Submarine evidence of the Aavatsmark and Dahl Glaciers fluctuations in the Kaffiøyra region, NW Spitsbergen

Marek GRZEŚ, Michał KRÓL and Ireneusz SOBOTA

Zakład Kriologii i Badań Polarnych, Instytut Geografii UMK, Gagarina 9, 87-100 Toruń, Poland
<gmark@umk.pl> <mkrol@stud.umk.pl> <irso@umk.pl>

Abstract: Although the terrestrial marginal zones of some glaciers on Spitsbergen are relatively well described, we are largely ignorant about the morphology of their submarine forefields. Initial reconnaissance of the forefields of the Aavatsmark and Dahl glaciers in the Kaffiøyra region and soundings made in that of the Hans Glacier (southern Spitsbergen) indicate the occurrence of sea-floor push-moraines which can be as much as 3 m high. Their lateral separation is considered to denote annual recession rates. They appear to result from cyclical annual advances of ice-cliffs during winters when the deposits are risen up at the contact of the ice with the sea-floor. The development of the major forms may be related to surge. There is some evidence that certain elements in the sea-bed morphology date from the Little Ice Age (LIA).

Key words: Arctic, Svalbard, glacier, push-moraines.

Introduction

There are two types of glaciers which terminate in the sea: floating and grounded (tidewater) glaciers. The former do not rest on the sea-bed but, at least for the greater part, float on the sea. Their proximal positions are referred to as ice-fronts. The latter forms when the glacier is more or less totally grounded on sea-bed sediments. The resulting ice margin is called an ice-wall, or more frequently, an ice-cliff (Marsz 1987). These two glacier types are associated with quite different glaci-marine deposits in the aquatic environment, a reflection of controls by both environmental and marine conditions (*e.g.* temperature, sea currents and salinity) upon glacial sedimentation (Powell and Domack 2002).

In the Arctic, most glaciers which terminate in the sea are grounded tidewater glaciers (Powell and Domack 2002) and it is generally assumed that the dynamic processes of the ice margins usually play a significant role in the modelling of the sea-floor glacial relief. In west Spitsbergen, the investigations of underwater marginal zones require the application of relatively expensive special techniques; hence, these have been somewhat infrequent (Boulton 1986; Kowalewski *et al*.
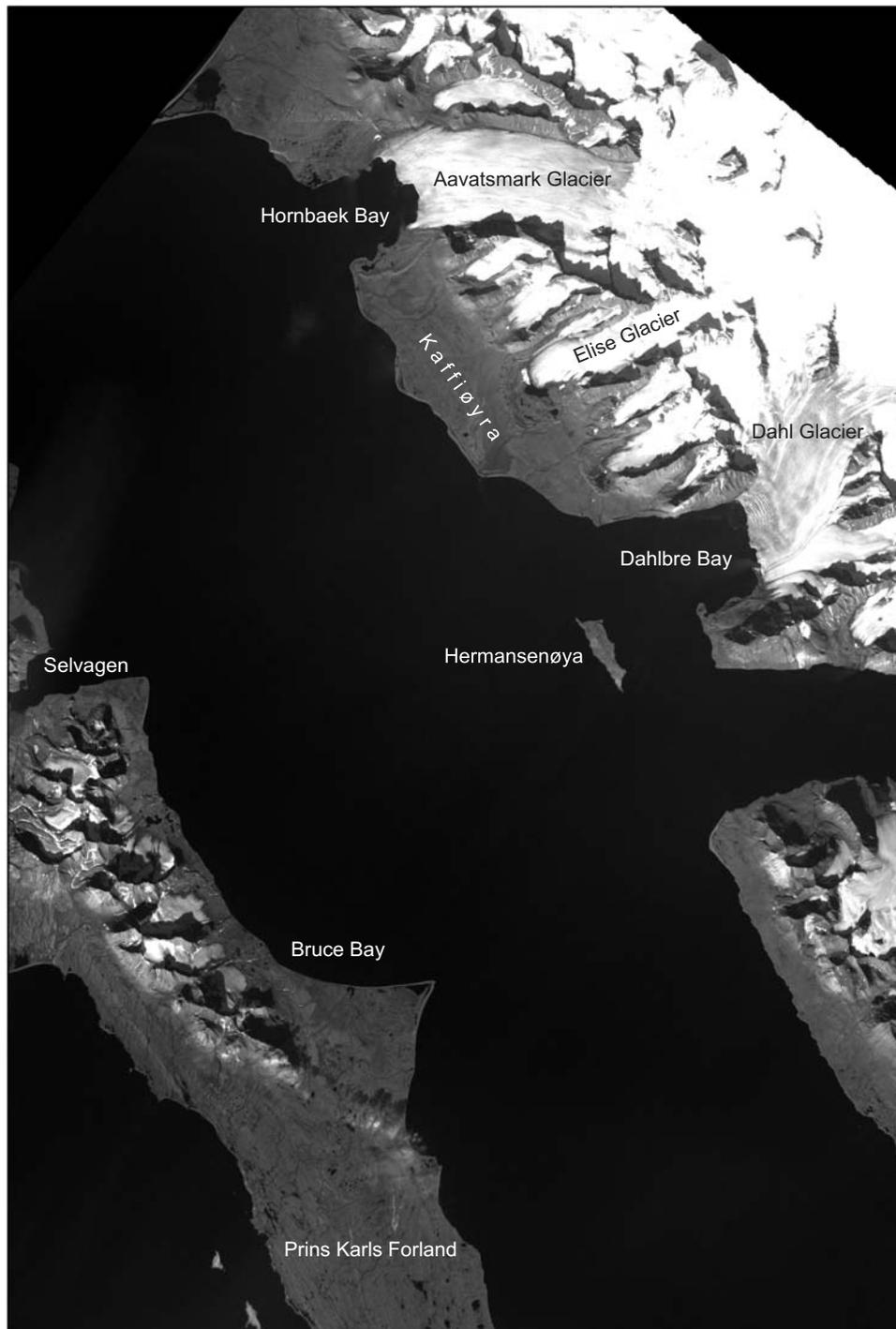
Fig. 1. Location map of the research area.

Fig. 2. Aavatsmark Glacier in 2008 (view from the north). *Photograph by P. Wyszyński*.

1987, 1990, 1991; Jania 1988, 1998; Marsz 1993; Rudowski *et al*. 1994; Rudowski 1998). The glaciers former extensions which terminated in the sea in historical times can be identified from archival maps, aerial photographs and by extrapolation, from the terrestrial parts of their marginal zones (Jania 1982, 1988, 1998; Landvik *et al*. 1987, 1992, 1998; Mangerud *et al*. 1998; Lankauf 2002).

In the summers of 2004–2006, echo soundings with trace logs were conducted in the forefields of the Aavatsmark and Dahl Glaciers in the Kaffiøyra region (Fig. 1). The Aavatsmark Glacier bounds the Kaffiøyra on its northern margin (Figs 2, 3). Its area in 1995 was 75 km^2 (Lankauf 2002). The Dahl Glacier drains an extensive area of highland ice caps and terminates in the sea an ice-cliff 35–40 m in height (Fig. 4). Together with its firn, the Løvenskioldfonna and lateral glaciers, it has an area of *ca* 132 km^2. Changes in the position of the Dahl Glacier front in the 20th century have been described by Lankauf (2002).

The main objectives of our research are to answer the following questions: (1) are the glaciers in the Kaffiøyra region now characterised by winter advances which lead to the development of the particular sea-floor relief forms, (2) do certain sea-floor forms correspond to the location of ice-fronts during known periods of advance, such as the LIA, the glacial episode (3.0–2.5 ka B.P.), or the Late Vistulian (13–10 ka B.P.), (3) which forms are related to the periods of significant glacier advances, (4) which forms develop as a result of annual winter advances of the glaciers, (5) which forms develop during a surge, (6) can the bathymetry of the bays in which the glaciers terminate impose a significant restrain to a glacier advance.

Fig. 3. Aavatsmark Glacier in 2008 (view from the south). *Photograph by A. Nitowski*.

Fig. 4. Dahl Glacier in 2003 (view from the south). *Photograph by S. Nowak*.

Methods

The echo soundings in the forefields of the glaciers were conducted by means of an echosounder which was integrated with a Garmin GPS Map 178C receiver. The potential error of any GPS measurement was 5 m. The long profiles were established by means of a trace record, and the data were compiled using GPS Utility 4,20.4 and Origin 7.5 software. The number of measurement points in a profile ranged from 300 (profile A-B of the Aavatsmark Glacier) to >600 (profile G-H of the Aavatsmark Glacier). Tidal changes which can be as much as 1–2 m in the sea level were disregarded.

The position of the Aavatsmark Glacier in 1936 was taken from topographic maps (Kongsfjorden 1989). The positions of the ice-cliff in all other years were taken from topographic maps of Lankauf (2002). A knowledge of the bathymetry of Hornbaek Bay makes it possible to establish the putative maximum extent of the glacier during the Late Vistulian (13–10 ka B.P.).

The locations of the front of the Dahl Glacier between 1936 and 1986 were taken from topographic maps (Prins Karls Forland 1989). The positions from the turn of the 19th century and from 1909 were taken form Lankauf (2002). The putative position of the ice-cliff in the Late Vistulian (13–10 ka B.P.) was taken from Forman (1987) and Niewiarowski *et al.* (1993).

Results

Aavatsmark Glacier. — In 2006, the recession was recorded on the entire length of the glacier front. In the years 2000–2006, the recession of the central parts of the front ranged from 360 to >540 m. In the lateral zones of the partly submerged ice-cliff, the retreat was >230 m in the northern part and >150 m in the southern.

Six ridges averaging 3 m (0.2–7.8 m) and, on average 35 m wide (19–73 m) were identified in the area vacated by the glacier between 1995 and 2000 (Fig. 5). The ridges are approximately 25–30 m distant from each other, *i.e.* nearly the same as the mean annual recession over a recent multi-year period. In the northern marginal zone, which was occupied by the glacier in the years 2001–2004, four ridges 1.7–2.4 m high and 25–32 m wide are present. The distance between individual ridges increased from 38 m (between 2001 and 2002) to 85 m (2003 and 2004), indicating that the glacier recession rate is accelerating here (Fig. 6). These forms appear to be push-moraines, which are generally believed to develop as a result of an ice-push of glaci-marine material during winter frontal advances (Boulton 1986; Jania 1998).

Lankauf (2002) claimed that, between 1978 and 1985, the Aavatsmark Glacier was in a phase of accelerated movement *i.e.* it was a surge. During that period, the ice-cliff moved forward, by as much as 250 m locally (Lankauf 2002). Our profiles, particularly those in the southern frontal part, show several forms which we regard as traces of the surge from the years 1978–1985. We assume that push-moraines might
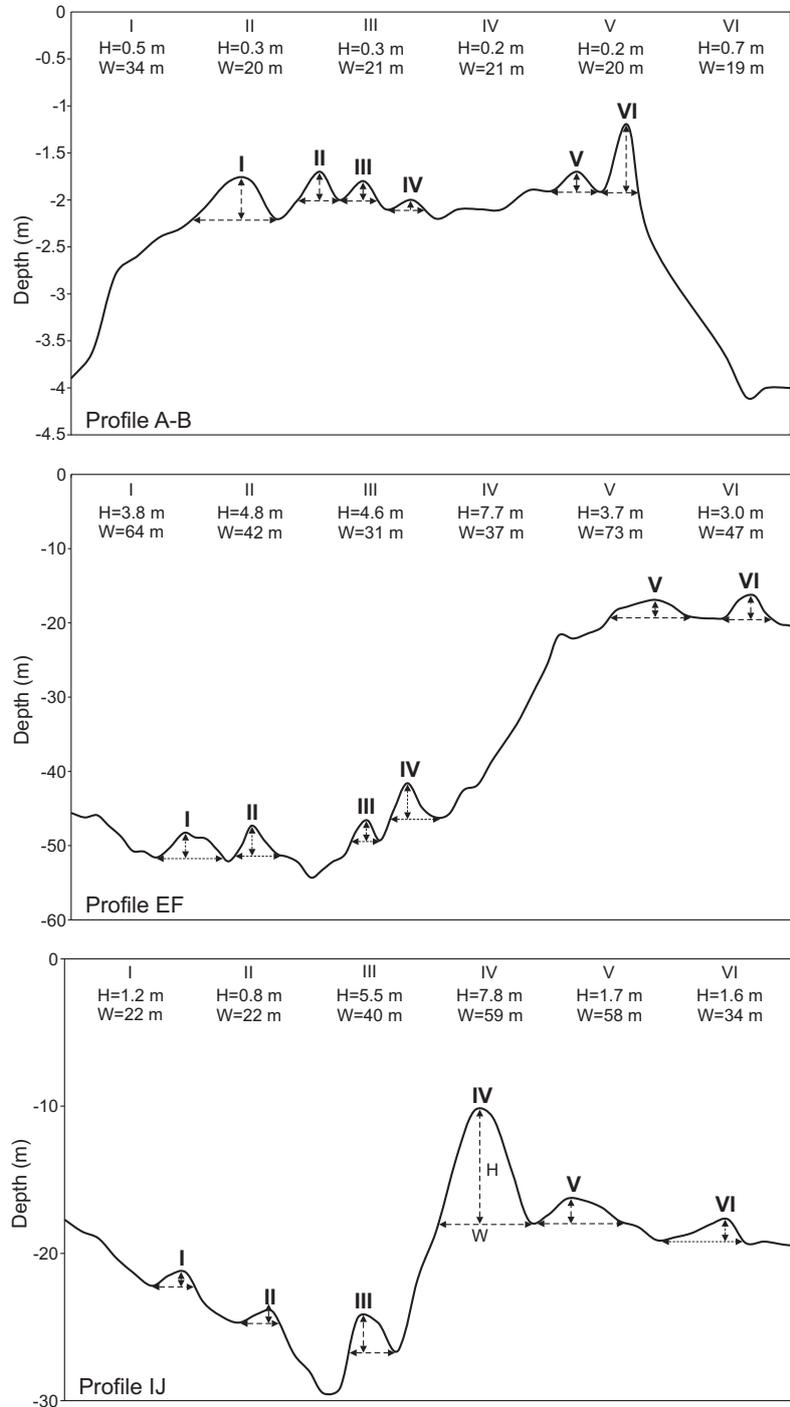
Fig. 5. Characteristics of the selected annual push-moraines of the Aavatsmark Glacier formed be-
tween 1995 and 2000 (H – height, W – width).
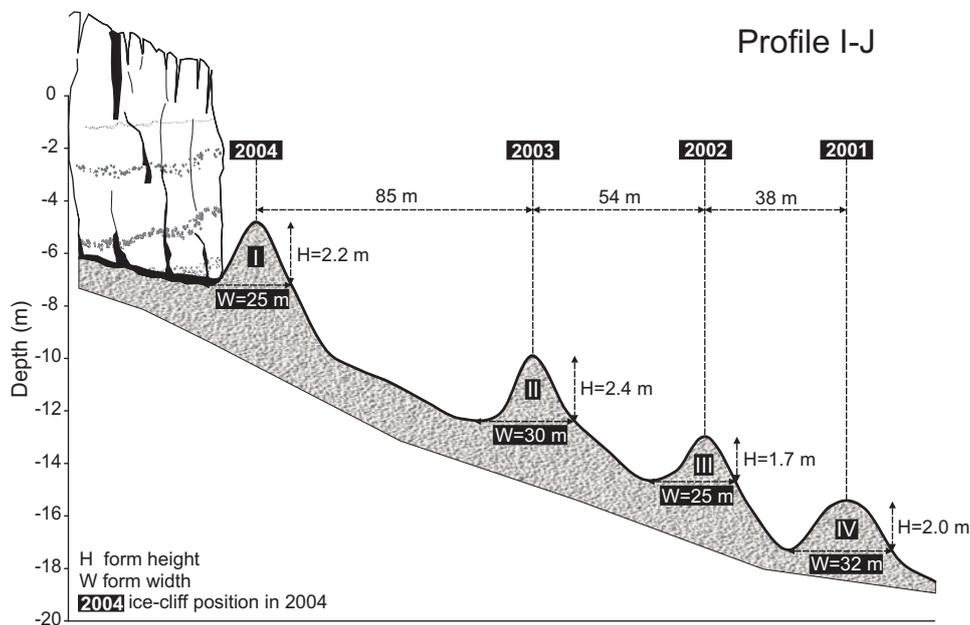
Fig. 6. Annual push-moraines of the Aavatsmark Glacier formed between 2004 and 2001 (H – height, W – width).

have developed as a result of this surge. Certainly, these forms are comparable with those resulting from the annual advance of the cliff during winters. However, owing to the limitations of our survey methods, it is not yet possible to distinguish the forms developed by surge from those resulting from annual advances of the ice-cliff.

From our profiles and the bathymetry plan (Fig. 7), the position of the front in 1936 may easily be proposed and we consider that the distribution and location of sea-floor moraines relate to the shape and position of the ice-cliff at that time (Figs 7–13).

The glacier left a very distinctive trace in the sea-bed relief in 1909 (Figs 9–13), the profiles showing distinct shallows from 6.5 m (profile CD) to 29 m (profile GH). The sea-bed relief forms relating to the 1909 front are complex. They include moraines which are typically 5–18 m high (profiles EF and GH) and 90–240 m wide (profiles CD and IJ). The LIA also left a clear trace in the sea-bed relief (Figs 7–13), which is easily correlated with terrestrial deposits dating from that time (Lankauf 2002). The bed relief in Hornbaek Bay possesses a notable succession of push-moraines which are 6–17 m high (profiles IJ and AB) and 160–220 m wide (profiles CD and IJ) (Fig. 8).

Dahl Glacier. — In the years 1986–2005, the Dahl Glacier recession averaged 310 m in the northern part of its front and 100 m in the central part of the glacier (Fig. 14).
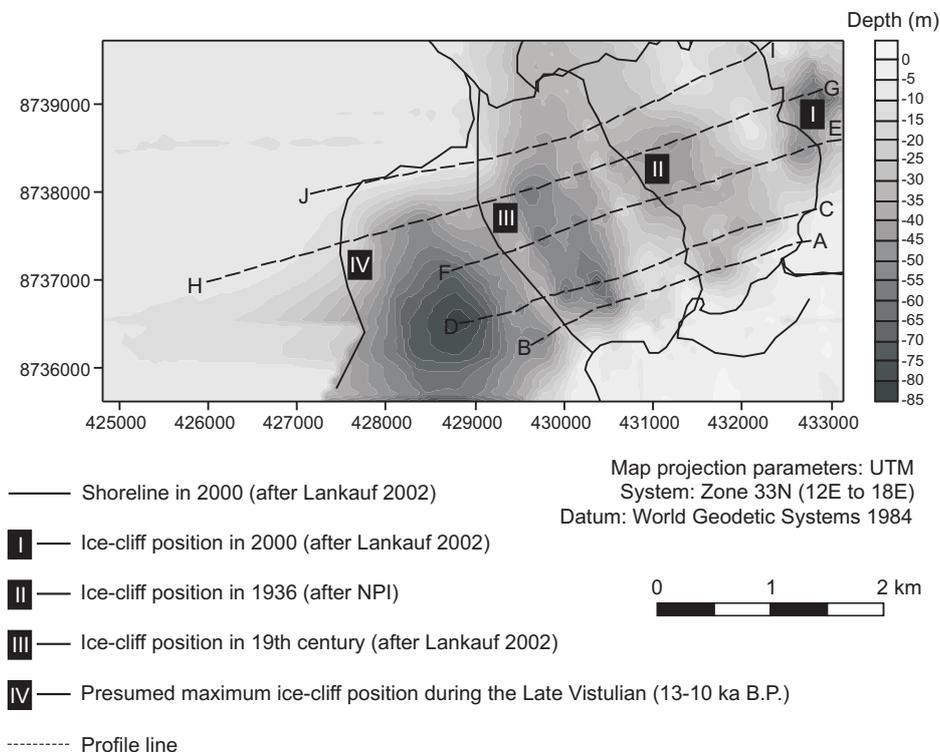
Fig. 7. The bathymetry of the Hornbaek Bay with adjoining basins of the Forland Strait and the ice-cliff position of the Aavatsmark Glacier in different periods.

Presented echo-sounder profiles from beyond the margin of Dahl Glacier strongly suggest the absence of small moraines on the sea-bed which were produced annually. In the sea-bed relief, the frontal position in 1936 is delineated by a distinct ridge which is 20 m high and 230–350 m wide. This shallow area is only 50 m (profiles AB and CD) and >70 m in depth (profile EF) (Figs 15–17).

The position of the glacier front from the 19th century (LIA), as determined by Lankauf (2002), is quite distinctive in the sea-bed relief. This local shallow area is only 10–30 m depth. Towards the southwest of this shallow, there is a sudden increase of depth up to 120 m. The record of the ice-cliff extent at the turn of 19th century is obvious (Figs 15–17). The glacier's recession from the period of the LIA to 2005 was >4 km (profiles A-B and C-D) and 4.7 km (profile E-F) (Fig. 14). Beyond the line which marks the frontal position at 19th century, there are relief forms which may indicate the glacier's position during the glacial episode of 3.0–2.5 ka B.P. These forms are impersistant ridges which are <40 m high (on average 17 m) and 160–300 m wide (Figs 15–17). Significantly, the presumed ice-cliff position during the glacial episode of 3.0–2.5 ka B.P. corresponds to the line of the Bregrunnen and Farmgrunnen shallows. From the period of 3.0–2.5 ka B.P. to 2005, the Dahl Glacier probably retreated by *ca* 5.5 km overall (profiles A-B and C-D).
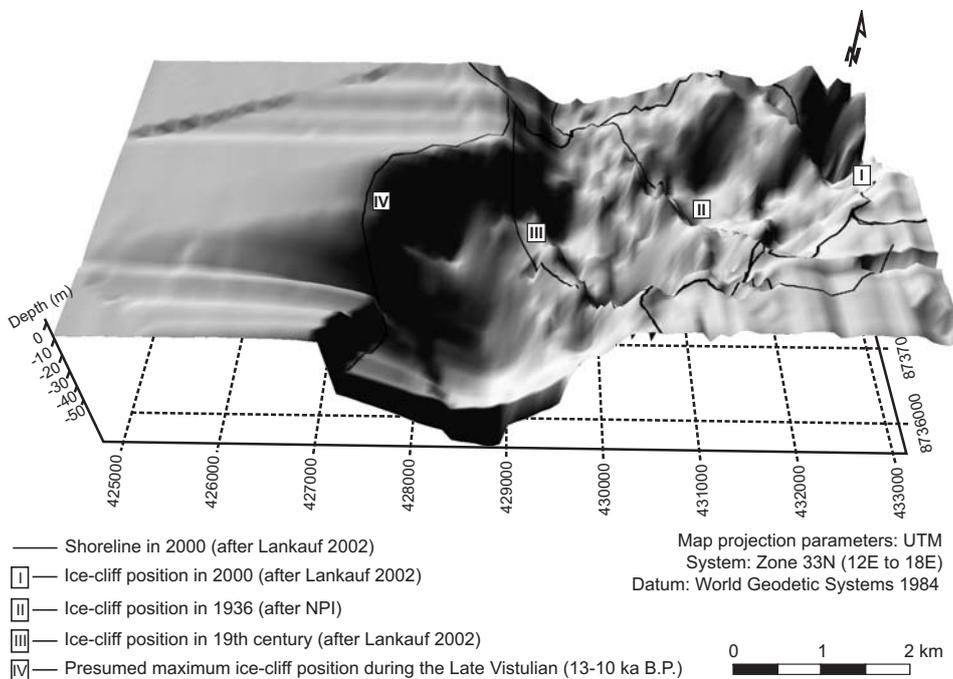
Fig. 8. Spatial drawing of the sea-floor relief in the forefield of the Aavatsmark Glacier and the ice-cliff position in different periods.

Discussion

On the basis of literature and archival cartography, it was possible to discuss changes in the extents of the Aavatsmark and Dahl Glaciers, while analysis of the forms of the bottom relief and terrestrial marginal zones led to an attempt of defining older extents of their ice-fronts. The initial results of our investigations indicate that the sea-floor relief in the forefields of these two glaciers is the product of the pushing up of glaci-marine materials in a relatively stable ice-cliff zone. Selected profiles drawn perpendicular to the fronts clearly show forms which are related to both seasonal and multi-year oscillations. The seasonal forms are 2–5 m high, and their deposition may have taken place during winter advances of the ice-cliff. Such a mechanism has been described by Boulton (1986) and Jania (1988, 1998). Marsz (1987) reasoned that, when the position of an ice-cliff was stable, a substantial terminal moraine formed at its base. However, our investigations in the submerged marginal zone of the Aavatsmark Glacier show that these moraines (which can be as much as 3 m high and 30 m wide) may form as a result of surging. They are otherwise similar to those which develop during annual advances of the ice-cliff during winter. Brown *et al*. (1982), Boulton (1986) and Jania (1986, 1998) showed that, regardless of the general retreat most of the Spitsbergen ice-cliffs, in such periods of time as a hydro-
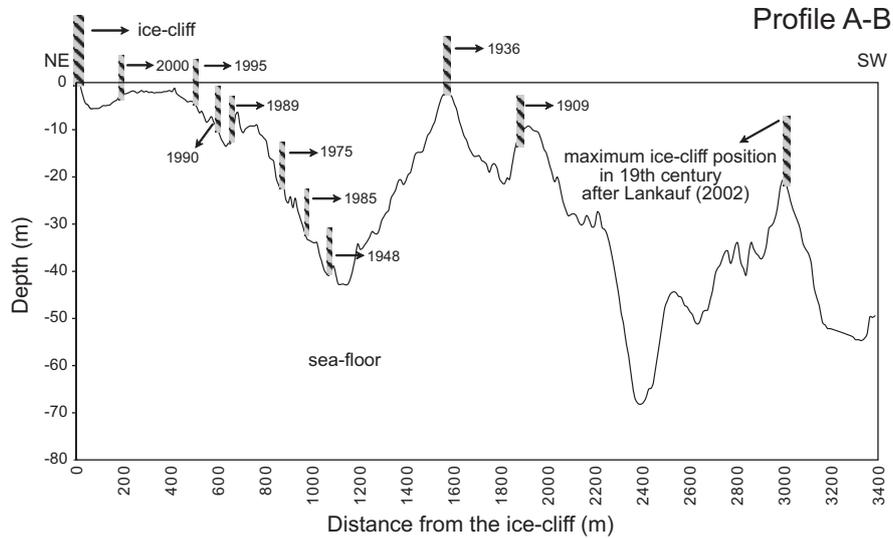
Fig. 9. Sea-floor relief in the forefield of the Aavatsmark Glacier, profile A-B.
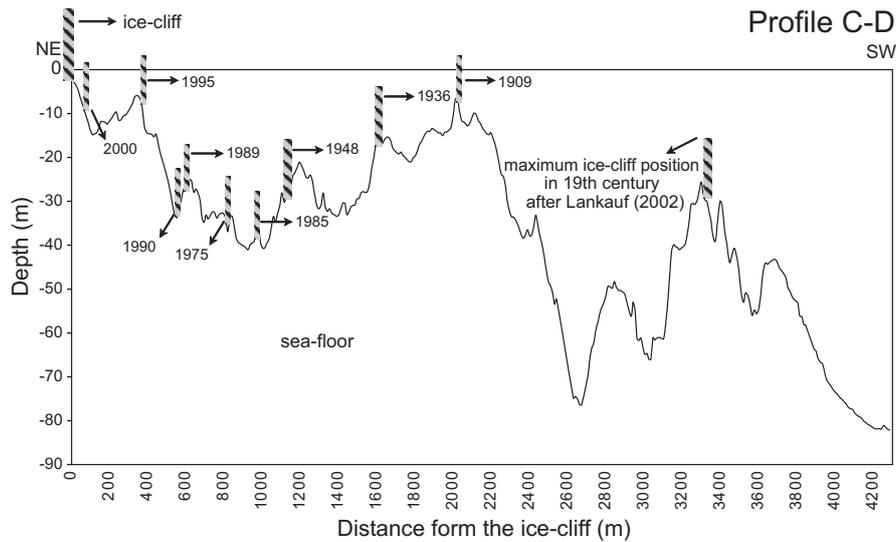
Fig. 10. Sea-floor relief in the forefield of the Aavatsmark Glacier, profile C-D.

logical year or a balance year, there are conspicuous oscillations which are difficult to explain as annual effects. Furthermore, these oscillations are several times larger than those attributable to a year-on-year recession. This implies that, during a balance year (most frequently in May and June), there is an advance of the glacier consequent upon winter accumulation and only limited calving of the ice-cliff at that time. Annual fluctuations of the stable ice-cliffs are recorded in the sea bottom relief. The pushing up of glaci-marine sediments at the contact of the ice with the sea-floor
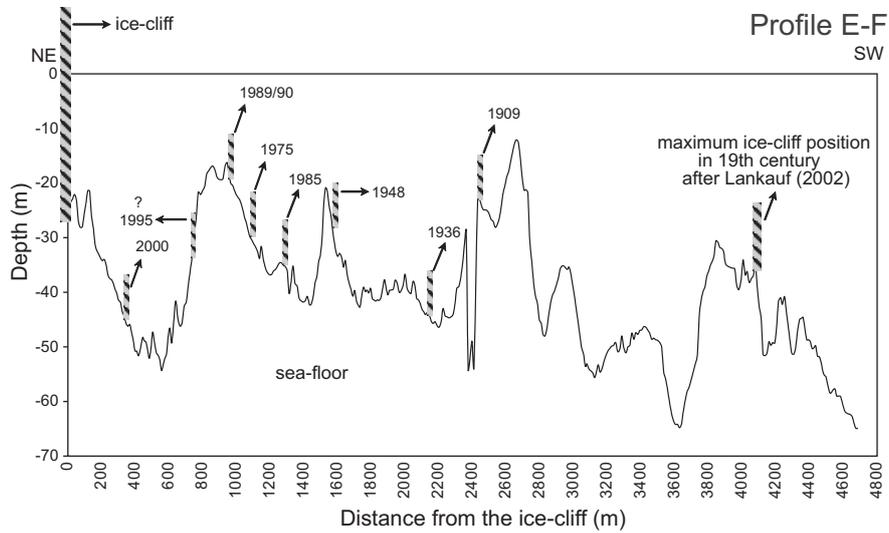
Fig. 11. Sea-floor relief in the forefield of the Aavatsmark Glacier, profile E-F.
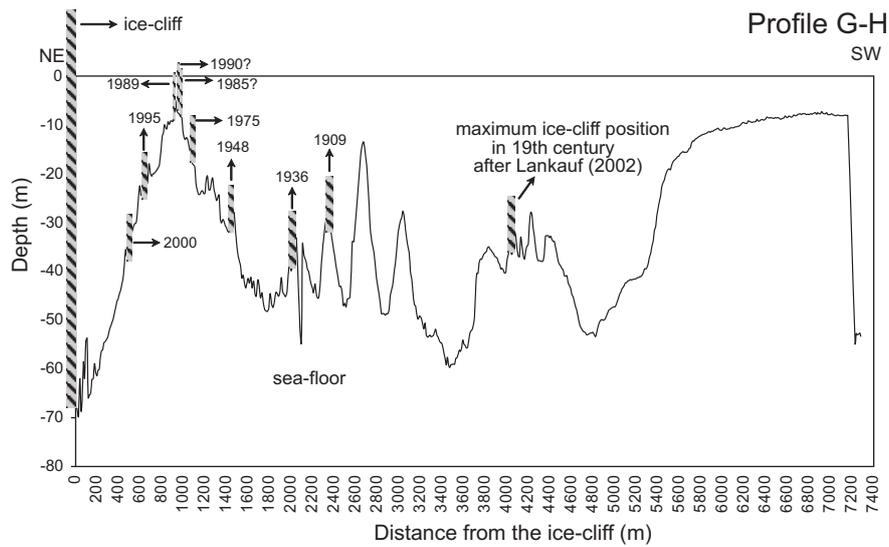
Fig. 12. Sea-floor relief in the forefield of the Aavatsmark Glacier, profile G-H.

seems to have the biggest impact upon the development of these relief forms, which result from both annual ice-cliff oscillations and surges. With regard to the grounded tidewater glaciers, during the winter advances, its adhering glacial and meltwater deposits are pushed to form small annual push-moraines, the distance between them being frequently equal to the mean annual recession. The accumulation of the larger moraines in both subareal and submarine conditions, is mainly related to the intense movement dynamics and a prominent surge (Jania 1988). According to Powell
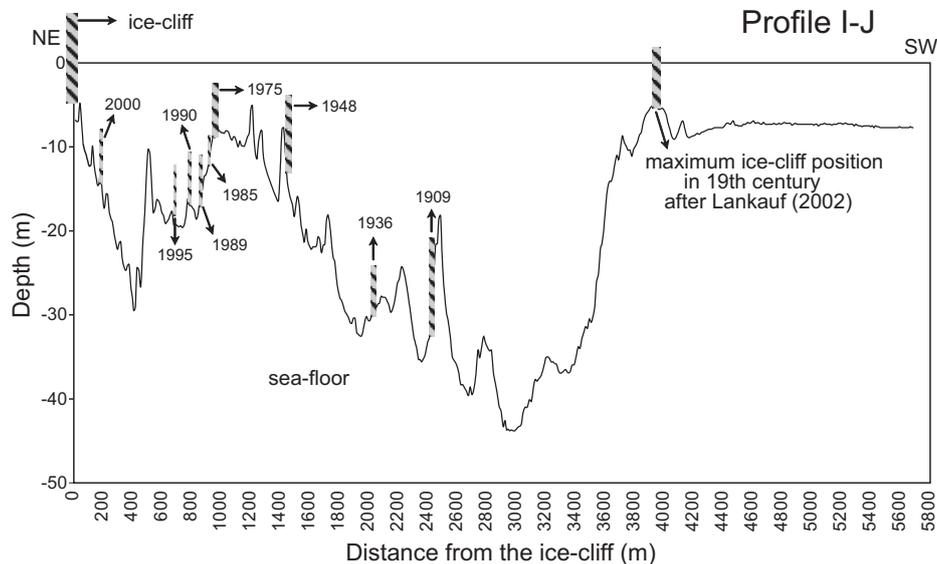
Fig. 13. Sea-floor relief in the forefield of the Aavatsmark Glacier, profile I-J.

(1984) and Smith (1990), the pushing up of glaci-marine deposits is directly associated with seasonal changes in the locations of the ice-cliff. Push-moraines may form both on land and in the marine environment (Boulton 1986; Jania 1998). Boulton (1986) included a detailed description of those forms and he quoted the Aavatsmark Glacier as an example of a glacier which had produced conspicuous push-moraines.

Soundings conducted in the forefield of the Hans Glacier (southern Spitsbergen) showed the existence of submarine annual push-moraines up to 2 m high. Their crests lie *ca* 40 m distant one from another (Jania 1998). According to Jania, these forms develop as a result of pushing deposits during the seasonal advance of the front during winters.

Rudowski (1998) discussed the problem of glaci-marine sedimentation in the fiords of Spitsbergen. On the basis of seismo-acoustic analysis of the Quaternary bed deposits, he distinguished four basic seismic units, by means of which he created a spatial model of the glaci-marine sedimentation environment here. With regard to the marginal zone of sedimentation, Rudowski (1998) considered that the sea-floor covers and ridges of terminal moraines were forms developed from the position of the glacial debris carried from the frontal part of the glacier by supraglacial, englacial and subglacial waters and mass gravitational movements. He also distinguished glaci-marine fan complexes formed from suspension flows. The sea-bed in the marginal zone is hummocky, with small (<5 m high) ridges of recessional moraines. He considered that beds of lodgement till present here are probably related to the reces-

Fig. 14. Changes in the position of the Dahl Glacier front and putative ice-cliff position during the →
glacial episode of 3.0–2.5 ka B.P. and the Late Vistulian (13–10 ka B.P.).

Profile A-B
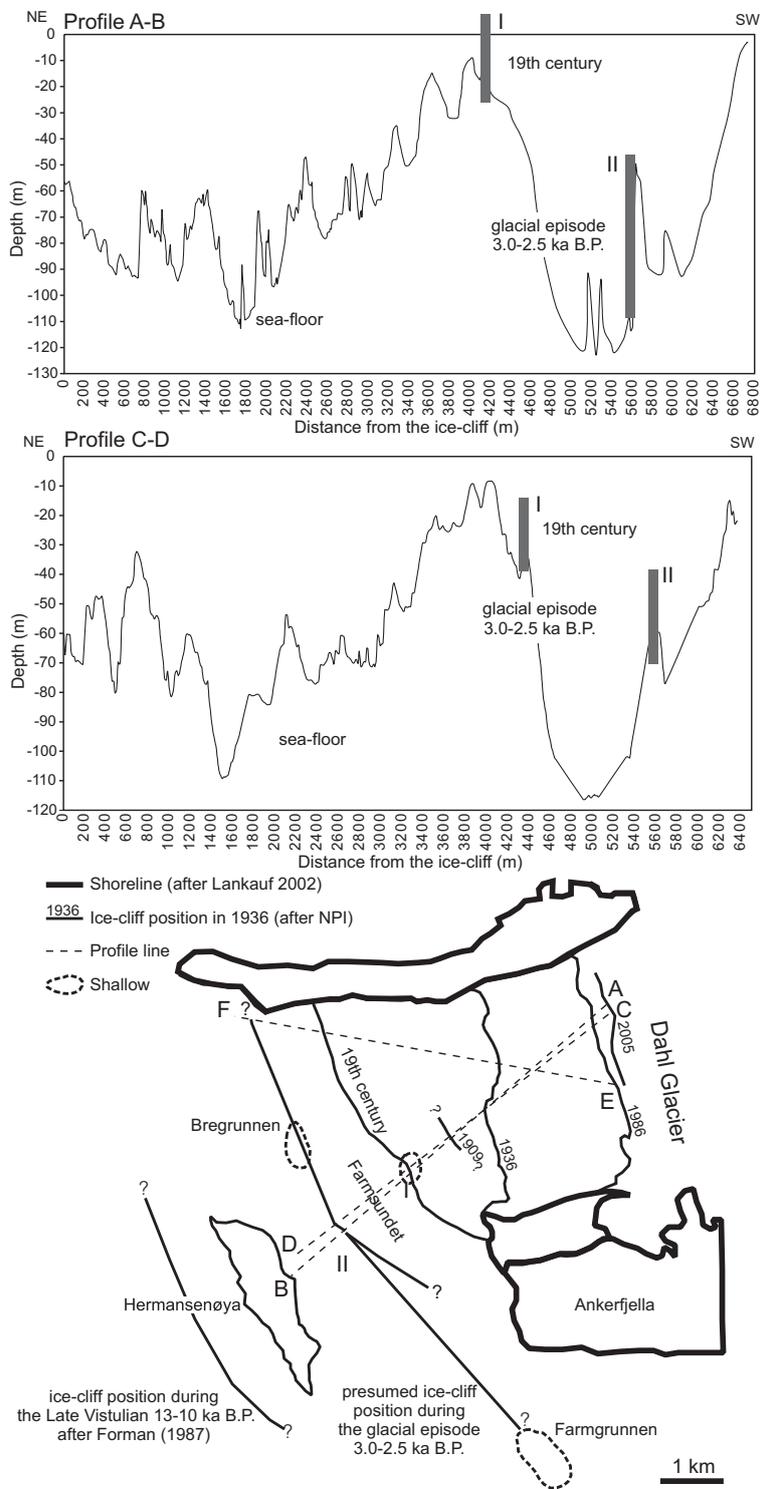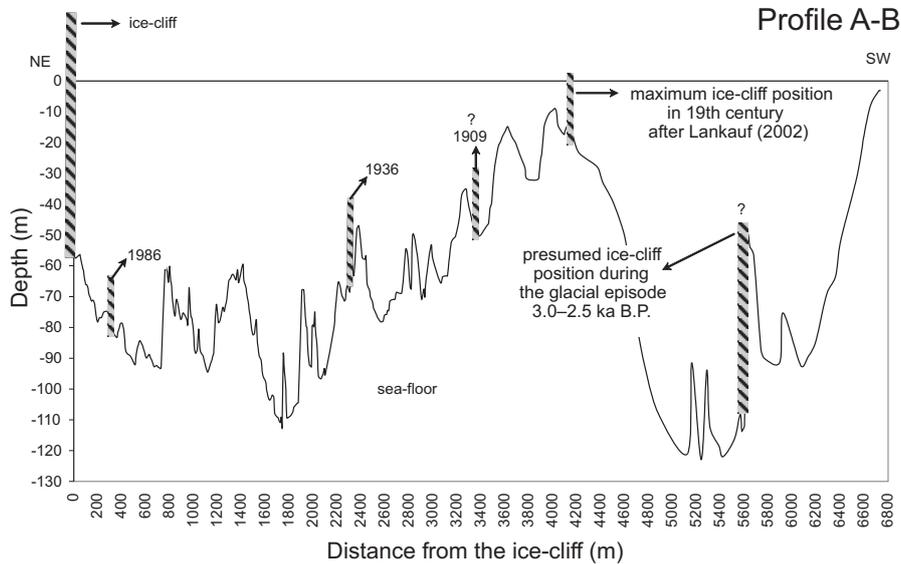
Profile C-D

Fig. 15. Sea-floor relief in the forefield of the Dahl Glacier, profile A-B.
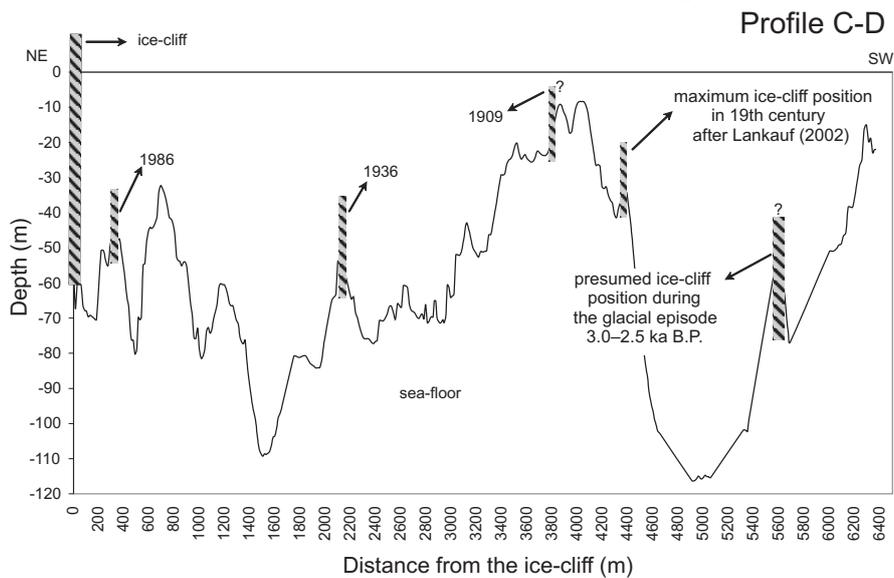
Fig. 16. Sea-floor relief in the forefield of the Dahl Glacier, profile C-D.

sional phase of the glaciers, which persisted until *ca* 9 ka B.P. Still preserved on the floors of the inner basins, marginal fan complexes evolved during the period 10–9 ka B.P. These were disturbed when glaciers advanced during the LIA. At that time, the glaciers reached the present border, where they formed a thick complex of morainic ridges (with heights up to 50 m) composed of glacial deposits. It was then that the marginal fan complexes are considered to have formed, that were as much as 200 m
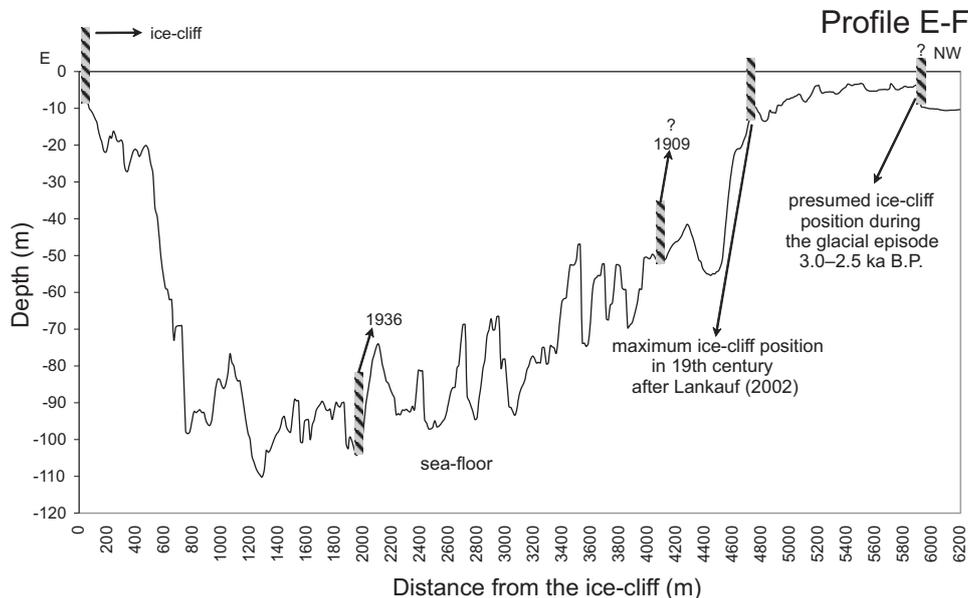
Fig. 17. Sea-floor relief in the forefield of the Dahl Glacier, profile E-F.

thick (Rudowski 1998). According to Rudowski (1998), these forms constitute the terminal moraines of the maximum extent of the glacier during the LIA. Many authors have drawn attention to the clearly very obvious traces of the LIA in the relief. Some (*e.g.* Kowalewski *et al*. 1987; Marsz 1993; Rudowski 1998) have attempted to explain the genesis of the older glacial deposits.

Forman (1987) conclude that during the Late Vistulian (13–10 ka B.P.) the Aavatsmark Glacier did not advance more than 1–3 km beyond its present location. If this is so, it is puzzling that the advance of the Aavatsmark Glacier was so limited when compared with that of the Dahl Glacier. We conclude that particular shape of the sea-bed relief in Horbaek Bay must have had a restraining influence on the advance of the Aavatsmark Glacier during the Late Vistulian, a critical factor have being the zone of sudden shallowing (3–10 m depth) at a distance of *ca* 1.3–1.5 km beyond the line of the glacier front during the LIA (Fig. 8). During the 19th century, the glacier grounded in this zone, and did not subsequently surmount it. During the Late Vistulian, this shallow zone also appears to have formed a natural barrier for the advancing glacier. The putative position of the glacier front in the period 13–10 ka B.P. was plotted (Figs 7, 8) on the basis of the form of this shallow zone. The glacier front so delineated the recession was *ca* 1.5 km (at the glacier's axis) implies that with respect to the LIA.

Our soundings clearly have a bearing on old controversies about the Dahl Glacier's range during the glacial episode of 3.0–2.5 ka B.P. The glacial deposits of the Dahl Glacier of the island of Hermansenøya which date from the Magdalenafjorden stadium (*ca* 2.5 thousand years ago) have been described by Szupryczyński (1983).

Thicker beds of the Late Vistulian and pre-Vistulian deposits discovered by Forman (1987) can be found in a 200 m long coast in the northern part of Hermansenøya, the age of which was determined as 9825 ± 90 B.P. However, nothing demonstrates younger than this has yet been discovered. Niewiarowski *et al.* (1993) considered that the Dahl Glacier advanced several times onto the island during the Vistulian. However, as with to Forman (1987), they could find no trace a glaciation which relates to the period 3.0–2.5 ka B.P. Such traces were found only in the marginal zones of the Aavatsmark and Eliza Glaciers. Niewiarowski (1982) conclude that considering the data from the forefield of the Elise Glacier it seems most probably that the relics of the older end moraines occurring in the forefield of the Aavatsmark Glacier come from the advance of about 3.0–2.5 ka B.P. They show that in the Kaffiøyra region an advance of greater glaciers at that time was of a similar extent as during the LIA. We discovered that, beyond the line which marks the frontal position at 19th century, there are relief forms which may indicate the glacier's position during the glacial episode of 3.0–2.5 ka B.P. These forms, which occur along the line of the Bre-Farm shallows, are impersistant ridges which are <40 m high (on average 17 m) and 160–300 m wide (Figs 15–17). Significantly, the position of the ice-cliff in the LIA also corresponds to this line of shallows (Fig. 14). It is concluded that the glacier could not have come closer to Hermansenøya during the glacial episode of 3.0–2.5 ka B.P. than 1 km. From the period of 3.0–2.5 ka B.P. to 2005, the Dahl Glacier probably retreated by *ca* 5.5 km overall (profiles A-B and C-D). However, during the Late Vistulian, there was an expansion of the glacier westwards, as indicated by the thicker beds of glacial deposits dating from 9825 ± 90 B.P. However, the exact extent and direction of this expansion are yet to be established. We consider that the Dahl Glacier probably terminated in deep water (>100 m) west of Hermansenøya (Fig. 14). It is, therefore, desirable that in order to look for evidence of Late Vistulian glacial expansion, our studies should be extended into this region.

Conclusions

The sea-floor relief appears faithfully to record changes in glacial dynamics. The sequence of forms and their concurrence with documented positions of the glacier front indicate their glacial genesis. The ice-cliff position in the LIA, 1909 and 1936 left a clear trace in the sea-bed relief, which is easily correlated with terrestrial deposits dating from that time. Within the sea-floor relief, it is also possible to distinguish forms which have resulted from the annual oscillations of the ice-cliff, and, more precisely, which have resulted from glaci-marine material which has pushed up at the contact of the ice with the sea-floor. These forms reach a height of 3–5 metres. The development of the larger forms may be related to surge advances. In the dynamic marine environment at the contact of the ice and the sea-floor, several processes are of important *e.g.* outwashing, lodgment of the

deposits on the grounding line, material melt-down and deposition directly from the ice-cliff or from disconnected icebergs. However these processes have a relatively minor impact upon the evolution of the sea-bed relief.

When the succession of glacial events in the Førland Strait is considered we presume that the outermost position of the Dahl Glacier front, as recorded in the bed relief, may have originated in the un-named glacial episode (3.0–2.5 ka B.P.). During this time, glacier could not have come closer than 1 km to Hermansenøya. We support Forman's (1987) views in respect of glaciation in this region during the Late Vistulian (13–10 ka B.P.). According to him, the Dahl Glacier extended to the island of Hermansenøya, as indicated by the glacial deposits present there which are apparently 9825 ± 90 B.P. old. By contrast the Aavatsmark Glacier did not apparently advance in the Late Vistulian further than 1.5 km from the line of frontal position during the maximum at 19th century. We consider that this is because the glacier was restrained by a continuous zone of shallow water (5–10 m deep), which can be found *ca* 1–1.5 km from the frontal line during the LIA. This formerly constituted a natural boundary for the expansion of the glacier, even during substantial advances.

The next logical step in these investigations would be to analyse cores of the Hornbaek Bay sea-bed sediment and apply geophysical investigations. However, this is, at present, beyond the resources of this project.

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