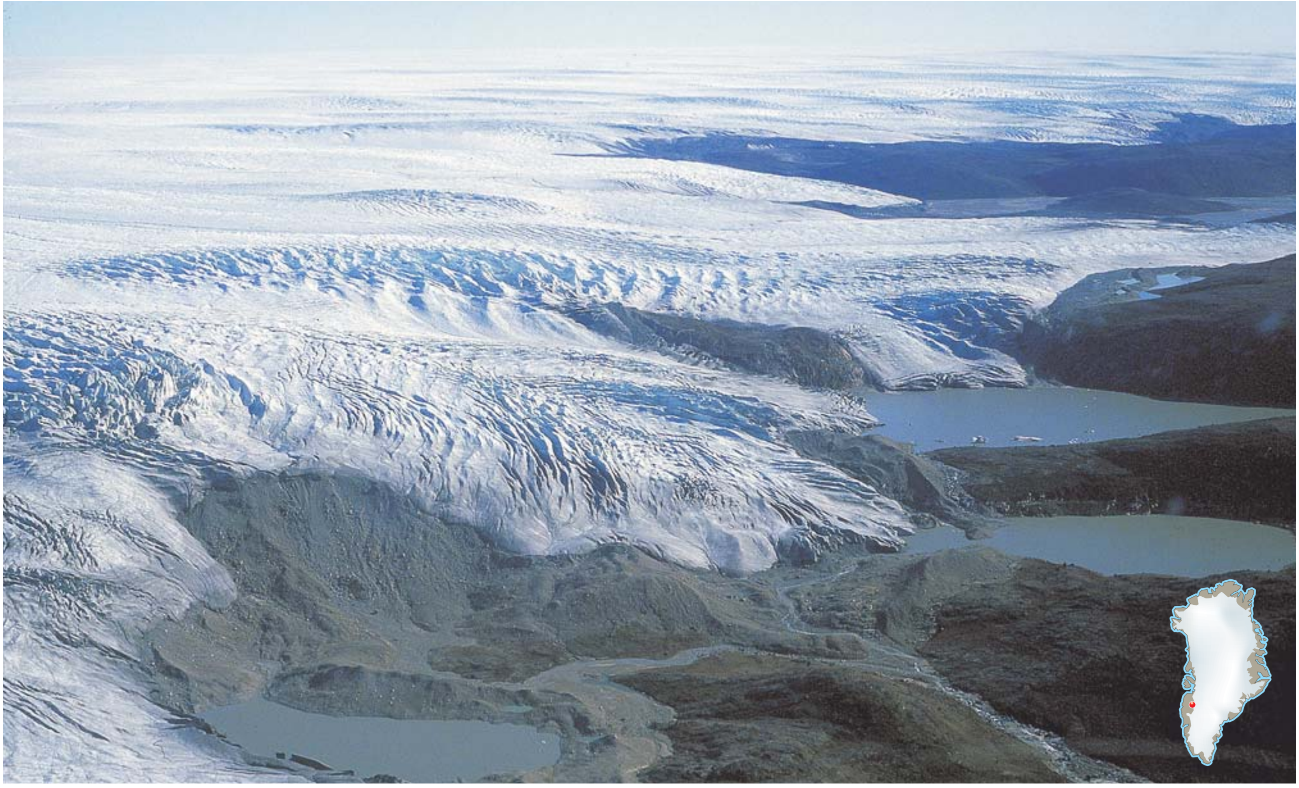


Fig. 46. Characteristic front of the Inland Ice abutting the ice-free land area, with moraines and small lakes. The distance from the bottom of the picture to the land area in the background is approximately 5 km. The locality is about 75 km north-north-east of Søndre Strømfjord airport, southern West Greenland, at *c.* 67°30′N. View is towards south. Photo: H. Højmark Thomsen.

Glaciology

The present ice cover of Greenland is a relic of the Pleistocene ice ages. It consists of the large continental ice sheet (the Inland Ice), and local ice caps and glaciers (Weidick 1995).

The Inland Ice has an area of *c.* 1 707 000 km^2 and reaches an altitude of 3230 m with a maximum thickness of 3420 m. The local ice caps and glaciers cover areas of *c.* 49 000 km^2 (Weng 1995). The volume of the Inland Ice has been estimated at 2 600 000 km^3, based on ice thickness measurements by airborne radio-echo sounding; a rough estimate of the volume of local ice caps and glaciers is 20 000 km^3. On the map, surface contours, isopachs of ice thickness and contours of the bedrock below the Inland Ice are shown.

Mean annual air temperatures on the Inland Ice range from –30°C over a large region in its central and northern parts to about –5°C in its south-western marginal areas. The temperature of the ice ranges between –32° and 0°C; with increasing depth, the temperature generally increases due to geothermal heat flux and internal heating caused by deformation. In some locations, the temperature at the base of the ice sheet may reach its melting point.

Mass balance

The mass balance (budget) of the Inland Ice is the difference between accumulation (of snow in the interior region mainly) and ablation by melting and calving of icebergs in the marginal areas.

The accumulation of snow decreases from south to north from more than 2000 mm water equivalent/year in coastal areas in the south-west to 100 mm water equivalent/year or less in interior north-eastern areas (Ohmura & Reeh 1991). Melt rates also decrease from south to north. Away from the coast in South-West Greenland, the annual melting of the ice at sea level probably reaches values near 10 000 mm water equivalent. However, even along the northernmost margins of the Inland Ice significant melting occurs; melt-rate models predict values near 2000 mm water equivalent/year at sea level. Calving glacier fronts producing icebergs are generally located at the heads of fjords at some distance from the outer coast. The most concentrated source region for icebergs is central West Greenland (Disko Bugt and the area between Nuussuaq and Svartenhuk Halvø) where about 100 km^3 of calf ice are produced annually.

The effects of climate change in recent years on the mass balance of the Greenland ice sheet have been documented by satellite gravity measurements. Over the four years 2004–2007 the ice sheet lost an average of *c.* 400–500 km^3 in the summer period of each year and only gained *c.* 250–350 km^3 of snow in the winter. The net result is a loss of *c.* 150 km^3/year from the beginning of the 20th century (Witze 2008), although some researchers estimate even larger figures for the present net loss.

Past climate and environment

Up to 2009 five deep ice cores have been retrieved by drilling through the Inland Ice (one drilling was only to a depth of 1400 m), and these have provided considerable information about climate and environmental variations during the past 150 000 years. The ice-core records indicate that in central Greenland the Inland Ice survived the last interglacial (the Eemian) which culminated about 125 000 years ago, without completely disappearing even when the climate was several degrees warmer than at present. However, according to ice-dynamic model calculations of the evolution of the Inland Ice, the ice cover in northern and southern Greenland was less extensive during the Eemian (Fig. 47).

The ice-core records indicate dramatic temperature fluctuations during the last ice age, which lasted from about 115 000 years ago to about 11 700 years ago. In the coldest parts of this period, temperatures in Greenland may have been 10–12°C colder than now, whereas temperatures in other periods of the ice age were only about 5 degrees colder (Dansgaard 1997; Hammer 1997).

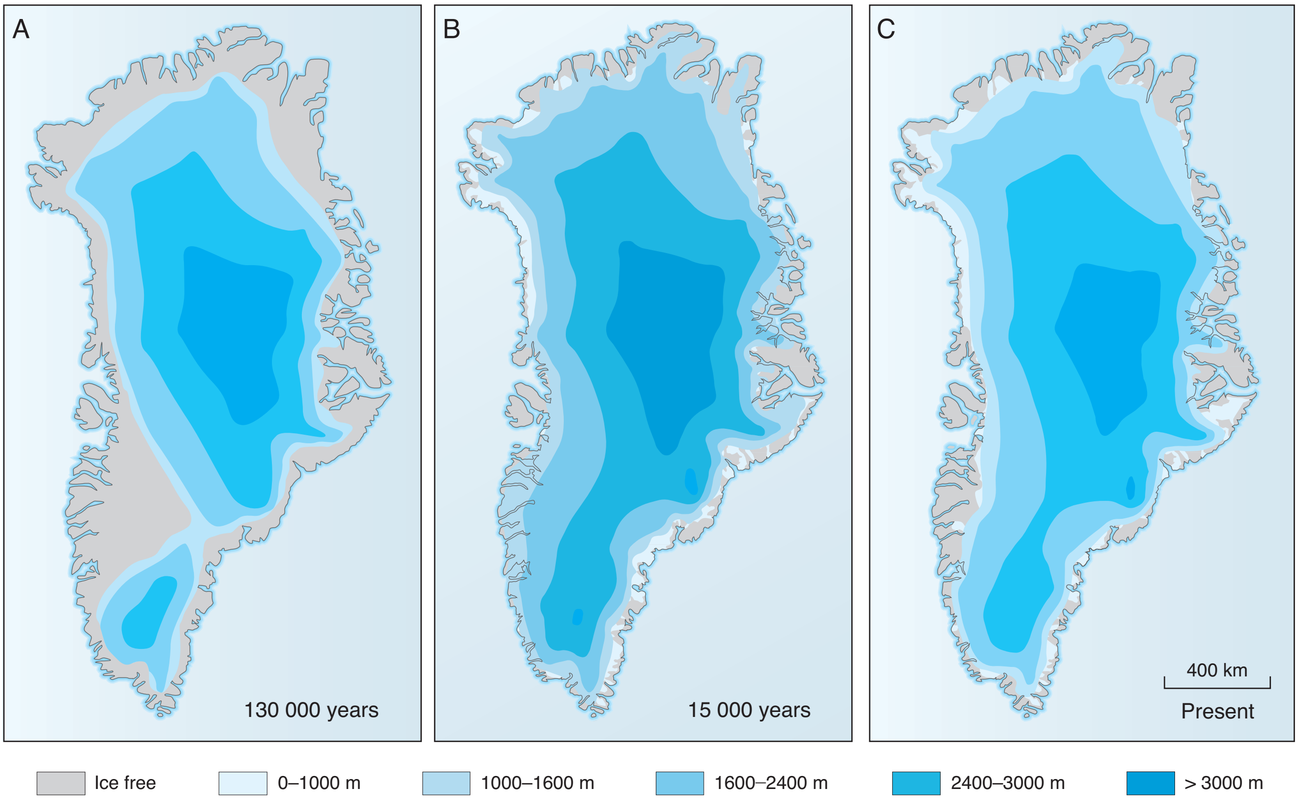


Fig. 47. Models of the Inland Ice with indication of thickness of the ice sheet in metres. **A**: The last interglacial (the Eemian) with a temperature 4–5°C higher than the present. **B**: During the late glacial maximum (Weichselian) with a temperature 10–12°C colder than at present. **C**: Under the present climatic conditions. From model calculations by Letréguilly *et al.* (1991). The models do not include the offshore extent of the ice, only that of present land areas.

Offshore geology

The interpretation of the offshore geology shown on the 1:2 500 000 scale map was based mainly on seismic surveys carried out between 1970 and 1992, supplemented by aeromagnetic and gravimetric data and, in the case of southern West Greenland, by data from five exploration wells drilled in 1976–77. Offshore South-East Greenland six holes were drilled in 1993 at *c*. 63°N, constituting Leg 152 of the Ocean Drilling Program (ODP; H.C. Larsen *et al*. 1994a); the positions of three of these wells are shown on the map. In 1995 three more holes were drilled off South-East Greenland as part of the aborted ODP Leg 163, but no results were available when the map went to press (H.C. Larsen *et al*. 1996). However, the coverage of geophysical data in different areas was, and still is, uneven, and is dependent on ice conditions. Off southern West Greenland, where there are only scattered icebergs and no pack ice in the late summer and early autumn, more than 37 000 km of seismic data were acquired by the oil industry in the 1970s and a further 10 259 km of non-exclusive data were acquired in this area in the years 1990–1994 (Pulvertaft 1997). In contrast, the often ice-infested areas off East, North-East and North-West Greenland were only covered by reconnaissance surveys, principally as a result of the KANUMAS and North Atlantic D (NAD) surveys. The KANUMAS project was a marine seismic reconnaissance financed by six major oil companies, with the Greenland-Danish national oil company Nunaoil A/S as operator (H.C. Larsen & Pulvertaft 1990; Pulvertaft 1997). In the 1990s KANUMAS surveys acquired *c*. 7000 km of seismic data off North-East and central East Greenland, and *c*. 4000 km of data off North-West Greenland; although these data are confidential company data, some results had been released in time to be included in the 1:2 500 000 map. The North Atlantic D project was a combined aeromagnetic and seismic survey of the East Greenland shelf carried out by Grønlands Geologiske Undersøgelse (GGU) in 1979–83. During this project *c*. 8000 km of seismic data were acquired off central East and South-East Greenland (Thorning *et al*. 1982; H.C. Larsen 1985). In Nares Stræde (Nares Strait) and off North Greenland, where no seismic data existed, interpretation of the geology was based on aeromagnetic and sparse gravity data alone. Aeromagnetic and shipborne magnetic data constituted the main source of information in oceanic areas.

The 1:2 500 000 map was designed to show two general aspects of offshore geology: (1) the extent of continental crust [**a**], oceanic crust [**c–g**], and of the intervening, poorly understood, transition zone [**b**] and (2) the distribution of sedimentary basins and major faults. Where extensive volcanic units are known to occur in areas underlain by continental crust, their distribution is also shown.

Since compilation of the 1:2 500 000 map a large amount of new data has been acquired in the maritime regions surrounding Greenland, both by the industry and by academic research institutes. It is clear from these data that the map is not correct in many places. In the following text, attention is drawn to the known errors in the map, and as more data are released, there will no doubt be need for further corrections. The distribution of crustal types offshore as now understood (2009) is shown in Fig. 49A, p. 68, while offshore and onshore sedimentary basins are shown in Fig. 56.

The continental margin off East and North Greenland

East Greenland south of 77°N

In general terms, the continental margin off East Greenland between the southern tip of Greenland and 76°N can be described as a volcanic rifted margin (H.C. Larsen 1990; H.C. Larsen *et al*. 1994a, b), formed when Greenland became separated from northern Europe at the start of sea-floor spreading in early Eocene time (magnetochron 24R). Between *c*. 68°N and the Jan Mayen Fracture Zone, however, Greenland remained attached to the Jan Mayen microcontinent (Fig. 49A, B) until Oligocene time when spreading shifted from the Aegir Ridge to the Kolbeinsey Ridge.

The position of the continent–ocean boundary (COB) was drawn on the basis of aeromagnetic data supplemented by characteristic features in the NAD reflection seismic data. The absolute seawards (eastern) limit of continental crust cannot overlap areas where linear magnetic anomalies characteristic of oceanic crust can be identified with confidence. Along the entire volcanic rifted margin seaward-dipping reflectors can be seen in the seismic data. These arise from subaerial lava flows or groups of flows that were erupted in the early stages of sea-floor spreading prior to differential subsidence below sea-level. In connection with the seaward-dipping reflectors, buried volcanic escarpments may occur. These are landward-facing escarpments formed at the landward end of the dipping reflectors, where lava flows interfinger with sedimentary rocks (Fig. 48; H.C. Larsen & Jakobsdóttir 1988; H.C. Larsen 1990).

The zone off East Greenland shown on the map as underlain by transitional crust [**b**] was drawn in a rather arbitrary manner, at least with regards to its width. This zone is thought to consist of attenuated and fragmented continental crust with increasing numbers of dykes and other intrusions as oceanic crust is approached. Much of the onshore coastal area around and south of Kangerlussuaq (68°N) is very intensely intruded by Palaeogene coast-parallel dyke swarms (Nielsen 1978; Klausen & Larsen 2002; not shown on the map). Aeromagnetic data indicate that these dyke swarms continue southwestwards under the shelf as far south as 63°N (H.C. Larsen 1978). This intensity of dyke intrusion suggests the proximity of the continent–ocean boundary, i.e. the outer edge of the transition zone.

Since the map was printed, results of intensive research carried out off southern East Greenland along Leg 152 of the Ocean Drilling Program (ODP sites 914–919, *c.* 63°N) have been published (H.C. Larsen & Saunders 1998; H.C. Larsen *et al.* 1998). Results of this research indicate that the continent–ocean boundary, defined here as the point at which thinned, intensely dyked continental crust finally gives way to a sheeted dyke complex, is situated in this area about 12 km landwards of the shelf break (H.C. Larsen & Saunders 1998 fig. 12). The shelf break here is the edge of a thick prograding wedge of glaciomarine sediments. The inner boundary of the continent–ocean transition zone, i.e. the point at which extensional faulting intensifies and marked attenuation of continental crust begins, lies 25–40 km landwards of the continent–ocean boundary (H.C. Larsen & Saunders 1998 fig. 12; H.C. Larsen *et al.* 1998 fig. 7). Thus the continent–ocean transition zone may be a little wider than shown on the 1:2 500 000 map, and the continent–ocean boundary probably lies about 25 km north-west (landwards) of the position shown on the map.

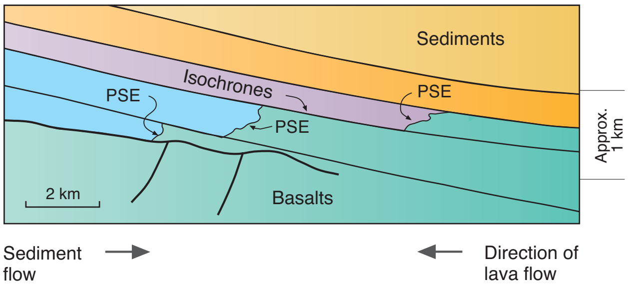
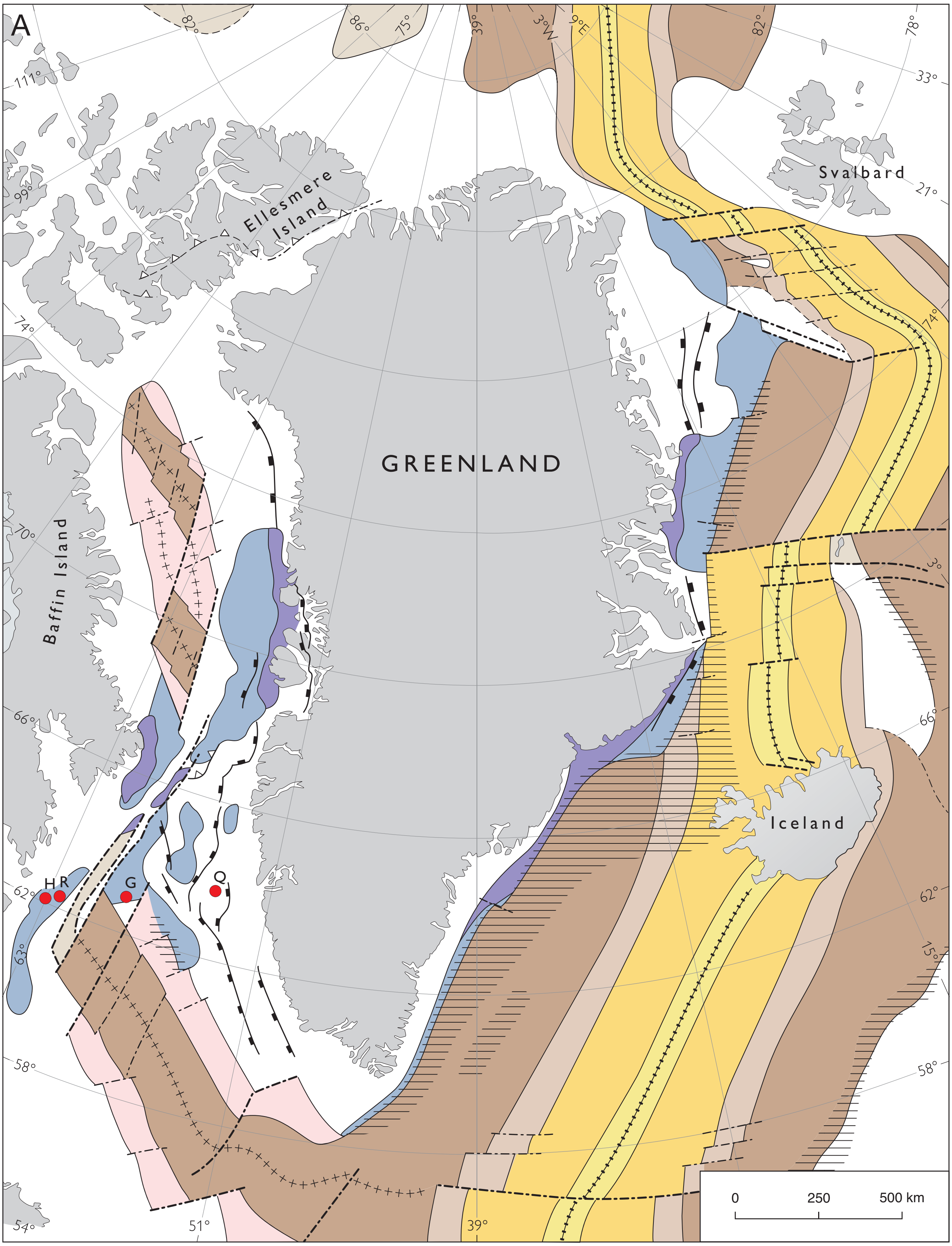


Fig. 48. Cross-section based on seismic section, illustrating the formation of so-called pseudo-escarpments (**PSE**) at the landward end of dipping basalts. **Sediments**: pale blue to light brown layers; **basalts**: green. Landward direction to the left. Slightly modified from H.C. Larsen (1990).

At *c.* 68°N the eastern margin of continental Greenland cuts obliquely across linear magnetic anomalies 24–13 in the oceanic crust. This was not originally regarded as the expression of a transform fault, but rather as an oblique ocean–continent transition along a former northward-propagating spreading ridge (H.C. Larsen 1988). However, prior to anomaly 13 time, the Jan Mayen microcontinent was attached to East Greenland between *c.* 68° and 72°N. A coast-parallel dyke swarm along Blosseville Kyst between 68°20′ and 70°N and voluminous intraplate volcanism in this region may reflect an unsuccessful attempt at continued Eocene spreading along an axis at about the position of the present coast (H.C. Larsen 1988). However, to find the 'missing' magnetic anomalies 24–13 (i.e. Eocene) oceanic crust, it is to the east of the Jan Mayen micrcontinent that one should look (e.g. Lundin & Doré 2002). During this period a transform fault must have linked the Reykjanes Ridge to the southern end of the Aegir Ridge – the Denmark Strait Fracture Zone (Lundin & Doré 2002). After anomaly 13 time spreading between Greenland and the Jan Mayen microcontinent propagated northwards, reaching the Jan Mayen Fracture Zone at about anomaly 6 time.

North of Jan Mayen Fracture Zone the 1:2 500 000 map shows the COB off East Greenland transgressing magnetic anomalies 24B, 24A and 23 at a low angle, indicating that here the spreading ridge propagated towards the south-west. Newer interpretations of the position of the COB here differ, not only from what is shown on the map but also from one another. Tsikalas *et al.* (2002) extend anomalies 24B and 24A into the shelf, the anomalies crossing the shelf edge at approximately 74°15′N and 73°55′N respectively. This implies that the shelf has prograded over oceanic crust here, and that there was no south-westwards progradation of the spreading axis in this region. However, refraction seismic data from the shelf between 72° and 74°N (Voss & Jokat 2007) show clearly that the boundary of true oceanic crust lies very slightly seawards of what is shown in the 1:2 500 000 map, while the transition zone, described by Voss & Jokat as "intruded and stretched continental crust", extends landwards 100 km from the COB. Farther


A
GREENLAND
Ellesmere Island
Baffin Island
Svalbard
Iceland
H
R
G
Q
82°
86°
75°
39°
3°W
9°E
82°
78°
33°
21°
111°
99°
74°
70°
66°
62°
63°
58°
54°
51°
39°
74°
3°
66°
62°
15°
58°
0
250
500 km

Legend, facing page:

- ⬜ Land areas. Maritime areas underlain by continental or transitional crust
- 🟨 Oceanic crust, Pliocene–Recent
- 🟧 Oceanic crust, Miocene
- 🟫 Oceanic crust, Oligocene
- 🟫 Oceanic crust, Eocene
- 🟪 Oceanic crust, Paleocene
- ⬜ Oceanic crust, Age uncertain
- 🟪 Palaeogene basalts exposed at seabed in areas underlain by continental or transitional crust
- 🟦 Palaeogene basalts buried under younger sediments in areas underlain by continental or transitional crust
- ☰ Seaward-dipping reflectors
- +++++ Oceanic spreading axis, active
- + + + Oceanic spreading axis, extinct
- –·– Transform fault
- ▬ Major basin – bounding fault; mark on downthrown side
- △ Compressional fault, thrust
- ● Exploration well; Q: Qulleq-1, G: Gjoa G-37, R: Ralegh N-18, H: Hekja 0-71

Fig. 49. **A** (*facing page*): Map showing distribution of crustal types and major structures offshore Greenland. Only basalts overlying continental crust are shown. **B** (*above*): Place names and names of structural features shown in **A** or mentioned in the text. **Lo.R.**: Lomonosov Ridge; **M.J.R.**: Morris Jesup Rise; **Y.Pl.**: Yermak Plateau; **S.F.Z.**: Spitsbergen Fracture Zone; **M.F.Z.**: Molloy Fracture Zone; **G.R.**: Greenland Ridge; **Jan Mayen F.Z.**: Jan Mayen Fracture Zone; **Jan Mayen Mc.**: Jan Mayen Microcontinent; **U.F.Z.**: Ungava Fracture Zone; **D.S.H.**: Davis Strait High; **'N.W.B.'**: 'North Water Bay'; **S.Sd.**: Smith Sound; **K.K.**: Kennedy Kanal; **B.P.** Bache Peninsula; **C.D.**: Cape Dyer.

north Berger & Jokat (2008), on the basis of reflection seismic data, place the landward boundary of oceanic crust a little seawards of the position shown on the enclosed 1:2 500 000 map, the difference between Berger & Jokat's interpretation and the 1:2 500 000 map diminishing south-westwards.

North-East Greenland (77–83°N)

At the Greenland Ridge (see Fig. 49A, B) a sliver of continental crust extends from the shelf south-eastwards along the fracture zone until this turns slightly anticlockwise. North-east of the fracture zone there is an area tentatively interpreted as extremely thinned continental crust (Døssing *et al.* 2008); this is included in the area shown as continental or transitional crust in Fig. 49A. North of the fracture zone the position of the COB steps to the north-west. Sea-floor spreading magnetic anomalies have lower amplitudes and are more ambiguous in the oceanic area north of the fracture zone and in the Fram Strait than in the oceanic region to the south. At the time the 1:2 500 000 map was compiled, there was no agreed interpretation of magnetic anomalies, and hence the age of oceanic crust between the Hovgård Ridge (position *c.* 0°W, 78°30′N, shown as transitional crust) and the Spitsbergen Fracture Zone (see Fig. 49A, B) is shown as [**g**] (age unspecified) on the map. Recently, however, Engen *et al.* (2008) have reviewed the geology and evolution of the Fram Strait and provisionally identified and numbered several sea-floor spreading magnetic anomalies in the seaway, so that the oceanic crust in the strait can now be subdivided chronostratigraphically as it is elsewhere (Fig. 49A).

Between the Greenland Ridge and the Molloy Fracture Zone (Fig. 49A, B) the oldest sea-floor spreading magnetic anomaly that can be identified with confidence is anomaly 18 (39 Ma; Middle Eocene) (Engen *et al.* 2008). Regarding the COB here, Døssing *et al.* (2008) place the seaward limit of continental crust slightly east of where it is shown on the 1:2 500 000 map, close to the shelf edge.

The north-east margin of continental Greenland, north of 79°N, has a very different character. It is shown on the enclosed map as a former intracontinental transform plate boundary. A consensus exists that a substantial dextral displacement of Svalbard relative to North-East Greenland took place along an intracontinental NW–SE megashear, the De Geer megashear (Harland 1969; Engen *et al.* 2008), in the time interval corresponding to magnetochrons 24R–13 (earliest Eocene – earliest Oligocene). In the early Oligocene, rifting began to take place along this zone, and a spreading ridge, the Knipovich Ridge, linking the Mohns Ridge and the Gakkel Ridge, developed along the site of the earlier transform fracture. Since this time, the flanking continental margins have developed as passive margins separated by an obliquely spreading ocean (e.g. Vogt & Tucholke 1989; Eldholm *et al.* 1990; Kristoffersen 1990; Engen *et al.* 2008; Faleide *et al.* 2008). Spreading propagated south-eastwards from the Gakkel Ridge so that the earliest identified magnetic anomaly between the Yermak Plateau and the Morris Jesup Rise is anomaly 7, while in the Fram Strait the oldest anomaly approaching the Spitsbergen Fracture Zone is anomaly 5 (Engen *et al.* 2008). It has also been suggested that mantle peridotite has been exposed in this part of the strait (Jokat *et al.* 2005). Whatever the case, the 1:2 500 000 map is incorrect with regard to the age of the oceanic crust in this area.

The Morris Jesup Rise and the Yermak Plateau

One particular outstanding problem here is the nature of the crust underlying the Morris Jesup Rise and its conjugate feature, the Yermak Plateau, north of Svalbard (Fig. 49A, B). Recent work has shown that the Yermak Plateau south of 82°N (the area shown without colour on the 1:2 500 000 map) is most likely underlain by continental crust, and that the south-west margin of this part of the plateau is a continental margin (Ritzmann & Jokat 2003; Engen *et al.* 2008).

North of 82°N the Yermak Plateau runs NE–SW; it is this outer part of the plateau that adjoined the Morris Jesup Rise prior to the opening of the Eurasia Basin and Fram Strait. The nature of the crust in the outer Yermak Plateau is still open to discussion. Jokat *et al.* (2008) favour a model of stretched and intruded continental crust, and Engen *et al.* (2008) argue against the earlier hypothesis that this part of the plateau formed by voluminous oceanic volcanism at an Eocene triple junction between Eurasia, Greenland and North America, pointing out that the termination of sea-floor spreading magnetic anomalies 23–13 at the north-east flank of the outer plateau favours a continental outer Yermak Plateau.

A limited amount of work has been done on the conjugate feature, the Morris Jesup Rise. Available data indicate that the feature is underlain by thinned and rifted continental crust (Ostenso & Wold 1977). As is also the case in the Yermak Plateau, high-amplitude and irregular magnetic anomalies suggest the presence of volcanic rocks, possibly of Cretaceous age (Engen *et al.* 2008). Dawes (1990) favoured the view that the Morris Jesup Rise has a complex structure and that it may contain appreciable continental remnants below a thick cover of volcanic rocks.

The continental margin off West Greenland

Southern West Greenland (58–64°N)

The continental margin off southern West Greenland also presents problems of interpretation, although more reflection seismic data are available from this area. There has also been some controversy as to when sea-floor spreading in the Labrador Sea began, as discussed in the following, but all agree that sea-floor spreading between Greenland and Canada began earlier than in the North Atlantic, and that it had ceased by magnetochron 13 time.

Distinct linear magnetic anomalies can be seen in the Labrador Sea off South-West Greenland (Srivastava 1978). The earliest magnetic anomalies trend NNW–SSE, while the younger anomalies (24 and younger) trend

NW–SE, parallel to an extinct spreading axis roughly midway between Canada and Greenland.

The oldest unambiguous magnetic anomaly in the Labrador Sea is anomaly 27. The crust landward of this is shown as transitional crust on the 1:2 500 000 map. Srivastava (1978) and Roest & Srivastava (1989) indicated that linear magnetic anomalies could be identified much closer to the continental shelf than anomaly 27, suggesting 31 or 33 as the number of the oldest anomaly. However, a substantial body of evidence has accumulated since 1991 that refutes the hypothesis that pre-anomaly 27R oceanic crust exists in the Labrador Sea. Modelling of the magnetic data acquired by the Bundesanstalt für Geowissenschaften und Rohstoffe (BGR), during seismic transects across the Labrador Sea in 1977, indicates oceanic crust with alternating strips of normally and reversed magnetised crust landwards as far as anomaly 26R or 27R (Chalmers 1991; Chalmers & Laursen 1995). Landwards of this, a model assuming thinned and rifted continental crust intruded by reversed magnetised igneous material provides the best fit with the observed data. South of 62°N it appears that serpentinised peridotite subcrops sediments in the outer part of the transitional zone and that the remnants of continental crust are very thin (Chian & Louden 1994; Chalmers 1997). Whatever the case, the seawards limit of normal continental crust off southern West Greenland lies well to the south-west of the continental slope, at water depths of more than 1500 m. This interpretation of the distribution of crustal types is supported by the structural pattern seen in the seismic lines. Both the normal continental crust and the zone of transitional crust show large tilted fault blocks overlain by syn- and post-rift sediments (Chalmers & Pulvertaft 2001). The oldest sediments are most likely of Early Cretaceous age (see later).

Definitive evidence concerning the crustal structure in the inner Labrador Sea was obtained when the Qulleq-1 exploration well was drilled in 2000 at 63°49´N, 54°27´W. According to Srivastava & Roest (1999 fig. 5), this should lie on magnetic anomaly 33 (73.6–79 Ma, middle Campanian; Cande & Kent 1995), but the well terminated in Santonian sandstone without reaching basement (Christiansen *et al.* 2001), showing that the basement here must be older than oceanic crust anywhere in the North Atlantic region north of 56°N. The crustal structure in the inner Labrador Sea has recently been studied in a reflection/refraction seismic transect running WSW from the Qulleq-1 well to the Gjoa G-37 and Hekja 0-71 wells off eastern Canada. This new seismic line shows that oceanic crust is restricted to two narrow zones at or close to the Ungava Fracture Zone, a major transform zone that transferred sea-floor spreading through the Davis Stræde (Davis Strait) into Baffin Bugt (Baffin Bay; Fig. 49A, B; Funck *et al.* 2007). Funck *et al.* believe that these zones of oceanic crust formed when phases of transtension along the fracture zone created gaps which were filled with melt that formed new oceanic crust.

Davis Stræde (*c.* 64–69°N)

The crust under Davis Stræde is estimated to be 22 km thick (Keen & Barrett 1972), which is intermediate between the thickness of normal oceanic and continental crust. There is a basement high within the strait (the Davis Strait High), where the sedimentary cover is thin and locally virtually absent. Palaeogene volcanic rocks have been recovered from the high (Srivastava 1983; Williamson *et al.* 2003), and it has been suggested that the high is a volcanic plateau formed by hotspot volcanic activity. Hotspot activity is also suggested by the occurrence of thick picritic lavas of Paleocene age in the Disko – Nuussuaq – Svartenhuk Halvø area in central West Greenland and at Cape Dyer (Fig. 49B) on the south-east side of Baffin Island in Canada (Clarke & Upton 1971; Clarke & Pedersen 1976). On the Greenland side of the strait, strata of Late Cretaceous age have been traced on seismic lines westwards from the Ikermiut-1 exploration well (66°56´N) onto the eastern flank of the high (Chalmers *et al.* 1995), indicating that the high is formed of pre-Late Cretaceous rocks. An E–W seismic profile across the high west of the Ikermiut-1 well shows dipping internal reflectors within the high (Gregersen & Bidstrup 2008), suggesting the presence a sedimentary unit. The very high abundance of Ordovician carbonates in dredge samples collected on the high suggests that this unit consists of Ordovician limestones (Dalhoff *et al.* 2006), a possibility supported by the occurrence of Ordovician carbonates in wells offshore Labrador (Bell & Howie 1990) and onshore at 'Fossilik' *c.* 65 km east of Maniitsoq in West Greenland (Stouge & Peel 1979; [9] on the 1:2 500 000 map). The crust under Davis Stræde is therefore interpreted by the Survey as being formed of thinned continental crust, in accordance with the interpretation of the distribution of crustal types farther south.

Baffin Bugt (69–77°N)

The distribution of crustal types underlying Baffin Bugt has not yet been mapped out with certainty, not least because linear sea-floor spreading magnetic anomalies are not distinct in this region. In the central, deep part of Baffin Bugt refraction seismic experiments have shown that the crust is very thin, the M (moho) discontinuity lying only 11 km below sea-level. The thickness of the cover of sediments exceeds 4 km in all but the southernmost part of this region. Seismic velocities in the crust below these sediments are in the range 5.7–7.0 km/sec (Srivastava *et al.* 1981; Balkwill *et al.* 1990), in agreement with those known from oceanic layers 2 and 3. Gravity and magnetic evidence (see next paragraphs) is also consistent with the interpretation of the central part of Baffin Bugt as being underlain by oceanic crust (e.g. Balkwill *et al.* 1990). However, geophysical evidence has been presented which indicates that in north-western Baffin Bugt continental crust is replaced oceanwards by a layer of serpentinised mantle, which would account for the lack of distinct linear magnetic anomalies in this area (Reid & Jackson 1997). The weakness of linear sea-floor spreading magnetic anomalies could also be attributed to very oblique spreading (see Roots & Srivastava 1984) and to the dampening effect of the thick sedimentary cover.

In spite of the weakness of the magnetic anomalies, Oakey (2005) has succeeded in interpreting linear anomalies between 69° and 73°N and divided these into two directions, NNW–SSE and NW–SE, corresponding respectively to the Paleocene and Eocene anomalies in the Labrador Sea. Oakey's interpretation has been accepted in the recently published 1:5 000 000 map of the entire polar region north of 60°N (Harrison *et al.* 2008), on which the oceanic crust in Baffin Bugt is divided into crust of Paleocene and Eocene age. The same interpretation is shown in Fig. 49A. The extinct spreading axis located by Chalmers & Pulvertaft (2001) is the axis of Eocene spreading. Oakey (2005) agrees with Chalmers & Pulvertaft (2001) that the major transfer faults along which the spreading axis was displaced, trend *c.* N–S.

In the absence of easily recognisable sea-floor spreading magnetic anomalies, the landward limit of proven oceanic crust cannot yet be placed with any degree of confidence. Existing released geophysical data are not sufficient to allow any interpretation of the position, width and nature of the continent–ocean transitional zone. On the 1:2 500 000 map the area underlain by continental crust was delineated on the evidence of crustal thickness and structural style (large extensional faults and rotated fault blocks), the latter being known from the KANUMAS reconnaissance seismic survey (Whittaker *et al.* 1997). In southern Baffin Bugt a recently acquired NW–SE refraction seismic line has revealed a material with high velocity (6.8 km/s) underlying sediments at *c.* 68°40´N, east of the N–S transform fault running N–S at *c.* 60°W (T. Funck, personal communication 2009). This could be oceanic crust, but at present neither the age nor the north–south extent of this high velocity layer is known.

Nares Stræde (77–82°N)

The nature of the geological structure underlying Nares Stræde, the linear seaway separating Greenland from Ellesmere Island (Fig. 49B), has for some time been a controversial subject (Dawes & Kerr 1982; Tessensohn *et al.* 2006). Geophysicists have argued that the strait is the site of a major transform fault with a left-lateral displacement of more than 100 km that accommodated the opening of the Labrador Sea and Baffin Bugt during the Palaeogene, and also of movement normal to the strait (e.g. Srivastava 1985). In contrast, geologists familiar with the surrounding onshore geology find that there is no significant lateral displacement in Smith Sund (Smith Sound). Several geological markers correlate perfectly from the Thule-Inglefield Land area in Greenland to south-east Ellesmere Island south of the Bache Peninsula (Dawes 2009a). However, in the Kennedy Kanal (kanal = channel) and Robeson Kanal and along the north-west coast of Kennedy Kanal, both thrusting and sinistral strike-slip faulting have been observed (Tessensohn *et al.* 2006), and a strike-slip displacement of up to 70 km is considered possible (Harrison 2006). The apparent contradiction between what can be interpreted in Kennedy Kanal and what is observed across Smith Sund can be resolved if south-east Ellesmere Island is geologically speaking part of the Greenland plate. Assuming this is the case, the tectonic junction between Greenland and Ellesmere Island runs along the north-west side of Kennedy Kanal as far south as 80°N and then turns inland just north of Bache Peninsula, where the strike-slip displacement in and along the north-west coast of Kennedy Kanal is transferred into thrusting along the southern front of the Eurekan Orogen (Fig. 49A). The cumulative shortening across the major thrust zones of the Eurekan Orogen has been estimated to be of the order of 100 km (De Paor *et al.* 1989).

Recent work in Baffin Bugt has diminished the likelihood that Nares Stræde is the site of a major transform fault, because 1) the direction of transform fracture zones and hence spreading movement in this area is almost north–south, i.e. at an angle of about 40° to Nares Stræde (see Fig. 49A, B; Wheeler *et al.* 1996; Whittaker *et al.* 1997; Chalmers & Pulvertaft 2001; Harrison *et al.* 2008) and 2) the width of the area of oceanic crust in Baffin Bugt may be much less than previously supposed (Oakey 2005; Harrison *et al.* 2008)). It may well be that Baffin Bugt spreading has been accommodated to a substantial degree by a series of rifts in the Canadian Arctic Islands together with compression in the Eurekan orogen, just as the Gakkel Ridge spreading axis terminates in a system of rifts and extension zones in the Laptev Shelf north of Siberia (Drachev 2000).

The continental margin off North Greenland, west of the Morris Jesup Rise

Ice conditions in Lincoln Hav (the Lincoln Sea) off North Greenland are the most severe anywhere in Greenland waters. Consequently, the only indications of what might underlie the sea were for many years provided entirely by airborne geophysical data. Summarising the results of earlier work in the region, Dawes (1990) concluded that while gravity data from Lincoln Hav suggest that the North Greenland margin is underlain by thinned continental crust, the crustal structure of the offshore region is conjectural. For this reason a wide area is shown without colour on the 1:2 500 000 map.

In 2006 geophysicists succeeded in acquiring two wide-angle refraction/reflection seismic lines on the sea ice north of Ellesmere Island and Greenland (Dahl-Jensen *et al.* 2006). These seismic sections show that thinned continental crust continues northwards from the inner Greenland shelf under a 2000 m deep channel to the southern end of the Lomonosov Ridge. Consequently it is now believed that the entire area shown without colour as 'type of crust unknown' on the 1:2 500 000 map is underlain by thinned continental crust. The >2000 m deep area shown as underlain by oceanic crust older than 45 Ma [**f**] on the 1:2 500 000 map is also likely to be underlain by continental/transitional crust, an inference supported by the fact that linear sea-floor spreading magnetic anomalies identified to the north do not continue into this area (Engen *et al.* 2008). The study by Dahl-Jensen *et al.* (2006) also showed that a previously suspected deep sedimentary basin underlying Lincoln Hav does in fact exist (see later).

Offshore sedimentary basins (Fig. 56)

Reflection seismic surveys have shown that large sedimentary basins occur offshore East Greenland between latitude intervals 67–72°N and 75–80°N. In the intervening area there are extensive Palaeogene basalts below which thick sedimentary successions have been tentatively interpreted, but these cannot be resolved in existing seismic data. Offshore West Greenland there are rift basins with substantial thicknesses of sediment as far north as 76°N, and also smaller basins in southern Nares Stræde. As just mentioned, under Lincoln Hav a major sedimentary basin has recently been identified (Dahl-Jensen *et al.* 2006; K. Sørensen, personal communication 2009).

North-East Greenland shelf (72–80°N)

The existence of thick sedimentary successions on the North-East Greenland shelf was first suggested on the basis of interpretation of aeromagnetic data (Thorning *et al.* 1982; H.C. Larsen 1984). Seismic and gravity data acquired as part of the KANUMAS project confirmed the existence of major sedimentary basins under this shelf, but at the time the 1:2 500 000 map was printed, very few details had been released. Later a summary of the structure and succession in the basins in this shelf has been published (Hamann *et al.* 2005).

The sediments on the shelf were deposited in two basins separated by a basement high, the Danmarkshavn Ridge (Fig. 50A). Judging from the known geology of the Barents Sea, the Norwegian shelf and onshore North-East Greenland, the age of these sediments is likely to be Devonian to Recent, with unconformities in the middle Permian and at the base of the Paleocene. In the inner basin the maximum thickness of the basin fill is *c.* 13 km, and the succession is thought to span the entire period between Devonian and Neogene. A profound unconformity separates the ?Devonian–Cretaceous section from the overlying Paleocene and younger units (Fig. 50B). The presence of a thick salt layer in the deep part of the basin is clearly shown by gravity lows that coincide with diapiric structures seen in the seismic sections (Hamann *et al.* 2005). This salt formation passes into time-equivalent carbonates deposited on the platform to the west. By comparison to the Nordkapp Basin, Norwegian Barents Sea, the salt is inferred to be of Late Carboniferous – earliest Permian age (Hamann *et al.* 2005). A very thick succession also occurs in the outer basin. Here, however, it has not been possible to interpret layers lower

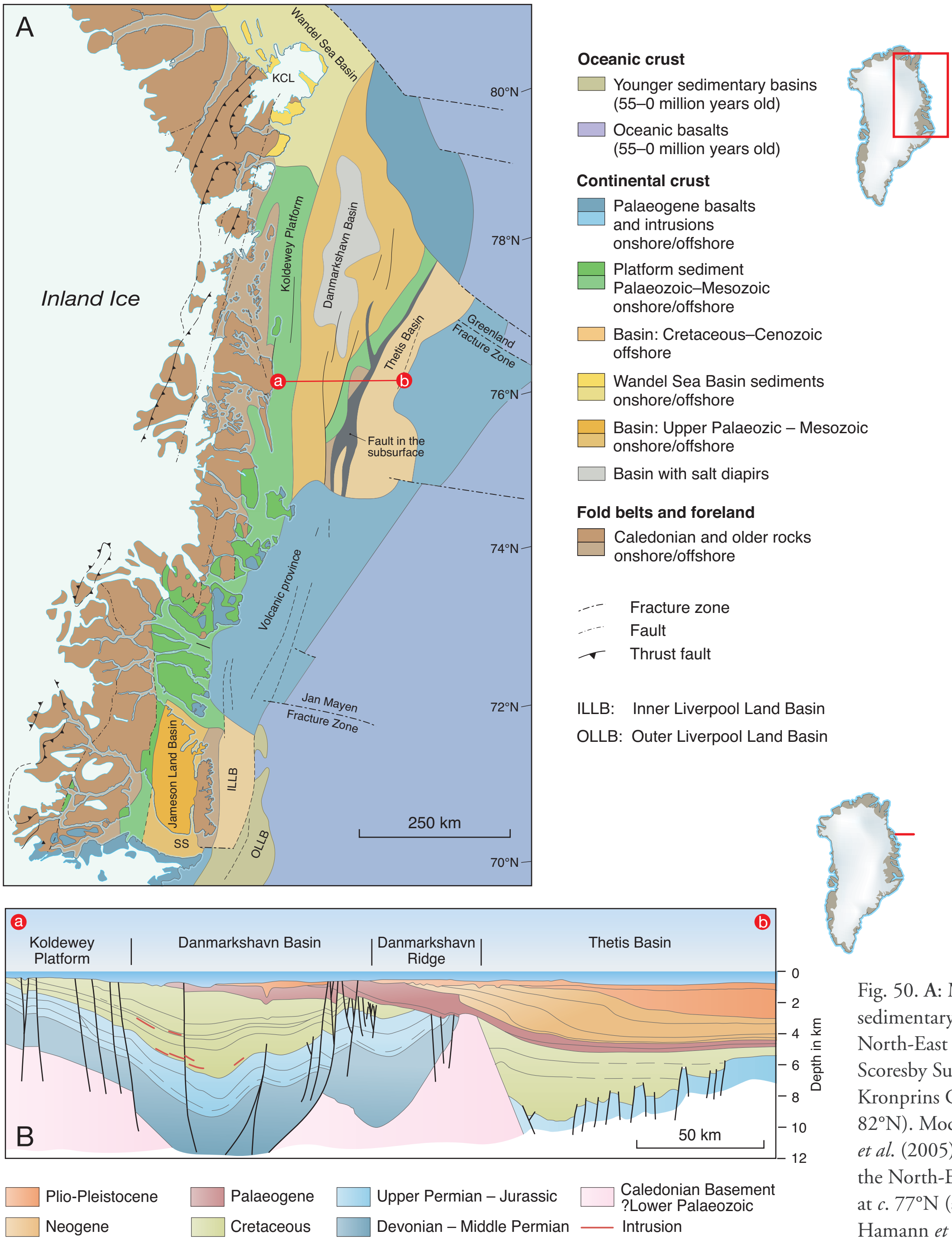


Fig. 50. **A**: Map showing the sedimentary basins offshore North-East Greenland, between Scoresby Sund (**SS**, 70°N) and Kronprins Christian Land (**KCL**, 82°N). Modified from Hamann *et al.* (2005). **B**: Cross-section of the North-East Greenland shelf at *c.* 77°N (a–b in **A**). From Hamann *et al.* (2005).

than ?Jurassic–Cretaceous, although there are indirect indications that older sediments occur here, just as they do in the inner basin.

On the shelf between latitudes 72°15′ and 75°30′N extensive volcanic rocks of presumed Palaeogene age have been interpreted from the aeromagnetic and seismic data (H.C. Larsen 1990). In the near-shore area these are exposed at the seabed; eastwards they become increasingly deeply buried under younger sediments. It is considered almost certain that the pre-Paleocene sediments interpreted to the north and south of this area continue beneath the volcanic rocks. High amplitude magnetic anomalies suggest the presence of igneous intrusions in the sedimentary and volcanic rocks just north of 72°N.

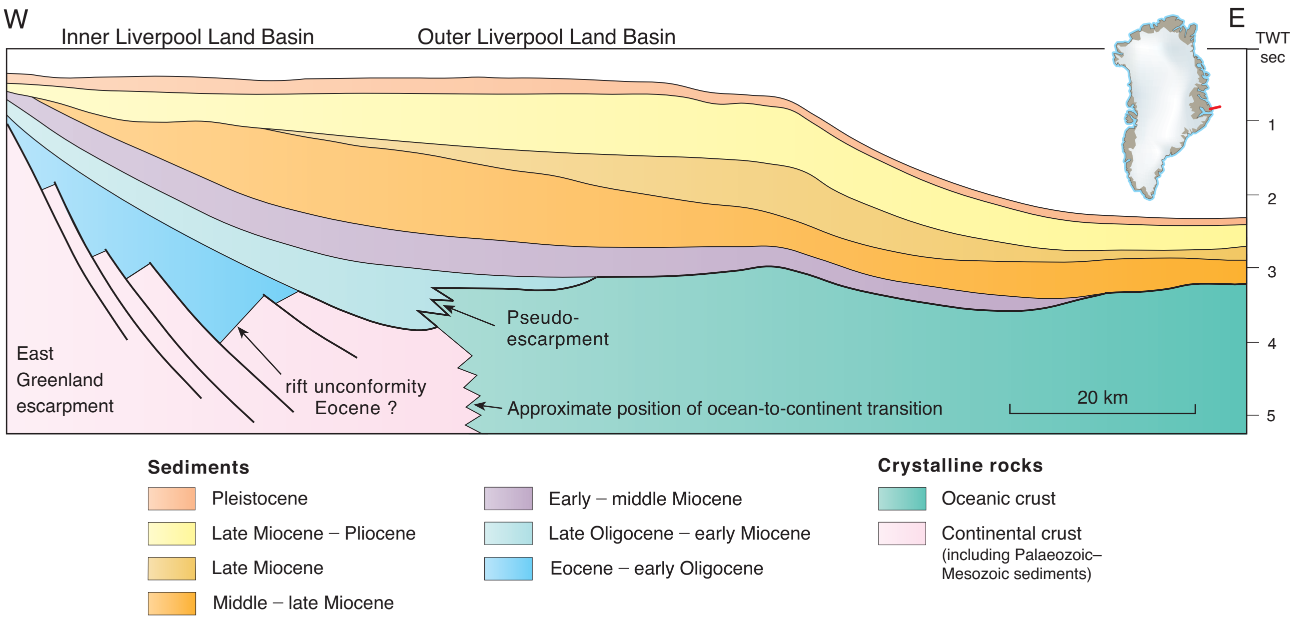


Fig. 51. E–W cross-section across the transition from continental to oceanic crust and the overlying Liverpool Basin at *c.* 71°N, East Greenland. From Larsen (1990 plate 6, profile E).

Liverpool Land Basin, central East Greenland (69°30´–72°N)

A very thick succession of sediments can be recognised in the seismic data offshore Liverpool Land. The sediments are particularly thick within the part of the area underlain by continental crust, where the base of the sediments cannot be identified on the existing data. The upper part of the sedimentary succession is a virtually complete Cenozoic succession up to 6 km thick; this formed a large prograding wedge that spread out across both continental and oceanic crust from the mouth of Scoresby Sund. In the part of the area underlain by continental crust the Cenozoic succession lies with angular unconformity on block-faulted and tilted sediments of pre-Paleocene (Late Palaeozoic – Mesozoic) age (Fig. 51). Where the Cenozoic sediments have prograded into the area underlain by oceanic crust, they are underlain by subaerial lavas seen as seaward-dipping reflectors in the seismic data (Larsen & Jakobsdóttir 1988; H.C. Larsen 1990).

Blosseville Kyst Basin, East Greenland (67–69°30´N)

More than 4 km of post-middle Eocene sediments occupy an elongate, coast-parallel sedimentary basin off the Blosseville Kyst. The sediments lie entirely on Palaeogene basalts. In the area underlain by continental crust there are almost certainly Mesozoic and Paleocene sediments beneath the basalts, as there are onshore and farther to the south (Ocean Drilling Program well 917A; H.C. Larsen *et al.* 1994a). However, it is not possible to interpret the geology underlying the basalts on the basis of existing seismic data.

Southern West Greenland (60–68°N)

Several large, more or less coast-parallel, rift basins occur offshore West Greenland between *c.* 62° and 68°N. Smaller basins south of 62°N have yet to be mapped properly. The earliest sediments that can be confidently interpreted in these basins are pre- and syn-rift sequences up to 3 km or more in thickness, the Kitsissut and Appat sequences (Fig. 52); by analogy with the Labrador Shelf, these are believed to be Early Cretaceous (Barremian–Albian) in age (Chalmers *et al.* 1993; Chalmers & Pulvertaft 2001). The abundance of Ordovician carbonates in dredge samples collected on the Davis Strait High (Dalhoff *et al.* 2006) indicates that Ordovician limestones overlie basement locally in the region. As noted above, this possibility is supported by the occurrence of Ordovician carbonates in wells offshore Labrador (Bell & Howie 1990) and onshore at 'Fossilik' *c.* 65 km east of Maniitsoq in West Greenland (Stouge & Peel 1979; [9] on the 1:2 500 000 map).

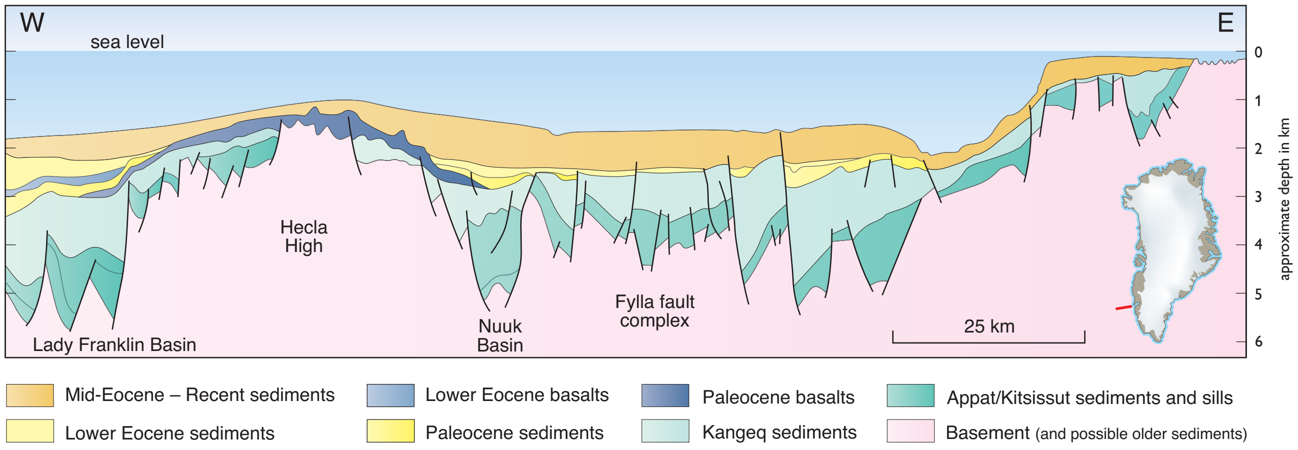


Fig. 52. Zig-zag cross-section through the area west of Nuuk at about 64°N based on interpreted and depth-converted seismic data from an unpublished interpretation by J.A. Chalmers.

The Appat sequence is overlain by a widespread Upper Cretaceous mudstone-dominated succession, the Kangeq sequence (Fig. 52), the upper part of which was penetrated in the Ikermiut-1 well (66°56´N, 56°35´W) and in the Qulleq-1 well (63°49´N, 57°27´W). A major hiatus spanning the interval Campanian – early Paleocene (Nøhr-Hansen 2003) probably reflects the same episode(s) of faulting, uplift and erosion as are recorded in the succession in the Nuussuaq Basin to the north (see pp. 57–60). Following these disturbances, fan sands intercalated with mudstones were deposited. Early Palaeogene volcanism gave rise to local 'shields' – the Maniitsoq and Hecla Highs.

Deposition of mudstones continued into the early Eocene, but from the middle Eocene sedimentation was dominated by coarser clastic sediments deposited mainly in southwards prograding sequences. A major middle Eocene (early Lutetian) hiatus occurs in the three northernmost West Greenland wells (Hellefisk-1, Ikermiut-1 and Kangâmiut-1; Nøhr-Hansen 2003), and another major hiatus spans at least the Oligocene (A.B. Sørensen 2006). During the Eocene, compressional structures developed in the area west of the Ikermiut-1 well as a consequence of transpression along the transform Ungava Fracture Zone, west of the 1:2 500 000 map boundary (Chalmers *et al.* 1993).

A study of dated minor intrusions onshore West Greenland has shown that intrusion forms and melt compositions changed with time, depending on increasing lithospheric attenuation (L.M. Larsen *et al.* 2009). Significantly, this study suggests that by Late Jurassic – earliest Cretaceous time thinning of the lithosphere had reached a stage when sedimentary basins could begin to form in the nascent Labrador Sea.

Central West Greenland (68–73°N)

The Palaeogene basalts exposed onshore in the Disko – Nuussuaq – Svartenhuk Halvø area continue offshore where they have been mapped from seismic and magnetic data over the entire region between latitudes 68° and 73°N (Gregersen & Bidstrup 2008). In the eastern part of this region the basalts outcrop at the seabed and have been sampled by dredging, but to the west they become increasingly buried under a cover of Eocene and younger sediments. While the upper surface of the basalts can usually be mapped easily from the seismic data, the base of the basalts is difficult to interpret and it is uncertain just where the basalts finally thin out and disappear to the west. Nevertheless, from newer seismic data acquired by the oil industry it can be seen that a substantial thickness of Mesozoic sediments underlies the basalts (Gregersen & Bidstrup 2008).

North-West Greenland (73–77°N)

North of 73°N seismic data acquired as part of the KANUMAS project confirmed the existence of a very deep graben or half-graben in the west and south-west part of Melville Bugt (Melville Bay; Fig. 53; Whittaker *et al.* 1997). This had earlier been outlined from aeromagnetic and gravity data acquired in the late 1960s and

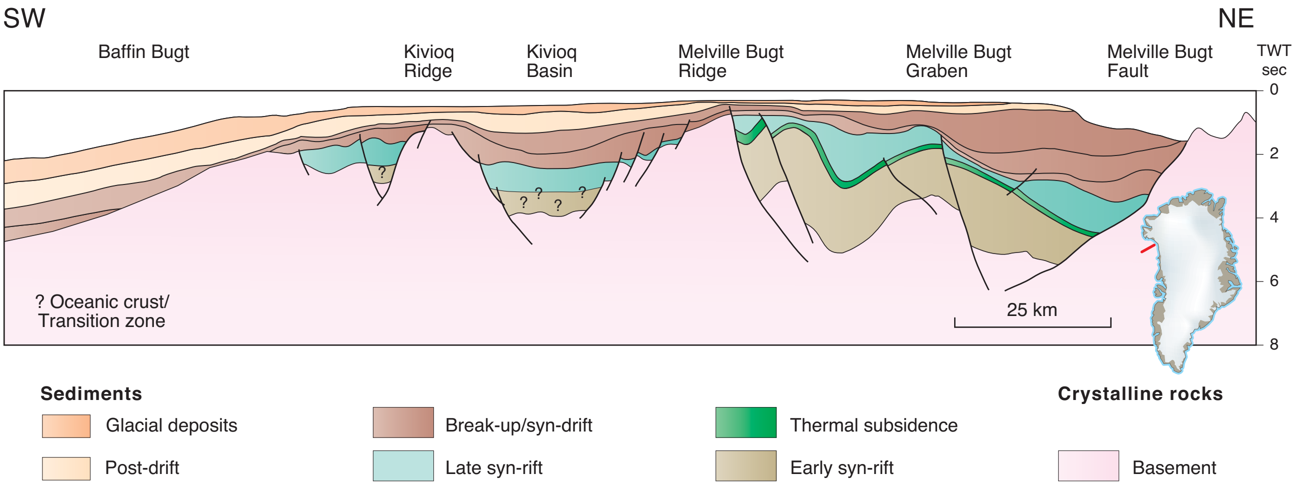


Fig. 53. Representative cross-section offshore North-West Greenland at *c.* 75°N compiled from seismic reflection data. From Whittaker *et al.* (1997).

early 1970s. The more recent data have also revealed several other graben and half-graben structures extending to the northern limit of the survey at 76°30′N. In the Melville Bugt graben the thickness of sediments exceeds 12 km. By analogy with the onshore geology of West Greenland and north-east Canada (Bylot Island), the main phase of rifting is thought to have taken place in the Cretaceous, prior to sea-floor spreading in Baffin Bugt. Later, northern parts of the area were subjected to marked inversion.

Small sedimentary basins with up to 4 km of ?Cretaceous–Neogene sediments occur in an area known as North Water Bay, between Kitsissut (Carey Islands) and Smith Sund (Neben *et al.* 2006). The two easternmost of these basins strike NW–SE; these are on line with extensional basins mapped farther south in northern Melville Bugt (Whittaker *et al.* 1997) and hence are not likely to belong to any pull-apart system (*pace* Neben *et al.* 2006). Sediments belonging to the Thule Supergroup underlie the younger sediments and appear to be continuous right across North Water Bay (Funck *et al.* 2006).

Offshore North Greenland; the Lincoln Sea Basin

As already mentioned, ice conditions in Lincoln Hav are the most severe anywhere offshore Greenland, and until recently airborne geophysical data provided the only hints that a major sedimentary basin lies underneath this sea (Sobczak & Stephens 1974; Kovacs 1982; McMillan 1982). Recently the two wide-angle refraction/reflection seismic traverses successfully carried out by Dahl-Jensen *et al.* (2006) not only established the continental nature of the crust between the Greenland margin and the southern end of the Lomonosov Ridge but also revealed an up to 14 km deep sedimentary basin consisting of two layers interpreted to be part of the Arctic Continental Terrace Wedge, under which there is a 9 km thick layer interpreted to be the offshore continuation of the Upper Palaeozoic – Mesozoic succession in the Sverdrup Basin in the Canadian Arctic Islands.