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In Search of the Wegener Fault: Re-Evaluation of Strike-Slip Displacements Along and Bordering Nares Strait

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Abstract: A total of 28 geological-geophysical markers are identified that relate to the question of strike slip motions along and bordering Nares Strait. Eight of the twelve markers, located within the Phanerozoic orogen of Kennedy Channel – Robeson Channel region, permit between 65 and 75 km of sinistral offset on the Judge Daly Fault System (JDFS). In contrast, eight of nine markers located in Kane Basin, Smith Sound and northern Baffin Bay indicate no lateral displacement at all. Especially convincing is evidence, presented by DAMASKE & OAKEY (2006), that at least one basic dyke of Neoproterozoic age extends across Smith Sound from Inglefield Land to inshore eastern Ellesmere Island without any recognizable strike slip offset. These results confirm that no major sinistral fault exists in southern Nares Strait.

To account for the absence of a Wegener Fault in most parts of Nares Strait, the present paper would locate the late Paleocene-Eocene Greenland plate boundary on an interconnected system of faults that are 1) traced through Jones Sound in the south, 2) lie between the Eurekan Orogen and the Precambrian shield on southeastern Ellesmere Island, and 3) follow the Judge Daly Fault System as far north as the Lincoln Sea. 350 km of sea floor spreading and continental margin extension in northern Baffin Bay are accounted for in the north by a series of synchronous tectonic motions including proposed northeast-southwest extension in Cretaceous – early Paleocene time on Ellesmere and Axel Heiberg islands (30 to 55 km), north-south extension during the Eurekan Orogeny in Lancaster Sound and Devon Island (55 km), dextral strike slip and extension in Jones Sound and on southeast Ellesmere Island linked to thrusting north of Bache Peninsula and sinistral strike slip on the JDFS (70 km) and, south-directed followed by southwest-directed shortening across the Sverdrup Basin (100 km +35 km). The latter deformation is attributed to the counterclockwise rotation and escape to the west of a semi-rigid northern Ellesmere block during collision with the Greenland plate.

Zusammenfassung: Die Korrelierbarkeit von insgesamt 28 geologischen Einheiten auf beiden Seiten der Nares Strait wird untersucht, um damit Argumente für oder gegen eine Seitenverschiebung in oder entlang der Nares Strait zu finden. Innerhalb des phanerozoischen Orogens der nördlichen Region Kennedy Channel – Robeson Channel lassen acht der untersuchten zwölf Einheiten einen sinistralen Versatz von 65-75 km am Judge Daly Störungssystem zu. Dagegen erlauben in der südlichen Region von Kane Basin, Smith Sound und der nördlichen Baffin Bay acht der neun Einheiten überhaupt keine Seitenverschiebung. Besonders überzeugend sind die Ergebnisse von DAMASKE & OAKEY (2006), dass zumindest ein basischer Gang neoproterozoischen Alters den Smith Sound von Inglefield Land bis zur Küste der östlichen Ellesmere Island ohne erkennbaren Versatz überquert. Diese Ergebnisse bestätigen, dass keine bedeutende Seitenverschiebung in der südlichen Nares Strait existiert.

Um dieser Situation Rechnung zu tragen, wird in diesem Beitrag vorgeschlagen, die paläozäne bis eozäne Plattengrenze Grönlands an ein System von Störungen zu knüpfen, das 1) durch den Jones Sound im Süden verfolgt werden kann, das 2) in der südlichen Ellesmere Island zwischen dem Eureka-Orogen und dem präkambrischen Schild verläuft und das 3) dem Judge Daly Störungssystem bis in die Lincoln Sea folgt. Die Öffnung der nördlichen Baffin Bay um 315 km durch Seafloor Spreading und Extension an den Kontinentalrändern wird im Norden durch eine Reihe von gleich alten tektonischen Bewegungen kompensiert. Diese umfassen die NE–SW Extension in Kreide und Paläozän auf Ellesmere und Axel Heiberg Island (30-55 km), N–S Extension während der Eureka Orogenese im Lancaster Sound und auf Devon Island (55 km), dextrale Seitenverschiebungen und Extension im Jones Sound und auf SE Ellesmere Island im Zusammenhang mit Überschiebungen nördlich der Bache Peninsula und Linksseitenverschiebungen am Judge-Daly-Störungssystem (70 km) und schließlich die S-, später SW-gerichtete Kompression des Sverdrup-Beckens (100 + 35 km). Die spätere Deformation wird auf die Rotation (entgegen dem Uhrzeigersinn) und ausweichende Westdrift eines semi-rigiden nördlichen Ellesmere-Blocks während der Kollision mit der Grönlandplatte zurückgeführt.

INTRODUCTION

It is apparent to both earth scientists and the general public that the shape of both coastlines and continental margins of western Greenland and eastern Arctic Canada provide for a satisfactory restoration of the opposing lands. The tectonic significance of this observation was first described by the Quaternary geologist Frank TAYLOR (1910). Alfred WEGENER (1912) successfully promoted the idea and applied it to all the major continents of the earth. A key element of the theory was that the sliding motion of North America necessary to form a rift system in Labrador Sea and Baffin Bay, also required some 330 km of sinistral strike slip on a fault system located in Nares Strait (DAWES & CHRISTIE 1982). From these humble beginnings, grew the compelling geological and later geophysical evidence that all the world's tectonic plates are mobile features and that arrangements of the continents and ocean basins have evolved more or less continuously throughout Earth history. In popular science literature the proposed transform plate boundary between northwestern Greenland and Ellesmere Island, originally named by Tuzo WILSON (1963), came to be known as the "Wegener Fault". All the standard plate reconstructions now place the Paleocene-Eocene plate boundary between North America and Greenland in the marine channel that separates the facing continental landmasses.

In the 1960s and 1970s, while the plate "tectonicians" were running to the ends of the earth to understand the nature and origin of the deep ocean basins, traditional onshore geologists were collecting field evidence and some used this to support (or refute) the new plate theory. One result of this activity in Arctic North America was the landmark publication entitled "Nares Strait and the drift of Greenland: A conflict in plate tectonics" (DAWES & KERR 1982). Briefly stated, the conflict arose from the fact that plate tectonic theory demanded (and some geological features were seen to be consistent with) about 250 km of sinistral strike-slip along Nares Strait (JOHNSON & SRIVASTAVA 1982). In contrast, geologists working along the strait saw geological continuity between Ellesmere Island and Greenland, were able to identify a wide range of specific features consistent with no lateral motion at all and, in the extreme, allowed for not more than 25 km of motion in a

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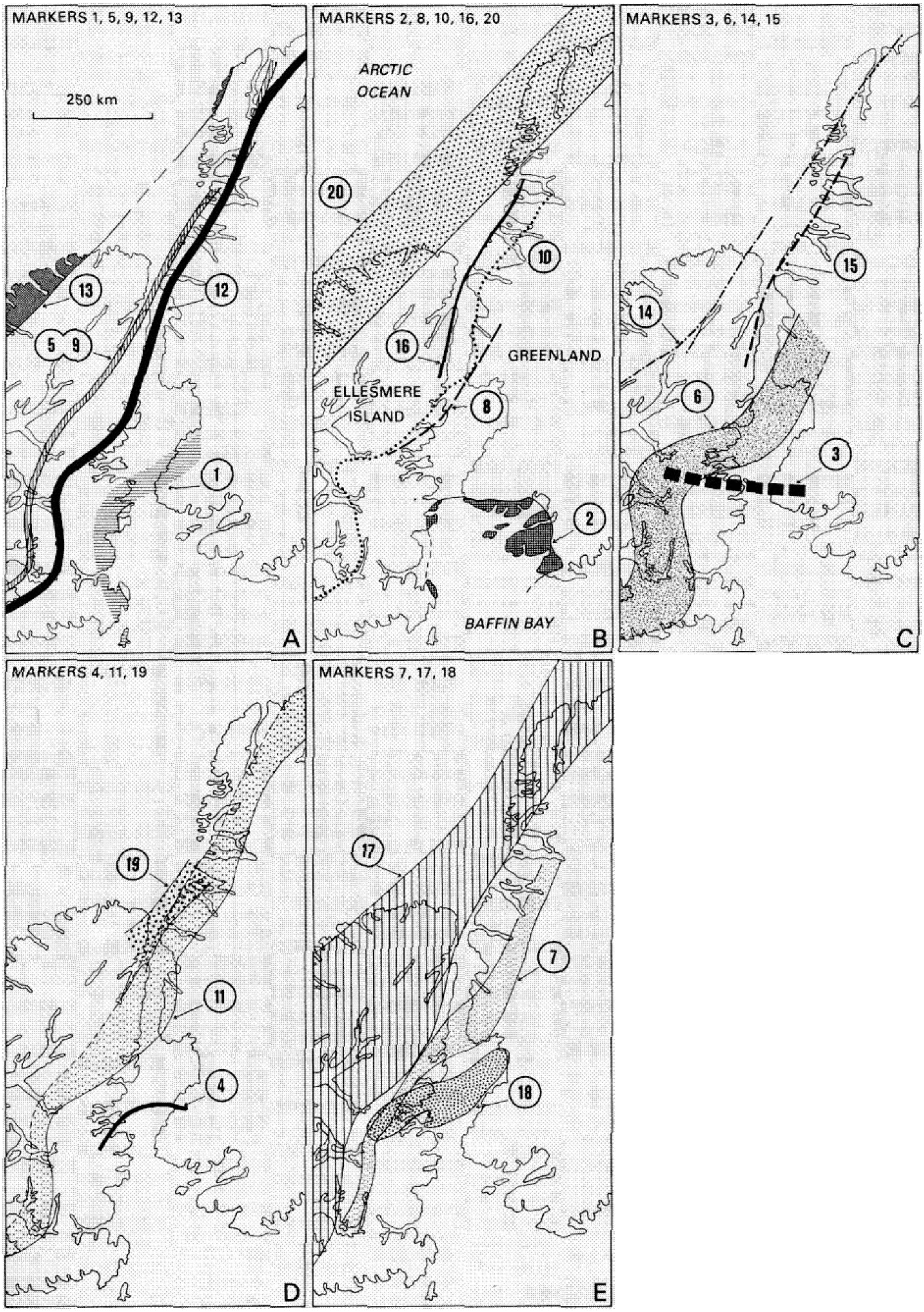


Fig. 1: The location of geological and geophysical markers that cross Nares Strait, reproduced from Figure 3 of DAWES & KERR (1982). These markers are re-assessed in the accompanying text and others, recently identified, are described for the first time (see also Tab. 1). (1) Archaean marble belt; (2) Proterozoic Thule Basin; (3) Proterozoic Bache Peninsula arch; (4) Zero-edge of Proterozoic Thule Group; (5) Cambro-Ordovician platform margin; (6) Lower Ordovician evaporite belt; (7) Middle Ordovician evaporite belt; (8) Cambro-Ordovician hinge line; (9) Ordovician-Silurian platform margin; (10) Lower Silurian facies contact, (11) Silurian carbonate build-up belt; (12) Southern boundary of the Ellesmere-Greenland fold belt; (13) Amphibolite facies metasediments; (14) Lake Hazen – Harder Fjord fracture line; (15) Judge Daly – Nyeboe Land fracture line; (16) Judge Daly – Wulff Land anticlinal zone; (17) Region of distinct magnetic character; (18) Kane Basin Magnetic anomaly; (19) Prominent steep gravity gradient; (20) Earthquake epicentre zone.

left lateral sense (Fig. 1; DAWES & KERR 1982).

While there have been some imaginative solutions to this, now classic, conflict in plate tectonics (SRIVASTAVA 1985, HIGGINS & SOPER 1989, OKULITCH et al. 1990), the fact remains that, right or wrong: 1) most earth scientists believe that a Cenozoic plate boundary probably exists in Nares Strait; 2) there is compelling geophysical evidence to indicate that young ocean crust (and or mantle-derived serpentinite) exists beneath the Labrador Sea and Baffin Bay; 3) local geological and geophysical evidence from the strait has not, until recently, presented a convincing case for a plate boundary in Nares Strait; and 4) published data are insufficient from most offshore areas to provide a completely satisfactory resolution to the Nares Strait controversy.

More than 20 years have now passed since the 1982 Nares Strait volume of DAWES & KERR (1982). In that time there have been new geological maps published from almost all onshore areas, industry reflection profiles have been acquired and made available to the science community, and potential field, bathymetry, refraction and some reflection seismic surveys have been acquired by the Canadian, Danish-Greenland and other government geological surveys. Many of the papers presented elsewhere in this volume provide the most recent data bearing on the Nares Strait controversy; most notably the results of the cooperative marine and onshore research effort in 2001 involving the German Geological Survey (BGR), the Geological Survey of Canada (GSC) and the Danish Lithosphere Centre (DLC) aboard CCGS "Louis S. St-Laurent" (JACKSON & SHIPBOARD PARTY 2003, REID et al. 2006). In light of all the new data, the present contribution attempts a critical re-assessment of the published geological-geophysical markers, most notably the twenty markers summarized by DAWES & KERR (1982). Eight additional constraints are presented for linear markers that have been documented since 1982.

Central to the present paper is the recent recognition of the extent of strike-slip displacements on northeast Ellesmere Island and the re-evaluation of the net accumulated motion on the Judge Daly Fault Zone of MAYR & DE VRIES (1982) and adjacent fault strands, here referred to collectively as the Judge Daly Fault System (JDFS). Drawing on a combination of features that intersect the JDFS, including Paleozoic folds and the displaced depositional limit of Lower Ordovician evaporites, HARRISON (in press) concluded that this fault system has experienced 69 ±5 km of sinistral strike slip motion. Radiometric and biostratigraphic evidence of ESTRADA et al. (in press) and LEE et al. (in press) indicates that this motion has occurred since the beginning of the late Paleocene and was therefore concurrent with sea floor spreading in Labrador Sea and Baffin Bay (CHALMERS & PULVERTAFT 2001).

Aeromagnetic data presented by DAMASKE & OAKEY (2006) reveal a clear step in magnetic field strength that coincides with the Judge Daly Fault Zone. The aeromagnetic step presents a clear contrast between magnetically more susceptible terrane to the southeast (presumably derived from basement rocks located beneath Paleozoic cover), and a magnetically less responsive succession located to the northwest (Fig. 2). The aeromagnetic step, traced from Carl Ritter Bay to Robeson Channel, coincides with a northeasterly-trending belt of discontinuous high amplitude magnetic anomalies related to the preservation of magnetically susceptible upper Paleocene volcaniclastic sediments (HOOD et al. 1985, DAMASKE & OAKEY 2006). Coinciding with the location of the aeromagnetic step, the high amplitude anomalies and associated volcaniclastic sediments are found on both sides of the JDFS in the north and between the Cape Back and Mount Ross faults in the Carl Ritter Bay area. The extent of the magnetic step and related high amplitude anomaly is unresolved south of Carl Ritter Bay but to the north, on the maps of HOOD et al. (1985) and MILES (2000a), is traced into the Lincoln Sea beyond 83 °N.

Significance of the JDFS is also indicated by colour alteration indices (CAI values) measured on conodonts collected from mostly Ordovician strata exposed on either side of the fault trace (HARRISON, NOWLAN et al. in press). Samples collected southeast of the fault possess CAI values that range from about 1 to not more than 2. In contrast CAI values of samples collected northwest of the fault are all 4 to 5+. South of Cape de Fosse the contrast in thermal maturity conditions coincides with the Mount Ross and Cape Back faults north of Carl Ritter Bay, and coincides with the Rawlings Bay Thrust as far south as Cape Lawrence at 80°25' N.

There are several important implications. First, the JDFS should be re-considered as a viable candidate for the plate boundary between Arctic Canada and Greenland during sea floor spreading in the Labrador Sea and Baffin Bay. Assuming this to be true implies that part of the Greenland plate is exposed on Ellesmere Island east of the JDFS where this fault system is located onshore on Judge Daly Promontory. The other implication is that the bedrock of Kennedy Channel, most of which lies entirely east of the JDFS, is merely the submerged western portion of the Greenland plate produced by differential glaciofluvial excavation of underlying Cenozoic sediments and Silurian flysch between the high standing carbonate terrains of eastern Ellesmere Island and Washington Land.

Southwards, beyond Cape Lawrence, there are two theoretical positions for the potential plate boundary: 1) it continues south southwesterly across Kane Basin and Smith Sound and into northern Baffin Bay following the hypothetical trace of the Wegener Fault, or 2) it follows the curvature of the Cenozoic orogen to the west where the JDFS may be kinematically linked with either the thin-skinned Parrish Glacier Thrust or with a potential westerly-striking (basement-rooted?) reverse fault that projects into Princess Marie Bay from western Kane Basin on the industry reflection profiles of RENDELL & CRAIG (1974). The evidence presented and interpreted in this paper will show that the only possible plate boundary in the Nares Strait region is one that follows the Judge Daly Fault System and the thrust-bounding curvature of the orogen onto central Ellesmere Island (Fig. 2). The general implications of this discovery for resolving the lingering Nares Strait controversy are summarized in the paper's concluding statements.

METHODS

The geological and geophysical markers described and tabulated by DAWES & KERR (1982) are reproduced in Figure 1 and are listed in Table 1. DAWES & KERR'S (1982) markers were

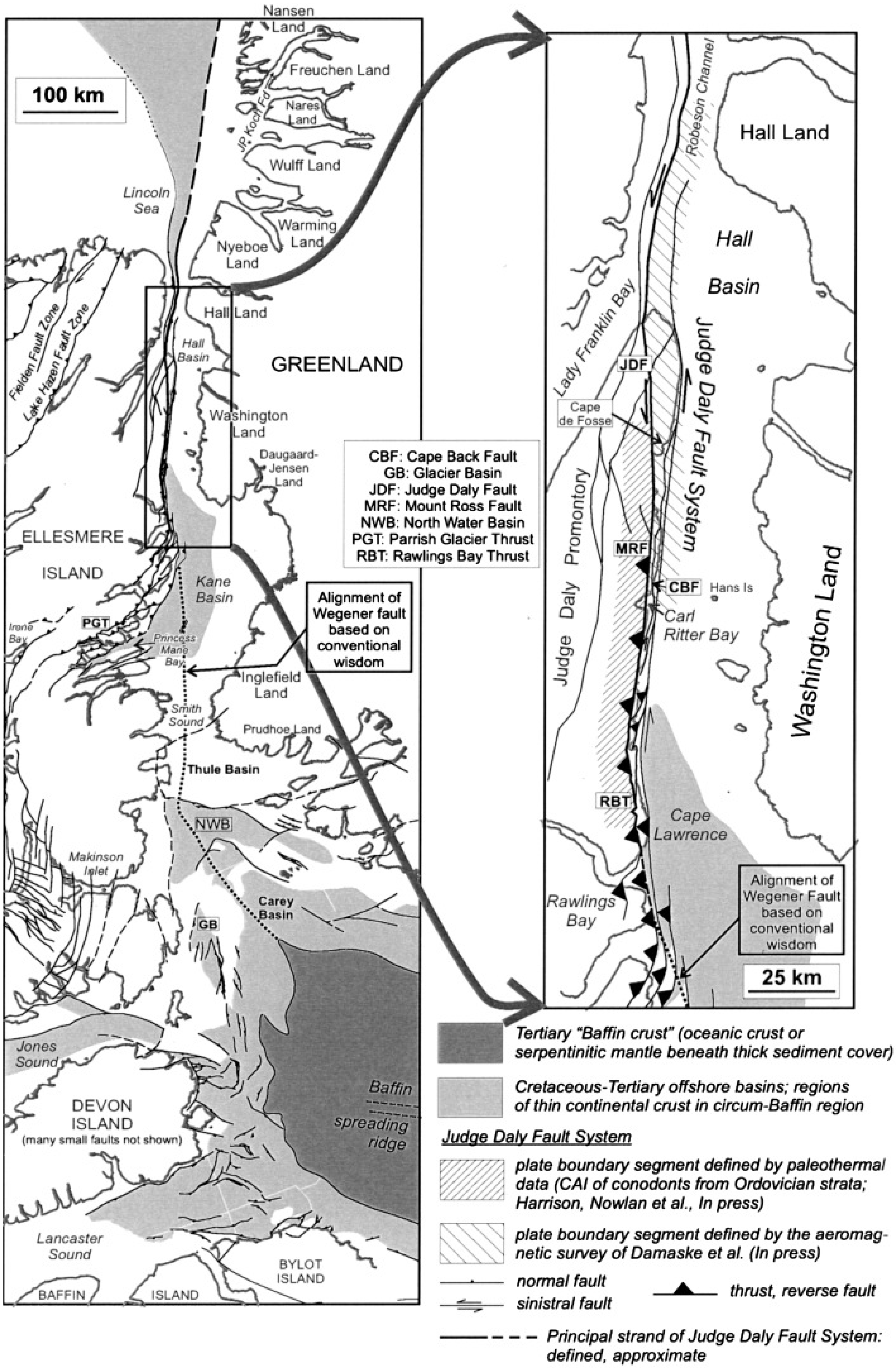


Fig. 2: The Judge Daly Fault System: the only candidate for a plate boundary between Greenland and Arctic North America in the Nares Strait region based on aeromagnetic surveys and thermal maturity data.

evaluated by them based on an assumption of zero offset of bedrock underlying Nares Strait. Their values are estimates of the maximum amount of sinistral strike slip that might be accommodated, rounded to the nearest 25 km, given uncertainties such as the width of Nares Strait and the limitations provided by lack of outcrop to constrain the position of onshore geological markers.

The present re-assessment has been accomplished by identifying specific onshore and offshore localities that constrain the geographic position of each marker. Many areas are covered by either water or glacier ice and in such circumstances a degree of semi-quantifiable uncertainty is introduced as to the extent of potential lateral offset. The degree of uncertainty is especially large in the strait itself where outcrop is limited to a small number of bedrock islands, and geophysical

Marker	Description	Allowable offset (DAWES & KERR 1982)	Offset estimate (this study)
1	Paleoproterozoic marble belt	0-100 km LL	20 ±20 km RL
2	Mid-Proterozoic Thule Basin (north edge)	0-75 km LL	0 ±15 km
3	Bache Peninsula arch	0-100 km LL	0 to 100 km LL
4	Zero-edge of Thule Supergroup	0-100 km LL	invalid marker
5*	Cambro-Ordovician platform margin	0-100 km LL	80 ±35 km LL
6*	Lower Ordovician evaporite belt**	0-50 km LL	69 ±5 km LL
7*	Middle Ordovician evaporite belt**	0-50 km LL	80 ±35 km LL
8*	Lower Cambrian hinge line**	0-150 km LL	75 ±25 km LL
9*	Ordovician-Silurian platform margin	0-100 km LL	80 ±35 km LL
10*	Lower Silurian facies contact	0-25 km LL	65 ±35-65 km LL
11*	Silurian carbonate buildup belt**	0-100 km LL	25 ±25 km LL
12*	Limit of Ellesmere-Greenland fold belt (Devonian)	0-100 km LL	45 ±25 km LL
13*	Amphibolite facies metasediments	0-100 km LL	not re-assessed
14*	Lake Hazen-Harder Fjord fracture line	0-150 km LL	invalid marker
15*	Judge Daly-Nyeboe Land fracture line	0-100 km LL	invalide marker
16*	Archer Fiord-Wulff Land anticlinal zone	0-100 km LL	60 ±10 km LL
17*	Region of distinct magnetic character	not estimated	not estimated
18	Kane Basin magnetic anomaly	not estimated	not estimated
19*	Judge Daly-Nyeboe Land 70 mgal contour **	not estimated	90 ±? Km LL
20	Seismicity belt	not estimated	not reassessed
New marker	Description (this study)		
21*	Upper limit of thermally mature sediments	not considered	135 ±45 km LL
22*	Judge Daly-Nyeboe Land 90 mgal contour	not estimated	60 ± km LL
23	Western transform margin of Baffin crust	not considered	220 ±90 km LL
24	Southeast limit of Paleozoic cover	not considered	0 ±10 km
25	Southeast limit of Paleozoic outliers	not considered	0 ±20 km
26	Archean-Paleoproterozoic contact	not considered	0 to 200 km LL
27	Trough axis of northern Thule Basin	not considered	0 to 15 km RL
28	Kap Leiper dyke (DAMASKE & OAKEY 2006)	not considered	0 ±1? km

Tab. 1: Geological-geophysical markers providing for estimates of strike slip motions in and adjacent to Nares Strait. The present study has been developed from the assumption that sinistral slip is constrained to the Judge Daly Fault System and, in the south, the hypothetical surface trace of the Wegener Fault based on conventional wisdom. Values from DAWES & KERR (1982) were based on markers that cross Nares Strait. LL = left lateral; RL = right lateral; * = Markers identified with an asterisk are located within the Paleozoic orogen of Ellesmere Island and North Greenland. ** = Marker has been re-defined in this study.

profiles are few. Unlike the somewhat fixist approach of DAWES & KERR (1982) the present re-assessment proceeds from the assumption that at least 69 km of sinistral strike slip are permissible for the onshore portion of the JDFS on northeastern Ellesmere Island. Considered here is the possibility that more or less than this amount of displacement is allowable, with estimates rounded to the nearest 5 km (see results in Tab. 1). For example, where a geological marker is extrapolated into the offshore it is assumed that curvature of the marker into exact alignment with the strait is a less viable solution than one which permits a degree of sinistral offset.

Markers that cross the strait between northern Kane Basin and the Lincoln Sea lie partly or wholly within the belt of mid-Paleozoic (Ellesmerian) and Cenozoic (Eurekan) folding and thrust faulting. In contrast, markers that cross the strait between central Kane Basin and northern Baffin Bay lie entirely beyond the Phanerozoic orogen. These areas are generally underlain by the Cambrian to Silurian Arctic Platform of Canada and Greenland, mid-Proterozoic strata of the Thule and Borden basins, and the underlying Archean to Paleoproterozoic supracrustal granulite and granitoid "basement" complex. It can be expected that the behaviour of strike-slip faults that penetrate stable continental lithosphere will contrast with the behaviour of similar faults that extend into an active orogenic belt. Thrust faults, for example, that delaminate the shallow part of the crust may override and hide older strike slip faults that are rooted in basement. Strike-slip motions can also occur during thrusting and in such circumstances the active fault planes may form a braided array of wrench tectonic slip surfaces interwoven with structures more specifically related to thrusting. For this reason, the present re-assessment is divided into two parts: 1) consideration of markers located partly or wholly within the Phanerozoic orogen where sinistral offset has been demonstrated along the JDFS; and 2) examination of markers located entirely beyond the orogen where the location of the Wegener Fault, not yet proven to exist, is based on conventional wisdom that places the plate boundary in southern Nares Strait more or less half way between Canada and Greenland.

GEOLOGICAL AND GEOPHYSICAL MARKERS THAT CROSS THE JUDGE DALY FAULT ZONE

Cambro-Ordovician platform margin (marker 5 of DAWES & KERR 1982).

Mapped units of the Cambro-Ordovician platform margin succession on the Canadian side of Nares Strait include the Scoresby Bay, Cass Fjord, Cape Clay, Ninnis Glacier, Bulleys Lump, Thumb Mountain, and Irene Bay formations (TRETTIN 1994, DEWING et al. in press). All these units grade northwestward to basin facies Hazen Formation within which there are five mapped units (TRETTIN 1994, DEWING et al. in press). On the Greenland side of Robeson Channel the Cambro-Ordovician platform margin succession includes the Brønlund Fjord, Ryder Gletscher and Morris Bugt groups on the map of HENRIKSEN (1989). The slope and basinal side of the facies change is represented by the Aftenstjernesø and Kap Stanton formations, and Amundsen Land Group (HENRIKSEN 1989, INESON et al. 1994).

The facies change is closely defined on north-eastern Judge Daly Promontory, particularly in the Middle Cambrian – Lower Ordovician interval (Fig. 3). However, the exact position of the Lower Cambrian facies change tends to be obscured by younger prograded components of the outer shelf. Similarly, the Middle through Upper Ordovician tends to be eroded above older basinal facies on the immediate basin-ward side of the facies change. For the remainder of the package, it is possible to identify the precise location where the facies

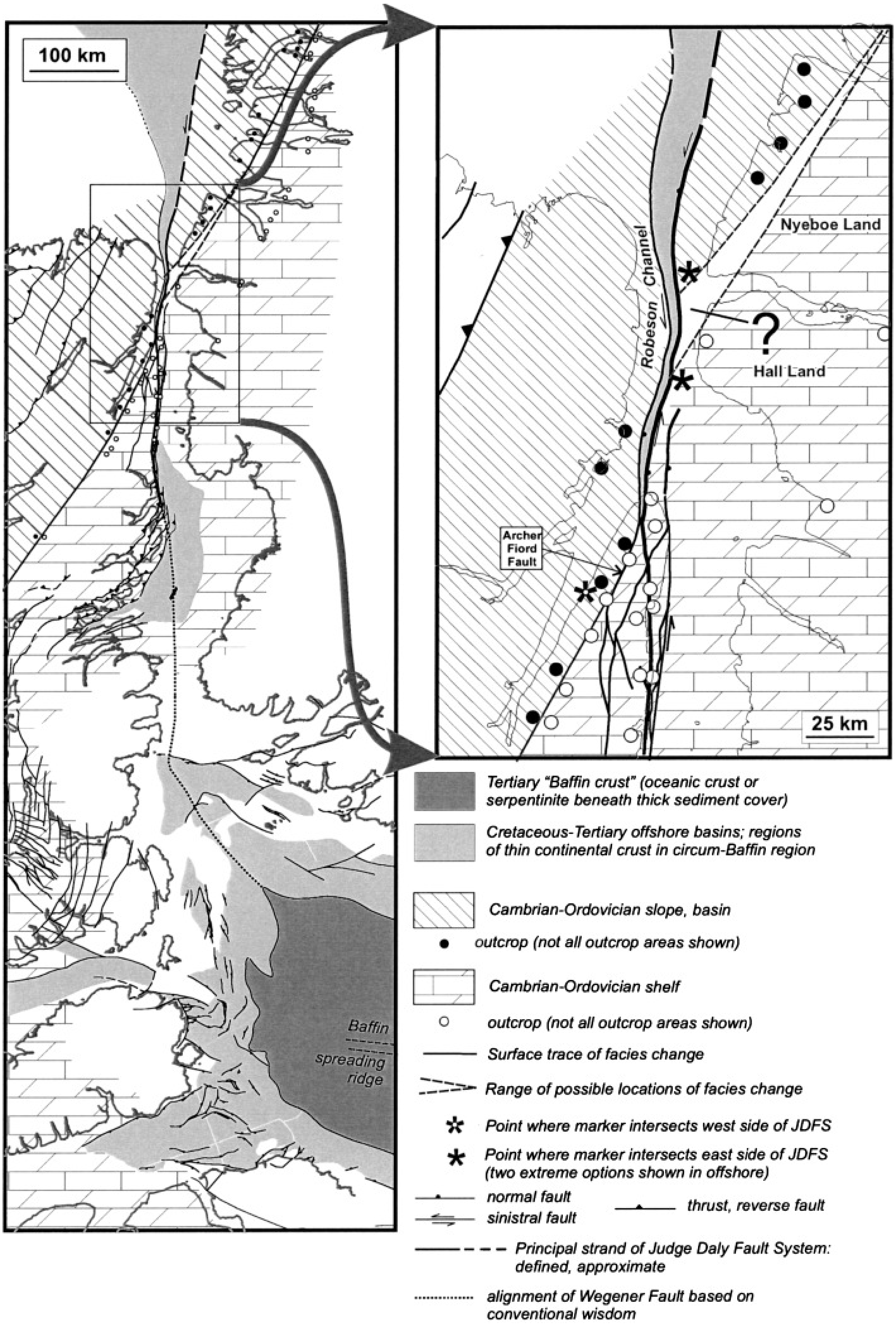


Fig. 3: Position of the Cambro-Ordovician and Ordovician-Silurian platform margin (markers 5 and 9 of DAWES & KERR 1982). Estimated sinistral displacement on the Judge Daly Fault System is 80 ±35 km.

front intersects the Archer Fiord Fault, the latter a prominent strand of the JDFS on northeastern-most Judge Daly Promontory (Fig. 3). From this point to the northeastern limit of outcrop on the promontory, the facies change is replaced by the surface trace of the Archer Fiord Fault, and basinal facies strata to the northwest are in continuous tectonic contact with mid- to outer shelf facies carbonates on the opposite side of the fault.

The Archer Fiord Fault carries a footwall syncline in Paleocene strata and in the hanging wall features cleaved southeasterly-facing overturned folds in the Hazen Formation that intersect the fault trace in a manner consistent with sinistral strike slip motion. The exposed fault surface dips to the northwest and has measured slickenline slip surfaces that are consistent with both thrust and sinistral slip motions (Piepjohn et al. in press). Based on these observation the maximum exposed left lateral motion on the Archer Fiord Fault strand is approximately 40 km.

The constraints are not as good on the Greenland side of Nares Strait. The northernmost exposures of the platform facies occur on coastal northern Hall Land. The westernmost and southernmost exposures of the basin facies lie on coastal north central Nyeboe Land. Intervening areas are covered by Silurian submarine fan deposits of the Peary Land Group which are found above both basin facies and shelf facies strata, and for this reason the unit is not useful for defining the position of the Cambro-Ordovician shelf margin. Extrapolation of the line of facies change from its narrowly defined position in Freuchen Land and Nares Land of eastern North Greenland (see Fig. 2 for location of some place names) through to possible locations on the sea floor trace of the JDFS in Robeson Channel yields a total sinistral offset of between 45 and 75 km. This assumes that there is no southerly deflection of the facies change in Robeson Channel. The combined total minimum and maximum potential sinistral displacements on the Archer Fiord Fault and principal strand of the Judge Daly Fault System is 45 and 115 km (or 80 ±35 km).

Lower Ordovician evaporite belt (re-defined marker 6 of Dawes & Kerr 1982)

Gypsum-anhydrite units in the Lower Ordovician in Washington Land are represented in the Poulsen Cliff and upper Nygaard Bay formations (Peel & Christie 1982, Higgins et al. 1991). The northernmost exposures of these formations are on the walls of Petermann Gletscher (Fig. 4; Jepsen et al. 1983). Farther east the correlative Lower Ordovician is represented by intertidal and subtidal algal laminates and pebble conglomerate of the Warming Land Formation (Higgins et al. 1991). The location of the contact between these belts, poorly constrained due to widespread cover by younger strata, is inferred to lie in the subsurface of Hall Land (Fig. 4).

Evaporitic rocks on north-eastern Ellesmere Island are represented by the Baumann Fiord Formation which includes a medial limestone unit sandwiched between two gypsum-anhydrite members. To the northwest, the Baumann Fiord Formation grades to correlative shelf carbonates of the Ninnis Glacier Formation. The northern limit of this facies change is tightly constrained on north-eastern Judge Daly Promontory (Fig. 4; Harrison, Mayr et al. in press), a line that provides for 53 km of left lateral strike slip motion on the principal strand of the JDFS which in this area includes the Judge Daly Fault Zone of Mayr & de Vries (1982) and the Mount Ross Fault to the southwest (Fig. 2; Harrison in press). To this is added an estimated 16 km of sinistral slip on the array of small mapped faults within the belt of Ninnis Glacier facies to the northwest (Harrison in press). Assuming that no significant slip is present in bedrock beneath the adjacent portion of Kennedy Channel (as indicated by the aeromagnetic and thermal maturity data of Damaske & Oakey 2006 and Harrison, Nowlan et al. in press, respectively) the locally assessed total sinistral displacement on all these faults, collectively referred to as the JDFS, is 69 ±5 km.

Middle Ordovician evaporite belt (re-defined marker 7 of Dawes & Kerr 1982)

Middle Ordovician evaporites occur in the lower part of the Bay Fiord Formation on Ellesmere Island and in the lower Cape Webster Formation in North Greenland (Fig.1; Peel & Christie 1982). Mapping on Judge Daly Promontory indicates that the gypsum-anhydrite facies is limited to the outcrop belt lying south of the Agassiz Ice Cap and along the coast adjacent to Kennedy Channel. However, the Ordovician shelf edge facies of the Bay Fiord Formation, adjacent to the Bulleys Lump carbonate buildup, contains little or no gypsum-anhydrite. Thickness of evaporites are also controlled by local structure, often thickening under anticlines and thinning to zero under adjacent synclines. In general, outcrop is inadequate to permit a reliable matching of the evaporitic facies across Nares Strait.

The maximum extent of the Cape Webster and Bay Fiord formations, regardless of the incidence of evaporites, coincides with that described for marker 5 (Fig. 3) which allows a sinistral offset of 80 ±35 km.

Cambro-Ordovician hinge line (re-defined marker 8 of Dawes & Kerr 1982)

Although the geographic position of enhanced thickening of Cambrian and Ordovician strata is drawn by Dawes & Kerr (1982) as a discrete thin line (Fig. 1), in reality this hinge would be expected to migrate basin-ward as younger deposits prograde across older sequences. To focus the discussion on a particular part of the Cambro-Ordovician succession, the present re-assessment considers only the thickness and facies associated with Lower Cambrian progradational clastic units of the report area (Fig. 5). Three facies belts are identified:
1) inner shelf quartz sandstones of the Dallas Bugt and Humboldt formations, 25 m up to several hundred metres thick in Inglefield Land, Washington Land (base covered) and on southeast Ellesmere Island (Peel et al. 1982, de Freitas et al. 1997, de Freitas et al. in press);
2) outer shelf facies (250-500 m thick) assigned to the Buen Formation across southern North Greenland (Higgins et al. 1991), and assigned to the source-proximal Ellesmere Group on northeast Ellesmere Island (Dewing et al. in press), and;
3) slope and trough facies clastics (500-2000 m thick) of the Polkorridoren Group in northernmost Greenland (Higgins et al. 1991), and the Grant Land Formation and distal Ellesmere

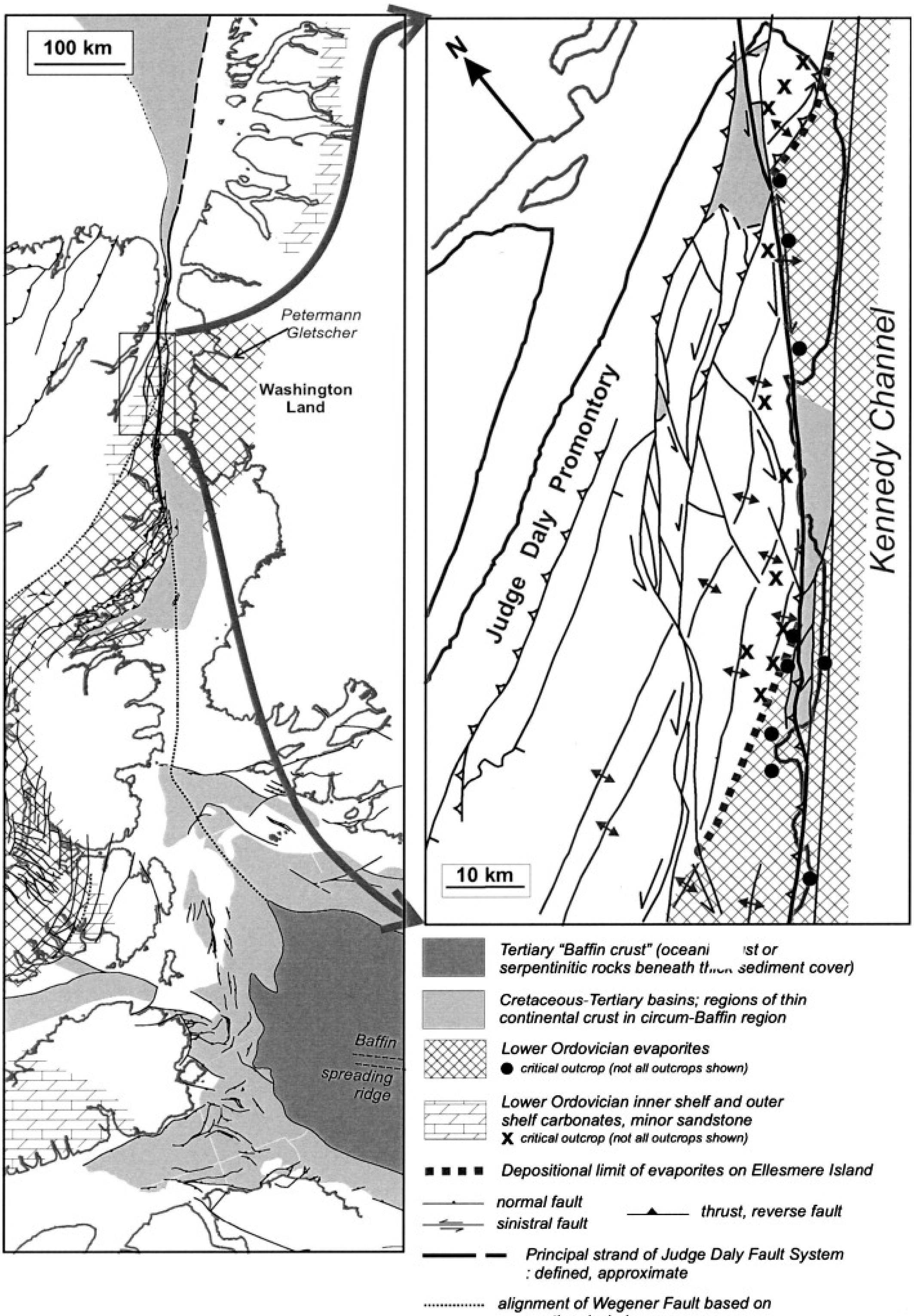


Fig. 4: Northern extent of the Lower Ordovician evaporite belt (re-defined marker 6 of DAWES & KERR 1982). Indicated total left lateral displacement for the Judge Daly Fault System (JDFS) is 69 ±5 km. This motion includes 53 km of sinistral slip on the principal strand of the JDFS, and 16 km of displacement on the array of smaller faults located to the northwest (HARRISON in press).

Group on northern Ellesmere Island (TRETTIN 1994, DEWING et al. in press).

The position of the inner to outer shelf transition is unconstrained (Fig. 5) and therefore cannot be considered a suitable definition of a depositional hinge. The boundary between outer shelf and slope facies Lower Cambrian is better defined on both sides of Nares Strait. Mapping on the Canadian side indicates that the facies change from Dallas Bugt Formation to proximal Ellesmere Group bends northwards as it approaches the Judge Daly Fault System from the west. The northernmost outcrops of Dallas Bugt Formation west of the fault lie between Carl Ritter Bay and Cape de Fosse. East of the JDFS, the Dallas Bugt Formation has been mapped to within five kilometres of the north-eastern point of Judge Daly Promontory.

Position of the shelf-slope break is less well constrained in Greenland as outcrop of Polkorridoren Group is unknown west of coastal north central Nyeboe land (Fig. 5). A reasonable

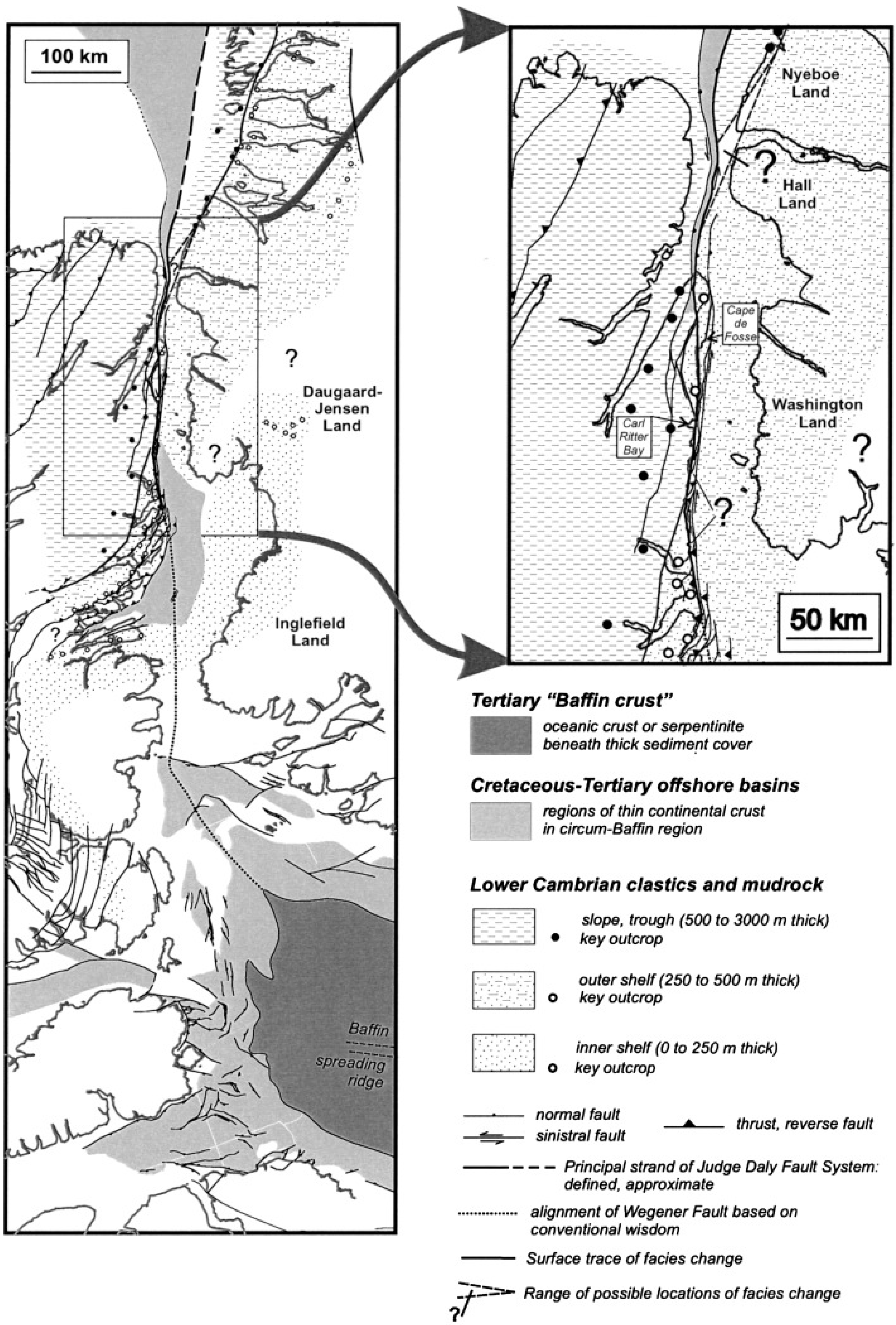


Fig. 5: Lower Cambrian depositional facies (modified definition of marker 8 of DAWES & KERR 1982). Estimated offset of the outer shelf to upper slope hinge line is 60 ±30 km left lateral.

estimate of sinistral displacement, based on this re-defined marker, is 75 ±25 km.

Ordovician-Silurian platform margin (marker 9 of DAWES & KERR 1982)

On northeast Ellesmere Island, the Ordovician-Silurian platform margin coincides with the facies change from shelf carbonate Allen Bay Formation (Ashgill to Llandovery age) to part of the upper member of the basinal Hazen Formation (TRETTIN 1994, DEWING et al. in press). In North Greenland this line coincides with the transition from shelf facies Aleqatsiaq Fjord Formation to basinal Amundsen Land Group (HIGGINS et al. 1991). The position of this line is similar to that of the Middle Cambrian through Lower Ordovician (marker 5, described previously) though less well constrained due to widespread erosion of Upper Ordovician and Lower Silurian strata on the basin-ward side of the facies front. Magnitude of strike slip offset on the JDFS on Judge Daly Promontory and

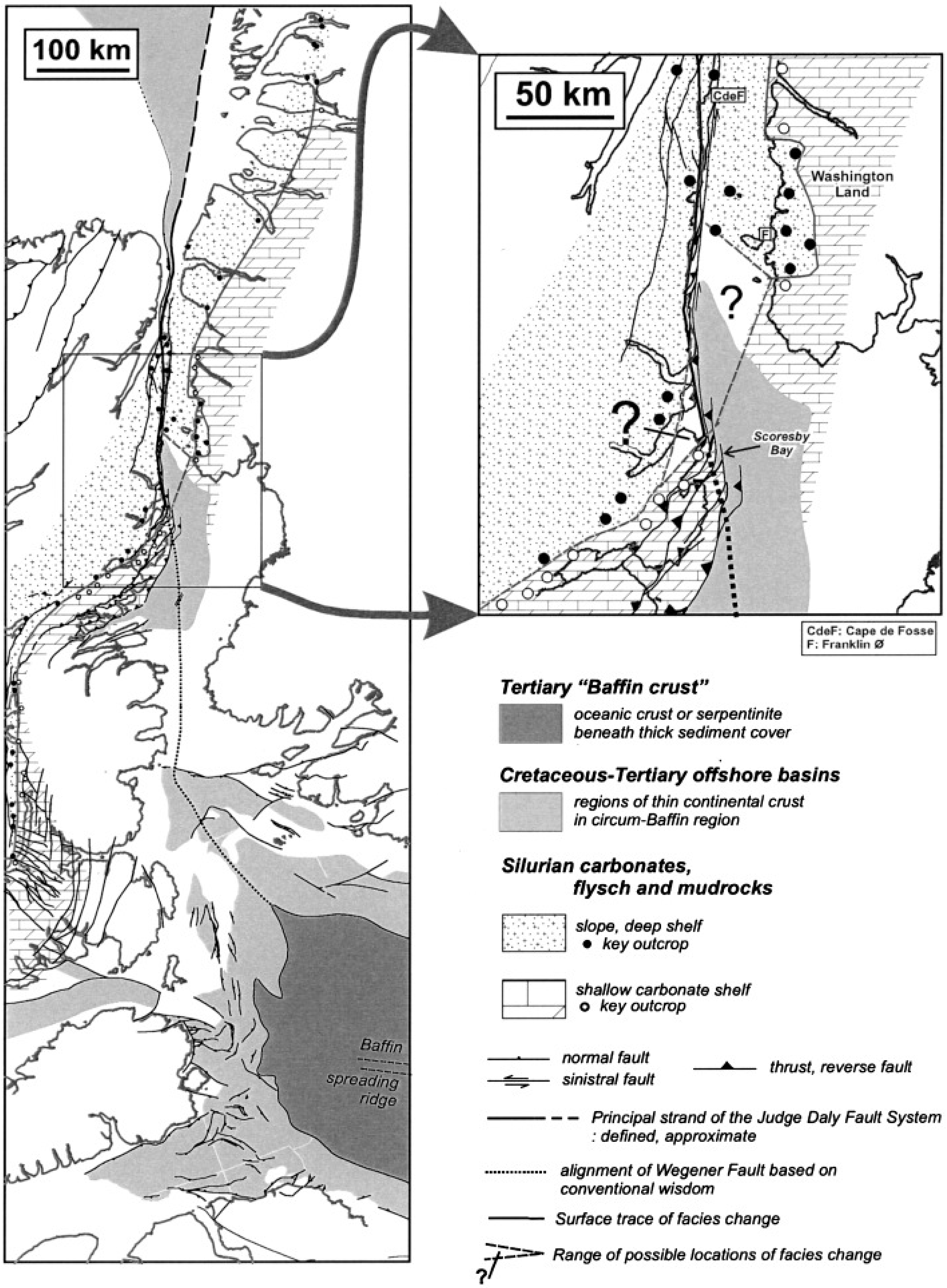


Fig. 6: Location of the line of Lower Silurian facies change (marker 10 of DAWES & KERR 1982). Estimate of sinistral offset of the line of facies change is 65+35-65 km.

in Robeson Channel is assumed to be the same as for marker 5 (80 ±35 km LL; Fig. 3).

Lower Silurian facies contact (marker 10 of DAWES & KERR 1982)

Graptolitic shales, deep water carbonates and submarine fan deposits extend across the Ordovician outer shelf throughout much of the Nares Strait region on both sides of the channel (HIGGINS et al. 1991, DE FREITAS et al. 1997, DEWING et al. in press, HARRISON in press). On Ellesmere Island the deep shelf graptolitic sediments are assigned to the Cape Phillips and Devon Island formations (DE FREITAS et al. 1997). The submarine fan deposits are included in the Danish River Formation (DEWING et al. in press). The southeastward transition to shallow marine carbonates is abrupt, and on the shelf-ward side of the facies front the succession includes Allen Bay, Cape Storm, Douro and Goose Fiord formations (DEWING et al. in press).

In Washington Land and to the northeast the deep shelf and submarine fan facies occur in the Peary Land Group. Correlative shallow marine carbonates are assigned to the Washington Land Group (DAWES et al. 2000, HENRIKSEN 1989).

The line of facies change is sinuous, especially in northern Washington Land, and the location of the facies front is unconstrained in Kennedy Channel south of Franklin Ø (Fig. 6). On northeast Ellesmere Island, Silurian strata south of Cape de Fosse are not exposed east of the Judge Daly Fault System. Likewise the location of the facies front cannot be precisely located immediately west of the fault and north of Scoresby Bay. For these reasons, this is a relatively poor marker. Assuming that the facies front has the same trend on both sides of the fault, a reasonable estimate of sinistral offset is 65 km. However, the range of left hand offsets could fall anywhere between 0 and 100 km (65 + 35-65 km).

Silurian carbonate buildup belt (re-defined marker 11 of Dawes & Kerr 1982)

Carbonate buildups and patch reefs occur throughout part of the Silurian deep shelf succession adjacent to the Silurian carbonate bank across much of North Greenland, on the islands in Kennedy Channel, in Washington Land (Dawes 1987, Henriksen 1989, Jepsen et al. 1983) and in parts of central Ellesmere Island (de Freitas et al. in press, Thorsteinsson et al. 1991, Figs 1, 7). The extent and distribution of carbonate buildups in Kennedy Channel and Hall Basin is also provided by the interpretation of new BGR-GSC reflection profiles acquired between Franklin Ø and Hall Basin (Harrison in press, Jackson et al. 2006). In general, the northwestern limit of the reef tract is poorly constrained

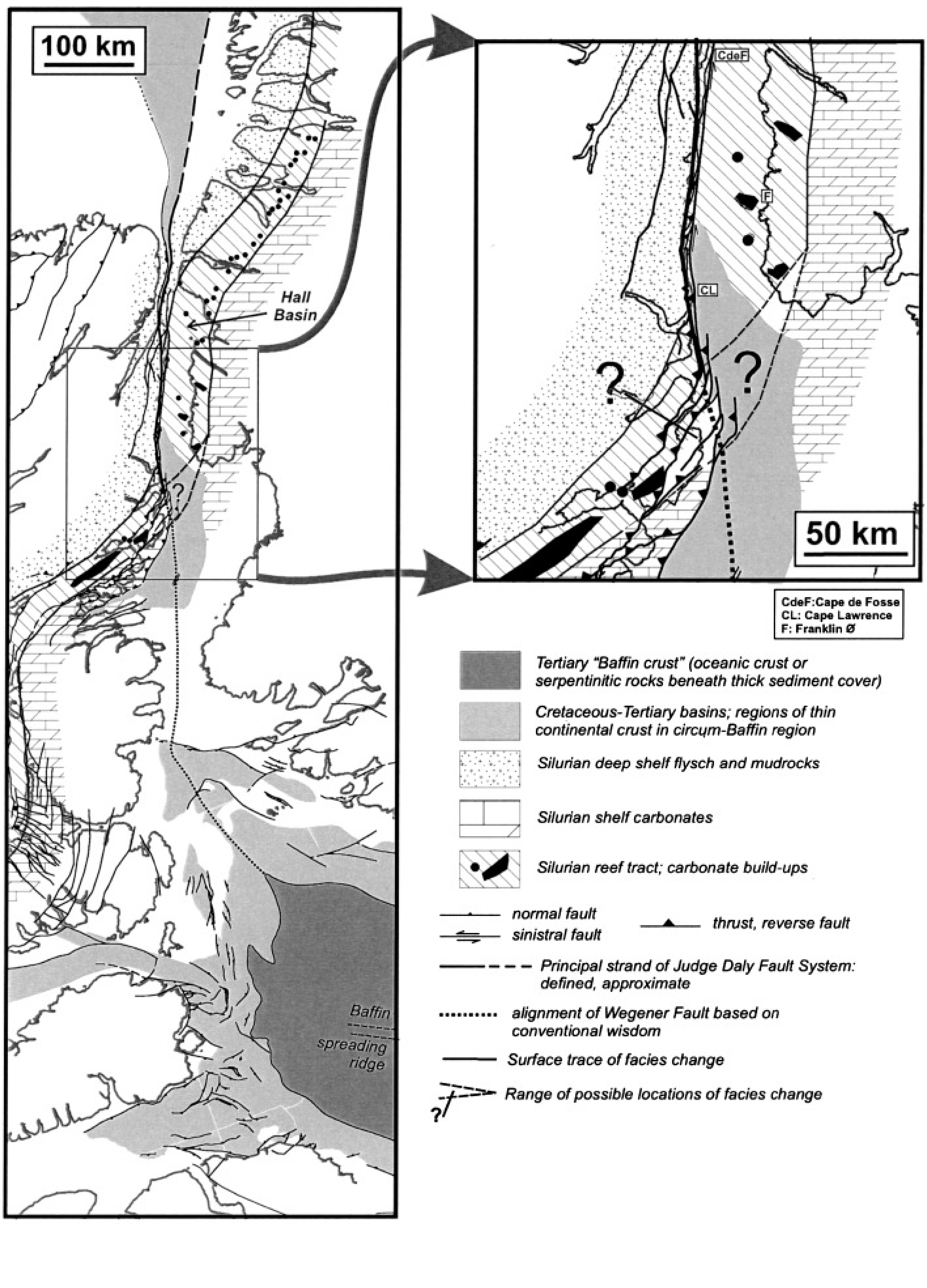


Fig. 7: Extent of the Silurian reef tract and carbonate build-up belt (marker 11 of Dawes & Kerr 1982). Estimate of sinistral offset of the southeastern limit of the reef tract is 25 ±25 km.

onshore on Ellesmere Island near Kennedy Channel and may coincide with coastal strands of the JDFS between Cape Lawrence and Cape de Fosse. Based only on the south-eastern limit of carbonate buildups (Fig. 7), the estimate of sinistral offset on the fault zone south of Cape Lawrence ranges to a maximum of 50 km with the potential for no displacement at all (25 ±25 km).

Southern limit of the Ellesmere-Greenland fold belt (marker 12 of DAWES & KERR 1982)

For the purposes of this account the Ellesmere-Greenland fold belt is considered the compressive deformation produced by the Ellesmerian Orogeny which occurred near the end of the Devonian (Fig. 8). The hinterland deformation and related crustal load on northern Ellesmere Island and in North Greenland has produced thermally overmature ranging to lower amphibolite grade regional metamorphism, folding at various scales, and intense axial planar cleavage fabrics in strata that

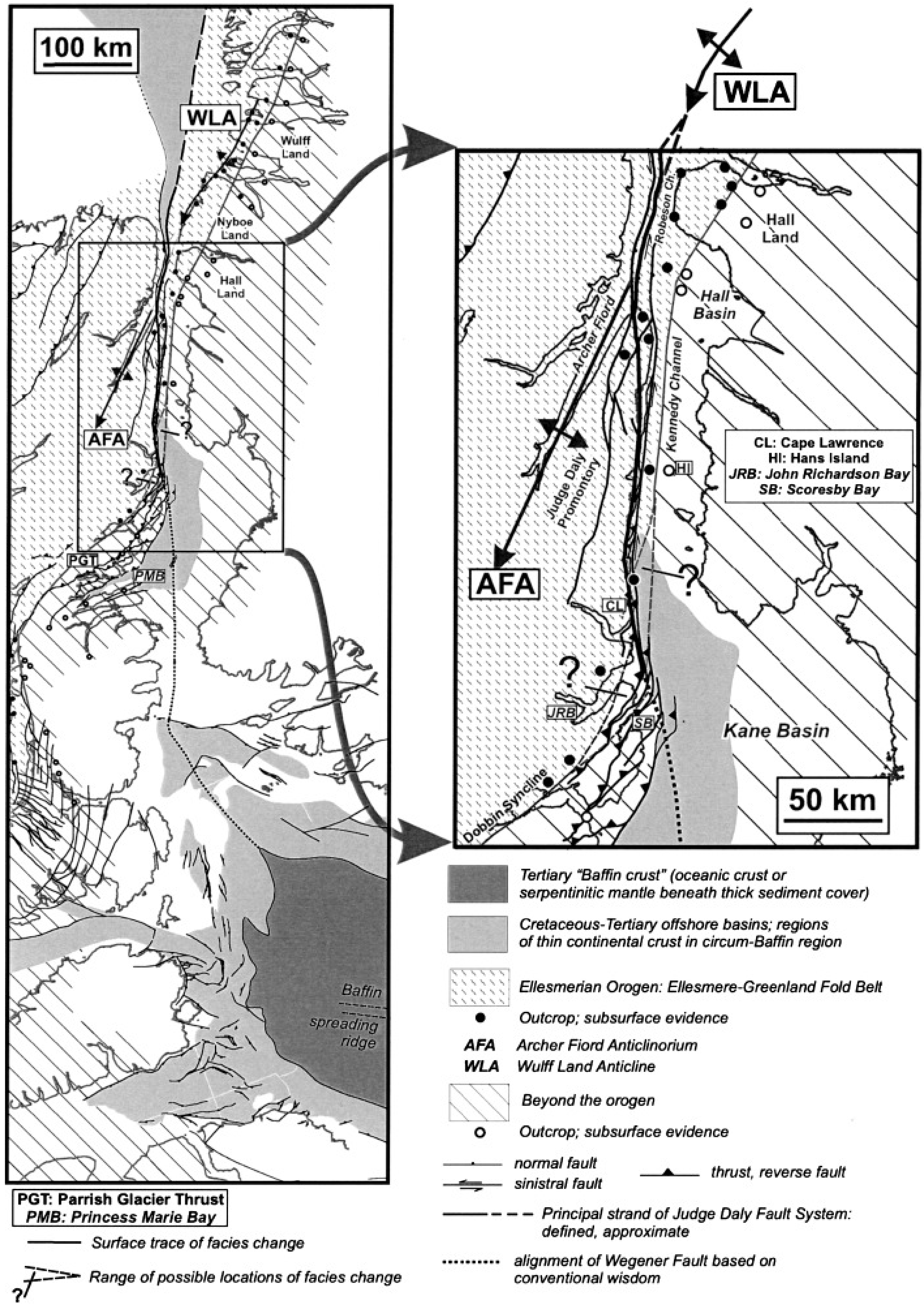


Fig. 8: Southeastern limit of the end-Devonian Ellesmerian Orogen (marker 12 of DAWES & KERR 1982) and the position of the Archer Fiord Anticlinorium – Wulff Land Anticline (renamed marker 16 of DAWES & KERR 1982). Respective sinistral offsets of these markers is 45 ±25 and 60 ±10 km.

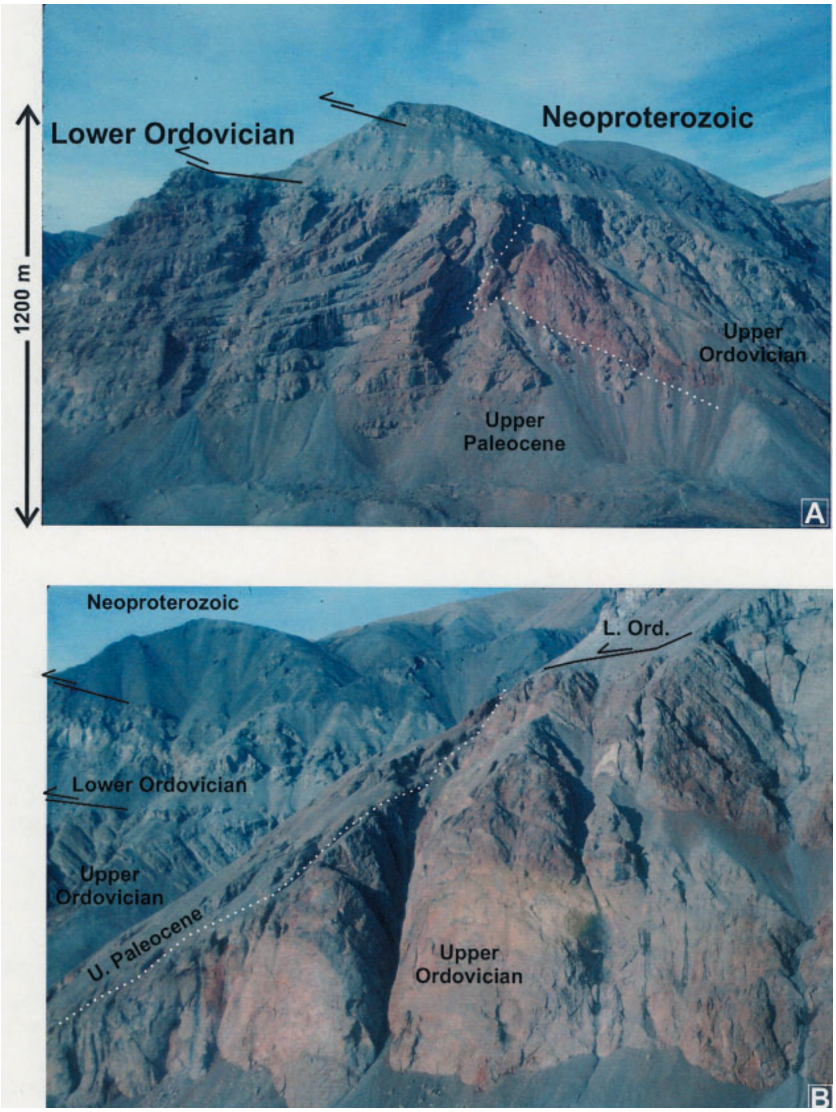


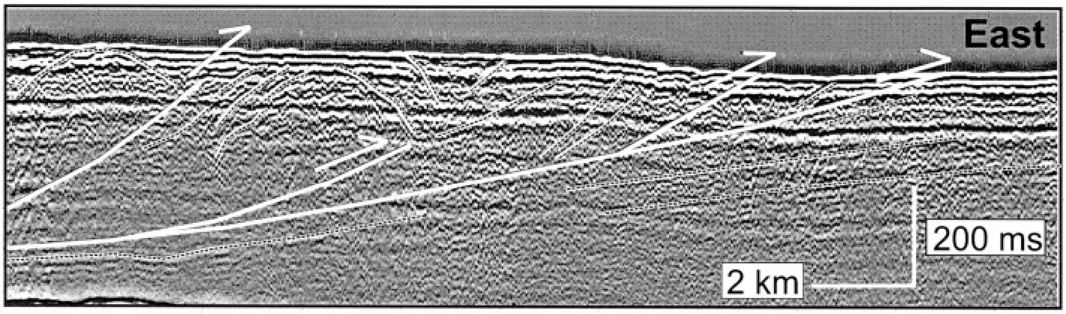
Fig. 9: Two views of the sub-Paleocene angular unconformity (white dotted lines) near Cape Lawrence. The unconformity surface (dotted white line) dips to the southeast and overlies near-vertically dipping Ordovician Thumb Mountain Formation. Thrust sheets in both photos carry Lower Ordovician Baumann Fiord Formation (evaporites) and Neoproterozoic Kennedy Channel Formation (dark coloured mudrocks and carbonates). Syntectonic conglomerate and breccia are dominant in the upper Paleocene section.

range from Neoproterozoic to Silurian in age (Surlyk 1991, Trettin 1994, Harrison in press). The foreland areas of the orogen range from immature to overmature paleothermal conditions (Harrison, Nowlan et al. in press), weak ranging to moderate cleavage development in Upper Devonian and older strata that are folded, mostly above the base of the Lower Ordovician (Baumann Fiord) evaporites on north-eastern Ellesmere Island (Harrison in press). The evaporite detachment has apparently been less active in North Greenland. In that area, and starting in the hanging wall of the Parrish Glacier thrust north of Princess Marie Bay (Fig. 8), there is the involvement of strata lying as deep as the Neoproterozoic (Soper & Higgins 1990).

West of John Richardson Bay the southern limit of Ellesmerian deformation lies along the axial trace of Dobbin Syncline (Fig. 8; Harrison in press, de Freitas et al. in press). Ordovician to Upper Silurian strata south of the syncline have no obvious cleavage fabrics and there are few regional scale folds. Position of the limit of deformation projects onto the offshore portion of the Judge Daly Fault System eastward from the north shore of Scoresby Bay (Fig. 8). However, extent of Ellesmerian deformation west of the fault is uncertain as far north as the north shore of John Richardson Bay.

East of the JDFS the southern limit of Ellesmerian deformation is indicated by a profound angular unconformity near Cape Lawrence that separates upper Paleocene sediments from vertical to overturned Upper Ordovician strata (Fig. 9). Farther north, the eastern limit of Ellesmerian deformation is recognized on BGR-GSC seismic profile 8-2001 located west of Hans Island (Fig. 10). Thin skinned folds in this area are interpreted to lie above a bedding parallel detachment located near the base of the Danish River Formation and above undeformed reflectors assumed to be generated from gently northwesterly-dipping Ordovician platform carbonates. The southeastern limit of this seismically-identified deformation intersects the sea floor in Kennedy Channel half way between Hans Island and the Ellesmere coast (Fig. 8).

In coastal western Hall Land of North Greenland the gradational limit of deformation is exposed in Peary Land Group strata, and the extent of mesoscopic folds in these strata can be outlined, somewhat ambiguously, on the detailed geological maps of Dawes (1987). A reasonable estimate for Cenozoic offset of the limit of Ellesmerian deformation is 45 ±25 km in a left lateral direction. This value has been determined on the offshore trace of the JDFS between Cape Lawrence and the mouth of Scoresby Bay (Fig. 8).



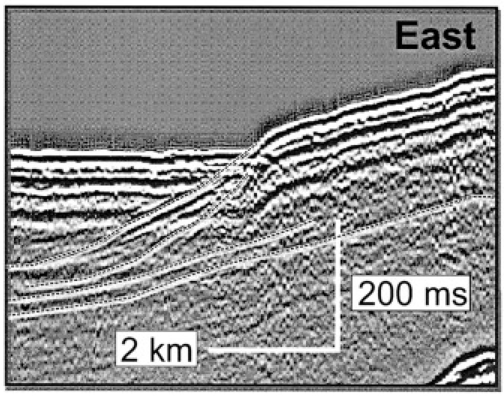


Fig. 10: Portions of two BGR profiles acquired in Kennedy Channel west of Hans Island area. Top = easterly-transported thrust folds developed in Silurian Danish River Formation. The through going basal reflection is interpreted to have been generated from undeformed Upper Ordovician strata. Below = down-lapping reflections at the Silurian carbonate shelf edge near Hans Island. Alternative interpretations of these profiles are provided by JACKSON et al. (2006).

Metamorphic isograds (marker 13 of DAWES & KERR 1982; new marker 21)

Throughout the Ellesmere-North Greenland region, regional metamorphism is restricted to lower Paleozoic and older strata (UTTING et al. 1989, SURLYK 1991, TRETTIN 1994). Above a profound regional angular unconformity there is a distinct break to lower grade thermal maturity conditions in upper Paleozoic and younger strata of the Wandel Sea Basin in Northeast Greenland (SURLYK 1991) and in the Sverdrup Basin succession of Ellesmere Island (UTTING et al. 1989). Marker 13 of DAWES & KERR (1982) considered the southern limit of lower amphibolite facies conditions on either side of Nares Strait (Figs. 1, 11). However, the shortest distance between amphibolite facies outcrops of northern Ellesmere Island and those mapped in northern Greenland is between 350 and 400 km. Precise understanding of the trend and position of the southern limit of amphibolite grade conditions is also hampered on Ellesmere Island by a lack of suitable rock types and by indications of metamorphic retrogression (TRETTIN 1994). For this reason there has been no attempt at re-assessment.

An alternative useful marker is provided by the southern limit of thermally over-mature conditions (new marker 21). This is a relatively well documented line on both sides of Nares Strait (Fig. 11). On Hans Island and throughout northeastern Ellesmere Island, thermal maturity has been estimated using CAI values measured on conodonts extracted from strata of mostly Lower Ordovician through Lower Silurian age (HARRISON, NOWLAN et al. in press). Conodonts with CAI values up to 3 have experienced immature and mature paleothermal conditions. CAI values of 3.5 and higher are considered over-mature (EPSTEIN et al. 1991). For North Greenland, the line separating mature and over-mature conditions has been obtained using an inferred 455° isocontour on the T_{MAX} contour map of Surlyk (1991). T_{MAX} values are obtained from geochemical analyses of dispersed organic matter, which in North Greenland appears to coincide with a pattern of sampling associated with various Silurian and Ordovician shelf carbonate formations.

Results, previously summarized in this account, indicate that the JDFS coincides with a dramatic break westward to higher paleothermal conditions. The 455° T_{MAX} isocontour in Hall Land projects across Nares Strait to north of Cape de Fosse, site of the northernmost sample locality having a CAI value of less than 3 (Fig. 11). West of the JDFS, the southern limit of over-mature conditions lies south of the head of Dobbin Bay and projects into the JDFS between the mouth of John Richardson Bay and Cape Lawrence. Implied sinistral offset is 135 ±40 km.

Although relatively large offsets are implied by these observations, it is reasonable to consider that the JDFS was the site of differential sediment accumulation during the peak of thermal metamorphism potentially coincident with the deposition of the Middle and Upper Devonian foreland clastic wedge. Portions of this wedge are preserved on central and southern Ellesmere Island (HARRISON, NOWLAN et al. in press, EMBRY 1991). Within the Nares Strait region, higher rates of sediment accumulation in the northwest, and greater total thicknesses of former sediment load in that direction could have created a distinct break in thermal maturity conditions in correlative rocks across a Devonian age fault having a down-to-the-northwest sense of slip during simultaneous sedimentation. The resultant thermal maturity step may therefore be all that remains of the former sediment load most or all of which would have been stripped during end-Devonian (Ellesmerian) and Cenozoic phases of orogeny (HARRISON, NOWLAN et al. in press).

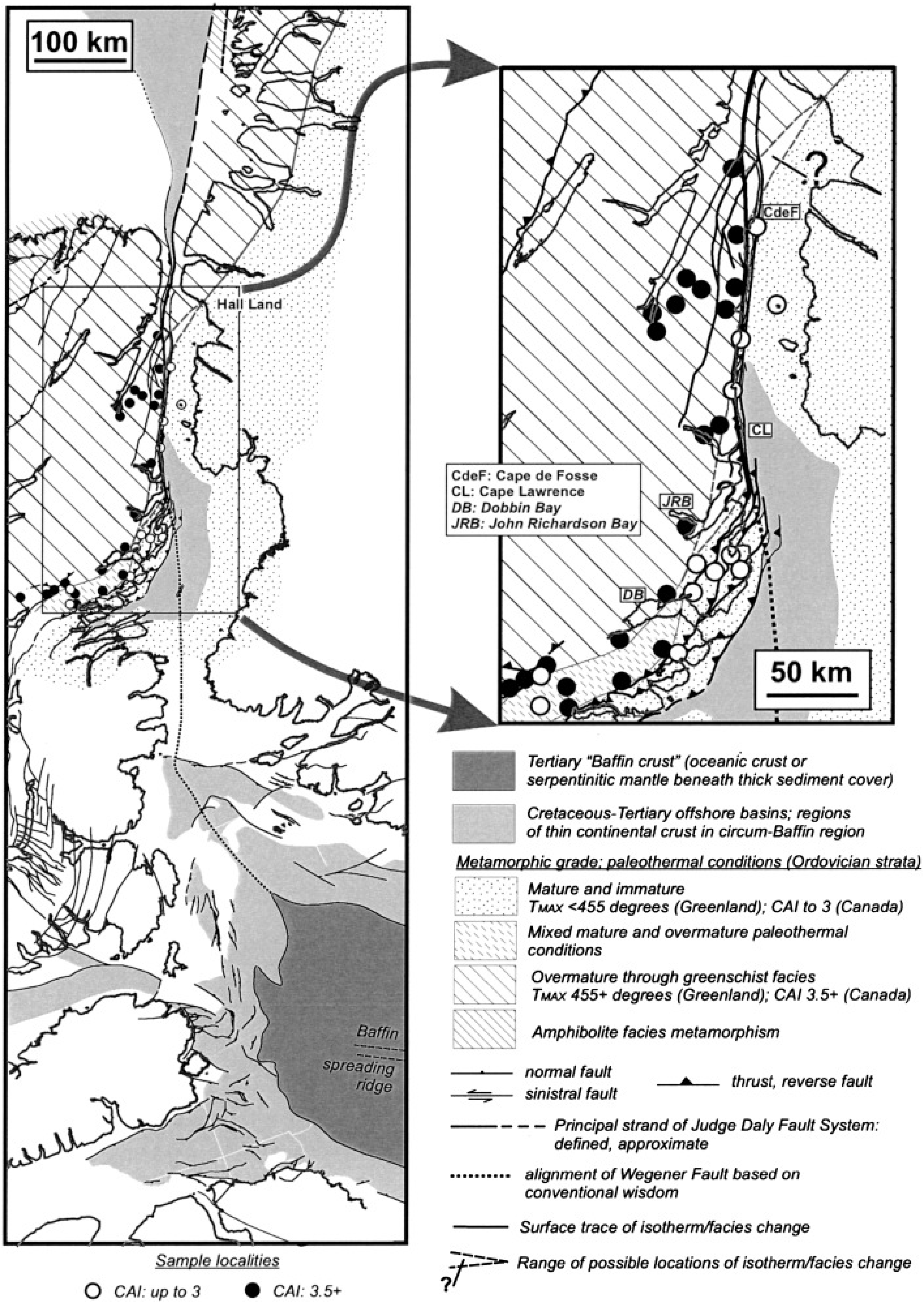


Fig. 11: Regional variation in metamorphic conditions within lower Paleozoic strata including the extent of amphibolite metasediments (marker 13 of DAWES & KERR 1982) and upper limit of thermally mature sediments (new marker 21, this study). Sinistral offset of the northwestern limit of mature paleothermal conditions in Ordovician strata is 150 ±45 km.

Lake Hazen-Harder Fjord fracture line (marker 14 of DAWES & KERR 1982)

The Lake Hazen – Harder Fjord fracture line (Fig. 1) has two components: the Harder Fjord Fracture Zone of North Greenland (HENRIKSEN 1989), and the Lake Hazen Fault Zone of Ellesmere Island (TRETTIN 1994). The Harder Fjord Fault Zone is traceable across North Greenland, bends southwards in the west and apparently dies out in east central Nansen Land north of J.P. Koch Fjord (HENRIKSEN 1989). Therefore, as mapped, the Harder Fjord Fault Zone cannot be continuous with the Lake Hazen Fault Zone of northern Ellesmere Island which extends into the Lincoln Sea near Alert (TRETTIN 1994). Likewise, there is no proven aeromagnetic linkage of these structures, and the intervening continental shelf is 230 km wide. Therefore, it remains to be established that the two mapped fault zones have ever been part of a single connected fracture system. In light of these uncertainties, the marker is considered invalid and no attempt has been made to re-assess the displacement on these separate fault zones.

Judge Daly-Nyeboe Land fracture line (marker 15 of Dawes & Kerr 1982)

Elements of the Judge Daly-Nyeboe Land fracture line (Fig. 1) include the westerly-striking Freuchen Land Thrust that extends from Freuchen Land to Nyeboe Land in North Greenland (features located on Fig. 2; Surlyk 1991), and the Judge Daly Fault Zone of Mayr & de Vries (1982) which is part of the JDFS of the present account. Within the Nares Strait region, the JDFS has been identified in this account as the only possible position for the plate boundary between Greenland and Arctic North America (see also Damaske & Oakey 2006). Although the Freuchen Land Thrust may extend westward into the JDFS, it cannot be considered part of a valid marker because it does not cross this fault system.

Archer Fiord-Wulff Land anticlinal zone (marker 16 of Dawes & Kerr 1982)

Wulff Land Anticline, the eastern half of the Archer Fiord-Wulff Land anticlinal zone in North Greenland (Surlyk 1991) is developed in Neoproterozoic strata, and basin facies Cambrian to Silurian strata (Figs. 1, 8). It plunges southwestward across northern Wulff Land and northern Nyeboe Land, and carries the Cambro-Ordovician shelf margin facies change on its southerly and south-easterly facing limb (compare Figs. 3, 8; Henriksen 1989).

The Archer Fiord Anticlinorium, the Ellesmere Island portion of the Archer Fiord – Wulff Land anticlinal zone, lies entirely within the Ellesmerian Orogen of north-western Judge Daly Promontory. This structure is mostly developed in Cambrian to Silurian slope and basin facies strata assigned to the Ellesmere Group, and Hazen and Danish River formations. The southeast-facing limb is faulted in the northeast and along strike to the southwest carries the Cambrian and Ordovician shelf margin facies change (Figs. 3, 8). The hinge of the anticlinorium plunges to the southwest and is exposed southwest of the head of Archer Fiord (Harrison, Mayr et al. in press). To the northeast the fold hinge intersects the JDFS in Hall Basin. Contained basin facies strata possess an intense axial planar cleavage characteristic of Ellesmerian deformation (Piepjohn et al. in press). Although there are Cenozoic faults on the south limb of Archer Fiord Anticlinorium, both this and Wulff Land Anticline are considered to be lower Paleozoic folds (Soper & Higgins 1990, Piepjohn et al. in press).

The precise location of the fold hinges in Hall Basin and Robeson Channel is unconstrained. If these two folds are displaced portions of a common lower Paleozoic structure, a reasonable estimate for the Cenozoic displacement of the fold hinge is 60 ±10 km left lateral (Fig. 8).

Region of distinctive magnetic character (marker 17 of Dawes & Kerr 1982)

A wide belt of distinctive magnetic character extends from north-westernmost Ellesmere Island to northernmost Greenland (Fig. 1), and approximately coincides with areas known to contain Cretaceous through lower Paleocene volcanic rocks and associated dykes. The youngest igneous rocks are the Kap Washington volcanics dated at 64 ±3 Ma (Larsen 1980). Volcaniclastic sediments and volcanic clasts (dated at about 60 Ma, Estrada et al. in press) in the upper Paleocene Pavy Formation of Judge Daly Promontory may also originate from aeromagnetically-defined features partly located in the region of distinctive magnetic character (Hood et al. 1985, Damaske & Oakey 2006). Thus the magnetic belt may have developed prior to strike-slip motion on the JDFS, or may be in part coincident with those motions.

Regardless of its origin, the North Ellesmere – North Greenland region of distinctive magnetic character is too wide and the boundaries are too diffuse to provide for a quantifiable estimate of post-magnetic offset (if any).

Judge Daly-Nyeboe Land gravity gradient (re-defined marker 19 of Dawes & Kerr 1982; new marker 22)

Contours of gravity in the Hall Basin and Robeson Channel region have been provided by G. Oakey (pers. comm. 2003), and are based on the regional compilation of Oakey et al. (2001). The Judge Daly – Nyeboe Land gravity gradient is most notably expressed in gravity contours that rise sharply toward the northwest from -100 to -20 mgal (Fig. 12). The steep gradient zone is 30-45 km wide on northern Judge Daly Promontory, 50 km wide across northern Nyeboe Land and, in both areas, appears to coincide with the Cambro-Ordovician shelf edge facies change (see Figs. 3, 12). The gradient is most notably expressed by the -70, -80 and -90 mgal contours. On the northwest side, these contours intersect the JDFS between central Judge Daly Promontory and the south shore of Lady Franklin Bay, and on the east side intersect the JDFS between the north tip of Judge Daly Promontory and Robeson Channel north of Hall Land (Fig. 12). The angle of intersection is 18° to 23°. Estimated left lateral offset of the -70 mgal contour is 90±? km (revised marker 19), and of the -90 mgal contour is 60±? km (new marker 22). Presently unknown is the concentration of data points that would serve to define the range of uncertainty in the position of these gravity contours.

Earthquake epicentres (marker 20 of Dawes & Kerr 1982)

Not re-assessed

TRANSFORM PLATE BOUNDARY IN BAFFIN BAY (MARKER 23) AND SOUTHERN NARES STRAIT

Regional free-air gravity data, interpreted by Chalmers & Pulvertaft (2001), reveal a northwesterly-trending spreading ridge running up the centre of Baffin Bay. The northwestern end of this linear gravity anomaly dies out in deep water east of the mouth of Lancaster Sound. However, projected to the northwest the spreading ridge must ultimately terminate at the eastern limit of Arctic North American continental lithosphere, a sinuous northerly-trending structure that is aligned with the southern end of Nares Strait (Fig. 2; Chalmers & Pulvertaft 2001, Harrison, Brent et al. in press). The precise position of the plate boundary in this area is not clearly defined by mapping using industry reflection seismic profiles. Various northerly-trending shelf margin basins and associated

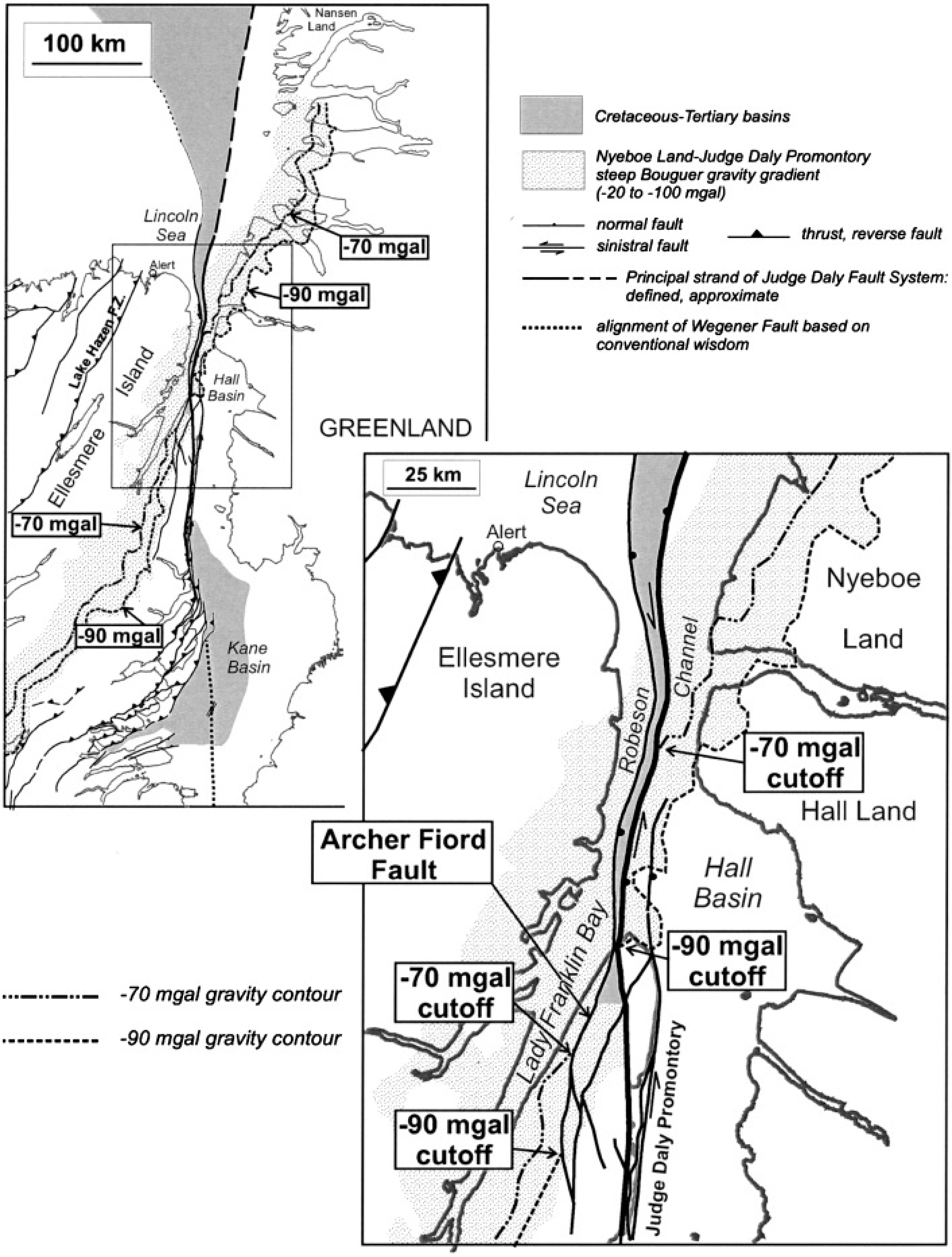


Fig. 12: Sinistral offsets of selected gravity contours from within the steep Bouguer gravity gradient zone include the -70 mgal contour (re-defined marker 19; 90 ±? km) and -90 mgal contour (new marker 22, this study; 60 ±20 km). Gravity contours provided by Oakey (pers comm. 2003) based on the compilation of Oakey et al. (2001).

short fracture systems could accommodate elements of strike slip motion including Glacier Basin, the west margin of Carey Basin and, farther north, the western margins of North Water and Thule Basins. Nevertheless, based on refraction profiles (Reid & Jackson 1997) and contoured bouguer gravity (Miles 2000b) that define the extent of oceanic (or "oceanized" serpentinitic) Baffin crust in central Baffin Bay, the total amount of Late Paleocene-Eocene sea floor spreading and coeval transform displacement measured at the northerly-trending eastern limit of continental lithosphere is about 260 km, perhaps half of this (or c. 130 km) acting in a sinistral sense north of the spreading ridge. To this must be added strike slip motions acting in similar directions during Cretaceous – early Paleocene extensional spreading of partly thinned continental lithosphere on the West Greenland and Devon Island – Baffin Island continental margins. With a maximum 50 % lithospheric thinning, the implied maximum additional strike slip motion is about 60 km, again perhaps half of this (or c. 30 km) acting in sinistral transform sense north of the spreading axis.

To summarize the above observations, the minimum sinistral displacement for the possible transform plate boundary at the northwestern limit of the Baffin spreading ridge is c. 145 ±15 for the case in which half of the available displacement is carried westward into the Arctic Islands or is accommodated on dextral transforms on the opposite side of the spreading ridge (new marker 23). The range in the minimum (130-160 km) takes into account the range of Cretaceous – early Paleocene transform slip and extension of the continental margin north of the spreading axis.

The maximum sinistral displacement for the possible transform plate boundary at the north-western limit of the Baffin spreading ridge is 280 ±30 km for the case in which all of the available displacement is carried northwards through Nares Strait. The range in the maximum (255-315 km) takes into account the range of Cretaceous – early Paleocene transform slip and extension of the continental margins on both sides of the spreading axis.

The sinistral displacement value adopted for this marker is the average between the above calculated minimum range and maximum range values (220 ±90 km)

Farther north, conventional wisdom places the plate boundary (Wegener Fault) in the middle of the channel between Ellesmere Island and Greenland. This necessitates finding a precise sinistral fault zone within or beneath: 1) Paleogene(?) strata that underlie north-western Kane Basin; 2) Cambro-Ordovician platformal strata which likely reach the sea floor in southern Kane Basin; 3) granulite facies plutonic and metamorphic supracrustal rocks believed to underlie Smith Sound; and 4) mid-Proterozoic sediments and volcanics of Thule Basin of Smith Sound and northern Baffin Bay (Harrison, Brent et al. 2006). These and other constraints, described below, indicate that there is unlikely to be any significant northerly-striking fault zone in southern Nares Strait.

GEOLOGICAL AND GEOPHYSICAL MARKERS THAT CROSS SOUTHERN NARES STRAIT

Erosional limits of the Paleozoic (new markers 24, 25)

The erosional limits of the Paleozoic (marker 24), and also many of the Paleozoic outliers (marker 25), occur as high standing features that, in the absence of bounding faults, present a cliff-forming escarpment above low-standing basement gneisses (Fig. 13). This escarpment, exposed throughout Bache Peninsula on eastern Ellesmere Island, includes Lower Cambrian Dallas Bugt Formation and, at the base, a tongue of Thule Supergroup sediments containing Neoproterozoic sills (Peel et al. 1982, de Freitas et al. in press). Eastward in southern Kane Basin these sills have an aeromagnetic expression (Hood et al. 1985, Okulitch et al. 1990) that serves to indirectly define the sea floor position of the basal Cambrian unconformity (Fig. 13).

In Greenland an escarpment also occurs on the southern edge of Daugaard-Jensen Land. Here the lowest scarp face exposures are Lower Cambrian Humboldt Formation (a stratigraphic equivalent of the Dallas Bugt Formation; Jepsen et al. 1983, Dawes et al. 2000, Dawes & Garde 2004), and the Greenland ice cap presumably covers low standing Precambrian basement. The escarpment and low relief basement complex are represented in the offshore as a bathymetric rise and deeper water moat, respectively, that together form a continuous submarine physiographic feature that extends from Bache Peninsula and Buchanan Bay through southern and eastern Kane Basin to the northern margin of Humboldt Gletscher and Daugaard-Jensen Land (Harrison, Brent et al. 2006). Offset in Smith Sound is estimated at 0 ±10 km.

The sub-Cambrian unconformity generally rises to the south on both sides of Smith Sound and a point is reached beyond which only Precambrian rocks are represented on all the higher ground. The line defining the onshore southeastern limit of mapped Paleozoic outliers is potentially continuous across the intervening channel without offset. Offset is estimated at 0 ±20km.

Although these markers describe physiographic elements of unknown age, probably modified substantially in the Quaternary, a key reason for including them in the present account is that they are devoid of features that should indicate a major fault zone. Major strike slip fault systems are typically associated with profound local differences in topography. Classic examples include the Great Glen Fault in Scotland, the Cabot Fault in the Canadian Maritimes, and the Tintina Fault and Rocky Mountain Trench system in the Yukon and British Columbia (Wilson 1963, Gabrielse et al. 1991). Even the Judge Daly Fault Zone where it is exposed onshore on Judge Daly Promontory, has dramatic physiographic expression (Mayr & De Vries 1982, Harrison in press). All this indicates that basement and cover on either side of Smith Sound and southern Kane Basin must be part of a common crustal block. Patterns of uplift and differential erosion are entirely inconsistent with the existence of a through-going young fault system.

Contact between Archean and Paleoproterozoic rocks (new marker 26)

A recent review of the major Paleoproterozoic and Archean geological features of northeastern Arctic Canada and western Greenland is provided by Jackson (2000). Related major features of the Nares Strait region are illustrated on Figure 14. Southern areas around northern Baffin Bay are included with the Neoarchean age Committee Belt (2.9-2.5 Ga granite-greenstone terranes with Paleoproterozoic reworking). These terranes are bound to the northwest by the Thelon Tectonic Zone (TTZ) of northern Devon Island, south-eastern Ellesmere Island and Inglefield Land. The TTZ is a wide but lengthy belt of Paleoproterozoic or older(?) supracrustal rocks (i.e. marble, metasulphide facies iron formation, quartzite, and aluminous metapelite) and 1.7 to 2.0 Ga charnockite granitoid intrusions that has been traced from subsurface western North America, east of the meso-Archean Slave Craton to north-eastern Arctic Canada and Greenland (Jackson 2000).

Basement crystalline rocks in the Victoria Fjord area, located between the ice cap and the flat-lying Cambrian cover in North Greenland (Fig. 14), have provided zircon U-Pb ages in the 2.9 to 3.2 Ga range (Hansen et al. 1987) and also indications of a Proterozoic metamorphic overprint. This is the exposed part of a potentially much larger Mesoarchean cratonic block that, like the Slave Craton, appears to be bound on the south and east by the Thelon Tectonic Zone.

The southern contact of the TTZ with the Committee Belt is drawn through Prudhoe Land north of Inglefield Bredning in Greenland (Jackson 2000), a position taken from the 1:500 000 bedrock geology map of the Thule area (Dawes 1991). However, unpublished radiometric ages recently obtained from Prudhoe Land now indicate that the boundary between the Neoarchean and Paleoproterozoic is gradational in this area (Dawes pers. comm. 2004)

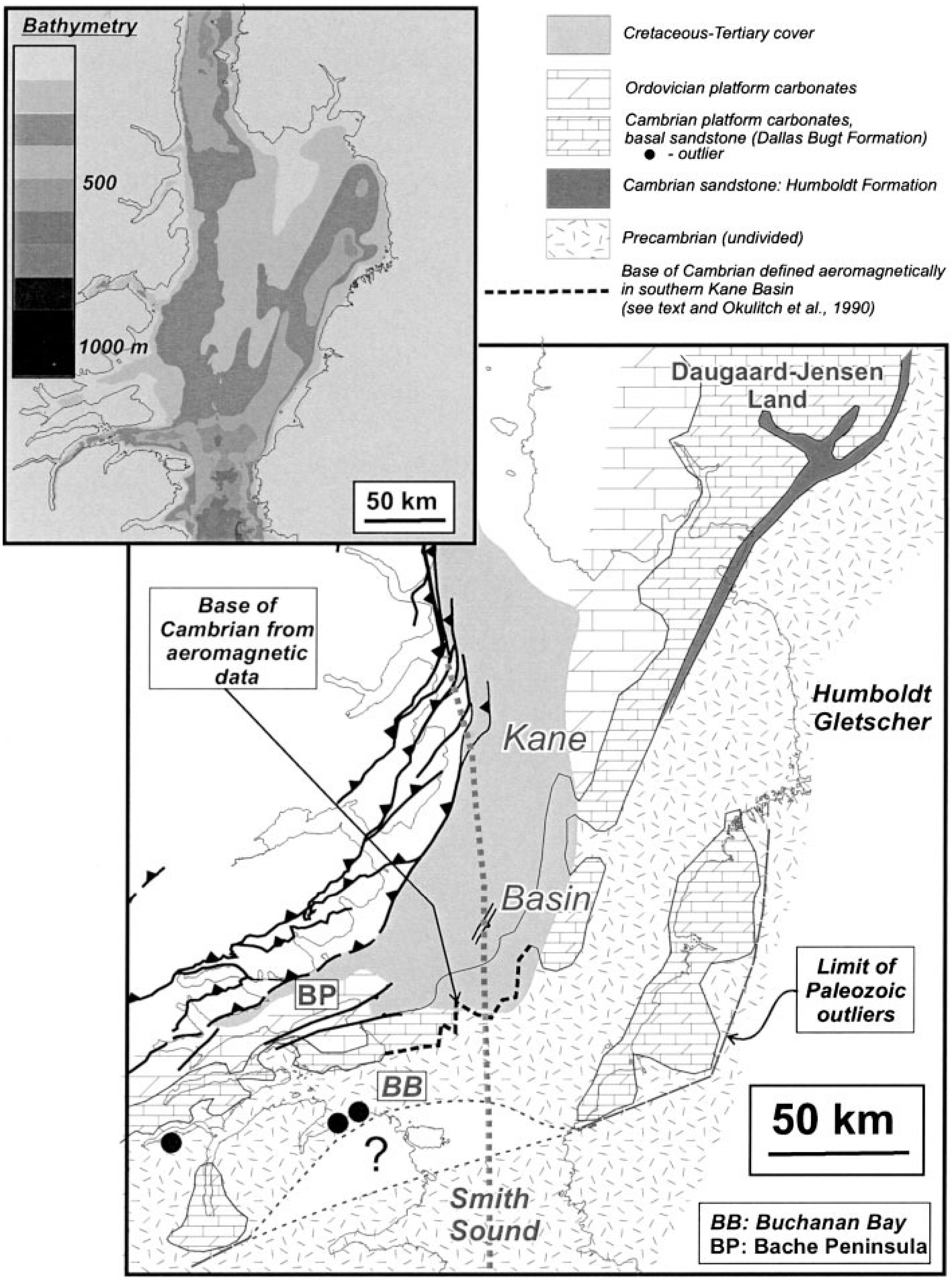


Fig. 13: Southeastern limit of continuous lower Paleozoic cover and limit of lower Paleozoic outliers in the Kane Basin and Smith Sound region (new markers 24 and 25). Estimated potential offset is 0 ±10 and 0 ±20 km for these markers, respectively. Although these bathymetrically and topographically defined features are of unknown age, there is no indication that a major plate boundary fault is present in this area.

On southern Devon Island the northern limit of the Archean against the Thelon zone has been defined aeromagnetically and drawn by JACKSON (2000) as a major westerly-striking dextral sense shear extending into the central Arctic Islands beneath widespread Paleozoic cover. Minor shear zones are common in basement rocks of eastern Devon Island, including at least one west – southwesterly-striking mylonite-pseudotachylite zone hundreds of metres wide in a Precambrian basement inlier (see NTS 48G, Bear Bay East map sheet of HARRISON 1984). However, bedrock geology maps, prepared by Petro-Canada staff in the early 1980s, indicate that Proterozoic metasulphide facies iron formation and marble, common on south-eastern Ellesmere Island, grades southward to quartzite and marble. Together these meta-sediments form a linked depositional facies belt that is continuous through Coburg Island to northern Devon Island. The Petro-Canada maps and report indicate that the quartzite may unconformably overlie gneisses of Archean age in the south of this belt and that Archean gneissses and Paleoproterozoic meta-sediments are tectonically interleaved throughout eastern Devon Island (HARRISON 1984).

The extent of the TTZ, Committee Belt and intervening zones of gradational character within the present report area is illustrated on Figure 14 together with the trend of linear aeromagnetic anomalies from MILES (2000a) and the strike of Paleoproterozoic foliations recorded on the maps of HARRISON (1984) and FRISCH (1988) for Ellesmere and Devon islands,

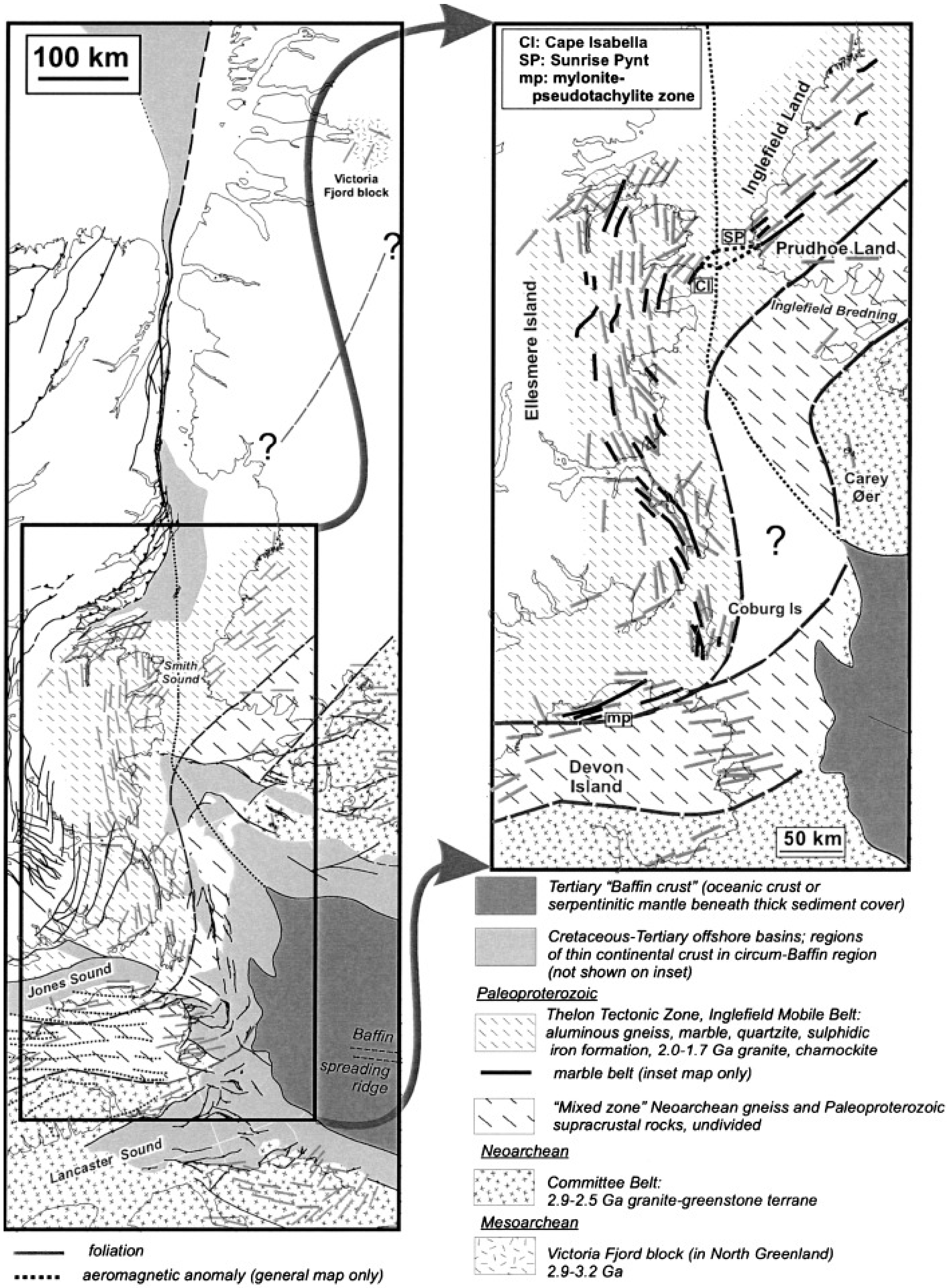


Fig. 14: Regional tectonic features of the Neoarchean Committee Belt and Paleoproterozoic Thelon Tectonic Zone (Inglefield Mobile Belt) of the Smith Sound and northern Baffin Bay region. Paleoproterozoic marbles and the major Precambrian crystalline terrains can extend from Canada to Greenland without appealing to large fault motions in the offshore (marker 1 of DAWES & KERR 1982; new marker 26, this study).

and DAWES (1991) and DAWES & GARDE (2004) for Greenland. Tectonic trends within the Thelon Zone are 1) east north-easterly on Devon Island and on the south coast of Ellesmere Island (consistent with a regional dextral shear zone to the south), 2) northerly on south-eastern Ellesmere Island and on the Carey Øer (DAWES 1991), and 3) easterly to north-easterly in Inglefield Land (DAWES & GARDE 2004). The northerly strike of foliations on the Carey Øer has also been confirmed by this author during a brief sampling visit to the westernmost part of those islands in 2001 (HARRISON 2003).

Two reconstructions might be considered. 1) Inglefield Land geology of the Thelon Tectonic Zone in Greenland (= Inglefield Mobile Belt of DAWES et al. 2000) is continuous with the TTZ rocks on eastern Ellesmere Island directly west of Smith Sound. 2) The Inglefield region of the TTZ in Greenland is continuous with the TTZ on eastern Devon Island.

The first solution is simplest as it requires no Cenozoic strike slip motion in northern Baffin Bay and Smith Sound. The uncertainty, however is very large (at least ±100 km) because the shape of the Precambrian crystalline belts is unconstrained in continental offshore areas of northern Baffin Bay. The second solution implies up to about 230 km of sinistral slip but is rejected as it is inconsistent with other independently derived markers that cross southern Nares Strait without apparent offset (see below).

Paleoproterozoic marble belt (marker 1 of DAWES & KERR 1982)

Within the Thelon Tectonic Zone, the belt of marble-bearing rocks of FRISCH & DAWES (1982) is shown extending from Coburg Island in the mouth of Jones Sound northwards through Ellesmere Island, across Smith Sound and Inglefield Land (Fig. 1). Specific marble units exposed at Cape Isabella on the Ellesmere coast are correlated by FRISCH & DAWES (1982) with similar marbles that extend across Inglefield Land and into the offshore at Sunrise Pynt (Fig. 14). Close examination of the correlated marble belts indicate, that the Cape Isabella marble strikes N50°E into the offshore (FRISCH 1988). The thickest of the Sunrise Pynt marble belts, traceable over a distance of about 100 km across Inglefield Land, strikes toward S75°W where it extends into the offshore (as measured on the geology map of DAWES & GARDE 2004). Offshore curvature of the correlated marbles would appear likely and zero offset is therefore possible. Maximum permissible offset, based on no curvature of the extrapolated onshore trends and a through going fault in mid-channel, is 20 km. Zero offset is more likely. The latter conclusion is an especially compelling argument in the light of other constraints from this same area, described below.

Mid-Proterozoic Thule Basin (marker 2 of DAWES & KERR 1982, new marker 27)

Mid-Proterozoic sedimentary and volcanic rocks of the Thule Supergroup are described by DAWES (1997). The nature and distribution of these strata in the offshore areas of northern Baffin Bay are described by NEBEN et al. (2006) and the distribution of correlative strata, including occurrences of northern Baffin Island, are illustrated on the geology map described by HARRISON, BRENT et al. (2006). Important features that distinguish Cretaceous-Tertiary sedimentary basins from those associated with the Thule Basin include:
1) refraction and reflection stacking velocities of less than 3.5 in the Cretaceous-Tertiary, but mostly above 4.0 in the Proterozoic;
2) a recognizable seismic stratigraphy in the Cretaceous-Tertiary, including distinctive onlap and progradational features, all of which are absent from the Thule Basin;
3) listric normal faults in the younger succession but planar normal faults in the older succession;
4) bedding attitudes and regional structural trends that are north-westerly in the Cretaceous-Tertiary but more westerly in the Thule Basin;
5) synclinal shape of the Thule Basin - two parallel regional synclines with intervening high - but half-graben structural style of the younger succession, and;
6) recessive weathering characteristics for the younger rocks but extremes of differential erosion in the Thule Basin resulting from the contrasting erosional properties of contained mafic sills and flows, cemented and uncemented sandstones and high-standing basement-rooted horst blocks.

These characteristics establish the correlation of onshore Thule Supergroup and basement gneiss exposures in the Thule area, around Inglefield Bredning, and on the offshore islands with related seismic features on GSC-BGR and Industry seismic profiles of northern Baffin Bay (Fig. 15). More specifically, Thule Supergroup sediments are thickly developed on BGR profile 4-2001 which indicates westward continuity of the northern part of the basin (and of the bounding basement high blocks) to within 27 km of the east Ellesmere coast.

The seismically-defined, eastward-trending, trough axis located within the northern part of Thule Basin (Fig. 15) is also aligned with an east-plunging syncline axial trace drawn on the Ekblaw Glacier map sheet (NTS 39F) of HARRISON (1984) and located within outcrop of the Thule Basin exposed between Gale Point and Cape Dunsterville on the Ellesmere coast (see also DAWES 1997). If the reflection seismic syncline and onshore synclinal trace are common structural elements, the maximum implied offset on an intervening north-striking strike slip fault would have experienced about 15 km of dextral sense motion (new marker 27, this study)! However, it is more likely that the seismically identified fold has an element of northward curvature and zero offset in the direction of the Ellesmere coast.

A final argument in support of minimal offset in southern Nares Strait is observed near the northern margin of the Thule Basin in western Inglefield Land and south of Cape Isabella on eastern Ellesmere Island. In both these areas the Thule Basin succession is locally down-faulted against and parallel to foliations in the Paleoproterozoic marble-granitoid-granulite complex. Near Sunrise Pynt (Fig. 15) a fault has been mapped by DAWES & GARDE (2004) that strikes toward S15°W, displaces Thule Basin rocks, and is located 1200 m south of the Sunrise Pynt marble belt described in marker 1, above. At Cape Isabella, the Thule Basin succession is also apparently down thrown against the Paleoproterozoic (see FRISCH 1988). However, Recent glacial outwash gravels lie on the probable fault trace. Trend of the up-thrown basement escarpment is toward S5°W. The Cape Isabella marble belt (of marker 2, above) is within 500 m to the north of the assumed fault trace, and there is good general alignment of both post-Thule strata faults/lineaments without the need for appealing to a young sinistral fault system in the intervening channel. Allowing for the possibility that the Sunrise Pynt and Cape Isabella faults do not represent the northern margin of Thule Basin, the allowable offset of this marker is 0 ±15 km (marker 2 of DAWES & KERR 1982).

Bache Peninsula arch (marker 3 of DAWES & KERR 1982)

The Bache Peninsula arch (Fig. 1), originally named by KERR (1967), was identified as a high block underlain by crystalline rocks of the Precambrian shield that separates Thule Basin in the south from thickening Cambrian and younger strata of the Franklinian Basin to the north. PEEL et al. (1982) identified this as a marker extending from Ellesmere Island to Greenland in the vicinity of Smith Sound. DAWES & KERR (1982) suggested a maximum possible sinistral offset of 100 km.

Field work has subsequently shown that the shield was tectonically elevated and associated with the accumulation of redbeds and proximal coarse clastics on central and southern Ellesmere Island in Late Silurian and Early Devonian time ("Inglefield Uplift" of SMITH & OKULITCH 1987, "Bache Uplift" of DE FREITAS et al. 1997). This is a distinctly different northerly-trending basement structure similar in some ways to the Boothia Uplift of the central Arctic Islands. Regardless of these new contributions, there is no reason to reject the marker

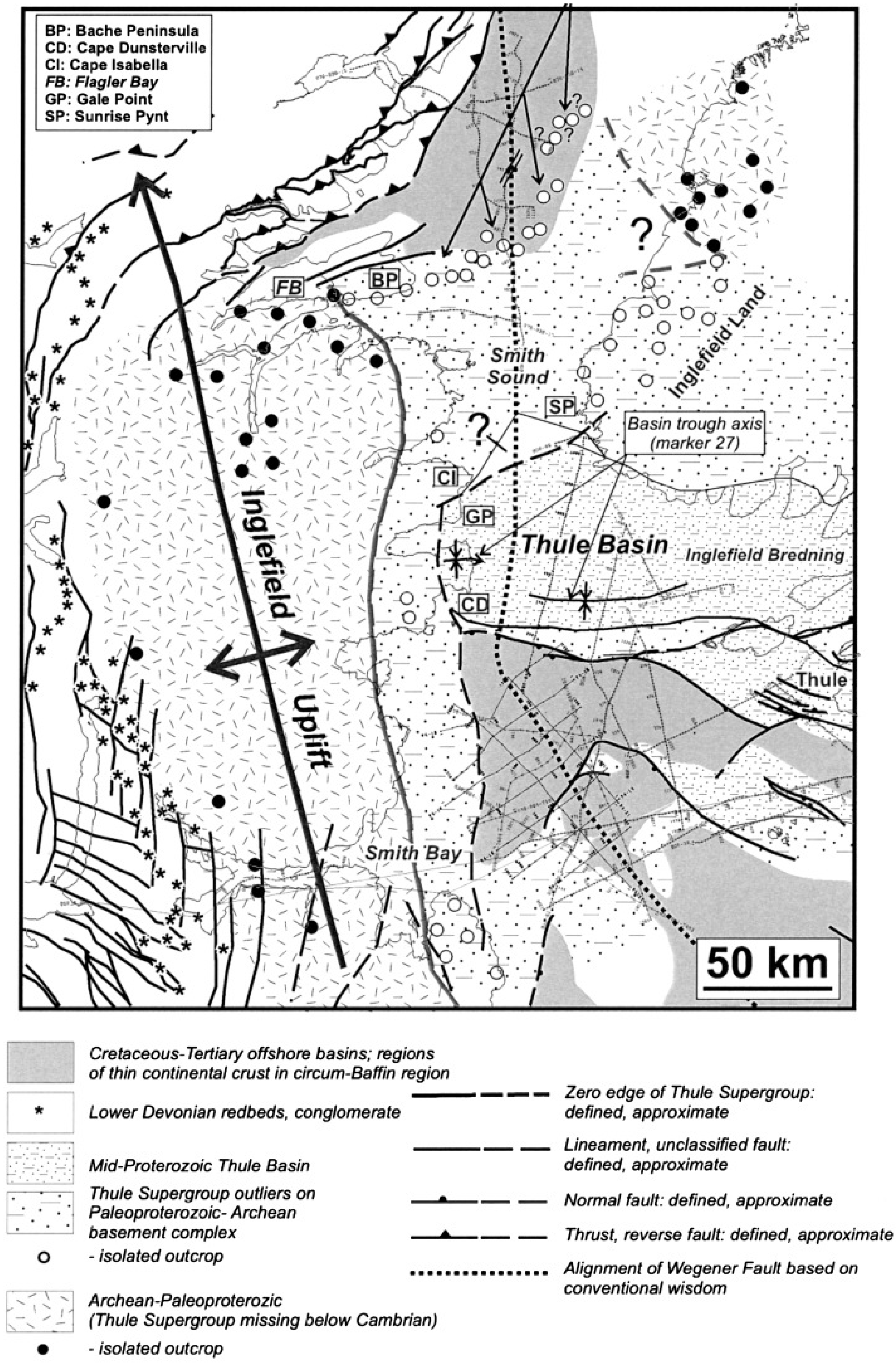


Fig. 15: Geological-geophysical markers provided by the Thule Basin and Thule Supergroup of northern Baffin Bay (marker 2 of DAWES & KERR 1982; new marker 27). Estimates of respective offset for these markers are 0 ±15 km and 0 to 15 km right lateral. The zero-edge of the Thule Supergroup can not be proven to cross Nares Strait (marker 4 of DAWES & KERR 1982).

identified by KERR (1967) and PEEL et al. (1982), nor to question the conclusions of DAWES & KERR (1982), who suggested that the position of the marker could accommodate anything up to 100 km of sinistral displacement.

Zero edge of Thule Supergroup (marker 4 of DAWES & KERR 1982)

Distribution of Thule Supergroup outliers is illustrated on Figure 15. This map also shows the distribution of Phanerozoic outliers present on the shield where intervening Thule Supergroup sediments are missing. Contrary to the alignment of this marker drawn by DAWES & KERR (1982; see Fig. 1), it is now apparent from the mapping by Frisch (1988) that the western limit of the Thule Supergroup is actually a northerly striking line that extends from south of Smith Bay on south-eastern Ellesmere Island northwards to the mouth of Flagler Bay on the south coast of Bache Peninsula (Fig. 15). Near Bache Peninsula the zero-edge of the Thule Supergroup actually strikes away from Nares Strait and to the north the extent of these rocks in the subsurface is unknown.

In Inglefield Land, the limit of the Thule Supergroup beneath overstepping Lower Cambrian Dallas Bugt Formation is a poorly constrained line that could trend anywhere from westerly to north-westerly (Fig. 15; Dawes & Garde 2004).

Aeromagnetic data supplied by Hood et al. (1985), most recently re-interpreted by Okulitch et al. (1990), provide an indication of the extent of Neoproterozoic sills in the Thule Supergroup east of Bache Peninsula in Kane Basin (Figs. 14, 15). These authors argue that positive aeromagnetic anomalies derived from the offshore sills carry evidence of two to four strike slip faults having a total of 25 km of sinistral motion. Not at all considered in those papers is the possibility that large apparent sinistral displacements in shallow dipping strata could also be generated from very small amounts of vertical throw. The interpretation adopted in the present paper is that the anomalies arise from shallow dipping Neoproterozoic sills and that sea floor outcrop forms a sinuous belt controlled by sea floor topography.

The conclusion to be drawn from all the above observations is

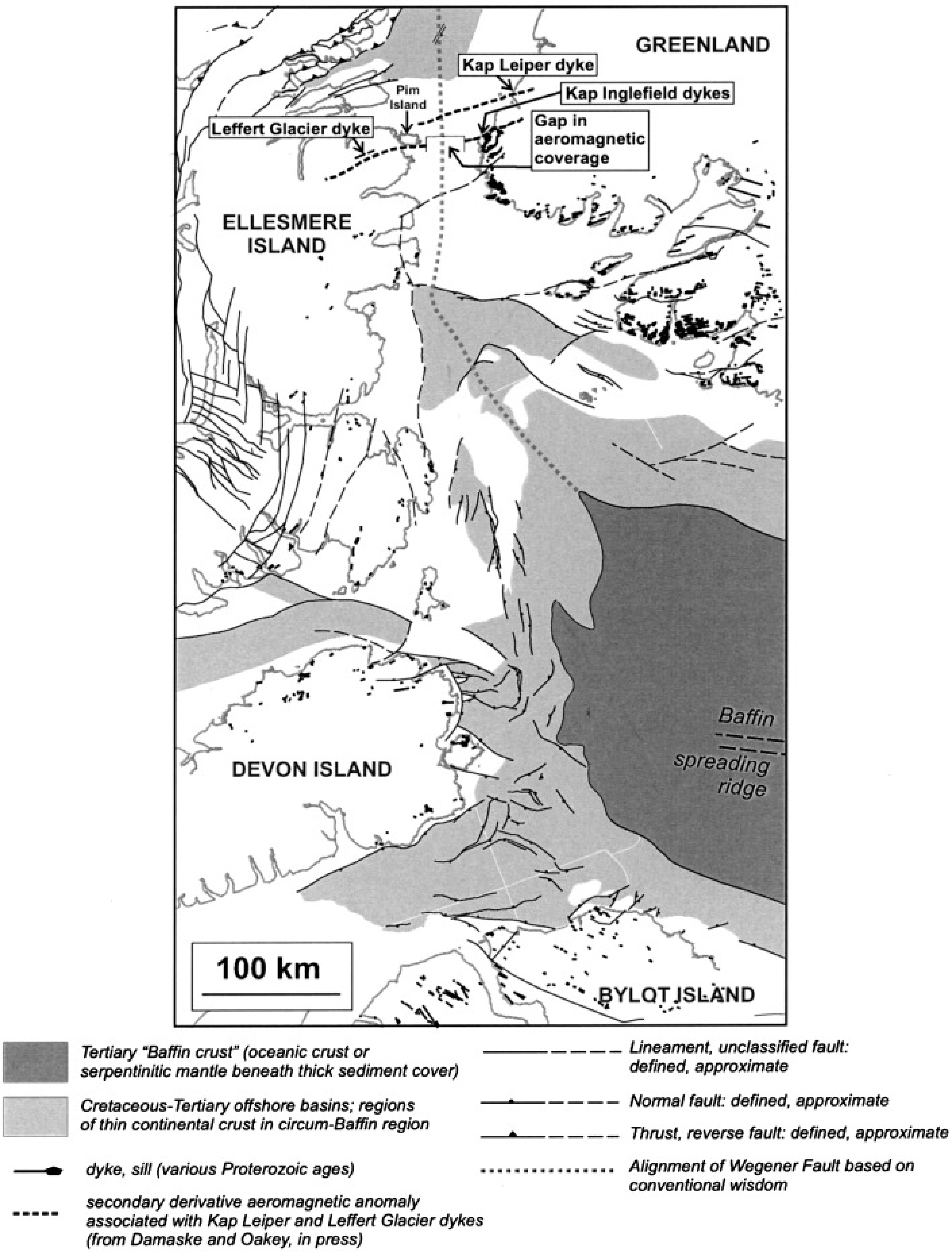


Fig. 16: Basic dykes and sills of southern Nares Strait and northern Baffin Bay region, and second derivative linear aeromagnetic anomalies in the Smith Sound area of southern Nares Strait (modified from Damaske & Oakey 2006). Indicated displacement of the Kap Leiper Dyke and related anomaly (marker 28) is 0 ±1(?) km.

that the zero-edge of the Thule Supergroup does not cross southern Nares Strait at any point and, for this reason, the marker must be rejected as a constraint for motion on a southerly-striking fault zone potentially located in the channel.

Kane Basin magnetic anomalies (marker 18 of Dawes & Kerr 1982, new marker 28)

A region of distinctive magnetic anomalies (Fig. 1) was identified in the Dawes & Kerr (1982) summary paper as extending from eastern Kane Basin to the Bache Peninsula region of eastern Ellesmere Island without obvious offset. Subsequent analysis of these anomalies by Hood et al. (1985) and Okulitch et al. (1990) indicated the possibility of a 25 km sinistral offset on two or more northerly striking fault planes proposed to exist east of Bache Peninsula in southern Kane Basin (see Fig. 15). The problem with these analyses is the fact that bathymetry was not taken into account. Not considered was the possibility that magnetic anomalies may be generated from shallow dipping beds having a sinuous surface trace around sea floor topographic features. Also not considered was the possibility that large apparent offsets could be generated by small amounts of vertical displacement.

In light of the important results arising from the detailed BGR-GSC aeromagnetic survey presented by Damaske & Oakey (2006), it is clear that at least one Neo-Proterozoic dyke (the Kap Leiper Dyke, named and described by Dawes et al. 1982) can now be traced aeromagnetically from a coastal outcrop in western Inglefield Land to the offshore Pim Island area, close to the Ellesmere coast (Fig. 16, new marker 28). Other aeromagnetic lineaments, notably one associated with the Leffert Glacier Dyke, may also extend across the channel and link with either of two mafic dykes located near Kap Inglefield in western Inglefield Land (Fig. 16; Dawes & Garde 2004). However, there is a critical gap in the aeromagnetic coverage (Damaske & Oakey 2006). Since the Kap Leiper Dyke is dated to the Neoproterozoic (Dawes et al. 1982), allowable strike slip offset through this part of southern Nares Strait must be close to zero since that time with little room for uncertainty or widely distributed strain.

INTERPRETATION

The results of the present re-assessment of the Dawes & Kerr (1982) geological-geophysical markers are illustrated on Figure 17, and summarized in bar graph form in Figure 18. The maps show the position of all quantitatively evaluated markers where each intersects, on the Greenland side, the trace of the Judge Daly Fault System in the north, and the hypothetical trace of the Wegener Fault running through Smith Sound and Kane Basin in the south. While the northern part of the fault system is tightly constrained by aeromagnetic data and onshore geological maping on northeast Ellesmere Island, fault alignment in the south is based entirely on conventional wisdom which locates the fault in mid-channel. However, compelling evidence has been presented in this paper, and elsewhere in this volume, to indicate that there can be no actual fault zone running through Smith Sound.

Displacement on markers that cross the Judge Daly Fault System (JDFS)

The maps and bar graph clearly indicate that the northern markers, all twelve of those located north of Kane Basin (marker 11 and others to the left on Fig. 18), including the corresponding ten markers discussed by Dawes & Kerr (1982), allow for some sinistral strike slip motion on the JDFS. The upper limiting value of 25 km, estimated by Dawes & Kerr (1982) for marker 10, is raised to 65+35-65 km in the present re-interpretation due primarily to the highly sinuous nature of the Silurian facies change. Therefore, within the ranges of assessed uncertainties, at least eight of twelve northern markers (5-10, 12, 16, possibly also 19 and 22) permit between 65 km and 75 km of sinistral strike slip motion on the significant mapped faults located within the Cenozoic and Paleozoic orogens of the northern Nares Strait region.

It may be significant that there is a progressive decrease in the determined sinistral displacement in a south-westward direction measured over a distance of approximately 180 km. This is observed on four markers (6, 10, 11 and 12) between northern Judge Daly Promontory, where there is 69 ±5 km of offset on marker 6, to the mouth of John Richardson Bay where left-hand displacement is 25 ±25 km on marker 11 (Fig. 18). The trend is complete near the foreland limit of the Ellesmere-Greenland fold belt and Cenozoic orogen; a point located northeast of marker 24 where there is potentially zero lateral displacement on the limit of continuous Paleozoic cover. Space does not permit the full evaluation of the significance of these observations. However, Harrison (in press) has suggested that Cenozoic sinistral slip motions on the JDFS are balanced by south-easterly directed restraining bend thrust displacements north of Bache Peninsula on central Ellesmere Island, and that there is a progressive decrease in lateral offset on the JDFS as slip is transferred westward on various thrusts.

For the thermal maturity data (marker 21), it is important to recall from the previous description of these data that the large sinistral displacements implied by the upper limit of thermally mature conditions in Cambrian and Ordovician strata (Fig. 11) could also be produced by differential subsidence across the JDFS during the peak of burial metamorphism, presumably having occurred in the Middle and Late Devonian. The result would have been a differentially thickened load of Devonian foreland clastic formations in the northwest, direct evidence of which has been entirely stripped from the report area. The possibility that the JDFS is an ancient and periodically reactivated structure is also indicated by the substantially greater thickness of Lower Cambrian clastic formations in all areas northwest of this zone (see marker 8, above), and the fact that other linear features, notably the Silurian facies change (marker 10), and the extent of the Ellesmerian fold belt (marker 12) may have been indirectly influenced by this zone of weakness. These results confirm and support earlier reports of Okulitch et al (1990), and Soper & Higgins (1990) who suggested that faults and lower Paleozoic facies changes along and parallel to Kennedy Channel are part of a long-lived zone of weakness, possibly initiated during extensional development of the Franklinian basin in the Neo-Proterozoic.

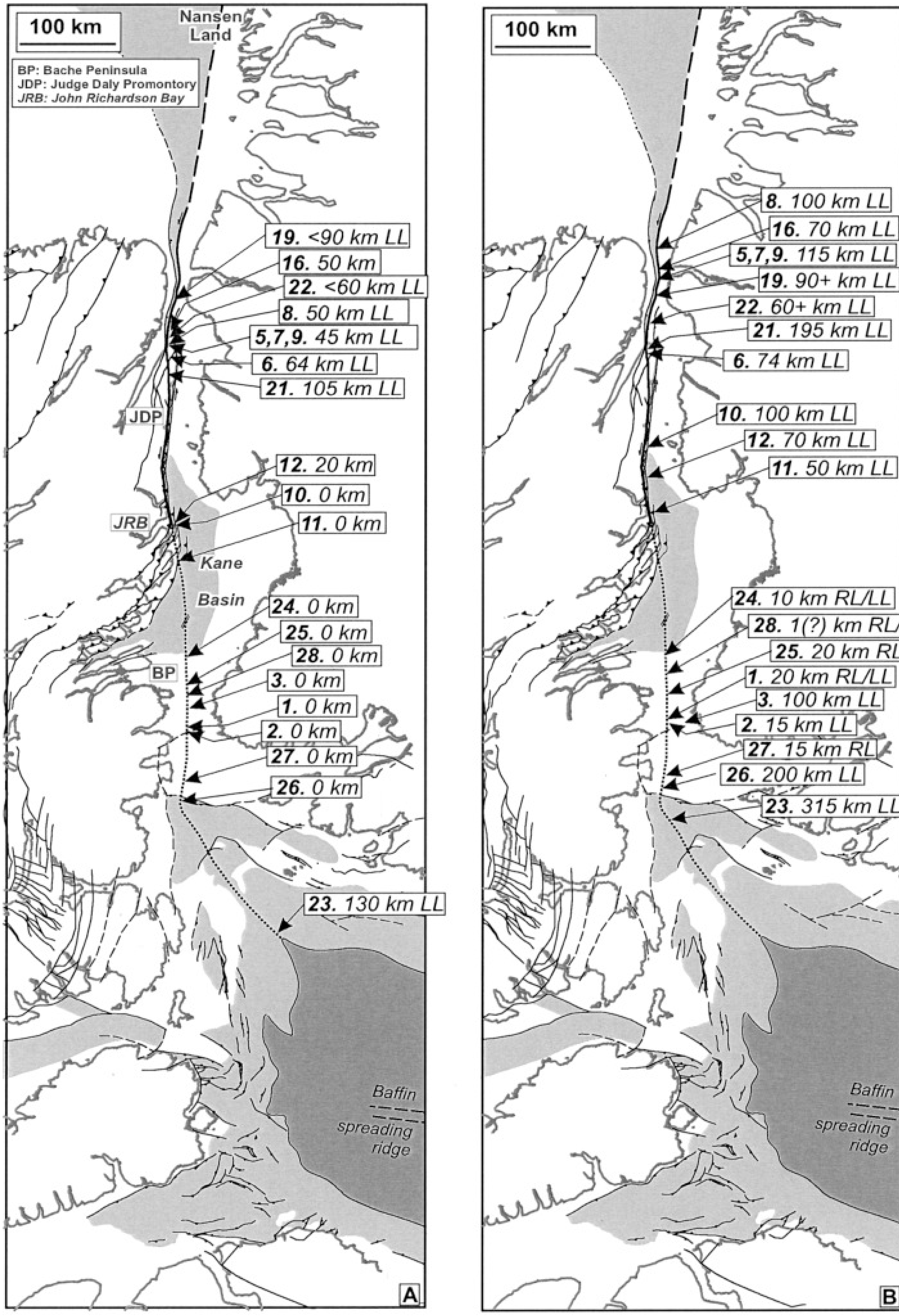


Fig. 17: Minimum (A) and maximum (B) possible offsets of quantitatively evaluated geological-geophysical markers, plotted where these markers intersect the Judge Daly Fault System and the hypothetical surface trace of the Wegener Fault based on conventional wisdom.

Displacement on markers that cross southern Nares Strait

Eight of the identified nine markers, here quantitatively examined (or re-examined) for lateral offsets in Smith Sound and northern Baffin Bay, allow the possibility of no displacement. Six of these eight indicate less than 20 km of displacement in any direction. Especially convincing is the evidence presented by Damaske & Oakey (2006), that at least one Neoproterozoic mafic dyke (marker 28) extends from Greenland to inshore Ellesmere Island without any recognizable post-emplacement displacement. Considering these data, together with observations provided by three other key Precambrian markers (1, 2, 27), there is only one conclusion. No through-going strike-slip fault exists between Ellesmere Island and Greenland in southern Nares Strait.

The last marker to be considered (marker 23) is the maximum and minimum amount of transform plate boundary strike-slip motion demanded by the width of oceanic (or "oceanized") Baffin crust as measured between the ocean-ward limits of

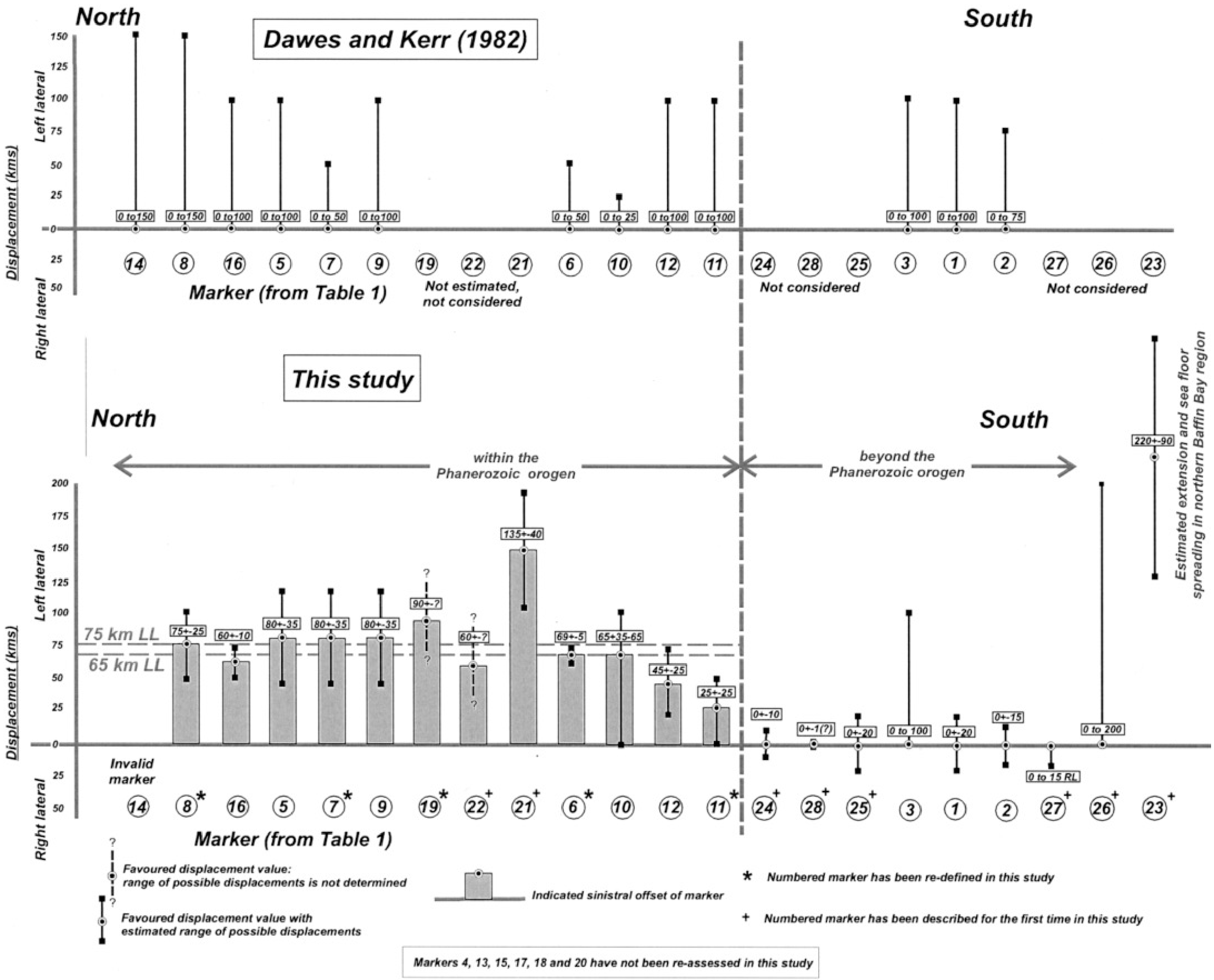


Fig. 18: Bar graph summarizing the average apparent horizontal offset (measured in km) of quantitatively evaluated geological-geophysical markers.

stable continental lithosphere. As previously discussed (marker 23, above) these constraints must account for 220 ±90 km of sinistral strike slip motion between Canada and Greenland at the northwest limit of the Baffin spreading ridge east of Devon Island. It would now appear to be geologically impossible to transfer any of this displacement through Smith Sound.

Where are the plate boundaries?

If the plate boundary between Arctic Canada and Greenland does not pass through Smith Sound then where is it to be found? The spreading ridge in central Baffin Bay, defined by free-air gravity, is approximately aligned with the northern shore of Lancaster Sound (Fig. 2; CHALMERS & PULVERTAFT 2001). Although at least one major extensional (or transtensional) fault defines the northern limit of Cretaceous-Tertiary strata in Lancaster Sound there is no obvious link between this

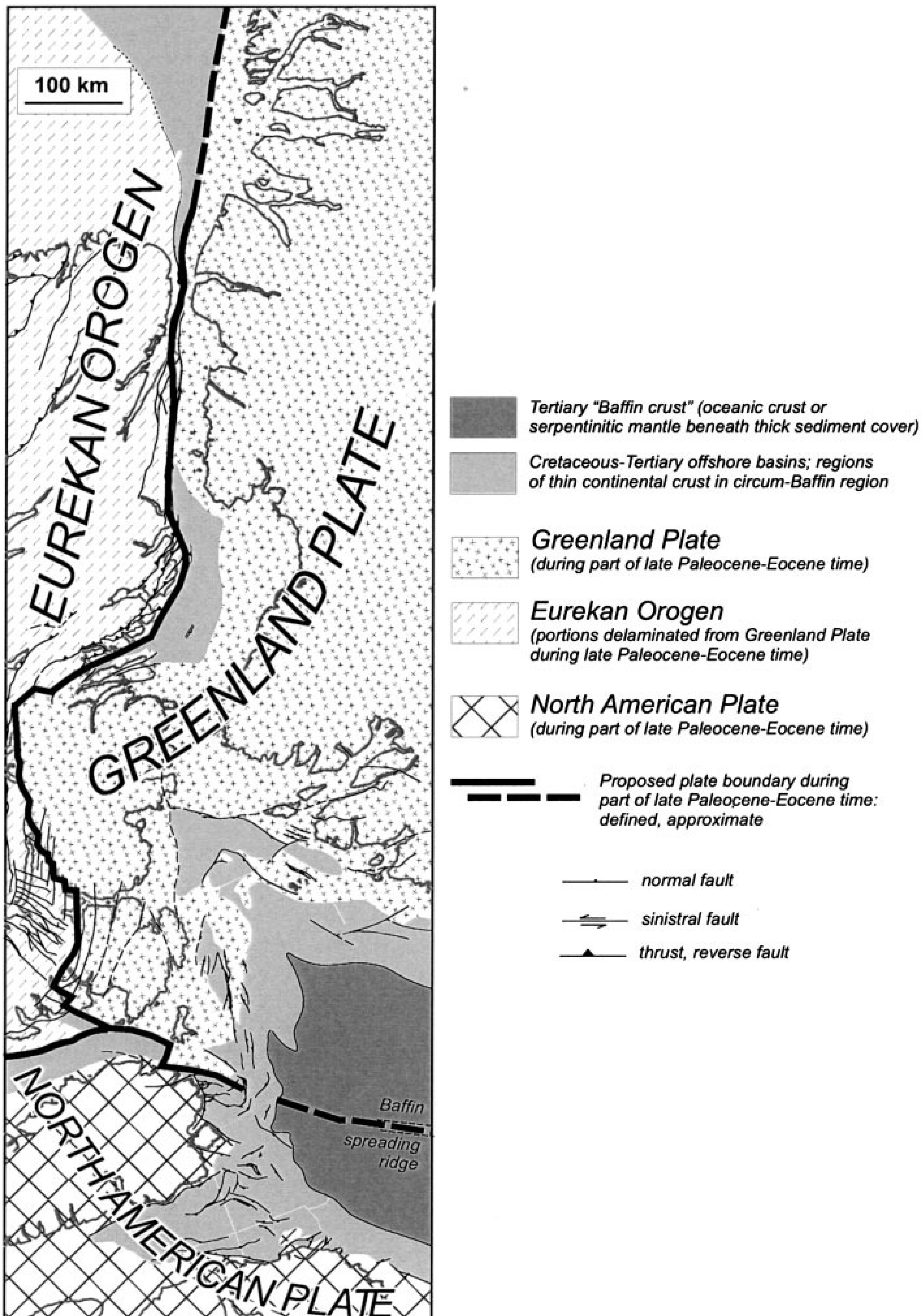


Fig. 19: Suggested tectonic boundaries for Greenland, North America and the Eurekan Orogen during part of late Paleocene – Eocene time.

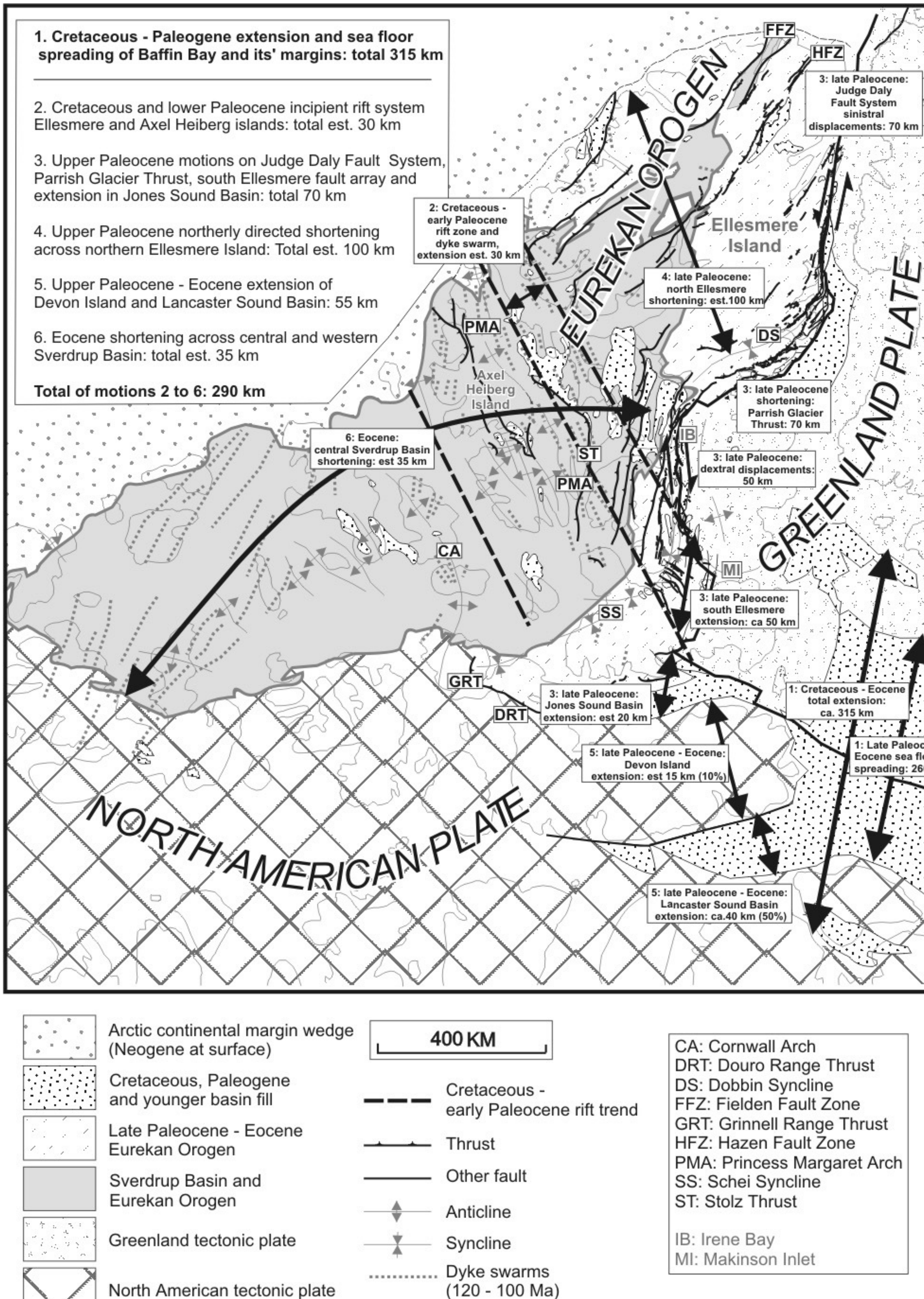


Fig. 20: Generalized reconciliation of plate motions between Greenland and North America during the opening of Baffin Bay and construction of the Eurekan Orogen. The Cretaceous (120-100 Ma) dyke swarm is reproduced from Buchan & Ernst (2004).

structure and the Eurekan Orogen which lies almost entirely north of Jones Sound (Fig. 20).

Jones Sound provides a second candidate alignment for the North America-Greenland plate boundary. A Cretaceous-Tertiary basin exists in this channel (Fig. 2; Harrison, Brent et al. 2006). However, there are very little seismic data here and significant features, such as extent and thickness of basin fill, remain uncertain (Harrison, Brent et al. 2006). Consideration of residual gravity data, after removing the marine layer, indicates that Jones Sound Basin probably has less accumulated sediment fill than Lancaster Sound Basin (G. Oakey pers comm. 2004). Likewise, Paleoproterozoic marbles, quartzites and related gneissic foliations of southeastern Ellesmere Island and Coburg Island appear to form a continuous uninterrupted arc across the mouth of Jones Sound to northern Devon Island (Fig. 14; Harrison 1984, Frisch 1988). Nevertheless, there is a better connection of basin-bounding faults in Jones Sound to the Eurekan Orogen. The linkage is either through an array of north-westerly-striking faults exposed on southern Ellesmere Island (Okulitch et al. 1990), or via western Jones Sound to the Douro Range and Grinnell Range thrusts of north-western Devon Island (Fig. 20; Eisbacher 1998). From the fault array of southern Ellesmere Island, there are direct spatial links to a system of upper Paleocene and younger thrusts and en echelon strike slip faults of the Eurekan Orogen (Figs. 2, 20). These easternmost Eurekan structures have been mapped in detail from the head of Makinson Inlet northwards along the west side of the shield to north and east of the head of Irene Bay at the western end of the Parrish Glacier Thrust (Thorsteinsson et al. 1991, Harrison et al. 1991). It is compelling to believe that a plate boundary may link the Baffin spreading ridge to the JDFS via faults of southern Ellesmere Island and the Parrish Glacier Thrust. This would then place on the Greenland Plate: 1) all of south-eastern Ellesmere Island, 2) bedrock of Smith Sound, Kane Basin, and Kennedy Channel and 3) a coastal sliver of Judge Daly Promontory (Fig. 19).

One problem with these ideas is that the JDFS can only accommodate between 20 and 50 % of the sea floor spreading identified in Baffin Bay. A second problem is that the northwesterly-striking fault zone of southern Ellesmere Island, being only about 100 km wide and generally empty of Cretaceous-Paleocene rift-fill, is unlikely to accommodate more than about 50 km of extension. A solution may follow from ideas partly developed by Harrison et al (1999) and Harrison (in press). Briefly stated, the Baffin rift and transform fault system is seen to have originally extended to Axel Heiberg Island and north-western Ellesmere Island beginning in the Cretaceous (Fig. 20). The evolution of the Eurekan Orogen then witnessed the successive footwall emergence of thrust structures within the dextral transpressive Fielden Fault Zone (Figs. 2, 20) followed by those located in the Lake Hazen Fault Zone and finishing with thrust strands in the JDFS (Rawlings Bay Thrust, Scoresby Bay Thrust and finally the Parrish Glacier Thrust). This model indicates that a much larger part of Ellesmere Island may have once been part of the Greenland Plate prior to the Eurekan Orogeny but that crustal delamination caused the plate boundary to shift southeastward and down-section through time.

There are independent lines of evidence to indicate that convergence of Greenland with Arctic North America also produced a counter-clockwise rotation of northern Ellesmere Island (Wynne et al. 1983, 1988, Harrison in press) possibly related to a pole of rotation on south-eastern Ellesmere Island. This was an idea first entertained by Kerr (1981). Early formed thrusts, such as the Rawlings Bay Thrust, are understood to have evolved into surfaces of sinistral strike slip during the later stages of the Eurekan Orogeny (Harrison in press). Similarly, the north-western end of the Cretaceous rift-transform system may have closed as northern Ellesmere Island escaped to the west between the dextral-sense Fielden Fault Zone and the sinistral JDFS, and impinged on the buttressing western shoulder of the former Cretaceous rift under Axel Heiberg Island (i.e. Princess Margaret Arch, Fig. 20). The semi-independent counter-clockwise rotation of northern Ellesmere Island can also account for the north-westerly trending fan of regional folds in the western Arctic Islands (Okulitch & Trettin 1991). Many of these folds are detached on Carboniferous evaporites, a unit of profound structural weakness that has permitted a greatly expanded Eurekan Orogen from Axel Heiberg Island to the western Sverdrup Basin.

Implicit in this model is that the north-westerly-trending Baffin rift system in the Arctic Islands was subjected to partial or widespread inversion throughout the late Paleocene and into the Eocene. The maximum 50 km of extension across southern Ellesmere Island can then be understood as the net result of a larger component of early extension now lost to subsequent thrust inversion. Can inversion of this rift zone account for the 220 ±90 km of extension required to explain the width of upper Paleocene-Eocene oceanic crust in Baffin Bay and the Cretaceous-early Paleocene extension of its' continental borderlands? An attempted accounting is summarized in the following conclusive statements.

PROPOSED REGIONAL SOLUTIONS

1. This paper has stated that the width of upper Paleocene-Eocene ocean floor indicated across northern Baffin Bay is 260 km (Fig. 20). To this is added 55 km of Cretaceous – early Paleocene extension having occurred on the north Baffin and northwest Greenland continental margins. The sum of these values (315 km) is to be sought by reconciling Cretaceous through Eocene plate motions within the Arctic Islands from Lancaster Sound to northern Ellesmere Island and westward to Axel Heiberg Island across the Eurekan Orogen. Here we ignore all extension which may have occurred south of Lancaster Sound to Hudson Strait south of Baffin Island as this deformation need only be only identified with the reconciliation of spreading estimates for Labrador Sea.

2. The northwest-trending Cretaceous-lower Paleocene continental rift-transform system is assumed to have been once continuously active from south-eastern Ellesmere Island through central Ellesmere Island to eastern Axel Heiberg Island and north-westernmost Ellesmere Island (Fig. 20). Related features at the north end of the rift system include a north northwesterly-striking120 to 100 Ma tholeiitic dyke swarm and voluminous mafic to alkaline volcanic flows in Lower and Upper Cretaceous strata (Buchan & Ernst 2004, Embry & Osadetz 1988).

Measured eastwards from the axial line of the Princess

Margaret Arch, the proposed continental rift zone is estimated to be currently 130 km wide. It is assumed to have been formerly 160 km wide and to have accommodated 30 km of Cretaceous – early Paleocene extension to partly match the 55 km of extension estimated for the north Baffin – West Greenland continental margins around northern Baffin Bay. This allows 25 km of progressive reduction in extension between the north Baffin and Axel Heiberg – northwestern Ellesmere portions of the rift, assuming the rift zone once had an overall northward narrowing tendency from its widest extent in southern Labrador Sea.

The remainder of the model, below, attempts to reconcile the observed 260 km of late Paleocene-Eocene sea floor spreading in northern Baffin Bay.

3. The Judge Daly Fault System (JDFS) can accommodate 70 km of sinistral strike slip motion (Fig. 20). This is matched by 70 km of north – south shortening on the Parrish Glacier thrust and all related restraining bend Eurekan thrusts located between Bache Peninsula and the axial line of Dobbin Syncline (Harrison in press). It is assumed that this was coincident with and kinematially linked to 50 km of dextral wrench faulting and normal faulting on the array of northerly-striking strike slip faults and northwesterly-striking normal faults that lie onshore on Ellesmere Island southeast of the Schei Syncline (Thorsteinsson et al. 1991, Okulitch et al. 1990); add to this a further 20 km (or about 30 %) kinematically linked extension in Jones Sound Basin.

Subtracting this (50 + 20 = 70 km from the 260 km that need to be found, leaves a remaining shortfall of 190 km.

4. The Sverdrup Basin is approximately 400 km wide as measured from its southern margin on south-western Ellesmere Island to its northern margin at the north point of Axel Heiberg Island. There are no westerly-striking Eurekan thrusts along this transect and therefore no obvious north-to-south telescoping of the basin. In contrast, the width of the Sverdrup Basin is only about 300 km wide as measured from its southern margin north of Bache Peninsula to its northern margin located near the northwest coast of Ellesmere Island. Westerly-striking structures on this transect include numerous thrust faults and the Lake Hazen Fault Zone which can probably accommodate 30-35 km of south-directed shortening. If the basin was originally 400 km wide along this second transect, then there has been 100 km or 25 % shortening of the Sverdrup Basin during the Eurekan Orogeny (Fig. 20). An admissible structural cross-section would help constrain this estimate.

Combining this 100 km of shortening with the 70 km previously identified from Jones Sound Basin, southeast Ellesmere Island and Judge Daly Fault System leaves a remaining shortfall of 90 km from the total 260 km that are required.

5. Late Paleocene – Eocene extension is also carried westward into the central Arctic Islands through Lancaster Sound and Devon Island (Fig. 20). For Lancaster Sound, 40 km of extension is assumed or about 50 % of the present 80 km width of the basin. The Devon Island array of easterly-striking faults can probably accommodate a further 15 km of extension or about 10 % measured over a north-south distance of 130 km. Devon Island and Lancaster Sound Basin can, therefore, together account for 55 km of extension. This leaves only 35 km unaccounted from the original total.

6. Counter-clockwise rotation of northern Ellesmere Island during the progressive indentation of Greenland caused elements of the north Ellesmere block to escape to the west. This was facilitated by dextral sense slip on the Fielden Fault Zone in the north and sinistral sense motion on the Judge Daly Fault Zone in the southeast (Fig. 20). Compressional structures related to the escape motion of northern Ellesmere Island include the entire fan of northerly- to northwesterly-striking folds from western Axel Heiberg Island to the western end of the Sverdrup Basin (Okulitch & Trettin 1991). These folds, however, can account for only limited shortening. More significant are northerly- to northwesterly-striking thrusts that have been mapped throughout eastern Axel Heiberg and northwestern Ellesmere Island (Okulitch & Trettin 1991). The largest of these is the Stolz Thrust which can accommodate about 25 km of east-directed shortening (Fig. 20). This leaves 10 km to be found on all other folds and small thrusts.

Some of the north-south thrust faults are candidate structures for the proposed Cretaceous – lower Paleocene continental rift-transform fault system, once continuous through to the continental margins of Baffin Bay and Labrador Sea. In this case, much of the east-directed shortening resulting from the counter-clockwise rotation of northern Ellesmere Island may have served to invert the Cretaceous – lower Paleocene rift.

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