Glaciology

From the geological section, it is clear that a presentation of the history of the Jakobshavn Isbræ ice stream must also include the history of the surrounding parts of the ice sheet. In the same way, the present major ice streams are a part of the local glaciation history of Disko Bugt.

Subsurface of the ice-sheet margin

The first mapping of large parts of the ice sheet was made in the period 1949–1951 during the 'Expéditions Polaires Françaises' using seismic methods (Holtzscherer & Bauer 1954). Their surveys covered the southern parts of the Inland Ice (south of *c*. 70°N in West Greenland and south of *c*. 72°40′N in East Greenland) and indicated a depression below present sea level in the central part of the ice sheet as well as a drainage channel towards Disko Bugt (Figs 4, 17). Later airborne radar surveys between 1968 and 1976 provided a measure of the thickness of the entire ice sheet (Gudmandsen & Jakobsen 1976; Overgaard 1981). These data provided a realistic impression of the gross features of the subglacial terrain, including highlands, uplands, lowlands, and drainage areas (Figs 4, 17). More recent data on the subsurface of the ice sheet are steadily improving our knowledge of the subglacial landscape (Bamber *et al.* 2001). In general, the elevation of the subglacial terrain falls from the marginal areas towards the central depression of the ice sheet, and as noted above, a depression connects Kangia with the interior region, lying at or below sea level, (Fig. 4).

Ice streams have been defined as "part of an ice sheet, in which the ice flows more rapidly and not necessarily in the same direction as the surrounding ice" (Armstrong *et al.* 1973, p. 26). The strong flow of ice streams, relative to the slower moving ice on either side, leads to strong and chaotic break-up of the ice in ice streams. This limits penetration by radar waves, and hence the determination of ice thickness below ice streams. However, the thickness of Jakobshavn Isbræ has been determined by seismic methods (Clarke & Echelmeyer 1996). In the central parts of this ice stream, the ice thickness varies from *c*. 1.9 km near the grounding zone (cf. Fig. 35) with the ice surface at an altitude of *c*. 500 m a.s.l., to about 2.5 km at a distance of 40 km behind the grounding zone where the ice surface is *c*. 1000 m a.s.l. Thus the bottom of the subglacial trough is found at a depth of *c*. 1.5 km below sea level; the trough can be described as a canyon-like feature about 7 km wide,

surrounded by a hilly subglacial landscape with elevations close to sea level. Much of the basal interface is probably underlain by compacted, non-deformable sediments (Clarke & Echelmeyer 1996). Farther inland, the trough beneath the ice stream gradually levels out. About 120 km east of the grounding zone, the subsurface depression can be interpreted as a shallow subglacial valley. The subsurface map of Fig. 4 also shows that other subglacial areas near the present ice margin in West Greenland are dominated by uplands, with no clear indication of other ice streams of the size of Jakobshavn Isbræ.

Supplementary depth soundings have provided more detailed information around the deep drilling sites in the central parts of the ice sheet and in marginal areas studied in connection with possible exploitation of hydropower resources. For example, a *c*. 800 km^2 area of the ice margin in the Paakitsoq area, north of Ilulissat, was mapped at a scale of 1:250 000 with 100 m contour intervals (Thomsen *et al.* 1988). Other less detailed maps cover smaller areas of the ice margin at Tissarissoq (south of Ilulissat), Alanngorliup Sermia and Saqqarliup Sermia (Thorning *et al.* 1986; Thorning & Hansen 1987; Weidick *et al.* 1990). One revelation from this detailed mapping was that even for ice thicknesses of 600–800 m, the depressions and elevations of the subsurface are reflected in the topography of the ice surface, although in a smoothed and somewhat distorted form. It thus appears that topographical features of the ice-free marginal areas bordering the ice sheet continue beneath the ice, and are readily discernable on Landsat images with a low sun angle (Fig. 36). The deep trough of Jakobshavn Isbræ is clearly visible from its abundant crevasses (dark colour, relative high ablation), in contrast to the other less important ice streams draining into Disko Bugt (Sermeq Kujalleq and Sermeq Avannarleq in Torsukattak icefjord). The continuation of Torsukattak icefjord beneath the ice sheet rapidly levels off into the subglacial uplands of the area, as seen from the depth of the present fjord (*c*. 600 m) and radar soundings 25–30 km from the front (Overgaard 1981); this picture is confirmed by the image of Fig. 4. Restricted ablation and high relief may explain the local high production of calf ice from the glaciers draining into Uummannaq Fjord farther to the north (Fig. 5). Snowfall here is also heavier than in the interior parts of Disko Bugt (Ohmura & Reeh 1991).
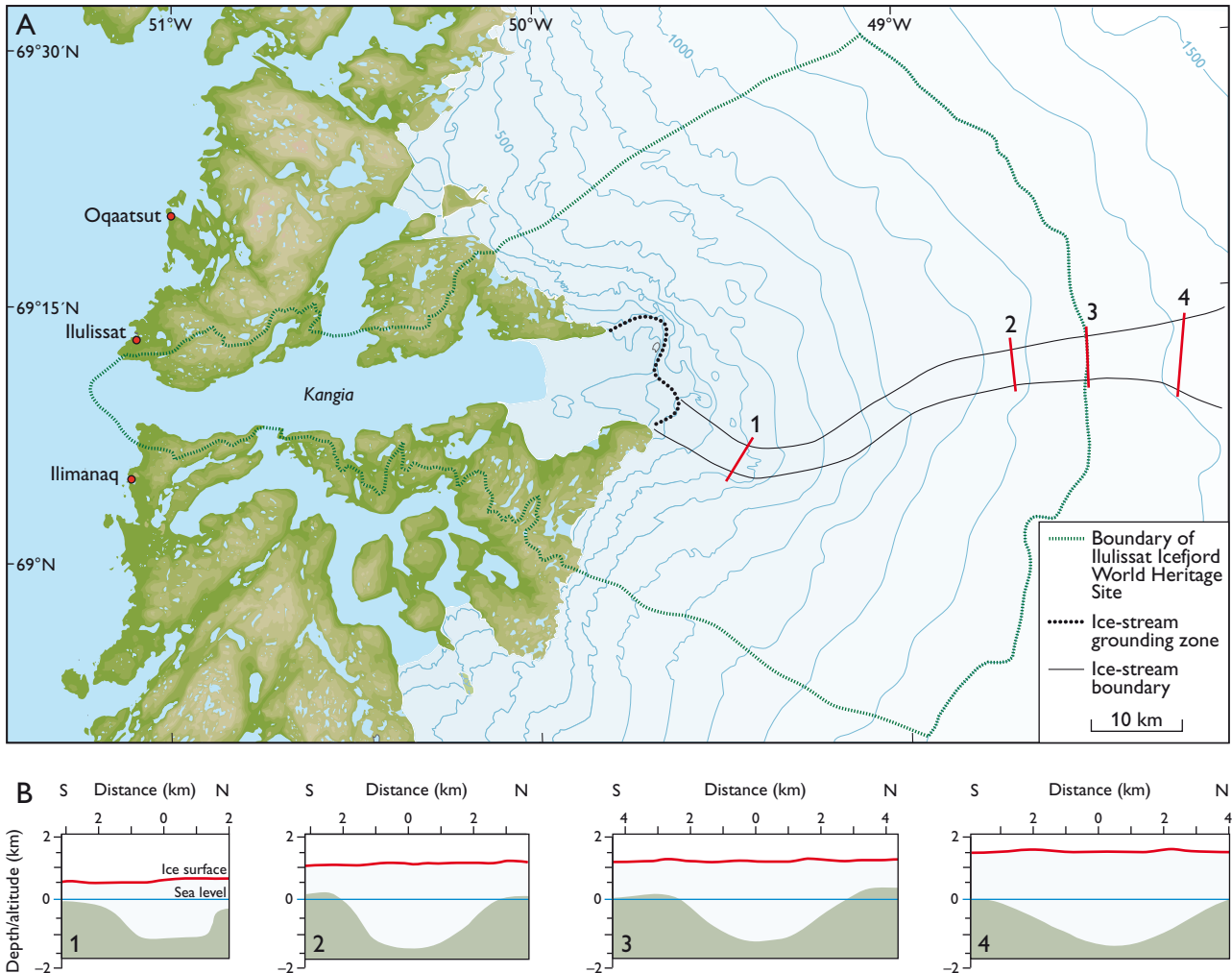
Fig. 35. **A:** Map of Kangia and the ablation zone of Jakobshavn Isbræ showing the positions of cross-sections **1–4** illustrated below. The ice-free areas have been mapped by GEUS, and the Inland Ice by H. Brecher (in: Fastook & Hughes 1994 and Fastook *et al.* 1995); supplementary data from the Inland Ice have been provided by S. Ekholm (National Survey and Cadastre, Copenhagen). Map compilation by Willy Weng and Annette T. Hindø (GEUS). The borders of the Jakobshavn ice stream (black lines) and the grounding zone *c.* 1985 (dotted black line) are from Echelmeyer *et al.* (1991, 1992). Ice-margin data from 1994; note that the present margin of Jakobshavn Isbræ, following the dramatic recent recession (Fig. 13), broadly coincides with the grounding line shown here. **B:** The four profiles of the ice stream (**1** to **4**) are based on seismic studies (Clarke & Echelmeyer 1996). **N** and **S** on the profiles indicate the north and south ends.

Present ice-margin surface

A topographic map of Greenland at a scale of 1:2 500 000 with 250 m contour intervals, including the entire ice sheet, was published by KMS [National Survey and Cadastre] in 1994, and the same topographic base was used in the Geological Map of Greenland at the same scale (Escher & Pulvertaft 1995). These maps show a low surface gradient from the highest central part of the ice sheet outwards towards the ice margin, reflecting in some respects the lowland areas underlying the ice (Fig. 4). Detailed surface maps were produced in connection with the investigations along the EPF-EGIG line at Eqip Sermia (Holtzscherer & Bauer 1954), over the ice sheet margin at Paakitsoq (Thomsen *et al.* 1988), and around Jakobshavn Isbræ (Fastook & Hughes 1994; Fastook *et al.* 1995). The map around Jakobshavn Isbræ was used to delineate the 'Ilulissat Ice-
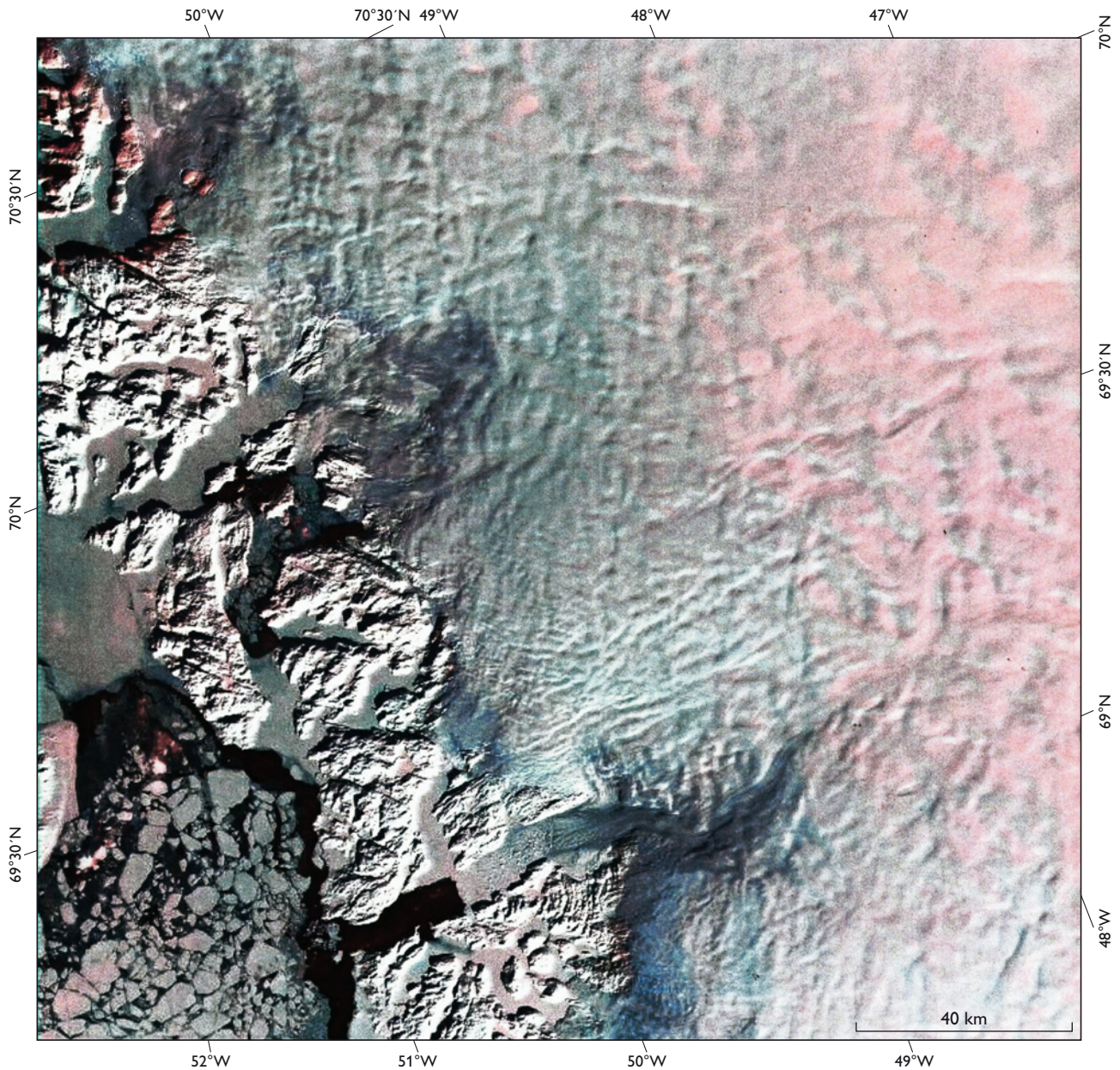
Fig. 36. False-colour satellite image (Landsat-7) of the interior of Disko Bugt and adjacent parts of the ice sheet, from 4 March 2002. The outer part of Kangia, and coastal areas north and south of the mouth of Kangia, are ice-free (black areas in lower part of the image). The morphology of the snow-covered, ice-free land continues under the surface of the ice margin, as can be seen from the surface features of the ice.

fjord' World Heritage proposal that was included on the World Heritage List in 2004 (Mikkelsen & Ingerslev 2002). This map is reproduced here (Fig. 35), and clearly shows the trend of Jakobshavn Isbræ.

In the Disko Bugt region, annual mass-balance field measurements have been carried out along the EPF-EGIG line (Holtzscherer & Bauer 1954; Bauer *et al.* 1968b, Ambach 1977), the GGU stake line (Fig. 15; Thomsen *et al.* 1988; Braithwaite *et al.* 1992) and a line along Jakobshavn

Isbræ (Figs 37, 38; Echelmeyer *et al.* 1992). The altitude of the equilibrium line, where the annual mass balance is 0, shows great annual variations, from *c.* 1000 m a.s.l. (extremely cold budget years) to more than 1200 m a.s.l. An in-depth discussion of climatic factors determining the variations in annual mass balance is given by Echelmeyer *et al.* (1992), who also record the albedo changes (increased ablation) due to inblown dust from the extensive trimline zone around Jakobshavn Isbræ. Calculated variations in
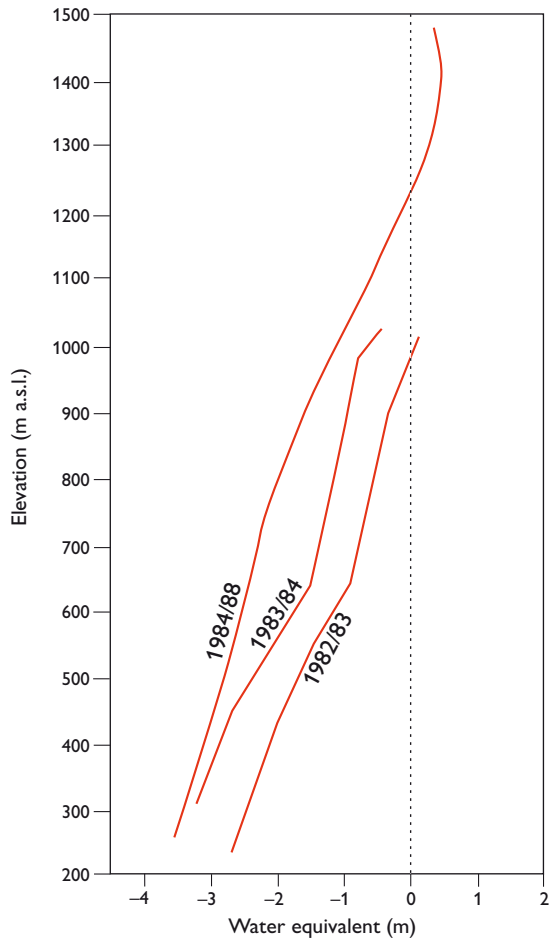
Fig. 37. Annual mass balance in relation to elevation along profiles in the Ilulissat area. 1982/83 and 1983/84 designate profiles along the GGU stake line (see Figs 15 and 38 for location; from Thomsen *et al.* 1988). 1984/88 shows measurements for this time period along Jakobshavn Isbræ (from Echelmeyer *et al.* 1992).

the annual ablation of the region for the period from 1961/1962 to 1989/1990 are reported by Braithwaite *et al.* (1992).

The trend of the annual mass balance along the GGU line, measured in 1982/83 and 1983/84 (Thomsen *et al.* 1988), is compared in Fig. 37 with the mass balance along Jakobshavn Isbræ, measured from 1984 to 1988 (Echelmeyer *et al.* 1992). The trends of the curves (the ablation gradient) are similar, but the difference between the equilibrium line altitudes (ELA) mainly illustrates the variations of the annual climatic conditions in the area, rather than variations due to the distance between the locations of the measurements (Fig. 38).

The ELA defines the upper limit of the ablation area, which is about 50 km wide. Above this, the accumulation zone extends to the top of the ice sheet around Summit. The decrease of intermittent summer-snow melt with increasing elevation is expressed by the division of the firn area into a region of superimposed ice over wet snow, a percolation facies and finally a dry snow facies. These facies were originally defined by Benson (1962), and have been modified by Williams *et al.* (1991) and Benson (1994, 1996).

The facies concept is applied in the current monitoring programme of the mass balance of the Inland Ice, which is based on data from satellites. The spectral variability of ice and snow surfaces is used to determine the different facies (Fausto *et al.* 2007).

The meltwater produced above the ELA during the summer is retained in the firn in increasing amounts with decreasing elevation. Refreezing of this meltwater leads to

Fig. 38. Map of the Jakobshavn Isbræ area, showing routes, stations and stake lines of the expeditions that have worked on the ice. **EGIG**, Expédition Glaciologique Internationale au Groenlande, mainly 1957–1960; **EPF**, Expéditions Polaires Francaises (Missions P.E. Victor 1948–1951); **GGU**, Geological Survey of Greenland, stake line 1982–1989. Modified from Wegener *et al.* (1930) and Weidick & Thomsen (1983).
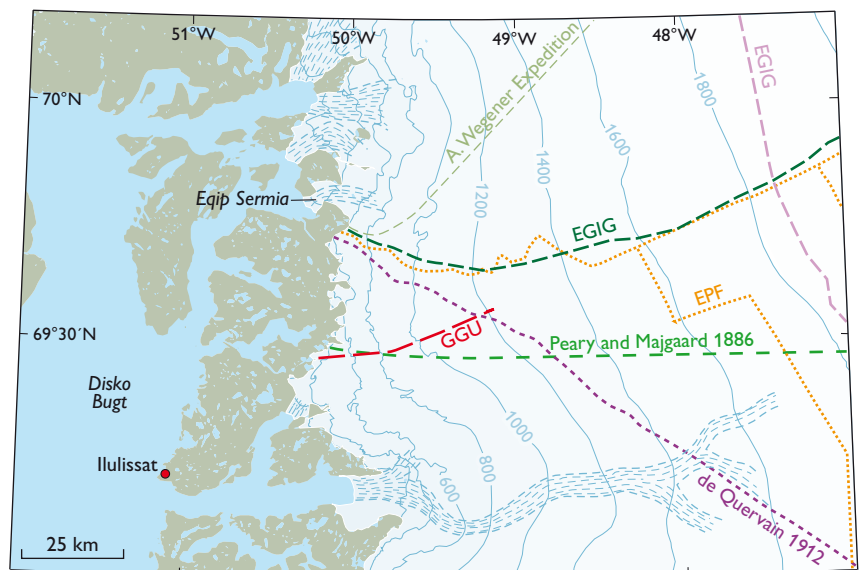
increased temperatures in the firn. In the ablation zone, the superficial meltwater drains through crevasses and sub-glacial tunnel systems, where refreezing and closure of the meltwater conduits can take place. The complex system of drainage that can occur in glaciers has been described by Roethlisberger & Lang (1987). Descriptions of the drainage in the Paakitsoq area (site of the GGU line) are provided by Thomsen *et al*. (1988) and Zwally *et al*. (2002), and for Jakobshavn Isbræ by Echelmeyer *et al*. (1992).

Drainage and thermal conditions at the ice margin

The large ablation area of a continental ice sheet such as the Inland Ice is characterised by extensive englacial and subglacial drainage. Superficial meltwater drains into sub-glacial channels and conduits that are fed from crevasses and moulins on the ice surface. At the base of the ice, the water can drain in thin water films or in subglacial chan-nels to the ice margin (Thomsen *et al*. 1988), and the water can emerge as upwelling plumes at the front of calving tidewater glaciers (outlets) such as Jakobshavn Isbræ (Echelmeyer *et al*. 1992). The contribution of basal melt-ing from the Jakobshavn Isbræ drainage basin has been cal-culated to be *c*. 20% of the total loss by ablation (Echelmeyer *et al*. 1992). The complexity of the drainage is influenced by refreezing or by internal as well as external heating due to the movement of the glacier.

Glacier-dynamic modelling has been applied to the so-called 'quiet' sector, located between Jakobshavn Isbræ in the south and the ice streams draining into Torsukattak icefjord in the north. The ice margin in this sector is bor-dered by land areas or lakes. Modelling of the bottom con-ditions by Radok *et al*. (1982) suggested that the basal ice reaches its pressure melting point 250–300 km from the ice margin. Based on a model for calculating the response of the marginal sector of the Inland Ice to mass-balance changes, the ice margin at Paakitsoq (Fig. 28) could be divided into three zones: (1) An inner zone more than 292 km from the ice margin where ice movement is dominated by internal deformation. (2) A zone between 292 and 18 km from the margin where the ice is at the pressure-melt-ing point, which leads to a significant degree of bottom slid-ing. (3) An outer zone, at 18–0 km from the ice margin, that is characterised by extensive sliding, high hydraulic pressure at the bottom of the ice, and conditions that allow for formation of cavities (Reeh 1983; Thomsen *et al*. 1988).

Ice-temperature measurements have been made at a few localities along both the EPF-EGIG and GGU lines. At Camp VI on the EPF-EGIG line, drilling from the surface
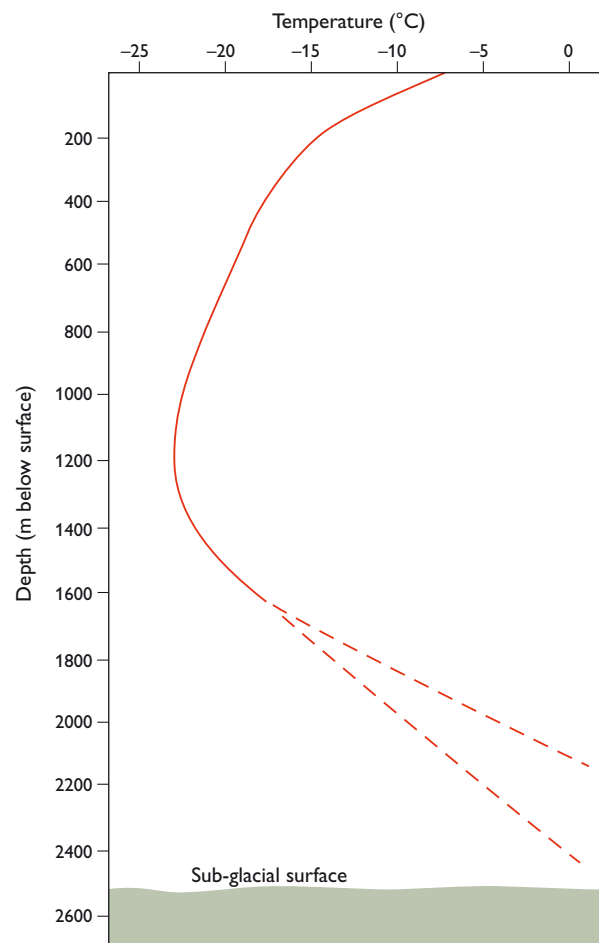

Fig. 39. Temperature profile through Jakobshavn Isbræ near its centre line, showing temperature versus depth below the ice surface (from Iken *et al*. 1993). The profile is situated about 50 km from the glacier front of 1988/89, at an elevation of 1000 m a.s.l., with the bottom of the ice at about 1500 m b.s.l. The dashed lines show possible limits for tem-peratures below the measured values. More recent modelling indicates that the thickness of the basal temperate ice (ice at the pressure-melt-ing point) is around 230 m in the ice stream, and *c*. 30 m in the adja-cent ice sheet (Funk *et al*. 1994).

at 1598 m a.s.l. extended only to a depth of 125 m; a tem-perature profile from this hole shows a decrease in tem-perature from *c*. –12.5°C to –16°C (Heuberger 1954; Robin 1983). The ice thickness here was about 1000 m. On the GGU line, temperatures were measured in a hole drilled at *c*. 500 m a.s.l. Temperatures were slightly nega-tive throughout the ice hole, decreasing to –2.1°C in the upper 50 m (Thomsen *et al*. 1991).

For Jakobshavn Isbræ, it has been calculated that 2–3 km^3/year of meltwater is generated by deformational heat-ing, although the meltwater does not seem to influence the movement of the ice stream (Echelmeyer & Harrison 1990; Echelmeyer *et al*. 1992).
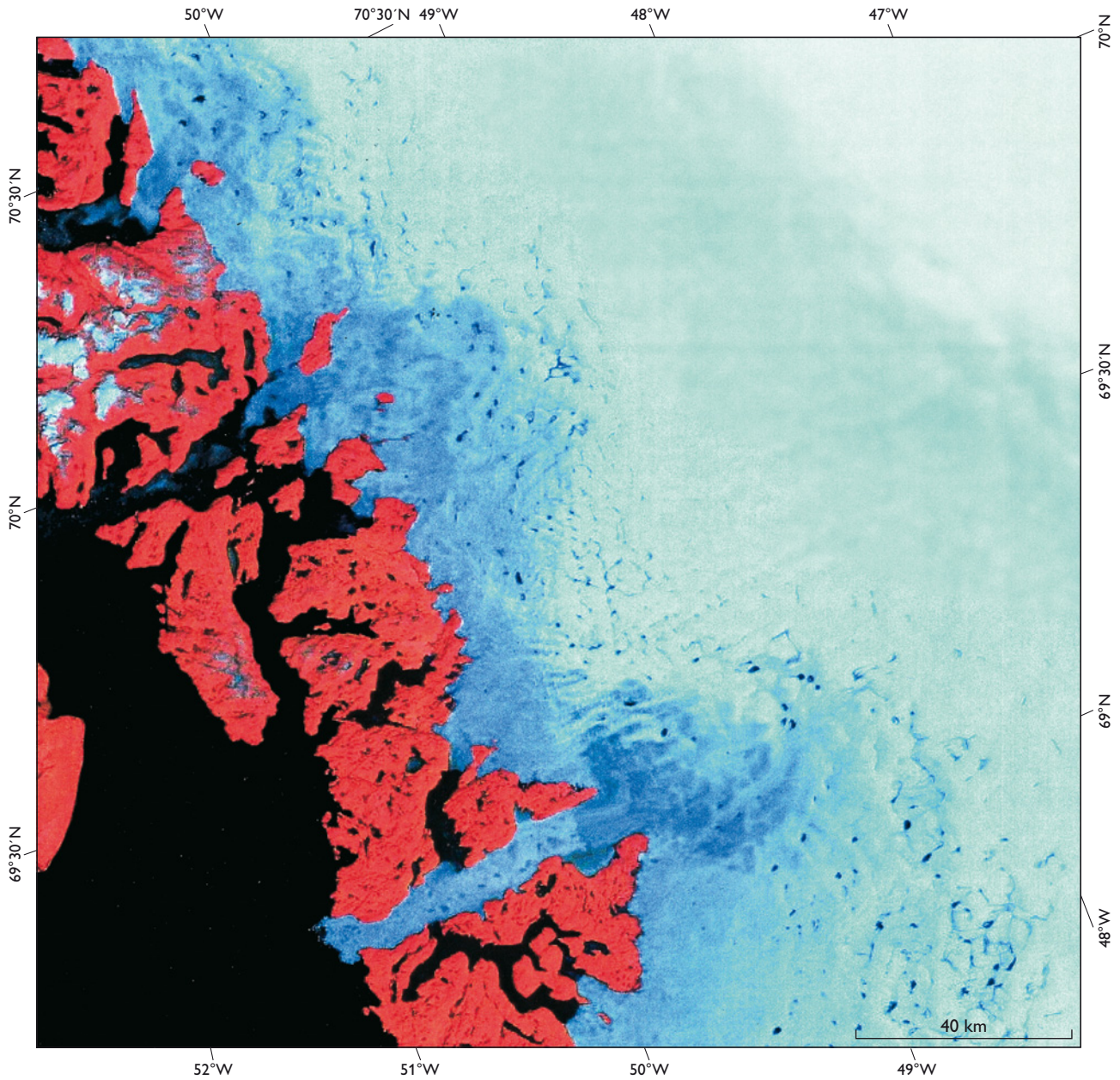
Fig. 40. False-colour satellite image (Landsat-7) of the interior of Disko Bugt and adjacent parts of the ice sheet, from 7 July 2001. The scene shows the zone with numerous lakes on the ice near the equilibrium line.

A temperature profile of Jakobshavn Isbræ was measured in 1988/1989 at a locality situated about 50 km from the glacier front at 1020 m a.s.l. The site is located at the centre line of the ice stream, where the surface ice moves at about 1 km/year, and the bed is situated at about 1500 m below sea level (Iken *et al.* 1993, Clarke & Echelmeyer 1996). The temperature profile (Fig. 39) shows a temperature minimum of –22°C *c.* 1200 m below the surface; this implies a thick, relatively warm and low viscosity bottom layer which is thought to facilitate the fast movement of

the ice stream. It has been suggested that this basal layer may contain Wisconsinan ice (Iken *et al.* 1993). This work was followed up by investigations of flow and temperature conditions at the transition between Jakobshavn Isbræ and the 'quiet' ice margin (Lüthi *et al.* 2002).

The surface features of the 'quiet' sectors of ice comprise lakes, rivers, crevasse formations and moulins, which have been mapped in detail and described along the lower parts of the EGIG line by Bauer *et al.* (1968a). Detailed maps and descriptions around the GGU line near Paakitsoq are

given by Thomsen *et al.* (1988). Thorough descriptions of Jakobshavn Isbræ and its surroundings are provided by Echelmeyer *et al.* (1992) and Fastook *et al.* (1995). The surface of the Inland Ice margin in the Disko Bugt region is characterised by numerous lakes up to an altitude of about 1400 m (Fig. 40). Above this altitude, corresponding approximately to the beginning of the wet snow facies of the accumulation area, lakes are present but are ice and snow covered and only faintly visible. The occurrence of numerous lakes extends down to about 1000 m a.s.l., corresponding to the zone of superimposed ice and the upper part of the ablation area. Lower down in the ablation area there are fewer lakes, due to the decreasing thickness of the ice cover that leads to increased crevassing during ice movement.

Where the ice margin is bordered by land, marginal lakes are common. Due to the bordering hilly upland, the occurrence and size of the lakes vary with changes in the position of the ice margin, just as the proglacial drainage shows variations. Changes in drainage patterns at Paakitsoq, between 1953 and 1959, can be documented by aerial photographs from the 1950s, 1960s and 1985 (Thomsen 1983).

Tininnilik, situated *c.* 40 km south of Jakobshavn Isbræ, is an ice-dammed lake at the ice margin that shows periodic drainage. The lake covers an area of 43 km² at its maximum extent and 20 km² at its minimum (Thomsen 1984). When the periodic drainage was first described in 1913 (Koch & Wegener 1930), it was stated that the filling/drainage cycle of the ice-dammed lake was about 10 years based on information from local people. This cycle seems to be more-or-less permanent, probably due to the nature of the damming glacier (Saqqarliup Sermia) which, in contrast to Jakobshavn Isbræ, has been almost stable over the last 150 years (Weidick 1994a, b). Braithwaite & Thomsen (1984) determined that during each drainage event nearly 2 km³ of water is released along Saqqarliup Sermia into the southern branch of the Tasiusaq fjord complex, and from there onwards to the south side of Kangia. Braithwaite & Thomsen (1984) also recorded the years of drainage for the period 1942–1983; the most recent drainage took place in 2003 (F. Nielsen, personal communication 2004).
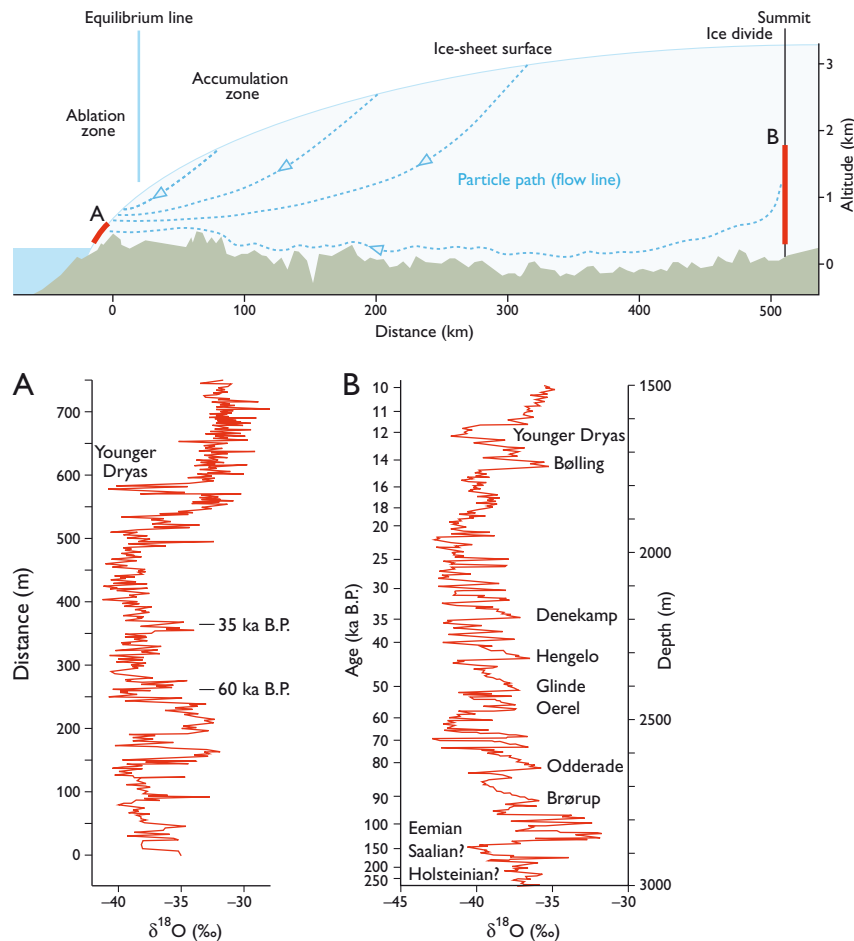

Fig. 41. Cross-section of the Inland Ice, showing schematic flow lines within the ice and the locations of (**A**) an oxygen isotope record near the margin and (**B**) the deep ice core at the summit of the ice sheet. **A**: Profile of the outer ice margin at Paakitsoq. The record covers major parts of the last ice age (Wisconsinan). From Reeh *et al.* (1991). **B**: A section of the ice core at Summit (Fig. 1) revealing successively older layers of snow and ice. The record covers about 250 000 years, and includes the Eemian and perhaps Holsteinian interglacials (Dansgaard *et al.* 1993).

The age of the ice margin has been investigated in detail along three profiles near Paakitsoq (Fig. 2). At certain localities along the ice margin, the ice stratification demonstrated from deep drill holes in the interior of the ice sheet can be preserved (Fig. 41; Reeh *et al.* 2002, Petrenko *et al.* 2006). Such localities, as at Paakitsoq, are characterised by a fairly smooth subsurface within so-called quiet marginal areas distant from larger ice streams. The profiles so far investigated at Paakitsoq cover a span of time from possibly 150 000 years B.P. to the present. These and similar investigations at the ice margin provide easy access to ice-age ice, a cheaply acquired supplement to the very expensive deep ice-core records, which can provide important information about past climates and the dynamics of the ice sheet.

Movement of the ice margin

The surface movement of the ice margin in Disko Bugt has been determined in the 'quiet' land-based area along the EGIG line north of Jakobshavn Isbræ, and in Fig. 42 it is compared with the horizontal movement of Jakobshavn Isbræ. The main differences are found in the areas around and below the equilibrium line. The thickness of the land-based ice is up to 700–800 m below the ablation zone along the EGIG profile (Holtzscherer & Bauer 1954; Bauer *et al.* 1968b). Along this profile, the horizontal surface movement increases westwards to a maximum of 100–200 m/year just above the equilibrium line, and then decreases towards the ice margin.

At the Swiss Station (for location, see Fig. 15), measurements of movements were made from 1996 to 1999 (Zwally *et al.* 2002). The station is situated near the equilibrium line at 1175 m a.s.l., at a site where the ice is 1220 m thick. The measurements show increasing ice velocities with increasing surface melt, indicating that bottom sliding is enhanced by rapid migration of meltwater to the bottom of the ice. This provides a mechanism for rapid, large-scale dynamic response of ice sheets to climate warming, for example during the transition from glacial to initial interglacial conditions.

With respect to calving tidewater outlets in fjords as well as calving outlets in proglacial lakes, it is generally known that the rate of movement increases towards the front reaching up to a few km/year. However, little information is available for calving outlets in the Disko Bugt region apart from at Jakobshavn Isbræ.
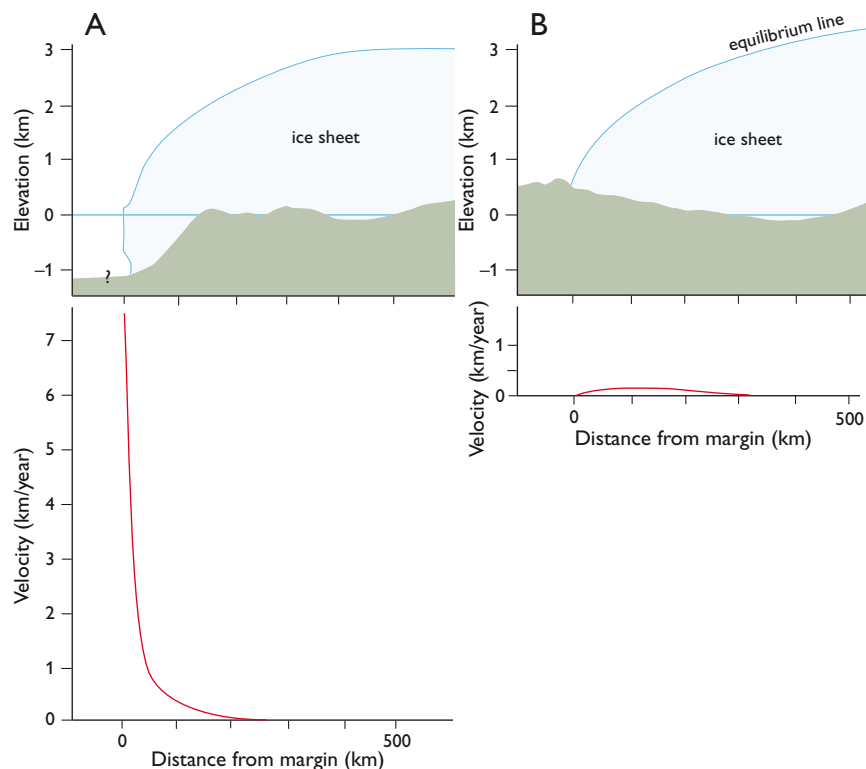

Fig. 42. **A**: Diagram showing a cross-section through Jakobshavn Isbræ together with a plot of the horizontal surface movement of the ice stream. Compiled from Fastook *et al.* (1995) and Joughin *et al.* (2004). **B**: Cross-section of the ice sheet along the EGIG line (Fig. 38) in a 'quiet' marginal area, about 60 km north of Jakobshavn Isbræ, in association with a plot of the horizontal surface movement of this part of the ice sheet, which has a maximum velocity of 200 m/year *c.* 100 km from the ice margin (from Bauer *et al.* 1968b).

Distance from flank of glacier (km)	Velocity		Comments
	m/24 h	km/y	
Alanngorliup Sermia, July 1875 (Helland 1876)			
<0.5		<0.2?	Velocity measurement uncertain
Jakobshavn Isbræ, 7–9 July 1875 (Helland 1876)			
From south side 0.40	14.7	5.4	Measurements close to front
0.42	15.4	5.6	Frontal height scarcely over 40 m a.s.l.
0.45	15.2	5.5	Width of the fjord estimated to 4.5 km (on present maps *c.* 7 km)
0.45	15.2	5.5	
1.05	19.8	7.2	
1.06	19.5	7.1	
Jakobshavn Isbræ, 22 March – 24 April 1880 (Hammer 1883)			
From south side 0.28	5.1	1.9	Measured *c.* 5 km behind glacier front
0.55	7.5	2.7	Maximum velocity in the central part of the glacier estimated to at
0.62	9.2	3.4	least 16 m/24 h. Height of glacier front *c.* 63 m a.s.l.
0.88	12.5	4.6	Width of fjord *c.* 7 km
0.87	12.3	4.5	
Jakobshavn Isbræ, July 1902 (Engell 1904)			
From south side 1.29	15.0	5.5	Velocity measured near the front
1.30	14.2	5.2	Width of fjord *c.* 7 km
1.87	19.8	7.2	
1.84	19.8	7.2	
4.26	22.8	8.3	
Jakobshavn Isbræ, 27–28 September 1929 (Sorge in Wegener *et al.* 1930)			
From south side 2.1–3.7 6.6–7.7	18–21	6.6–7.6	Preliminary calculations based on four points. Glacier *c.* 6 km wide, presumably measured near front
Sermeq Avannarleq in Torsukattaq fjord, 24–25 July 1875 (Helland 1876)			
From north side 0.21	3.8	1.4	Presumably measured near front
0.37	5.7	2.1	Recorded frontal height 15 m a.s.l.
1.93	8.8	3.2	Glacier *c.* 9 km wide
4.07	10.1	3.7	
4.94	10.2	3.7	
4.97	9.4	3.4	
Sermeq Avannarleq in Torsukattaq fjord (Steenstrup 1883a)			
From north side, 2.70	7.8	2.8	Measured near glacier front
5–7 May 1879 2.70	6.3	2.3	Width of glacier estimated to *c.* 8 km
From north side, 3.01	5.0	1.8	
21–22 May 1880 3.01	7.8	2.8	
2.17	5.0	1.8	
1.51	5.0	1.8	

Jakobshavn Isbræ is an extreme example, where the depth of the glacier is controlled by a subglacial trough reaching 80–100 km inland under the ice and with depths that reach down to 1.5 km below sea level in the outer parts (Clarke & Echelmeyer 1996). The trough can be envisaged as a continuation of the proglacial icefjord (Kangia), which is believed to be around 1000 m deep. Such a depth is extreme for the fjords in the area, where depths of 400–800 m are more usual (Weidick *et al.* 1974b). Fjord depths at the present fronts are not known, and can only be estimated from soundings made at some distance from the active fronts.

The empirical relationship between calving rate and water depth at the glacier front (Pelto & Warren 1991) is therefore difficult to establish, although an attempt was made by Long & Roberts (2003) for the deglaciation of Disko Bugt. Measurements of movement and thickness of all the tidewater glaciers in Disko Bugt and the Uumannaaq fjord complex were undertaken in 1957 and 1964 by the EGIG expeditions (Bauer *et al.* 1968a; Carbonnell & Bauer 1968), based on photogrammetric analyses of aerial photographs, and show a positive correlation between mean rate of movement and mean thickness of the glacier front.

With respect to the high movement rate of many of the calving glaciers in the Disko Bugt region, and variations of the velocity over time, a brief historical review is given below. The field conditions for the earliest measurements introduce an element of uncertainty, but do provide an order of magnitude of the possible variations in velocity over long time-spans. All the measurements given are related to the frontal areas, and for nearly all the glacier fronts the variations in position are within 2 km for the period since the end of the 19th century (Fig. 43; Weidick 1994a, b). The measurements of velocity at the front of most outlets were thus conducted at nearly the same location. The only exception is Jakobshavn Isbræ, which receded 26 km between *c*. 1850 and *c*. 1950 and has experienced continued recession since 2002/2003; in contrast to other outlets, the measurements were here undertaken from widely different positions of the glacier front during the recession.

Most early velocity measurements were carried out over short time intervals, and consequently where a low rate of movement was recorded, possible errors may be large, a fact that is stressed in some of the old descriptions (Helland 1876). In Table 5, older data are only given for the faster moving glaciers. Some original sources gave measurements in Danish feet (1 Danish foot = 0.31385 m), here converted to metres or kilometres. Rates of movement are given as m/24 h or km/year (Tables 1, 5). The latter is used for comparison with modern data and neglects possible variations in the course of the year.

Notes on individual outlets

The glacial histories of individual glacier outlets in the Disko Bugt region (Figs 2, 43) are summarised below.

Nordenskiöld Gletscher (Akuliarutsip Sermersua); 68°20′N, 50°51′W). This glacier does not drain into Disko Bugt, but its glacial history is closely related to the bay. The visits of the Nordenskiöld expeditions to the glacier in 1870 and 1883 gave rise to detailed descriptions of this glacier (Nordenskiöld 1885, 1886), but the velocity of the outlet was apparently first measured in July 1957 (Bauer *et al.* 1968a). The average velocity was found to be 3 m/24 h with a small production of calf ice (Fig. 5). From the descriptions, it appears that the build-up of a frontal moraine and proglacial delta hinders the production of calf ice. In the period 2000–2005, a slight increase in velocity was reported by Rignot & Kanagaratnam (2006).

Saqqarliup Sermia (68°54′N, 50°18′W) and Alanngorliup Sermia (68°55′N, 50°12′W). Alanngorliup Sermia was

visited in 1875 by Helland (1876), who found that the front had a height of *c*. 10 m. The velocity of the glacier was given as below 0.5 m/24 h, probably recorded in the lower reaches of the glacier although the location was not stated. Higher up in the glacier, where there is a tributary to Saqqarliup Sermia, the rate of movement is given as a maximum of 0.4 m/24 h. For both outlets, the mean frontal velocity in 1957 was measured at 0.9 m/24 h (Carbonnell & Bauer 1968). A slight velocity increase has also been reported for this outlet by Rignot & Kanagaratnam (2006).

Jakobshavn Isbræ (Sermeq Kujalleq; 69°11′N, 49°48′W). The first velocity measurements in 1875 were made along the southern side of the glacier. In the fast flowing, central part of the glacier, *c*. 1 km from the margin, a velocity of 19.8 m/24 h was measured (Table 5; Helland 1876). Helland also measured the movement of the glacier adjacent to the margin, where he found a velocity of not more than 0.02 m/24 h.

Subsequent measurements in 1880 recorded slightly lower velocities (Table 5; Hammer 1883; estimated maximum velocity >16 m/24 hours). The measurements appear to have been made *c*. 5 km behind the front, judging from a sketch map of measured points in Hammer's report. If the up-stream velocity decrease was similar to later values (Carbonnell & Bauer 1968), the frontal velocity may well have been of the same magnitude as given by Helland. However, the velocity does not always decrease immediately behind the glacier front; Joughin *et al.* (2004) reported that the marked decrease in speed of Jakobshavn Isbræ in the 1990s first occurred *c*. 14 km behind the front.

In 1902, the glacier was visited by Engell, who recorded a similar high velocity to that given by Helland for the central part of the glacier (*c*. 23 m/24 h; Engell 1904). Comparable values of between 18 m/24 h and 21 m/24 h were measured by Sorge in 1929 (Wegener 1930; Wegener *et al.* 1930).

Measurements in July 1958 gave a mean velocity of 13.1 m/24 h (Bauer *et al.* 1968a), whereas investigations in June 1964 gave a mean velocity of 19.1 m/24 h (Carbonnell & Bauer 1968). A very similar figure to the 'frontal velocities' of the central zone of the glacier given by Pelto *et al.* (1989): 21.1 (1964), 20.4 (1976), 21.0 (1978) 20.6 (1985) and 20.3 (1986) m/24 h. A decrease of velocity immediately behind the front was documented.

In a comment to the surprisingly large difference between the velocities in 1958 and 1964, it was pointed out that similar large changes were found at Rink Isbræ (Fig 1.), and that more measurements are needed to explain the difference (Carbonell & Bauer 1968, p. 77). Subsequent measurements of Jakobshavn Isbræ show velocity variations
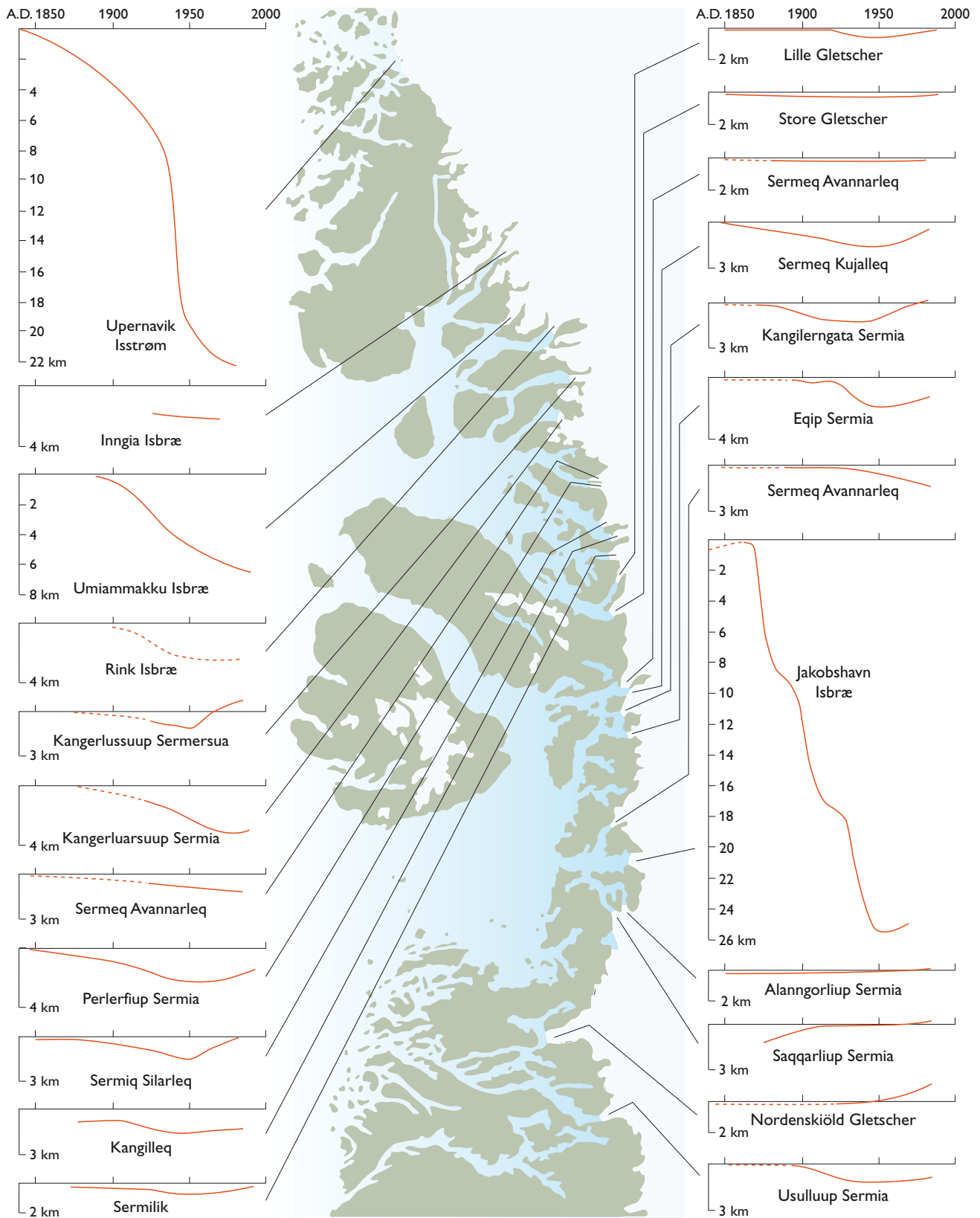
Fig. 43. Tentative reconstructions of fluctuations in the frontal positions of calving glaciers in central West Greenland from 1850 to 1985. Upernavik Isstrøm shows a recession of *c.* 23 km and Jakobshavn Isbræ shows a recession of *c.* 26 km (red lines), whereas the other glaciers show smaller fluctuations. From Weidick (1994b).

that are apparently related to the thickness of the ice margin. However, information is needed on the velocity in the time period between the measurements in 1929 and in 1948. It is also possible that the low velocity in 1958 could have been connected to a thickening of the glacier from *c.* 1950–2000, which was a period with a stable front. Studies of aerial photographs from the 1940s may provide data on thickness changes in the marginal parts of the ice sheet.

It is noteworthy that the maximum velocity, in spite of all the possible errors that may have affected the older measurements, has maintained a value of *c.* 5–9 km/year for more than a century. This is particularly relevant when considering the reasons for the marked velocity increase of the glacier after about 2000. Thus a 95% velocity increase took place between 1996 and 2005 according to Rignot & Kanagaratnam (2006). More details are provided below and in Fig. 44. In 2003, the velocity was 12.6 km/year and the discharge was *c.* 50 km^3 ice/year (Joughin *et al.* 2004).

Sermeq Avannarleq in Kangia (69°2′N, 50°18′W). The velocity of this glacier was first measured during the EGIG expeditions, who recorded an average velocity of *c.* 1 m/24 h (Carbonnell & Bauer 1968).

Eqip Sermia (69°48′N, 50°13′W). Velocity measurements of this glacier were first made from 31 July to 7 August 1912 by de Quervain, who recorded a maximum velocity at the front of 1.5–2.4 m/24 h (de Quervain 1925). Subsequent more detailed investigations on velocity and frontal fluctuations were made by Bauer during the EPF expeditions in 1948–1949 (Bauer 1955). A mean velocity of 3 m/24 h was recorded, and no difference was noted between September 1948 and June 1949.

The EGIG measurements from 12 to 17 July 1957 gave a mean velocity value of 3.1 m/24 h (Bauer *et al.* 1968a), whereas later measurements, from 5 to 9 July 1959 (Bauer 1968) and 29 June to 12 July 1964 (Carbonnell & Bauer 1968) both gave values of *c.* 2 m/24 h. The differences are related to insufficient measuring points for the old data (Bauer 1968, p. 10). Detailed records of the frontal fluctuations of the glacier in the 20th century are provided by Bauer (1955) and Nielsen *et al.* (2000), who indicate advances around 1920, and during the 1990s. Between 2000 and 2005, the velocity of Eqip Sermia accelerated by 30% (Rignot & Kanagaratnam 2006).

Kangilerngata Sermia (69°55′N, 50°17′W). The movement of this glacier was first measured from 7 to 17 July 1957, when the average velocity was given as 2.3 m/24 h (Bauer *et al.* 1968a). Subsequent measurements from 9 to 22 June 1964 gave 3.3 m/24 h (Carbonnell & Bauer 1968).

Between 2000 and 2005, the velocity of Kangilerngata Sermia also increased by 30% (Rignot & Kanagaratnam 2006).

Sermeq Kujalleq in the Torsukattak icefjord (70°00′N, 50°19′W). Velocity measurements of this large glacier started with the measurements of Bauer *et al.* (1968a) from 12 to 17 July 1957, which indicated a mean velocity of 7.2 m/24 h. From 9 to 22 June 1964, a mean velocity of 9.7 m/24 h was recorded (Carbonnell & Bauer 1968). Sermeq Kujalleq slowed down by 11% between 2000 and 2005 (Rignot & Kanagaratnam 2006).

Sermeq Avannarleq in the Torsukattak icefjord (70°04′N, 50°19′W). The velocity was first measured by Helland (1876) from 24 to 25 July 1875. The measurements were made from the northern side of the glacier, and extended 4 km into the central parts where a velocity of over 10 m/24 h was measured (Table 5). Subsequent measurements by Steenstrup (1883a) in May 1879 and May 1880, from nearly the same position, did not reach as far into the glacier, but a velocity of 8 m/24 h was recorded *c.* 3 km from the glacier margin. This can be compared with later measurements of the mean glacier velocity from 12 to 17 July 1957 of 6.4 m/24 h (Bauer *et al.* 1968a) and on 9 to 22 June 1964 of 5.2 m/24 h (Carbonnell & Bauer 1968). The velocity profile of the glacier is irregular. Maximum velocities of 9.5 m/24 h was found *c.* 1.5 km from the glacier margin (Carbonnell & Bauer 1968, fig. 38). Sermeq Avannarleq, as Sermeq Kujalleq, slowed down between 2000 and 2005 by 11% (Rignot & Kanagaratnam 2006).

Sermeq Kujalleq (Store Gletscher: 70°24′N, 50°32′W). Although situated somewhat farther to the north, north of Disko Bugt, this glacier ('Store Qarajaq Bræ' of Steenstrup 1883a and 'Grossen Qarajaq Eisstrom' of von Drygalski 1897) deserves mention because of its history of exploration. Its velocity was measured in August 1878. The maximum velocity of 12 m/24 h, measured *c.* 3 km from its north side by Steenstrup (1883a), is close to the mean velocity recorded by von Drygalski (1897) in 1893 of 12.9 m/24 h, of 11.6 m/24 h by Bauer *et al.* (1968a) for 1957, and by Carbonnell & Bauer (1968) of 13.4 m/24 h for 1964. These authors noted a significant upstream decrease in velocity for this glacier.

Historical information on the measurements of the calving tidewater glaciers in and around Disko Bugt leaves the general impression of high-speed behaviour of the major outlets to the fjords in the region; the scattered investigations lack sufficient details for further conclusions. The
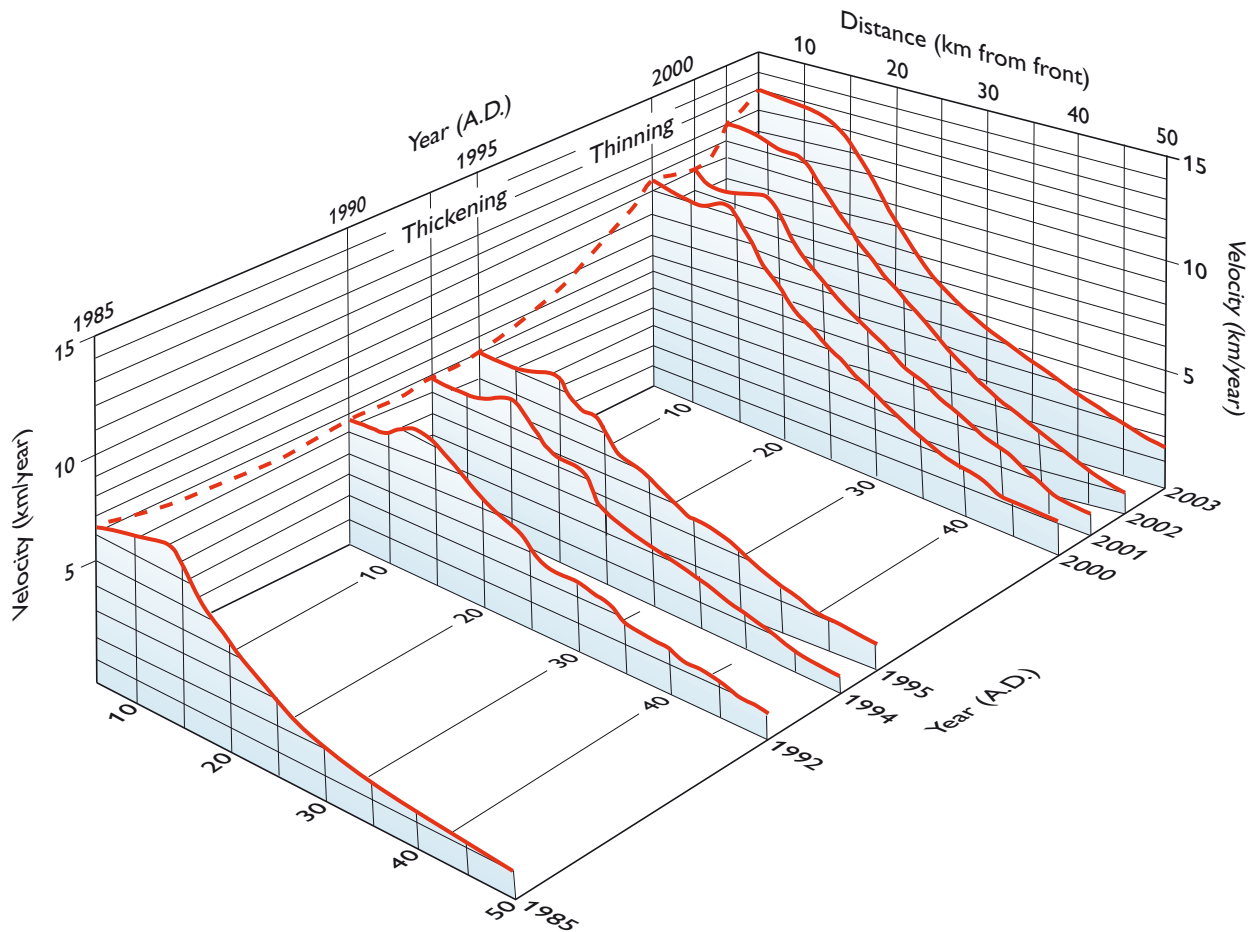
Fig. 44. Variations in the rate of movement of Jakobshavn Isbræ at distances *c*. 5–50 km behind the front. The marked increase in velocity that began in the late 1990s coincided with thinning of the front. Simplified from Joughin *et al*. (2004).

early measurements, especially those of von Drygalski's 'Grossen Qarajaq Eisstrom' and other outlets to Disko Bugt, reported the significant upstream decrease of the glacier velocities, although the investigations only covered the outermost 5–10 km of the glacier lobe.

Jakobshavn Isbræ is characterised by a high and rather constant flow rate at the glacier front, even though the position of the front has changed with time. Detailed coverage of the areal distribution of velocity is only known for Jakobshavn Isbræ, in the form of velocity contours (isotachytes of Ahlmann 1948). These cover an area of *c*. 80 × 80 km of the ice sheet upstream of the glacier front, with contours of 50 m/year, and record the situation at around 1985 (Fastook *et al*. 1995). The profile of Fig. 42 is derived from this source. A revision of the map for February 1992 and October 2000 (i.e. before the collapse of the front of Jakobshavn Isbræ in 2002/2003), is provided by Joughin *et al*. (2004), who describe the subsequent development illustrated by movement profiles from the front and reaching

50 km inland. Remote sensing data provide details of the frontal changes from 1985 to the present.

The velocity of Jakobshavn Isbræ has been more variable during the past few years than at any time since records began. A net thinning of the glacier of 200–300 m over the past 150 years, corresponding to an average thinning of 1.3 –2.0 m/year, is apparent from the elevation of the trimline zone around the glacier. This thinning has taken place over a time period when the glacier has had a rather constant velocity of 5 to 9 km/year. Detailed records covering the past few decades indicate that the 1984 velocity of 6.7 km/year had decreased to 5.7 km/year by 1992, and that this lower velocity continued until 1997. The velocity then increased sharply, and for the years 2000, 2002 and 2003 the glacier attained velocities of respectively 9.4, 11.9 and 12.6 km/year (Joughin *et al*. 2004). A thickening of *c*. 1 m/year was related to the period of low speed (1991–1997), whereas the subsequent higher velocities were related to a thinning of around 6 m/year, Fig. 44. The
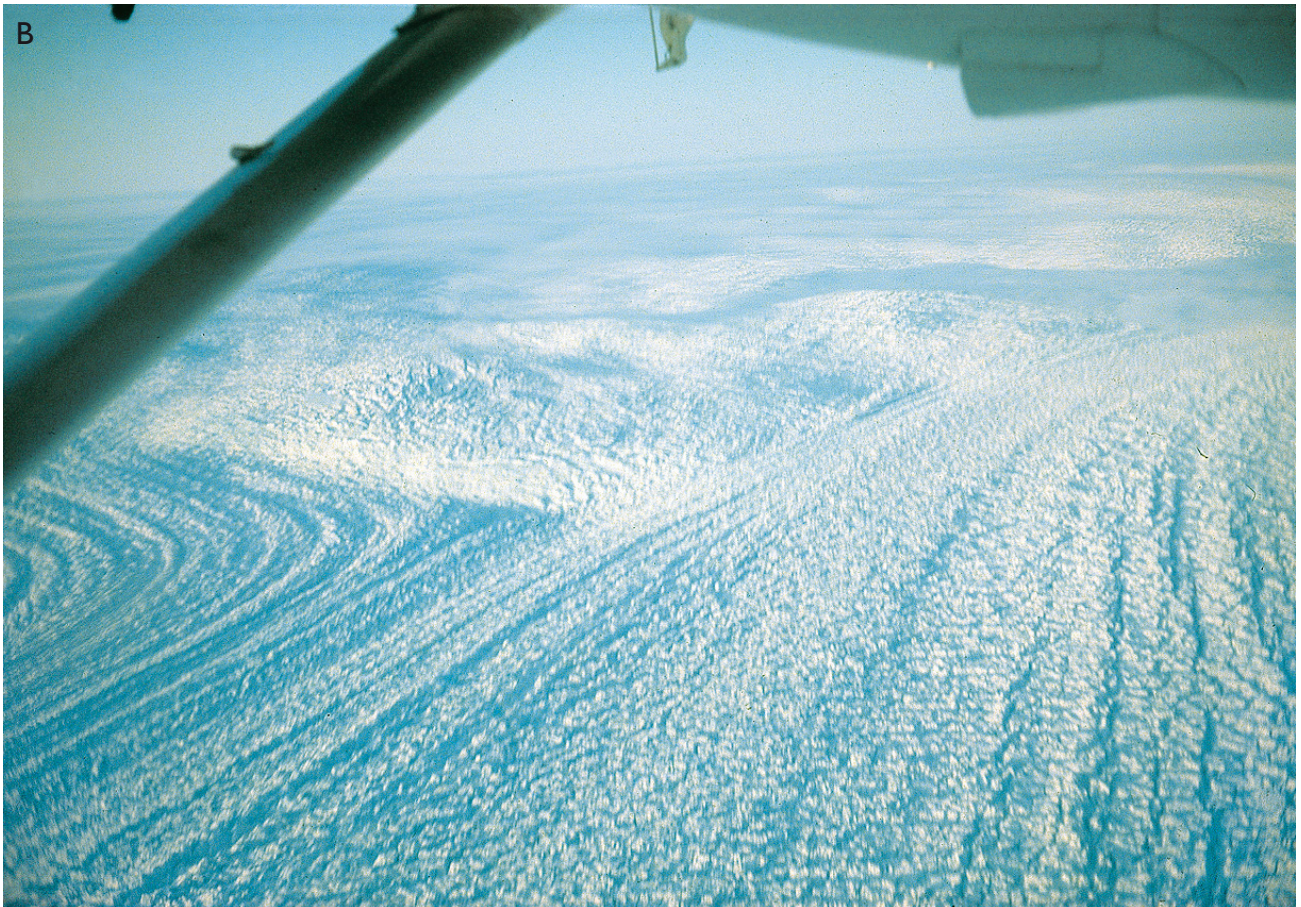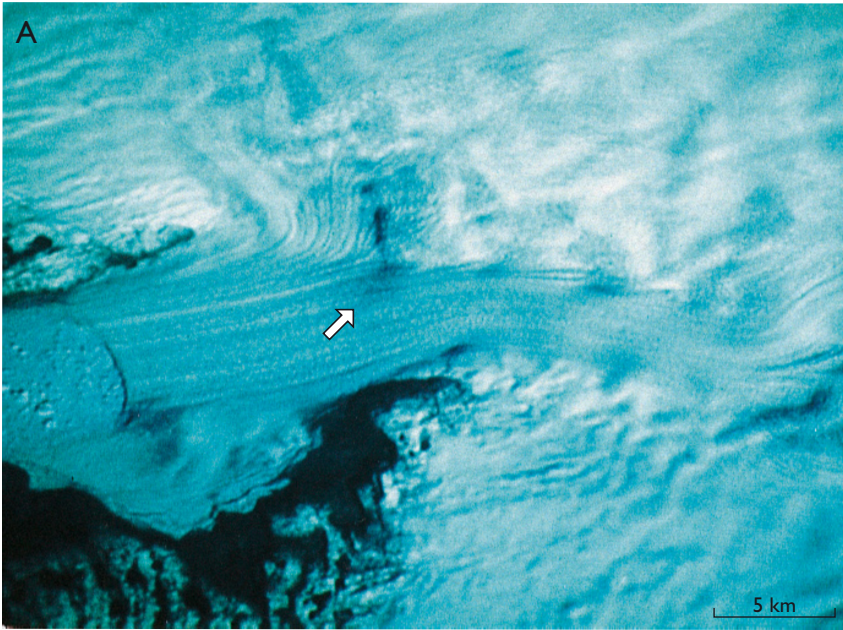
Fig. 45. Frontal area of Jakobshavn Isbræ. **A:** Landsat image of 27 September 1979. The glacier front and released icebergs can be seen near the left margin of the image. The ice stream is characterised by linear structures. A tributary from the north is separated from the main stream by a subglacial rumple. South of the front of Jakobshavn Isbræ is an area of stagnant ice (Tissarissoq, see also Figs 2, 25) that is separated from the ice stream by another rumple. The arrow indicates the view of the photograph below. **B:** Photograph of Jakobshavn Isbræ viewed from the WSW (see arrow on Fig. 45A) where the northern tributary joins the main ice stream. The bedrock topography under the ice is clearly reflected in the topography of the ice surface. Photograph by H.H. Thomsen, 1984.

sudden transition to rapid thinning that followed was at first confined to areas below *c*. 500 m a.s.l., but then spread inland and by about the year 2000 had reached up to 2000 m a.s.l. (Thomas *et al*. 2003).

The apparent quasi-stability of the front of Jakobshavn Isbræ from *c*. 1950 to 2000 has been related by Echelmeyer *et al*. (1991) to the presence of pinning points in the frontal areas of the floating front. However, little is known about the depths of the fjord below the floating outer parts of the glacier front that lie *c*. 22 km west of the lower seismic station of Clarke & Echelmeyer (1996), corresponding to profile 1 in Fig. 35. A possible threshold near the grounding zone (Figs 36, 45) may be viewed as a northern continuation of the curved bedrock lineaments on the south side of the glacier.

In contrast to the dramatic changes of the thickness and position of Jakobshavn Isbræ, the other calf-ice producing outlets to the north and south have shown only small changes in frontal positions during the past 150 years (Weidick 1994a, b). These small changes are perhaps linked to the shallower depths of these outlets demonstrated by the radar surveys of the Technical University of Denmark (Overgaard 1981) and apparent from low sun angle Landsat scenes.

For the areas around Jakobshavn Isbræ, the ice margin seems to have been nearly continuously receding over the period from *c*. 1850 to 1950. Marginal zones farther to the north and south of Jakobshavn Isbræ, however, show a slight advance during the last decades of the 20th century (Fig. 43). The regional monitoring of glaciers in West Greenland on the basis of aerial photographs stopped with the last full coverage flown in 1985; subsequent regional coverage has been based on satellite imagery. Modelling of present response patterns of the ice sheet to climatic changes seems to match the present patterns of recession and readvance, and an important thickening of the south-western parts of the Inland Ice seems to have taken place (Huybrechts 1994). In contrast to the modelling, monitoring of outlet glaciers and marginal elevation changes based on repeated surveys by laser altimetry in 1993/1994 and 1998/1999 has revealed a significant thinning of the surroundings of Jakobshavn Isbræ, whereas the thickness of Jakobshavn Isbræ itself was constant or increasing (Abdalati *et al*. 2001). However, it may be misleading to compare the very generalised trends of the change of the ice margin positions based on scattered historical information with the detailed trends revealed during the past few decades (see e.g. Thomas *et al*. 2003 and Joughin *et al*. 2004 for Jakobshavn Isbræ).

With respect to the response of ice streams from the Inland Ice to climatic forcing, a holistic modelling approach has been presented by Hughes (2004), which lengthens and lowers the profiles of the Greenland ice streams.