERIAL FM.
CANOL & HARE INDIAN FMS.
HUME FM. & OLDER DEVONIAN
Arnica Platform Dol.
10 km
sea level
1 km
2 km
3 km

Figure 27


West
East
Blueberry Creek K-53
KB 350 m
* .62
410 m
1300 m
2240 m
2361 m
TD 2965 m
LITTLE BEAR FM.
SLATER RIVER FM.
&
ARTIC RED FM
IMPERIAL FM.
CANOL & HARE INDIAN FMS.
HUME FM. & OLDER
DEVONIAN
Bluefish K-71
KB 498'
drift
600'
.46 to .52
* ** *
2030'
* .74
3921'
* .73 to .90
4339'
TD 5167'
SLATER RIVER FM.
Little Bear River
near Slater R. type-area
Windy Island A-53
KB 102 m
EAST FORK FM.
460 m
727 m
981 m
1430 m
TD 1503 m
FRANKLIN MT. FM.
SALINE RIVER FM.
BRACKETT FAULT ZONE
Mackenzie River
SUMMIT CREEK FM.
?
Ft. Norman K-14
KB 380'
900'
2254'
TD 2810'
POLICE ISLAND
STRUCTURE
TERTIARY-CRETACEOUS
UNDIVIDED
FRANKLIN MT. FM.
Old Fort Point E-30
KB 418'
910'
1985'
2144'
TD 2583'
CRET. ss
sea level
1 km
2 km
10 km

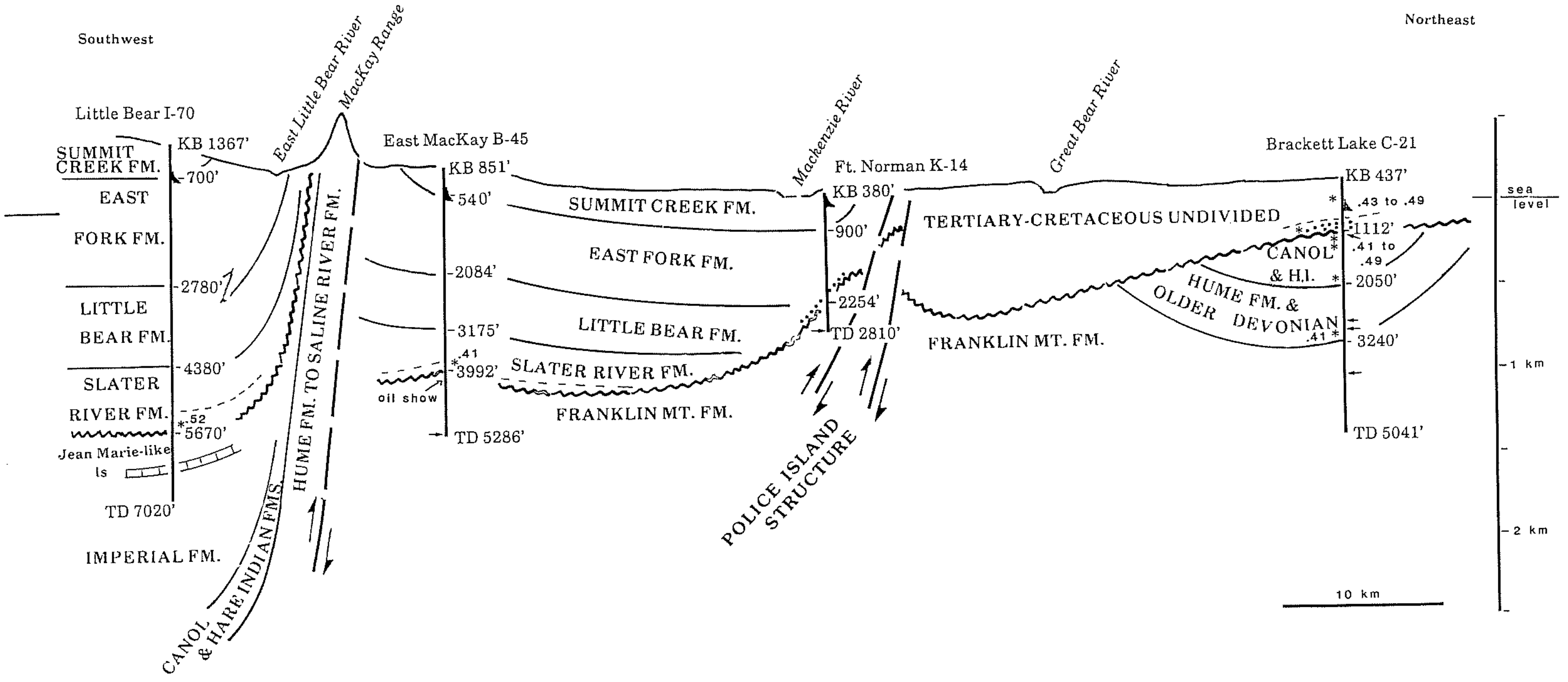

Southwest
Northeast
Little Bear I-70
KB 1367'
SUMMIT CREEK FM.
-700'
EAST FORK FM.
-2780'
LITTLE BEAR FM.
-4380'
SLATER RIVER FM.
.52
-5670'
Jean Marie-like ls
TD 7020'
IMPERIAL FM.
CANOL & HARE INDIAN FMS.
HUME FM. TO SALINE RIVER FM.
East Little Bear River
MacKay Range
East MacKay B-45
KB 851'
-540'
-2084'
-3175'
.41
-3992'
oil show
TD 5286'
SUMMIT CREEK FM.
EAST FORK FM.
LITTLE BEAR FM.
SLATER RIVER FM.
FRANKLIN MT. FM.
Mackenzie River
Ft. Norman K-14
KB 380'
-900'
-2254'
TD 2810'
POLICE ISLAND STRUCTURE
Great Bear River
TERTIARY-CRETACEOUS UNDIVIDED
FRANKLIN MT. FM.
Brackett Lake C-21
KB 437'
.43 to .49
-1112'
.41 to .49
CANOL & H.I.
-2050'
HUME FM. & OLDER DEVONIAN
.41
-3240'
TD 5041'
sea level
1 km
2 km
10 km

Figure 28

Figure 29

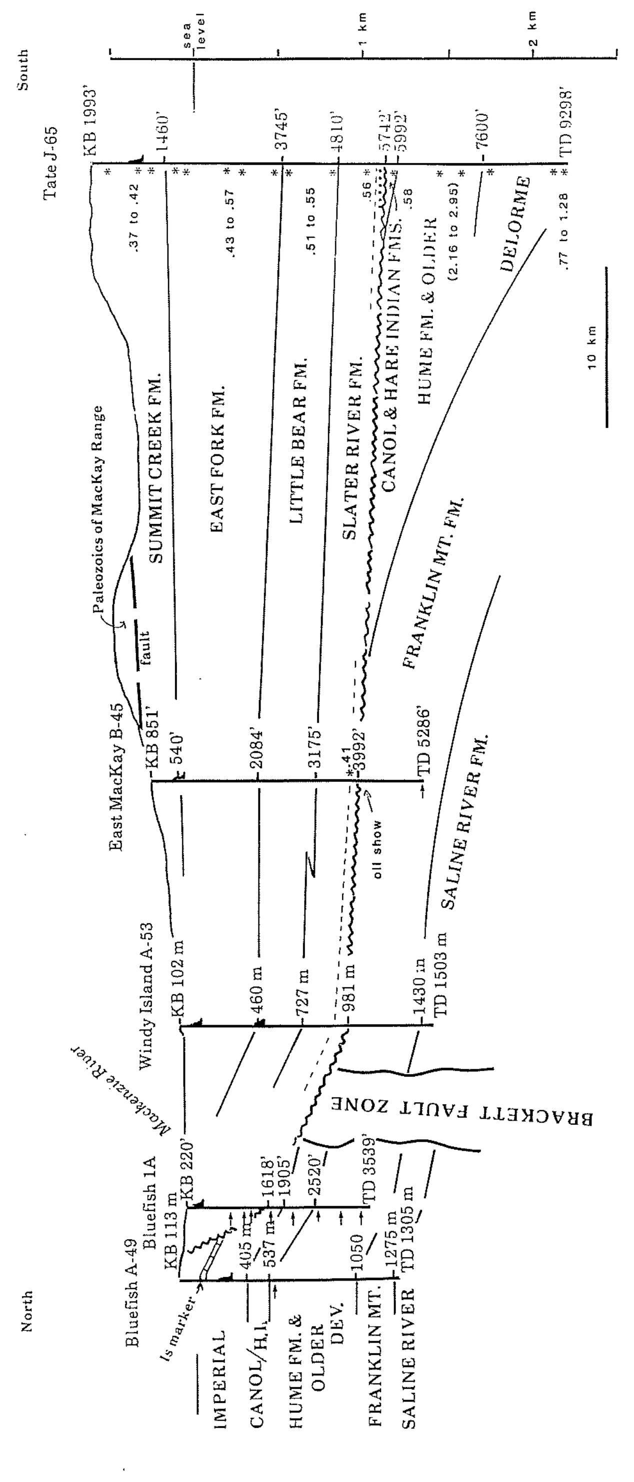

South
North
Tate J-65
KB 1993'
1460'
3745'
4810'
5742'
5992'
7600'
TD 9298'
sea level
1 km
2 km
10 km
.37 to .42
.43 to .57
.51 to .55
.56
.58
(2.16 to 2.95)
.77 to 1.28
SUMMIT CREEK FM.
EAST FORK FM.
LITTLE BEAR FM.
SLATER RIVER FM.
CANOL & HARE INDIAN FMS.
HUME FM. & OLDER
DELORME
FRANKLIN MT. FM.
SALINE RIVER FM.
Paleozoics of MacKay Range
fault
East MacKay B-45
KB 851'
540'
2084'
3175'
.41
3992'
TD 5286'
oil show
Windy Island A-53
KB 102 m
460 m
727 m
981 m
1430 m
TD 1503 m
Mackenzie River
BRACKETT FAULT ZONE
Bluefish 1A
KB 220'
1618'
1905'
2520'
TD 3539'
Bluefish A-49
KB 113 m
405 m
537 m
1050
1275 m
TD 1305 m
ls marker
IMPERIAL
CANOL/H.I.
HUME FM. & OLDER DEV.
FRANKLIN MT.
SALINE RIVER

Figure 30

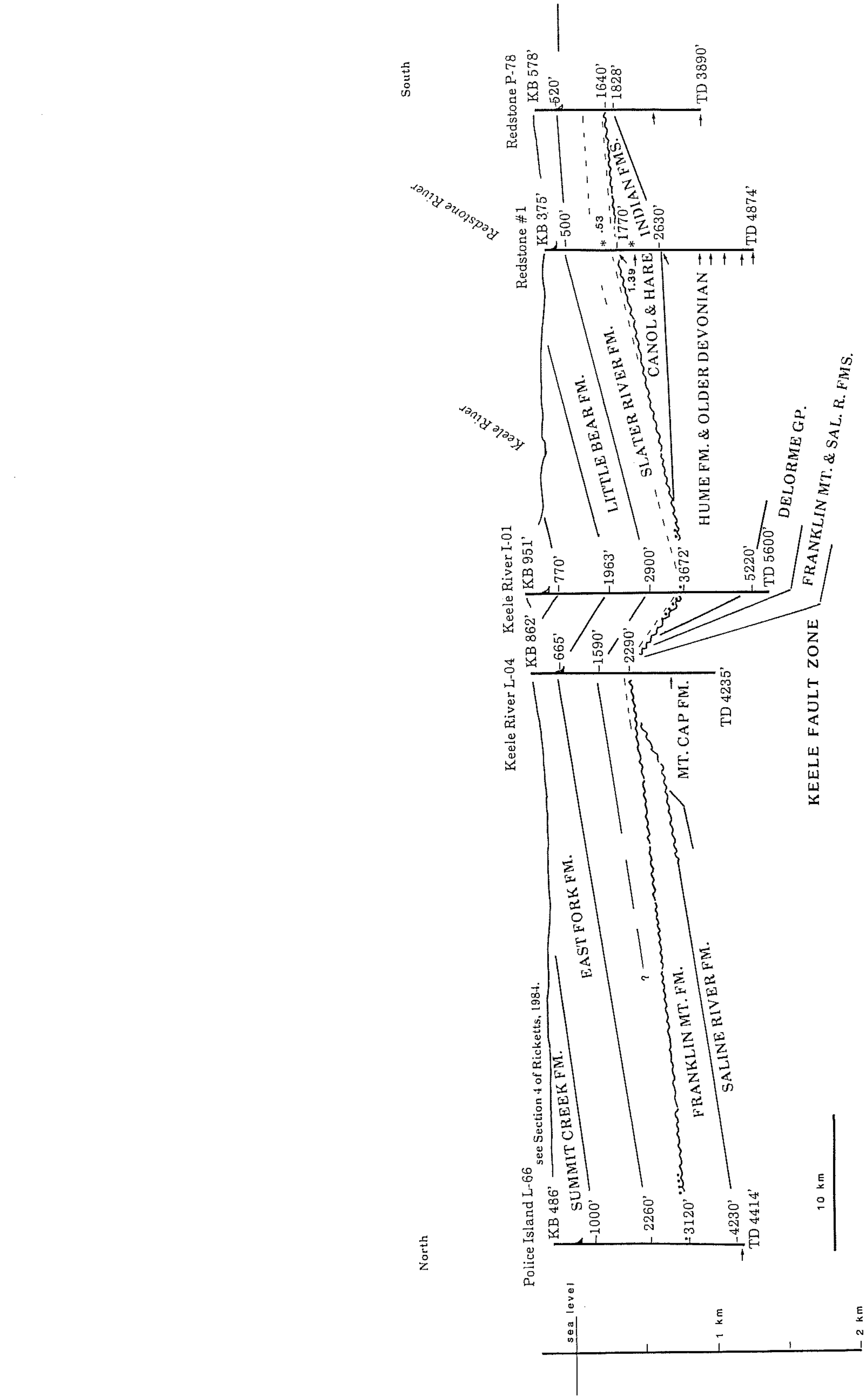
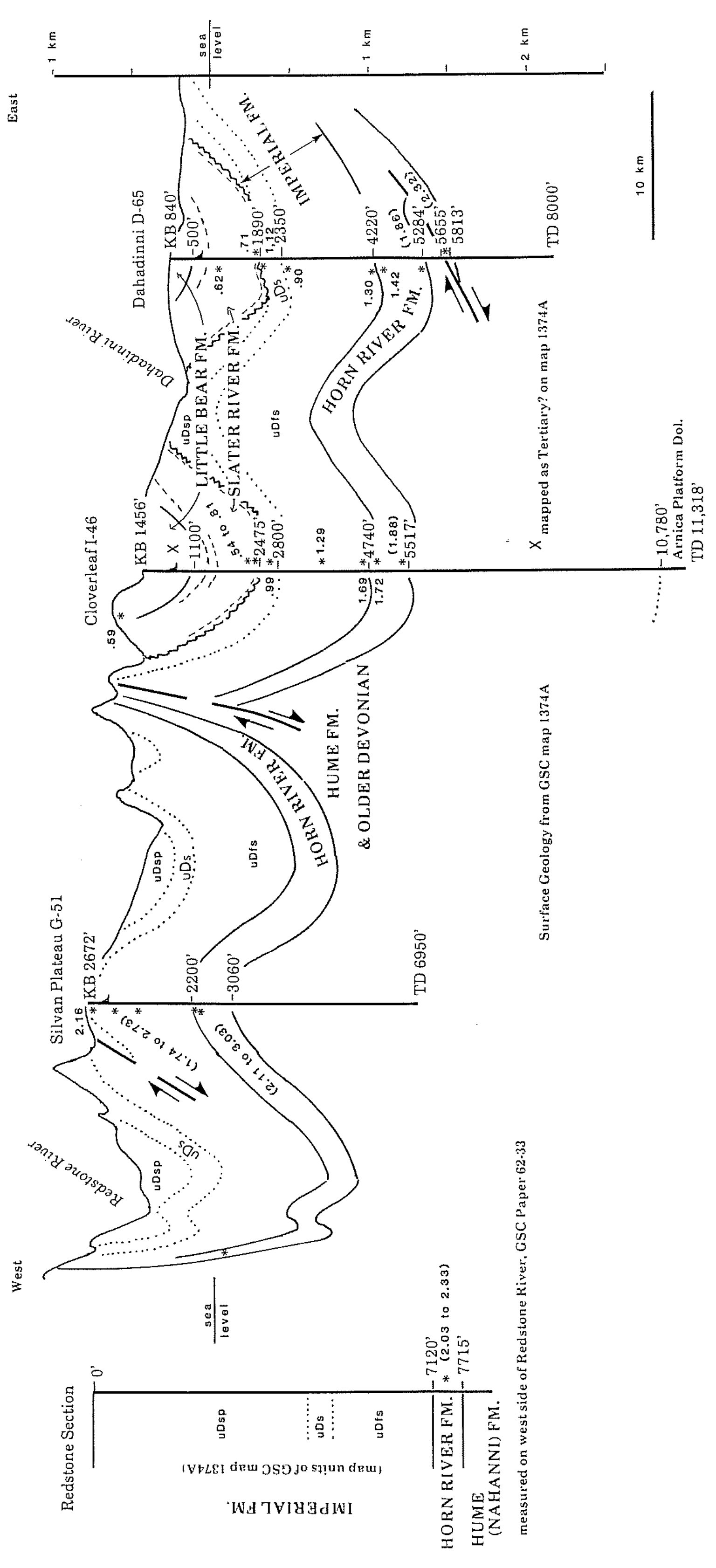
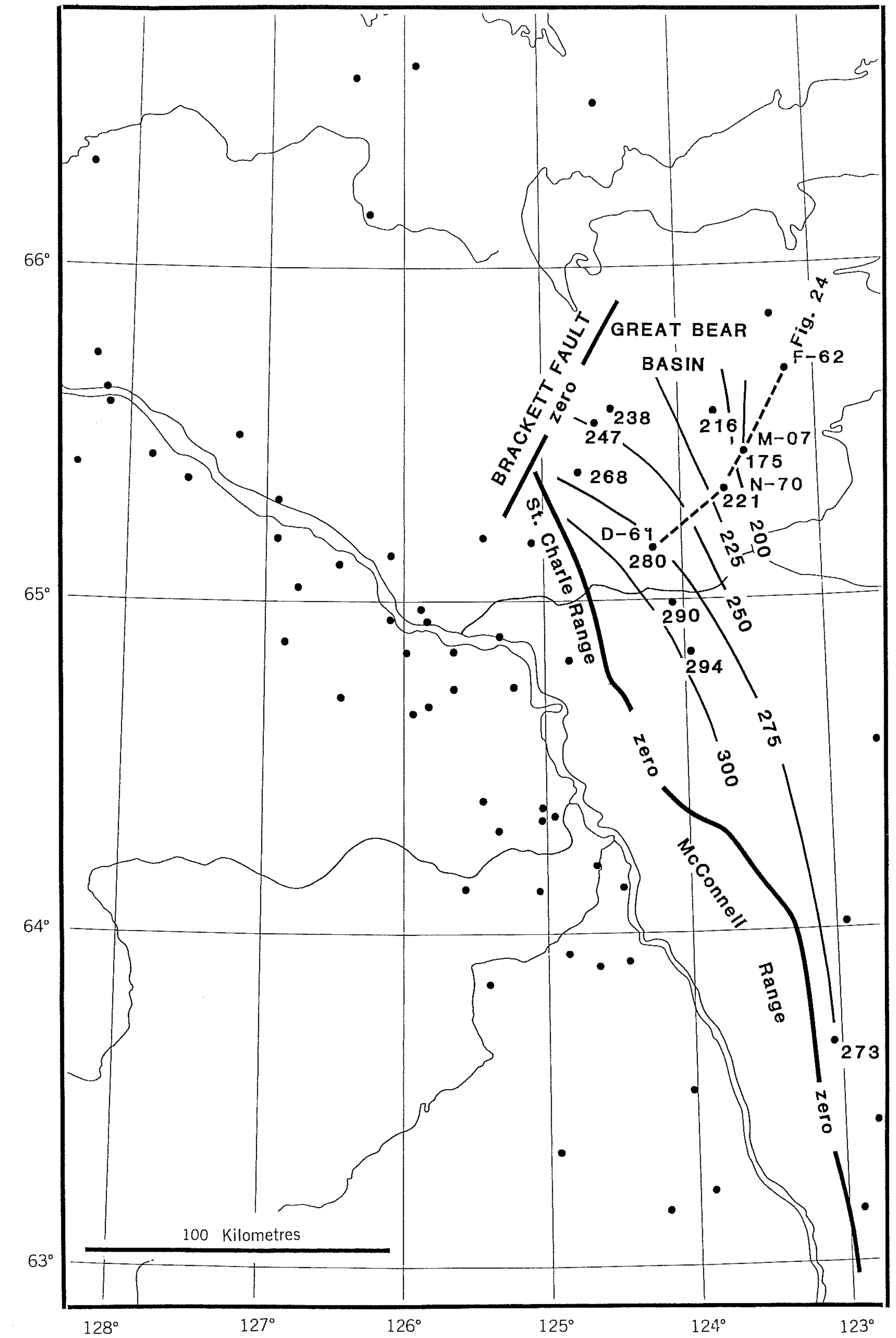

North
South
Police Island L-66
see Section 4 of Ricketts, 1984.
KB 486'
-1000'
2260'
3120'
-4230'
TD 4414'
Keele River L-04
KB 862'
-665'
-1590'
-2290'
TD 4235'
Keele River I-01
KB 951'
-770'
-1963'
-2900'
-3672'
-5220'
TD 5600'
Redstone #1
KB 375'
-500'
* .53
-1770'
1.39
* -2630'
TD 4874'
Redstone P-78
KB 578'
-520'
-1640'
-1828'
TD 3890'
Keele River
Redstone River
SUMMIT CREEK FM.
EAST FORK FM.
?
FRANKLIN MT. FM.
SALINE RIVER FM.
MT. CAP FM.
LITTLE BEAR FM.
SLATER RIVER FM.
CANOL & HARE
INDIAN FMS.
HUME FM. & OLDER DEVONIAN
DELORME GP.
FRANKLIN MT. & SAL. R. FMS.
KEELE FAULT ZONE
10 km
sea level
1 km
2 km

Figure 31

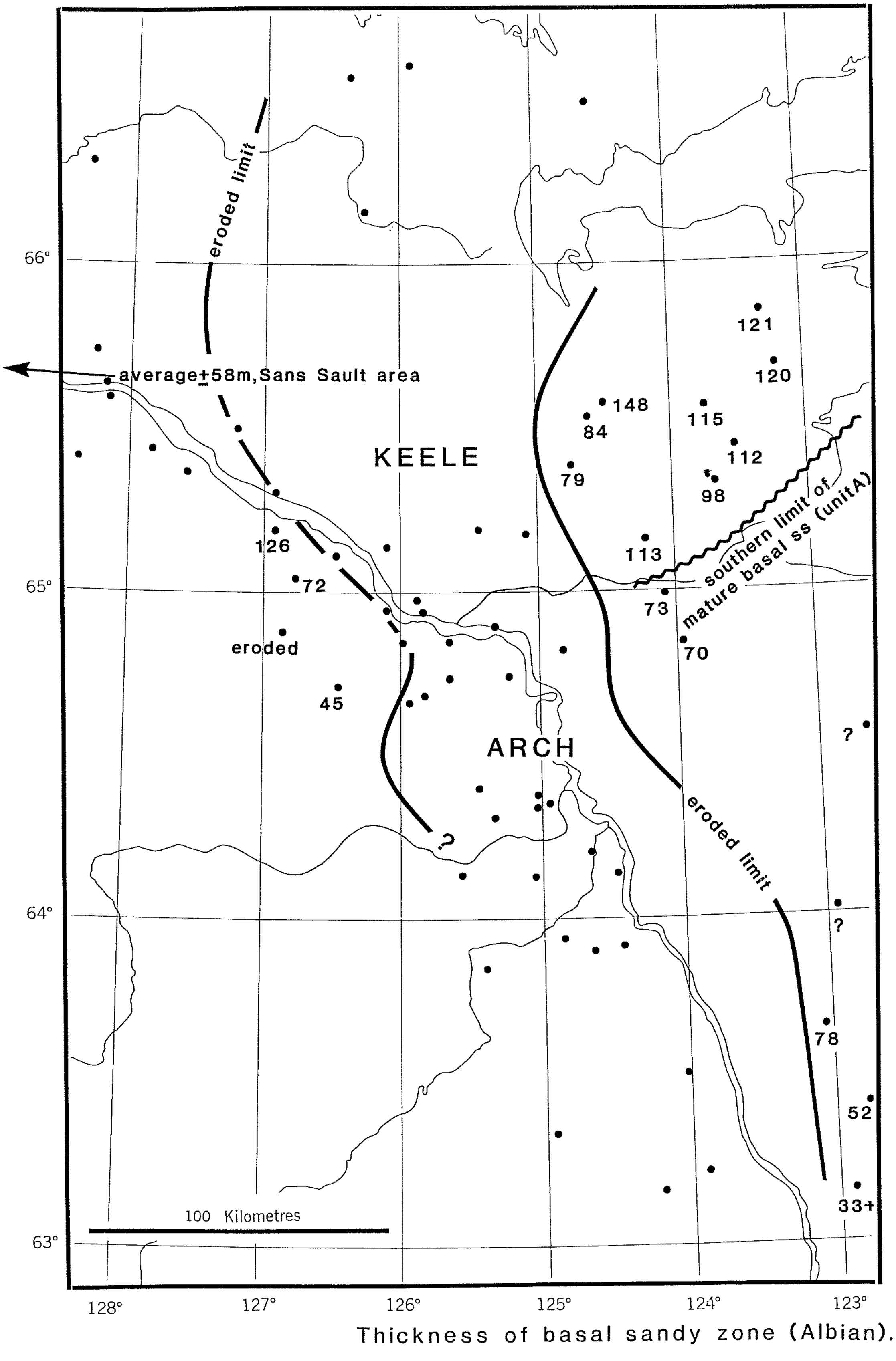


East
West
1 km
sea level
- 1 km
- 2 km
10 km
Dahadinni D-65
KB 840'
-500'
.71
*1890'
1.12
-2350'
-4220'
(1.86)
5284'
(2.32)
-5655'
5813'
TD 8000'
.62 *
uDs
.90
1.30
1.42
IMPERIAL FM.
Dahadinni River
LITTLE BEAR FM.
SLATER RIVER FM.
HORN RIVER FM.
uDsp
uDfs
Cloverleaf I-46
KB 1456'
X
-1100'
.54 to .81
*2475'
*2800'
* 1.29
*4740'
(1.88)
*5517'
.59
.99
1.69
1.72
-10,780'
Arnica Platform Dol.
TD 11,318'
X mapped as Tertiary? on map 1374A
HUME FM.
& OLDER DEVONIAN
Surface Geology from GSC map 1374A
Silvan Plateau G-51
KB 2672'
2.16
(1.74 to 2.73)
-2200'
-3060'
(2.11 to 3.03)
TD 6950'
Redstone River
Redstone Section
0'
IMPERIAL FM.
(map units of GSC map 1374A)
uDsp
uDs
uDfs
HORN RIVER FM.
-7120'
* (2.03 to 2.33)
-7715'
HUME (NAHANNI) FM.
measured on west side of Redstone River, GSC Paper 62-33

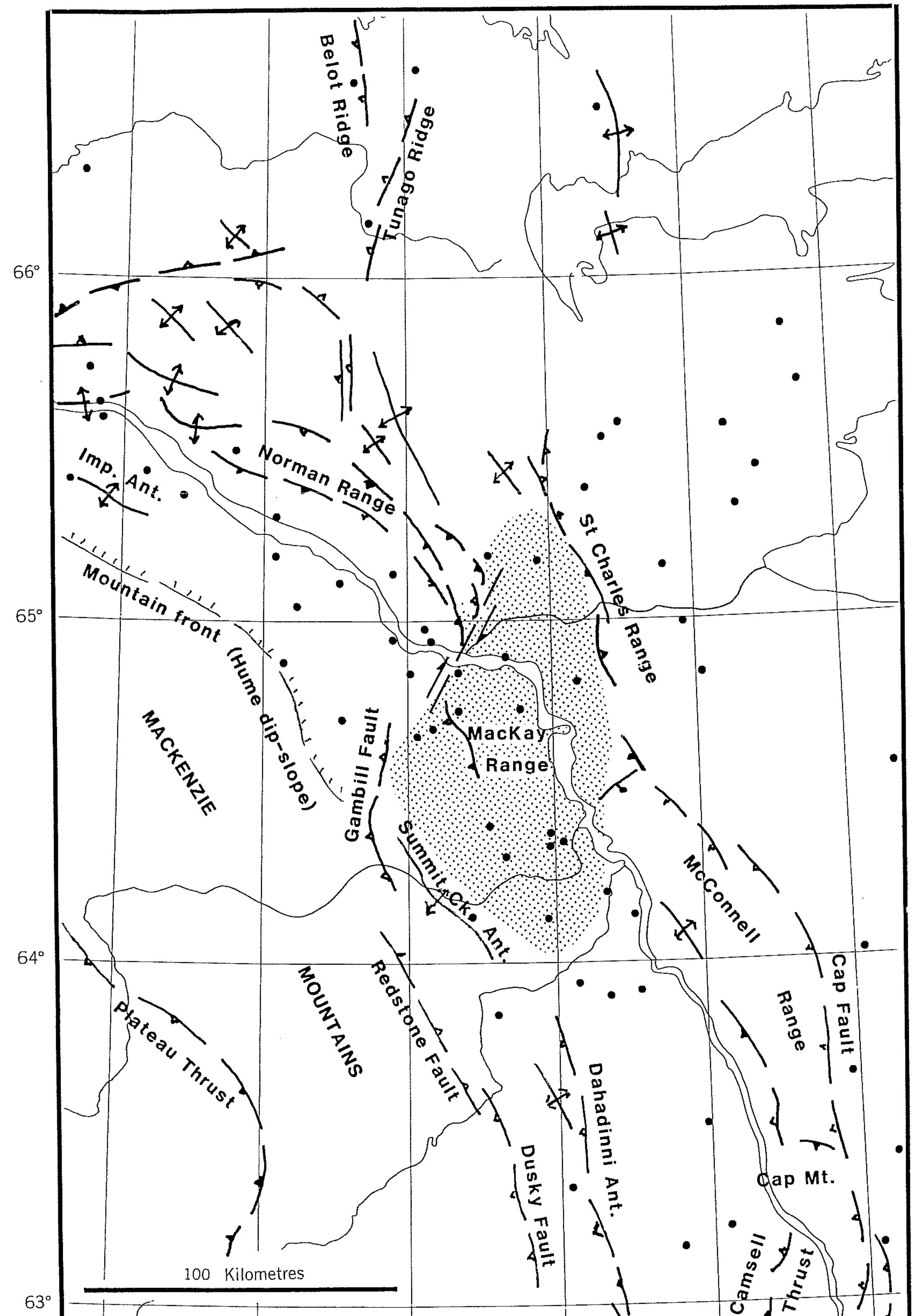


Isopachs, Cretaceous units C and D (part of Albian)

Figure 33

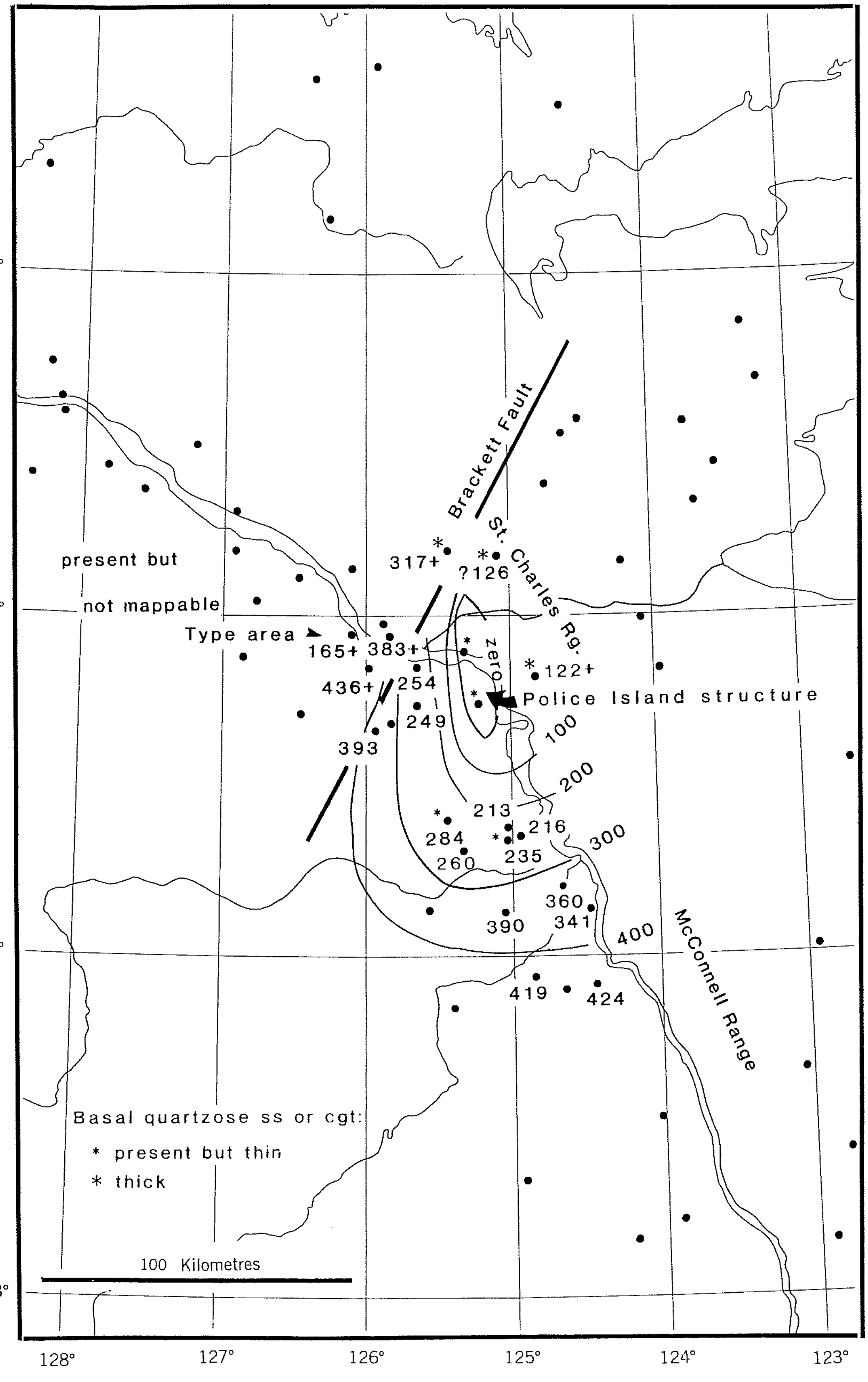


Thickness of basal sandy zone (Albian).



Brackett Basin (stippled) in relation to Laramide structures.

Figure 35



Isopachs, Slater River Formation.

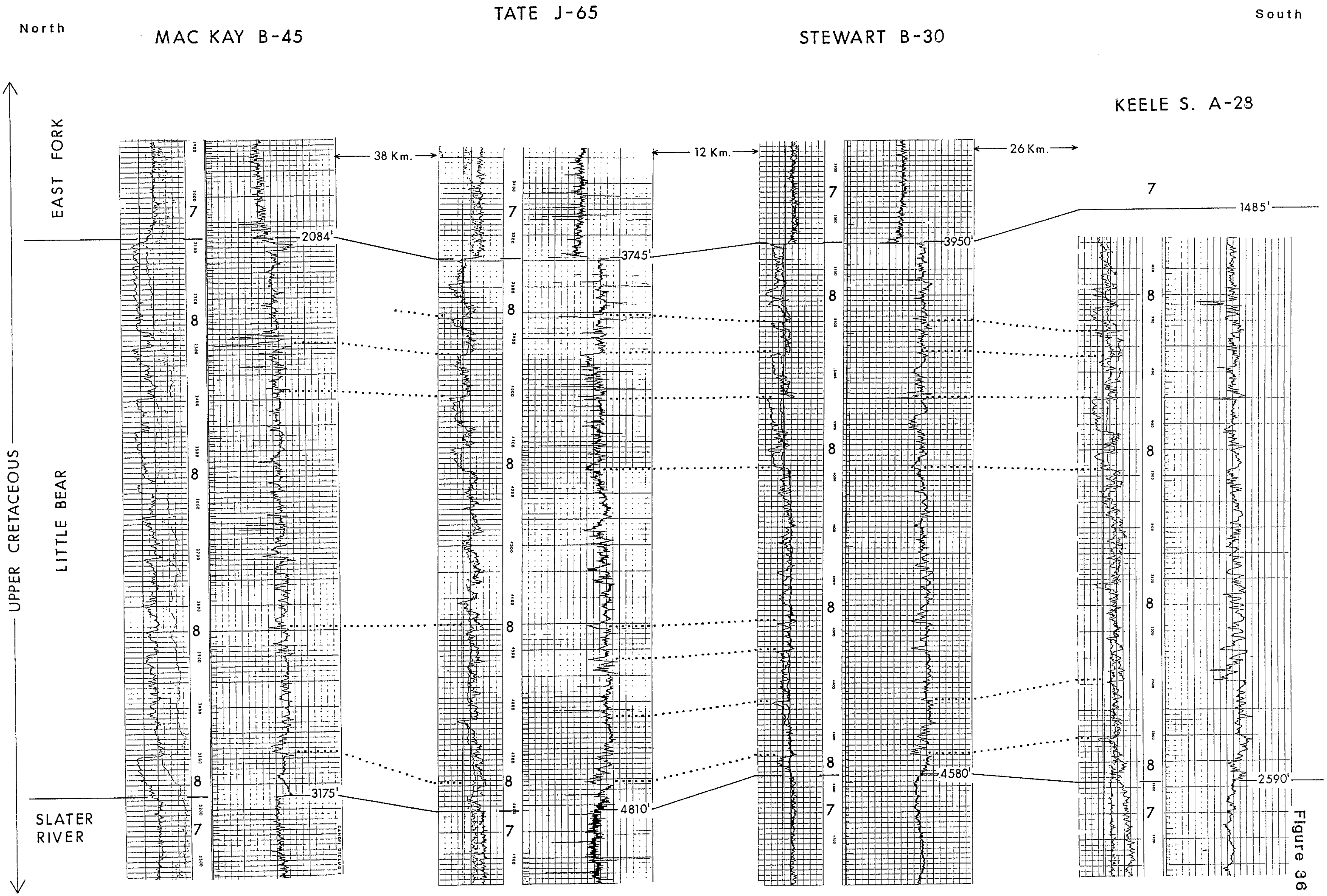
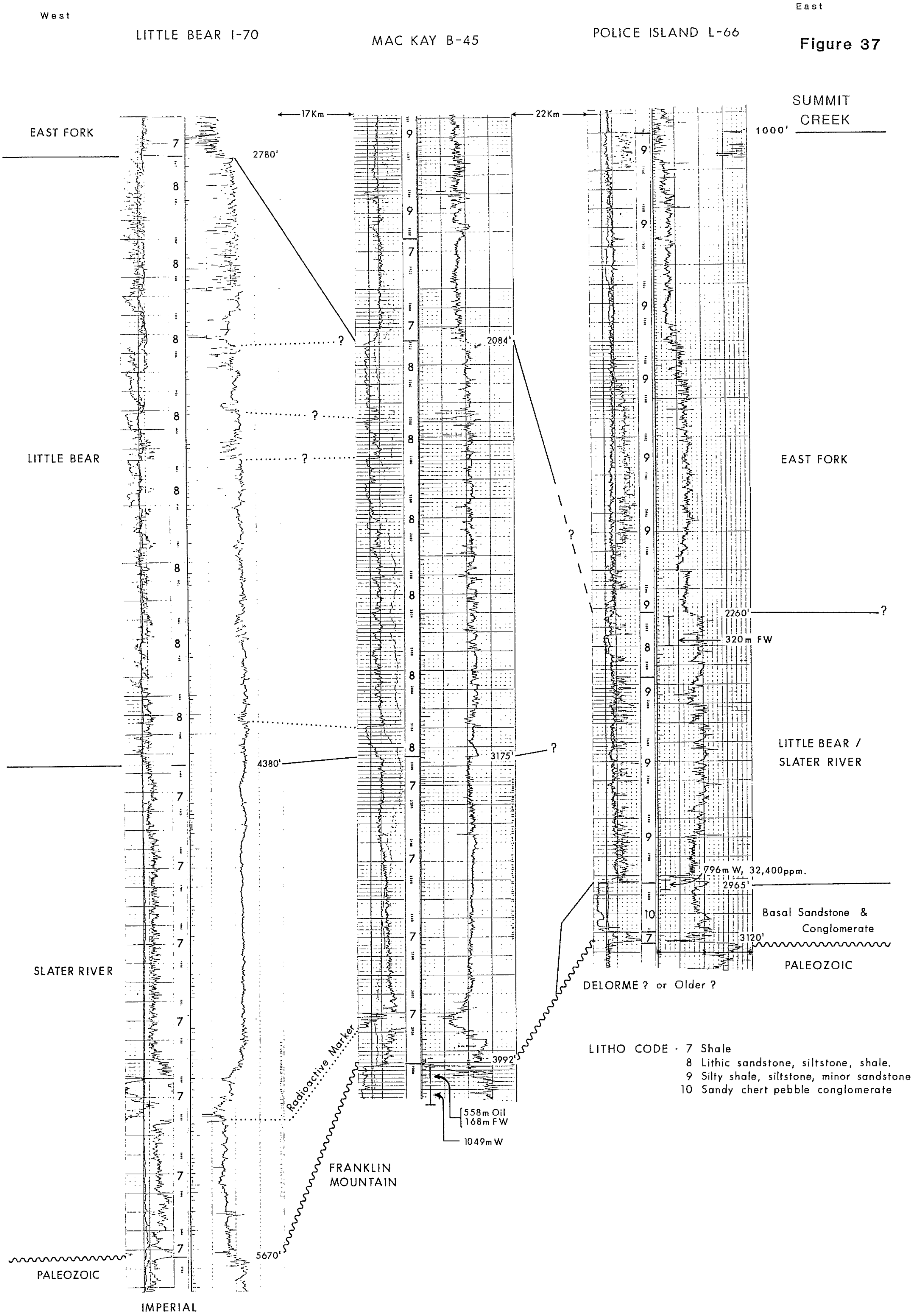
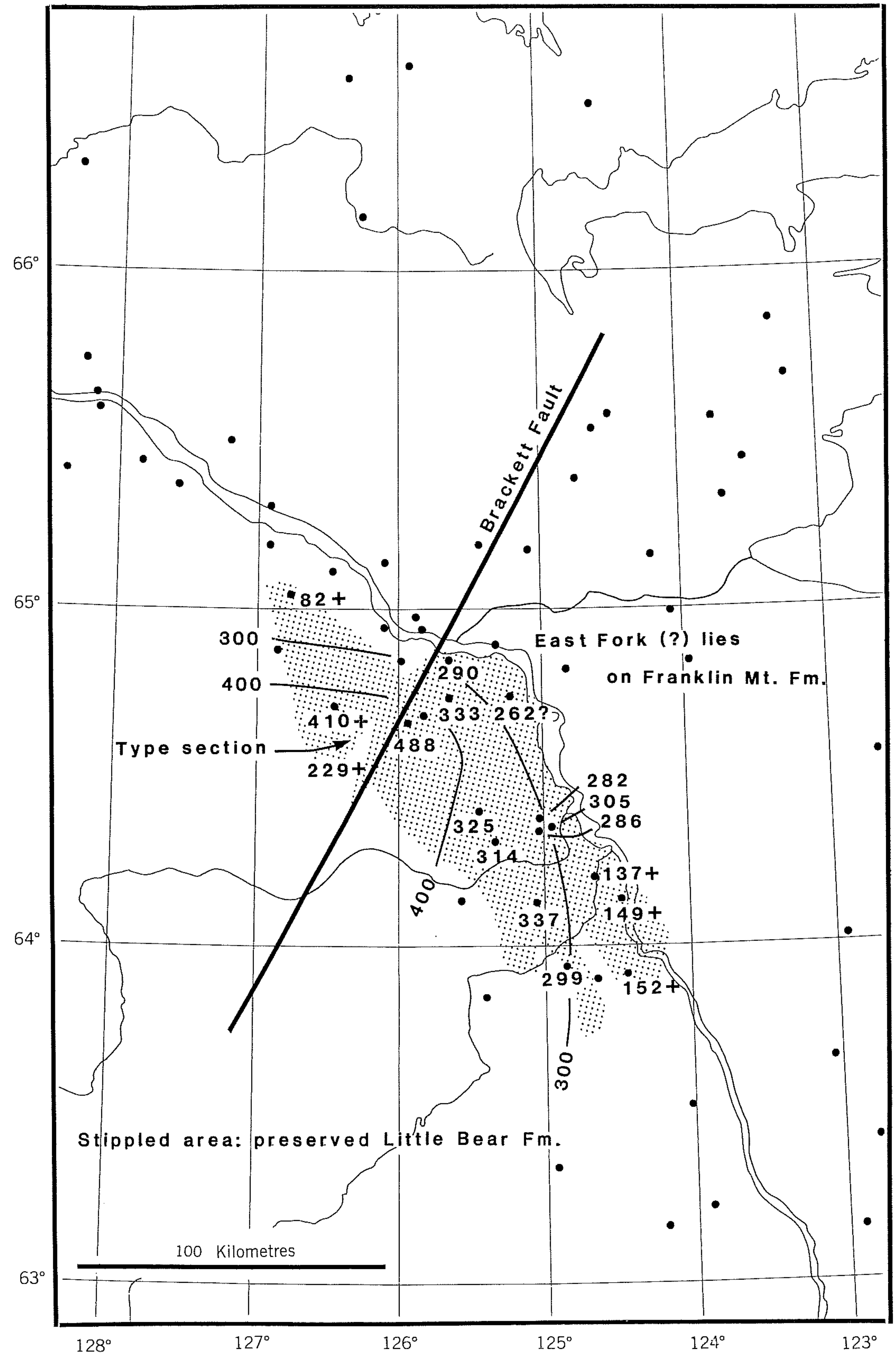


Figure 36

LITHO CODE: 7 Shale.
8 Lithic sandstone, siltstone, shale.

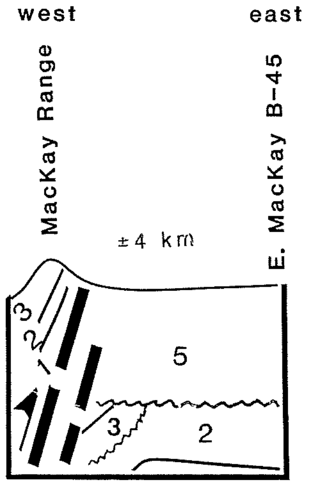
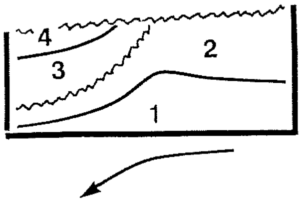
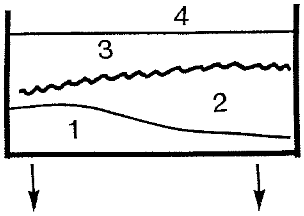
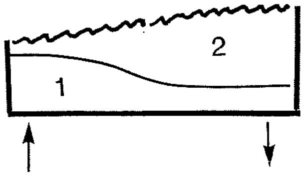
Figure 37


West
East
LITTLE BEAR I-70
MAC KAY B-45
POLICE ISLAND L-66
17Km
22Km
SUMMIT CREEK
1000'
EAST FORK
2780'
?
?
?
2084'
?
LITTLE BEAR
EAST FORK
2260'
?
320m FW
4380'
3175'
?
LITTLE BEAR / SLATER RIVER
796m W, 32,400ppm.
2965'
Basal Sandstone & Conglomerate
3120'
PALEOZOIC
DELORME ? or Older ?
SLATER RIVER
Radioactive Marker
3992'
558m Oil
168m FW
1049m W
FRANKLIN MOUNTAIN
5670'
PALEOZOIC
IMPERIAL
7
8
9
10
LITHO CODE · 7 Shale
8 Lithic sandstone, siltstone, shale.
9 Silty shale, siltstone, minor sandstone
10 Sandy chert pebble conglomerate



Isopachs, Little Bear Formation

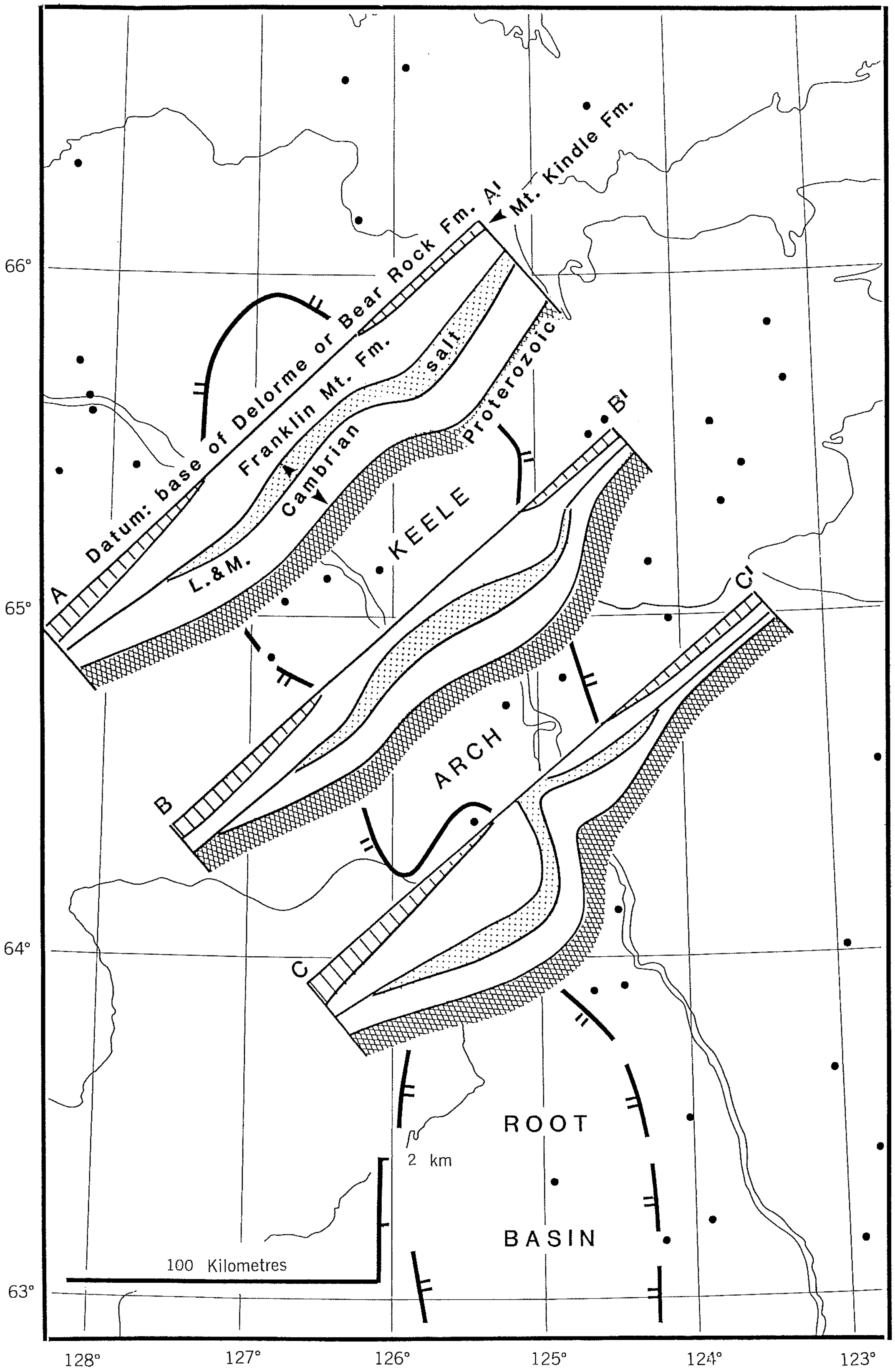
Figure 39.

TIME	EVENT	RESULT
LATE PALEOCENE	Mountain building, MacKay Range assumes it's present form, Franklin Mt. Fm. uplifted to about 350 m above the plains.	
PALEOCENE TO LATE CRETACEOUS	Sediments of the Brackett Basin, ±2 km.	
PRE-CRETACEOUS LACUNA	Tilt to the west, development of a sharp flexure, steepest dip on the west, deep erosion.	
LATE PALEOZOIC TO MIDDLE DEVONIAN	Onlap by Bear Rock & Hume Fms. (±600m) followed by at least 1 km. of younger Devonian sediments, later history unknown.	
EARLY DEVONIAN	Uplift & erosion, either an up on the west fault or a west facing escarpment, or both.	

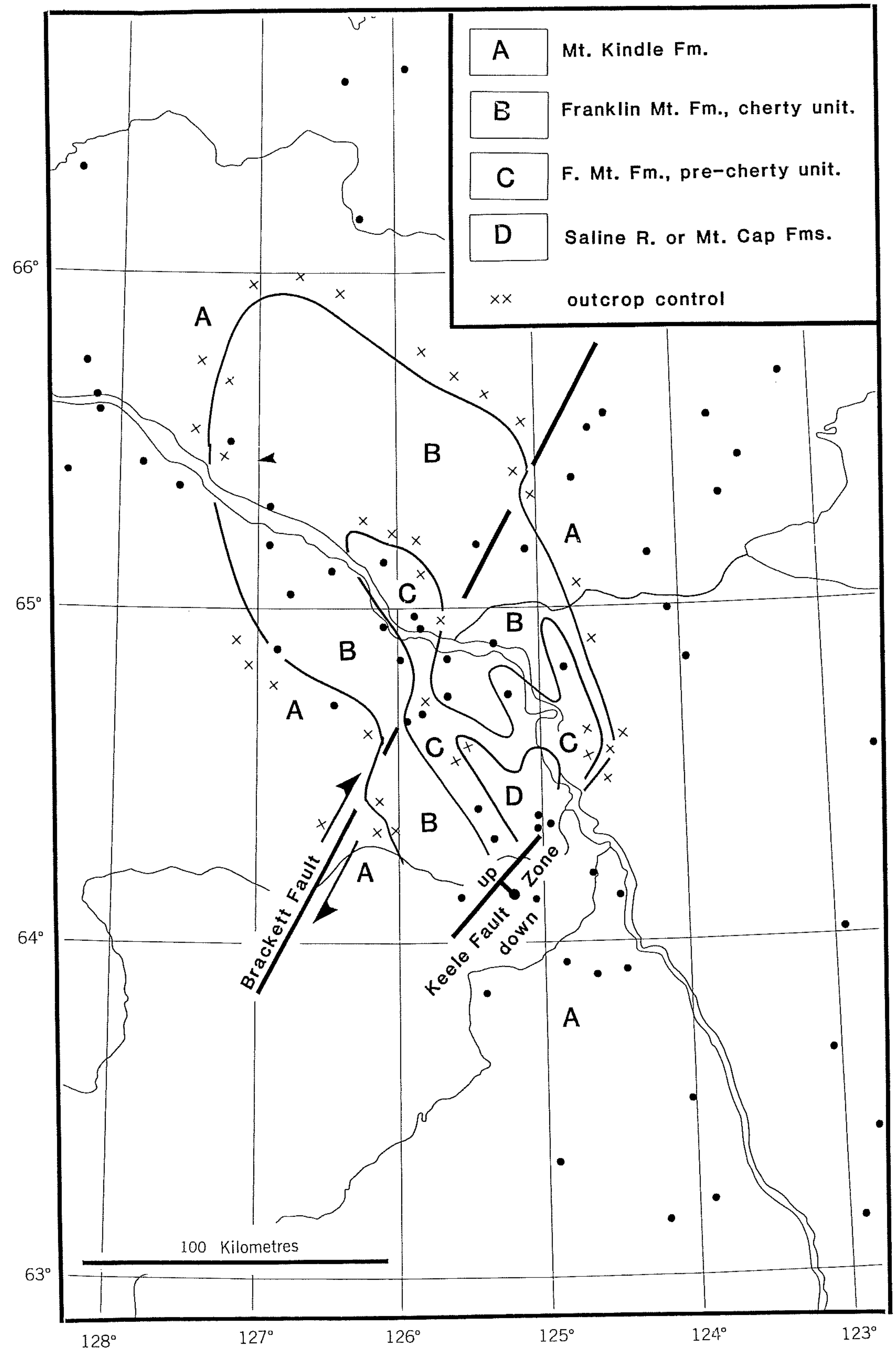
1, Saline River Fm. 2, Franklin Mt. Fm. 3, Hume and Bear Rock Fms.
4, post-Hume Devonian, 5, Late Cretaceous and Paleocene.

Stages in the development of the MacKay structure.

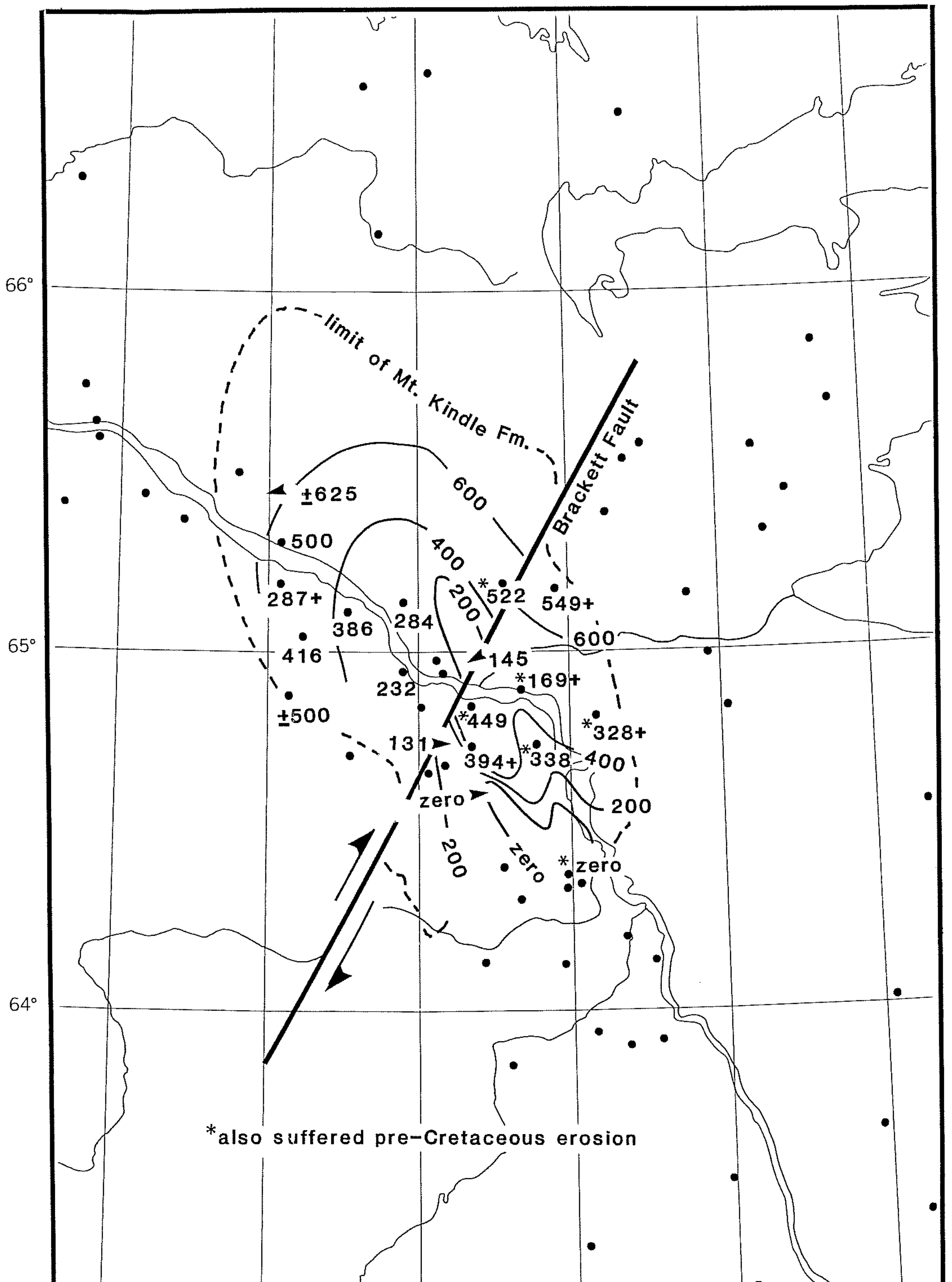
Figure 40



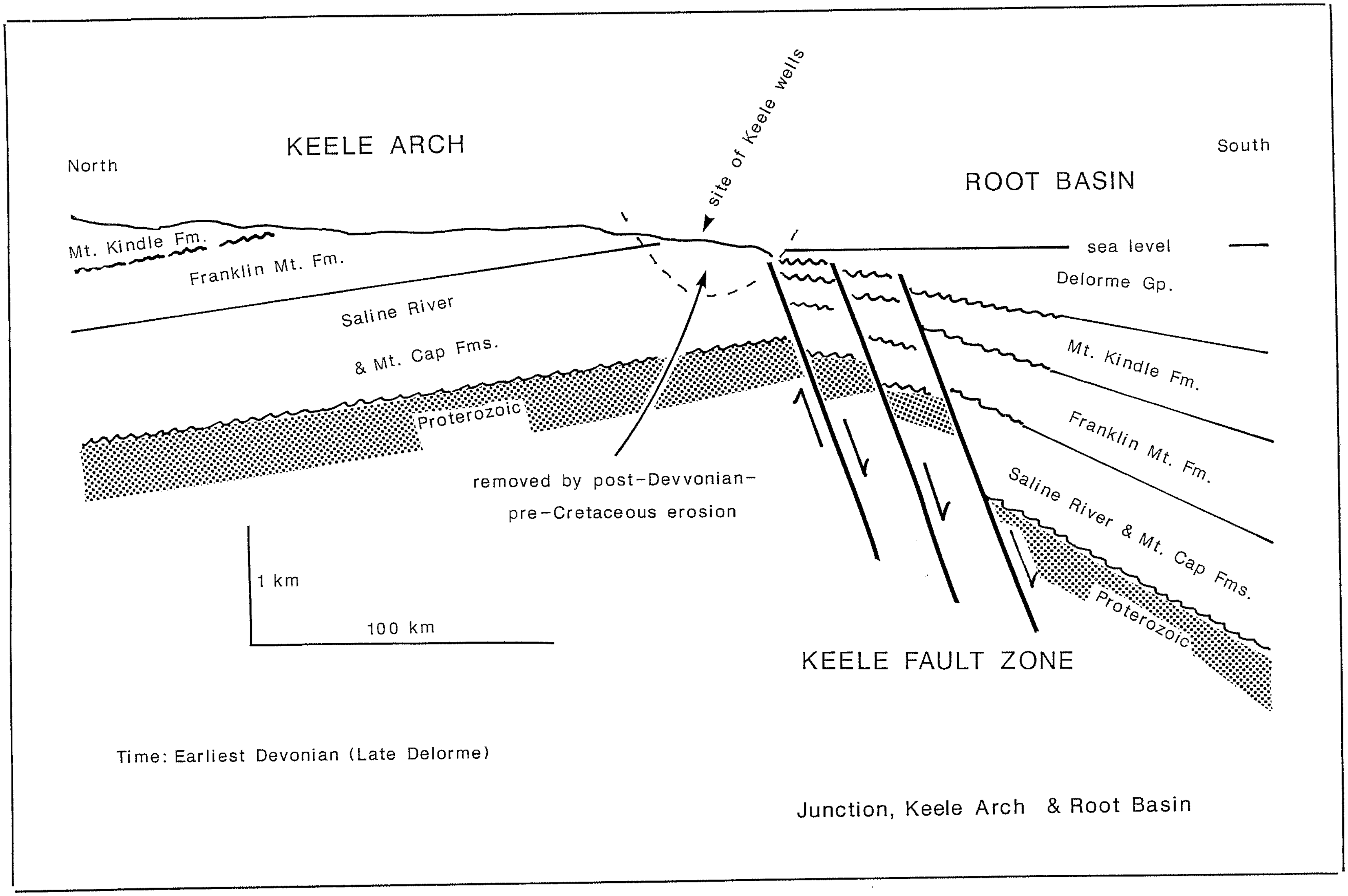
Cross-sections, Keele Arch, early phase



Pre-Bear Rock (or pre-Delorme) geology, Keele Arch, early phase.



Isopachs of remnant of Franklin Mountain Formation, Keele Arch, early phase.

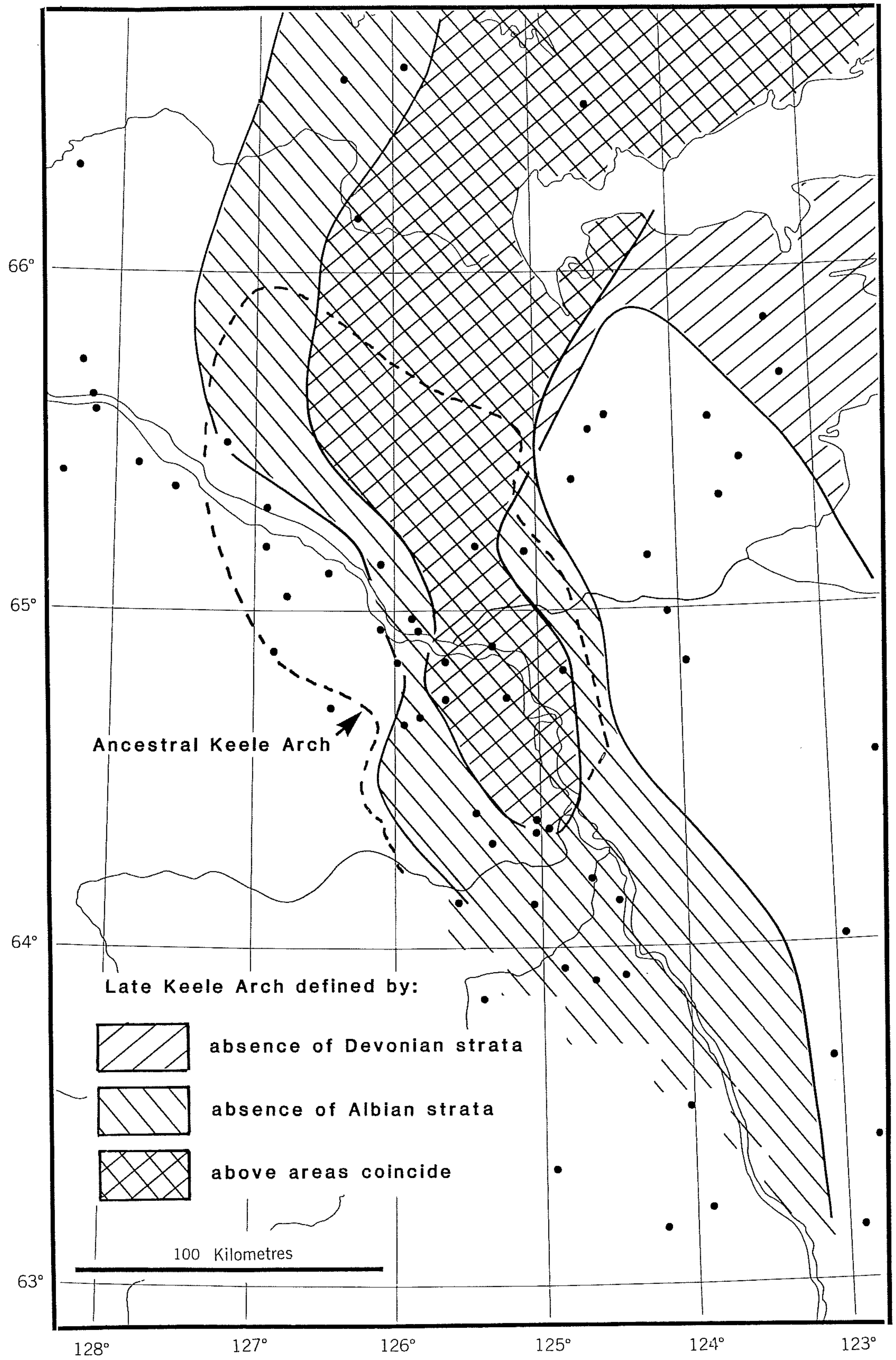


Time: Earliest Devonian (Late Delorme)

Junction, Keele Arch & Root Basin

Figure 43

Figure 44



Keele Arch, late phase (options).