

# Thermal regime of a valley glacier, Erikbreen, northern Spitsbergen

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The thermal regime of the valley glacier, Erikbreen, northern Spitsbergen (79°40'N, 12°30'E) was studied using radio-echo sounding and temperature measurements from eight boreholes ranging from 13.5 to 24 m. Radar images indicate a glacier with a two-layered thermal structure. A surface layer of cold ice, 20 to 60 m thick along the centre flow line, extends from an altitude above equilibrium line to the glacier front. This layer represents 20 to 35% of the glacier thickness, except at the floating front, where the cold layer is about 50%. The ice beneath the cold surface layer is interpreted to be temperate. Cold-based areas exist near the glacier margin and in some locations in the accumulation area; the ice is interpreted to be entirely temperate in central parts of the accumulation area at high altitude. Freezing of temperate ice at the base of the cold surface layer is probably the main mechanism of cold ice formation in the frontal parts of Erikbreen. Calculated heat fluxes based on the borehole measurements show that a steady state cold layer 25 to 30 m thick is likely, assuming a surface melting of 1.7 m/y and a maximum water content of 3%. In the frontal parts the calculated mean annual upward heat flux at 10 to 15 m depth is roughly 0.6 W/m<sup>2</sup>.

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## Introduction

Previous measurements of glacier ice temperatures have shown that several glaciers of the Svalbard archipelago have a cold surface layer. Schytt (1969) presents an early intuitive statement on the subpolar nature of these ice masses. Liestøl (1976) gives a general description of the thermal regime of glaciers, and relates the regime to permafrost conditions in Svalbard. Several 10 m depth temperatures available from the ablation area of Finsterwalderbreen (Fig. 1) demonstrate that the surface is cold (Nixon et al. 1985). Measurements in deep boreholes in Austre Brøggerbreen, Midre Lovénbreen and Kongsvegen near Ny-Ålesund (Fig. 1) in Spitsbergen document a cold surface layer overlying a temperate layer (Bjørnsson et al. unpubl.; Hagen, Norsk Polarinstittutt, unpubl.).

Radio-echo sounding (Dowdeswell et al. 1984a, b; Kotlyakov & Macheret 1987; Bamber 1987a, b; Bjørnsson et al. unpubl.) show that an internal reflection horizon is common on Svalbard glaciers. This horizon has been interpreted to be a boundary between surface cold ice and deeper ice containing liquid water. Water inclusions are

known to have a strong influence on the absorption and scattering of radio waves, depending on frequency and bandwidth (Smith & Evans 1972). In the most recent work by Bjørnsson et al. (unpubl.), the results from radio-echo sounding are compared with thermistor measurements in several deep boreholes in glaciers in the vicinity of Ny-Ålesund. These data indicate that the presence of temperate ice can be mapped based on the backscatter from water inclusions, and that the internal reflection horizon can be interpreted as the actual transition zone between cold and temperate ice. However, radio-echo data and temperature measurements from Austre Brøggerbreen indicate that it is difficult to detect a thin layer of temperate ice close to the bedrock. The interpretation of the radio-echo data presented in this paper is similar to interpretations from Storglaciären in Sweden (Holmlund & Eriksson 1989).

The aim of our study was to map the extent of cold ice in a polythermal glacier using radio-echo sounding and ice temperature measurements with thermistor cables. Based on the temperature data, the mean annual temperature gradient and heat flux at 10 to 15 m depth were estimated. This then

led to a discussion of the mechanisms controlling the extent and thickness of the cold surface layer of Erikbreen.

### Location and description of Erikbreen

Erikbreen is situated on the north shore of Liefdefjorden on the northern coast of Spitsbergen (79°40'N, 12°30'E) (Fig. 1). The nearest meteorological station is in Ny-Ålesund, 80 km to the south, where the mean annual air temperature is

-6.0°C (1971-1988) and average July and August temperatures are 5.2°C and 4.1°C, respectively. Permafrost depths are known to vary from 150 to 500 m (Liestøl 1976).

Glaciological investigations on Erikbreen were undertaken in 1990, 1991 and 1992. Radio-echo sounding took place in the spring of 1991. In addition to mapping the thermal regime, studies were also carried out on mass balance, ice surface flow, water discharge, suspended sediment load, dissolved material, and glacier channel geometry (Vatne et al. 1992; Etzel Müller et al. in press).

Erikbreen is a valley glacier of 9.0 km<sup>2</sup> (1990)

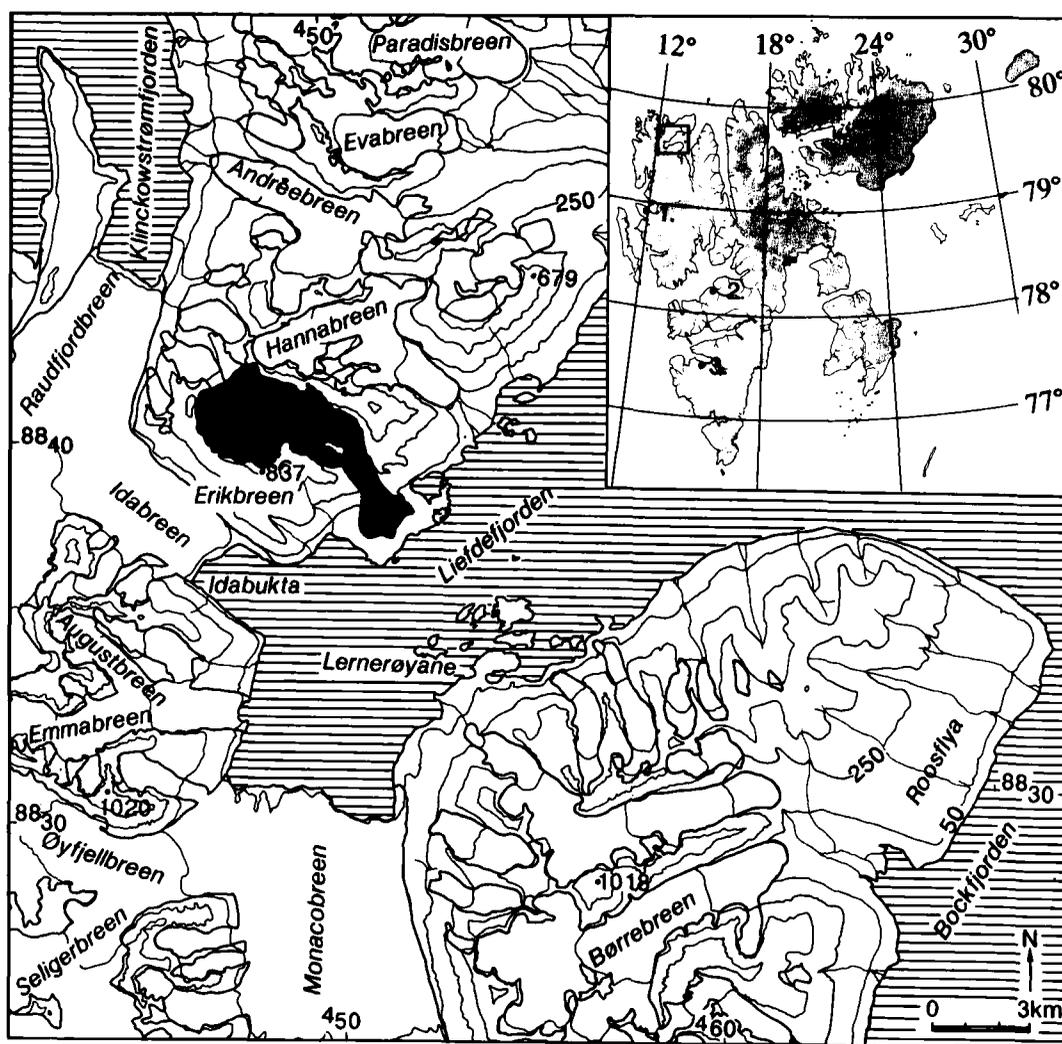


Fig. 1. Location of Erikbreen. Map compiled from base map 1:100 000, Norsk Polarinstittutt. Inset map: 1. Ny-Ålesund. 2. Longyearbyen. 3. Finsterwalderbreen.

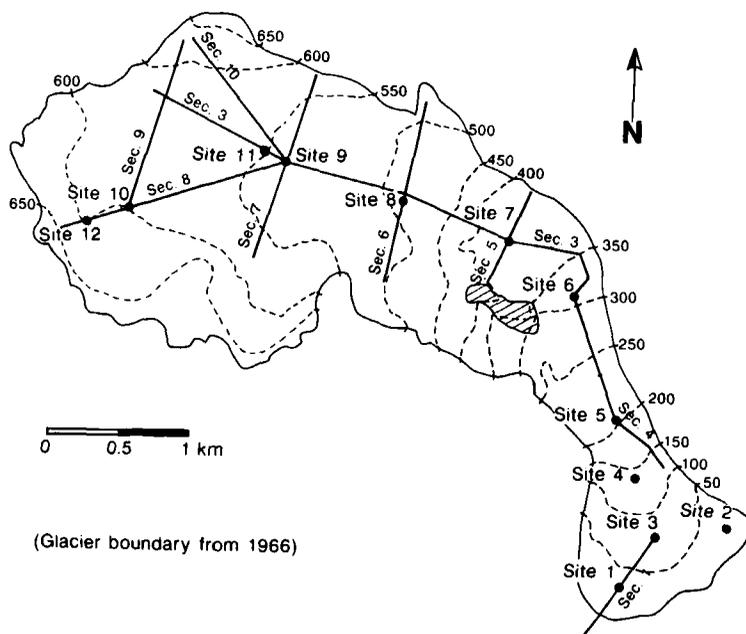


Fig. 2. Map of Erikbreen showing the radar sections and sites where measurements were made. Shaded area is a nunatak. Temperature measurements were made at sites 1, 2, 3, 7, 9, 10, 11 and 12.

with a piedmont-like front (Fig. 2). The total drainage basin area is 12.4 km<sup>2</sup>. The equilibrium line altitude was at 550 m a.s.l. in 1990 and at 340 m a.s.l. in 1991. Based on comparison with long-term measurements made on Austre Brøggerbreen near Ny-Ålesund, the summer balances in 1990 and 1991 were probably close to the maximum and minimum years respectively (Hagen & Liestøl 1990; Hagen pers. commun.). The summer balance from 11 July 1990 to the end of the season was -0.95 m. In 1991 the summer balance from the start of the melting season through 18 August was -0.46 m. Net surface melting of ice close to the terminus is 1.7 to 2.3 m/y.

Erikbreen has probably been retreating since the beginning of the century (Lefauconnier & Hagen 1990), and parts of its front now terminate in a lake dammed by large ice-cored moraines (Fig. 3). The glacier had a much larger extent in 1910 when the front was at the upglacier side of the ice-cored moraines (photograph, Norsk Polarinstittutt archives).

Maximum ice surface velocities measured in 1990 and 1991 were 0.16 m/day. This is about one order of magnitude higher than on other Svalbard glaciers of the same size, such as nearby Hannabreen as well as Austre Brøggerbreen and Midre Lovénbreen near Ny-Ålesund (Liestøl 1988; Etzelmüller et al. in press).

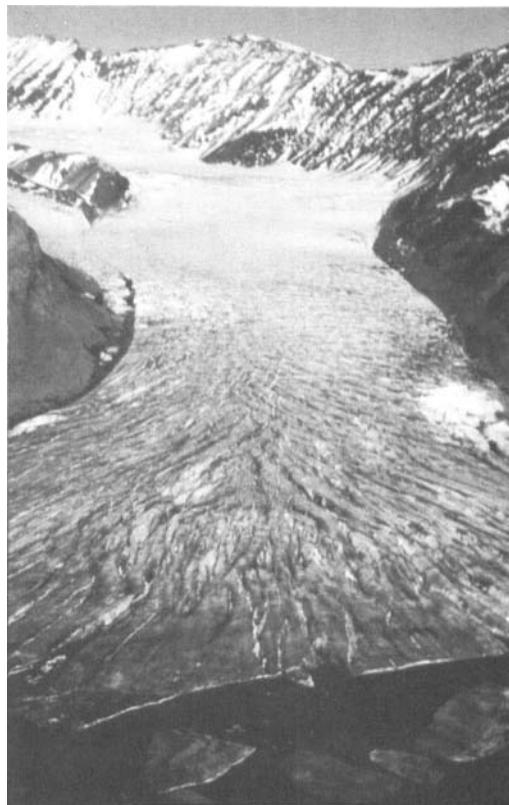


Fig. 3. Oblique air photo from the crevassed frontal part of Erikbreen (17 August 1990). Snow on the glacier is from a snow fall in early August (Photo: Norsk Polarinstittutt).

## Methods of mapping the thermal regime

The thermal regime was mapped with the use of radio-echo sounding and direct measurements with thermistors in eight boreholes 13.5 to 24 m in depth. In addition, temperatures at the interface between winter accumulation and ice or firn were measured at 21 sites along a profile close to the centre flow line before the start of the surface melt in the spring of 1991.

The radar system used is a time-gated synthetic pulse radar developed by the Royal Norwegian Council for Scientific and Industrial Research, Environmental Surveillance Technology Program (Hamran 1989). The radar and antennae were mounted on a sledge pulled by a snowmobile. The antennae were spaced approximately 2.5 m apart on a glassfibre rod attached to the sledge in a transverse position. The mapping speed was about 20 km/hour, depending on surface conditions. A portable generator was used as a power source.

The frequency bandwidth of the system is from 0.1 MHz to 3 GHz, thus covering the total frequency band needed for geophysical investigations. The antennae are the band-limiting factor for the sounding equipment. On Erikbreen, a 5–20 MHz resistively loaded dipole antenna was used to map the subglacial bedrock. Backscattering from bedrock was lost at depths below 230 m, probably due to the antenna-snowmobile-sledge configuration. This problem had never been observed before. To map water inclusions in the glacier, 320–370 MHz Yagi antennae were used. The horizontal distance between each measurement, controlled by a wheel on the sledge, was 4.2 m with the Yagi antennae and 8.4 m with the dipoles. To map the 1990/1991 snow accumulation and the upper firn layers, high resolution measurements were made with log-periodic antennae with frequencies from 1–3 GHz.

In the data reduction the dielectric constant of ice was assumed to be 3.2, giving a phase velocity of 168 m/μs. A phase velocity of 230 m/μs was used for snow and 202 m/μs for firn in the reduction of the 1–3 GHz data.

In addition to the radar survey, eight boreholes were drilled at altitudes ranging from 14 to 590 m a.s.l., and to depths of 13.5–24 m. A conventional steam drill was used with an optimal pressure of 3.5 bar. One to three individually calibrated thermistors, with an accuracy of approximately

±0.1°C, were placed in each borehole, and measured manually from 1 to 11 times between 15 July 1990 and 27 July 1992.

The basal temperature of the winter accumulation was measured at 21 sites at different altitudes in May 1991. The measurements were made with a thermistor attached to a stake, the end of which had a low thermal conductivity. The temperature under a thick snow cover is known to be stable in late winter until the start of the surface melt (Haerberli 1973; Haerberli & Patzelt 1982).

In permafrost mapping, such measurements are known as BTS-measurements (Basal Temperature of Snow). On Erikbreen the measurements were made to study the intensity of conductive cooling in late winter at the surface of the ice or firn.

## Results

Along its centre flow line 85% of the glacier has an internal reflection horizon. The horizon becomes more diffuse upstream from the equilibrium line altitude and disappears close to site 9 at a depth of 30–40 m. At higher elevations in the accumulation area an internal reflection horizon is not detected at the centre flow line, but distinct horizons are detected near the margins of the glacier along section 10 (Fig. 4) and the upper parts of section 8 (Fig. 5).

The largest depth of the internal reflection horizon along the centre flow line was 60 m, occurring at site 7 just down glacier from the equilibrium line altitude. North of the centre flow line in section 3 the maximum depth was 90 m and in cross-section 5 it was 100 m. In the ablation area below site 6 the depth of the horizon varies from 20 to 40 m. The maximum slope of the horizon in section 3 was 35°–40°.

Temperature measurements are given in Table 1 and Fig. 6. At the front of the glacier, at sites 1, 2 and 3, the measurements show a cold surface layer with a steep gradient at depth. Measurements from site 1 were not influenced by nearby crevasses. The temperature gradient between the two lowermost thermistors ranges from 0.18–0.36°C/m with an average of about 0.27°C/m. At site 3 a crevasse opened about one metre from the thermistor cable during the 1990/1991 winter. Interpretation of the site 3 data is therefore based solely on the 1990 summer temperature readings.

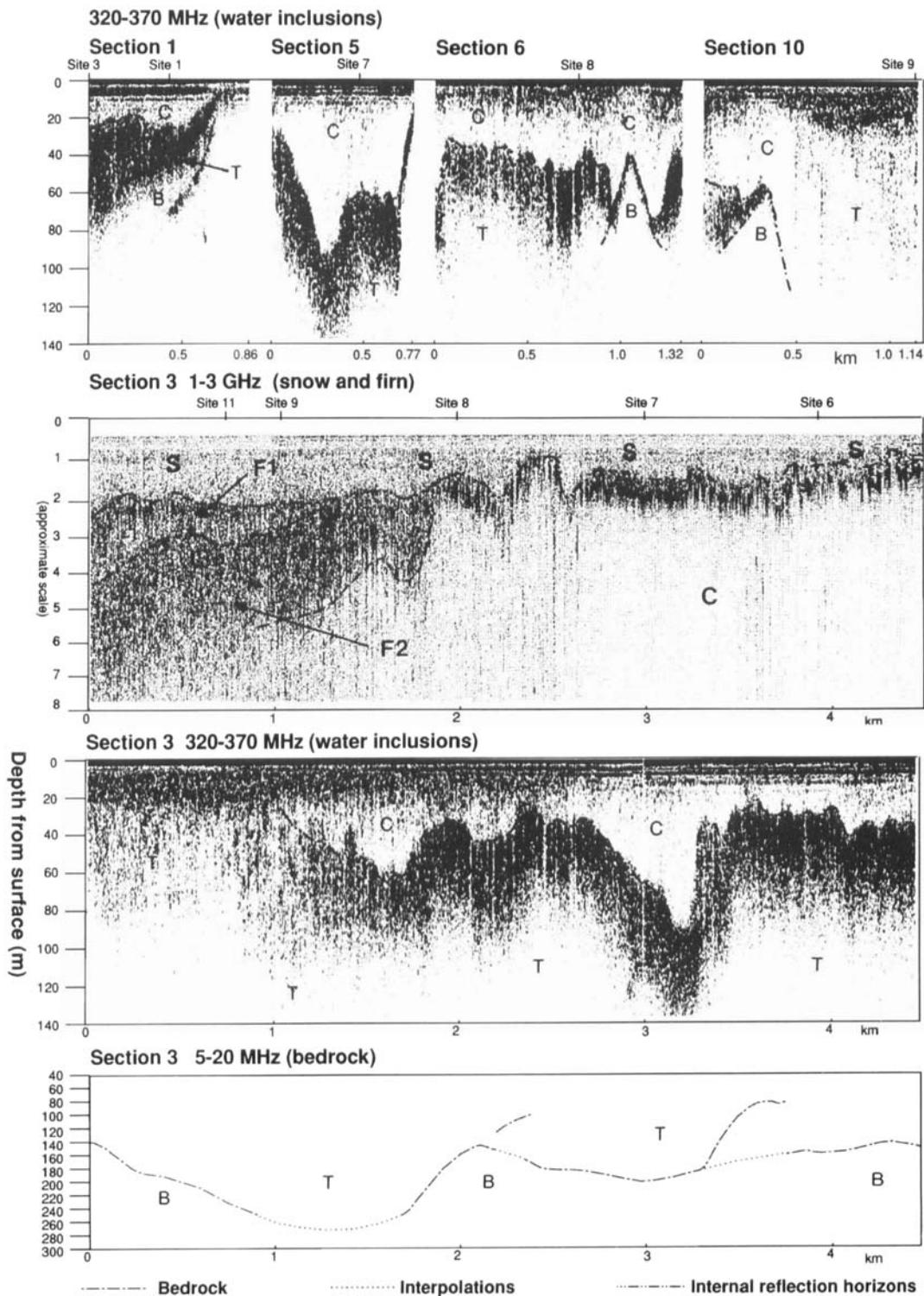


Fig. 4. Radar sections from Erikbreen with interpretations. C = cold ice. T = temperate ice. B = bedrock. S = snow. F1 = firn from summer 1990. F2 = older firn. The lower dashed line in the 1-3 GHz image between sites 8 and 9 is believed to represent the lower level of meltwater percolation. In the 5-20 MHz (bedrock) graph the shallow reflection at 2.25 km is interpreted to be reflection from the side of the valley. The shallow reflection between 3 km and 4 km is due to the radar section being measured north of the centre flow line close to the margin of the glacier (Fig. 2). Note different scales on ordinate.

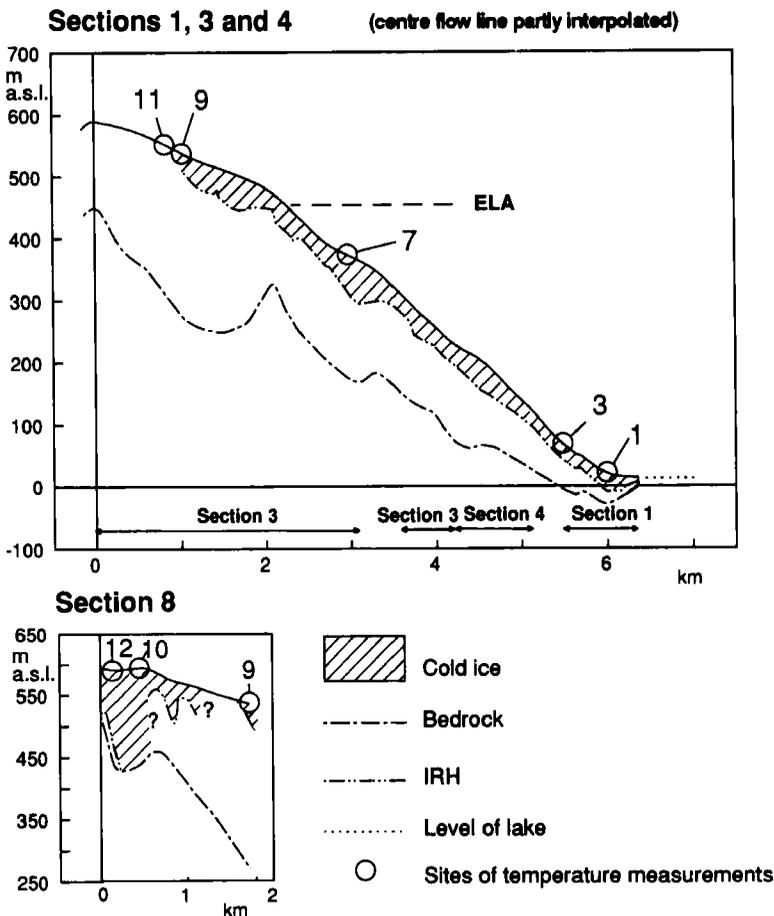


Fig. 5. Sections 1, 3 and 4 fall along the centre flow line of the glacier and show a layer of cold ice above a layer interpreted to be temperate ice. The centre flow line is partly interpolated (see lower part of figure). The interpolated parts are just downglacier site 7 and upglacier site 3. Section 8 is located across the western parts of the accumulation area and show a zone interpreted to be cold based at high altitude. ELA = equilibrium line altitude. IRH = internal reflection horizon.

At site 2 the deepest thermistor showed a rise in temperature during one year of  $0.5^{\circ}\text{C}$ . The temperature gradient between the two lowermost thermistors at 14 and 19 m depth ranged from  $0.14$  to  $0.18^{\circ}\text{C}/\text{m}$  in 1990 and from  $0.26$  to  $0.28^{\circ}\text{C}/\text{m}$  in 1991; such a rise was not detected at the other thermistor cables.

The temperature gradient at 10–15 m depth decreases significantly upglacier. At site 7, close to the 1991 equilibrium line altitude, the temperature gradient between 10 and 15 m depth varied from  $0.12$  to  $0.18^{\circ}\text{C}/\text{m}$ . Results from site 9 at 536 m a.s.l. show that the maximum temperature gradients at depths from 8.5 m to 13.5 m in the spring of 1991 and 1992 were  $0.06^{\circ}\text{C}/\text{m}$  and  $0.04^{\circ}\text{C}/\text{m}$ , respectively (Figs. 6 and 7). Late in the melt season the temperatures at these depths were about equal in all three years. At site 11, a single measurement in August 1991 showed a temperature of  $0^{\circ}\text{C}$  at 18 m depth. This is the

only measurement documenting ice at pressure melting point below 10 m depth.

A temperature of  $-5.4^{\circ}\text{C}$  was recorded at 10 m depth at site 12, which lies in a depression in the accumulation area at 590 m a.s.l. (Fig. 5). In contrast, at site 10, located 320 m from site 12 on a convex part of the glacier surface, a single measurement of  $-0.9^{\circ}\text{C}$  was found at 14 m depth.

The temperature at the base of the winter snow accumulation (BTS) in the spring of 1990 ranged from  $-5.4$  to  $-9.6^{\circ}\text{C}$  and did not vary systematically with the elevation. Snow depths were from 0.85 to 2.3 m (Table 2). Measurements were made along the centre flow line at intervals of 150 m or 300 m.

## Interpretation and discussion

The mechanisms controlling the thermal regime of a subpolar glacier are a complex interaction of

Table 1. Results from temperature measurements in boreholes.

Site number	Altitude m a.s.l.	Internal reflection horizon depth (m)	Approx. balance (m/y)	Mean annual gradient 10-15 m depth (°C/m)	Late summer gradient 10-15 m depth (°C/m)	Late winter gradient 10-15 m depth (°C/m)	Extrapolated mean surface temperature (°C)	Calculated upward heat flux 10-15 m depth (W/m <sup>2</sup> )
1	17	28	-1.7	0.27	0.18	0.36	-6.2	0.6
2	14	Not measured	-1.8	0.15	0.12	0.18	-3.8	0.3
3	68	25	-2.3	-0.32	0.24	Not measured	-6.2	-0.7
7	370	60	-0.4	0.14	0.16	0.12	-5.8	0.3
9	536	30-40	0.4	0.025	0	0.05	Not calculated	0.05
10	595	None	0.2	>0	>0	>0	Not calculated	>0
11	552	None	Not measured	0	0	0	Not calculated	0
12	590	110-120	Not measured	-0.06	0.06	Not measured	-6.0	-0.1

(1) radiation; (2) heat conduction, advection and production; and (3) non-steady state changes (Paterson 1972; Hooke et al. 1983). The release of latent heat by percolation and refreezing of meltwater is an important internal heat source, especially in the accumulation area (Sverdrup 1935; Schytt 1969).

In the interpretation of the radar images it is assumed that the internal reflection horizon represents the transition zone between cold and temperate ice (Holmlund & Eriksson 1989; Björnsson et al. unpubl.). Due to limitations of the drilling equipment, no temperature measurements were made at depths below the internal reflection horizon. Therefore, no definite conclusion can be made on this point on Erikbreen.

The results indicate that the glacier has a two-layered thermal structure. The cold surface layer starts upstream of the equilibrium line altitude and appears to extend continuously all the way to the front. It is underlain by a layer of temperate ice. The formation of cold ice by freezing of temperate ice at the base of the cold surface layer is probably the main mechanism of cold ice formation in the frontal parts of Erikbreen below site 6. In lower parts of the ablation area, at an

Table 2. Surface temperature of ice/firn 7/5-91 (Basal temperature of snow - BTS).

Approx. altitude m a.s.l.	BTS (°C)	Snow depth (m)	Interval between measurements (m)
590	-9.3	2.3	
576	-8.2	1.95	300
562	-8.3	2	300
548	-8.6	2.1	300
535	-7.7	2	150
527	-7.9	1.95	150
512	-5.4	1.85	300
497	-7.6	1.8	300
480	-7.3	1.75	300
430	-9.6	0.85	325
380	-7.3	1.5	300
369	-7.9	1.7	300
351	-7.5	1.5	300
337	-6.5	1.55	300
315	-6.6	1.45	300
270	-6.5	1.5	300
235	-8.0	1.2	300
221	-8.2	1.25	300
205	-8.1	1.15	150
163	-7.3	1.02	300
118	-7.7	0.95	300

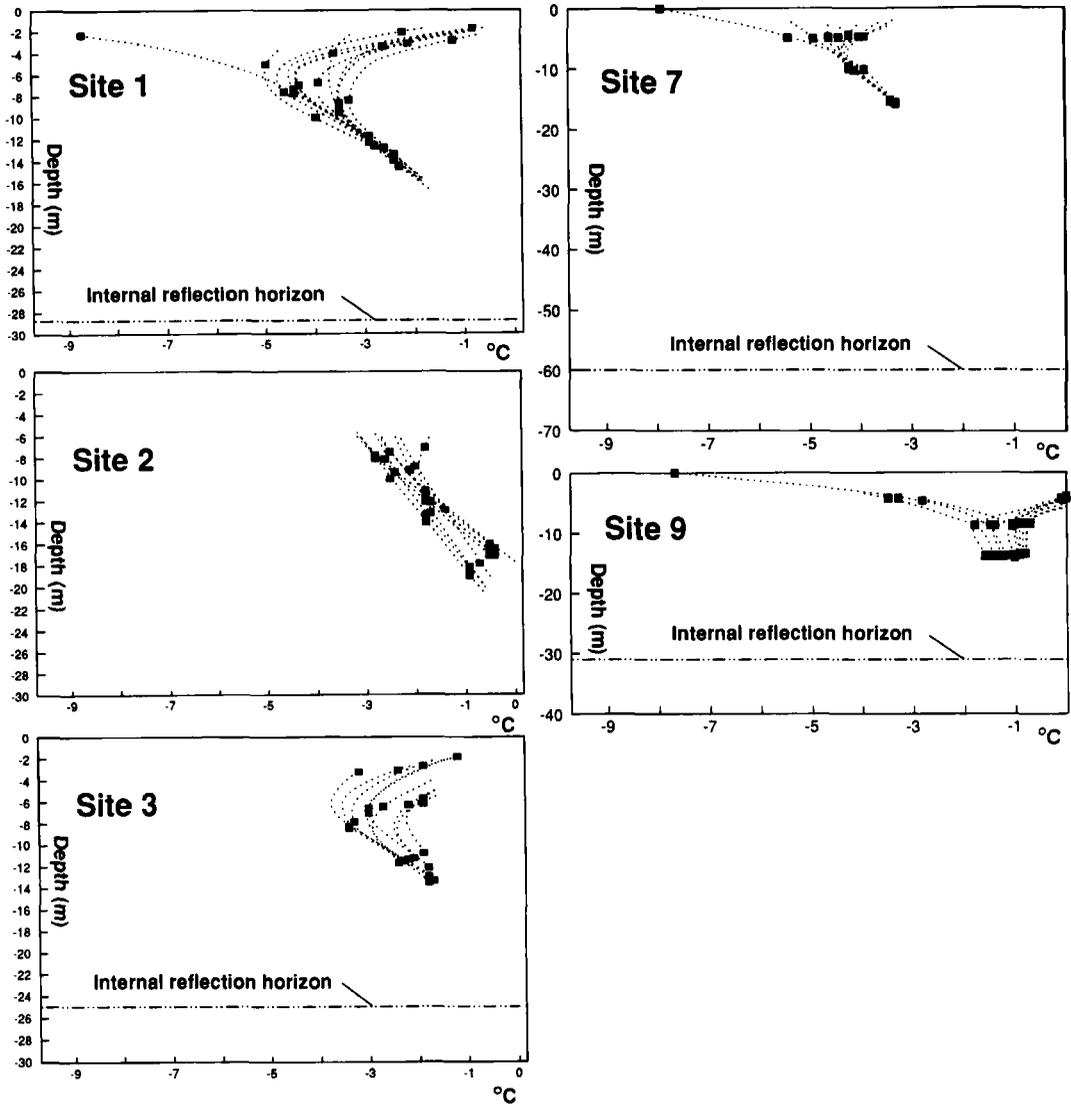


Fig. 6. Plots of temperature measurements in boreholes. Depths of measurements are adjusted for surface melt or accumulation, so they are approximate depths at times of measurement. The dotted lines are coarse interpolations and extrapolations. Dash double dot lines are internal reflection horizons believed to mark the base of the cold layer. Note different scales on ordinate. Original depth of thermistors: site 1: 4 m, 9.5 m and 14.5 m; site 2: 10 m, 14 m and 19 m; site 3: 3.8 m, 9 m and 14 m; site 7: 5 m, 10.25 m and 15.75 m; site 9: 4 m, 8.5 m and 13.5 m.

altitude below site 6 (Fig. 2), the continuous cold surface layer is interpreted to be 20 to 40 m thick. The surface melt is about 1.0 m/y at site 6, increasing to 1.7 m and 2.3 m at the terminus; surface ice velocities are about 30 to 45 m/y. The rates of surface melt, surface ice velocities and the measured ice temperatures indicate that the thermal conditions in the frontal parts are at or close to equilibrium. Assume that the thickness of the

cold layer is controlled by mean surface temperature, surface melting and water content in the transition zone between cold and temperate ice. In such a model a steady state is possible, where the conductive heat transfer in the cold surface layer is sufficient to freeze temperate ice at depth at the same rate as surface melting.

For example, at site 1 the lower thermistor shows a decrease in temperature of 0.6°C between

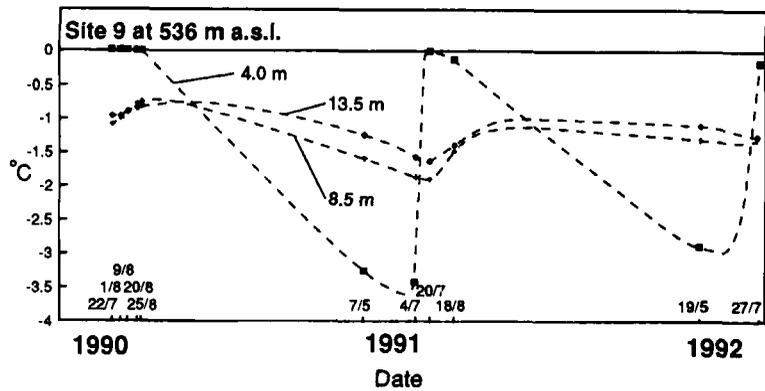


Fig. 7. Time series of the temperature measurements at site 9. Dashed lines are coarse interpolations.

1990 and 1991; the mean ablation rate is about 1.7 m/y. The yearly average gradient between the lowermost two thermistors was  $0.27^{\circ}\text{C}/\text{m}$ . The thermal conductivity of ice is  $2.1 \text{ W}/\text{mK}$  (Paterson 1981). Thus the mean annual upward heat flux is about  $0.57 \text{ W}/\text{m}^2$  at 10 to 15 m depth. For the base of the cold layer to move downward at a rate of 1.7 m/y, equal to the ablation rate, the water content of the ice would have to be 3.1%.

Bamber (1987a) states that a thermally regulated boundary would be expected to be smooth on a 10 m scale. The irregular appearance found on Erikbreen can be explained by local variations in water content, which have a strong influence on the propagation of a freezing front.

If the gradient at depth is extrapolated to the surface, the mean surface temperature can be calculated to be about  $-6.2^{\circ}\text{C}$  and the depth of the cold layer 23 m. These are probably underestimates. Because ice is constantly cooled as surface melting causes its depth in the glacier to decrease, the amount of heat moving upward through the ice increases towards the surface (Hooke 1977). Thus the temperature gradient is expected to decrease with depth. Radio-echo data indicate a cold ice thickness of 28 m, which also implies a lower gradient at depth. Since a lower gradient is expected at depth, the water content calculated above is a maximum.

Measurements of water content on Svalbard glaciers are few. In a deep borehole on Fridtjovbreen an internal reflection horizon was associated with temperate ice with a water content of 1–2% (Glazovsky & Moskalevsky 1989). Bamber (1987a) suggests a water content of  $>3\%$  based on interpretation of radio-echo data.

At site 3 the measured late summer temperatures in 1990 suggest a gradient of about  $0.24^{\circ}\text{C}/\text{m}$  compared with  $0.18^{\circ}\text{C}/\text{m}$  at site 1. A

steeper gradient at site 3 is consistent with the larger surface melt at this site.

Observations at site 2 do not fit into the steady state model. A rise in temperature of  $0.5^{\circ}\text{C}$  at 17–19 m depth from 1990 to 1991 cannot be explained by heat conduction from the surface. The rise could be due to crevasses that formed in 1991, admitting meltwater, or the melting of cold ice under the floating front of the glacier. This site probably began to float during the 1990 season. Radar-echo sounding was not attempted in this area. The depth of the lake in front of site 2 was 25 m in 1990, which gives an approximate maximum thickness of the glacier. The bottom of the glacier was not reached by drilling to 19 m depth in 1990.

Just below the equilibrium line altitude there are low rates of surface melt. Here a non-steady state is expected with an increasing cold layer thickness. An increase in thickness from 30 to 60 m is found just upglacier of site 7. The temperature gradient at 10–15 m depth is about half the values at sites 1 and 3. These observations may be a result of freezing at the base of the cold layer, but the local ice flow regime is probably of importance. The latter is supported by the steep slope of the internal reflection horizon in this area.

That the cold surface layer extends upglacier from the equilibrium line altitude is supported by temperature measurements at site 9 (Fig. 7), where cold ice was detected in the melting season in 1990, 1991 and 1992 at depths of 8.5 m and 13.5 m (Fig. 7). The diffuse backscatter at 30–40 m depth could be explained by a wide transition zone from cold to temperate ice. Observations suggest a mean annual temperature gradient at 10–15 m depth of just  $0.02\text{--}0.03^{\circ}\text{C}/\text{m}$ . Thus freezing at the base of the cold surface layer is a less

important process here, and the thickness of the cold layer is probably mainly controlled by the actual accumulation of cold ice and vertical ice movements.

Cold ice cannot survive in the accumulation area if summer meltwater percolates through the level of the winter cold wave and refreezes. Based on the thermistor measurements and BTS-measurements at site 9 in the spring of 1991, the energy needed to bring the surface layer of firn to 0°C can be calculated. Beneath a 2 m snow cover the temperature was -7.7°C, at 4 m depth -3.3°C, and at 8.5 m depth -1.5°C. A rough calculation including a snow pack of 2 m with a temperature of -15°C shows that the snow pack and the upper 8.5 m of firn/ice could have frozen 0.20 m to 0.25 m of meltwater in 1991. The total ablation at this site in 1991 was 0.34 m. Thus even in a year of minimum surface melt such as this, latent heat released from refreezing of available meltwater is potentially more than that needed to bring the surface layer to 0°C. The high altitude areas of Erikbreen are probably not in the cold firn zone as are high altitude areas in the Alps (Haeberli 1976).

Low permeability at depth is probably responsible for temperatures below zero at low altitudes in the accumulation area. Low permeability can be caused by a fast transformation of snow to ice by melting and refreezing or by formation of ice lenses. The former is supported by the high resolution 1–3 GHz radar image of the surface layer (Fig. 4) that shows a change in the surface conditions along the section from site 8 to site 9. The diffuse scatter at depths from 2 to 8 m above site 8 is believed to represent firn. At site 9 the scatter from firn disappears below 5 to 6 m depth. Ice of low permeability is interpreted to exist below this depth.

The 320–370 MHz radar images indicate a complicated thermal regime at higher elevations in the accumulation area. Cold-based ice is interpreted to exist locally in sections 8 and 10 (Figs. 4 and 5). Thus within a few hundred metres of each other there are areas interpreted to be temperate throughout, cold throughout and temperate with a cold surface layer. In the upper parts of section 8 the interpretation of the radar image indicates cold-based ice with a minimum ice thickness of 150 m (Fig. 5). The ice surface gradients and the subglacial topography indicate stagnant ice or very small movements transverse to the radar section.

## Conclusions

The cold surface layer on Erikbreen extends from the terminus upglacier to the lower parts of the accumulation area. A local zone of cold surface ice is also documented at higher elevation in the accumulation area. Cold ice may be formed in the accumulation area because of the low permeability of the surface layer, which inhibits percolation of summer meltwater.

The interpretation of the internal reflection horizon as the top of the temperate ice (Holmlund & Eriksson 1989; Björnsson et al. unpubl.) is supported by the observed temperatures on Erikbreen, but no measurements were made beneath the internal reflection horizon.

Measured temperature gradients and calculations of heat flow show that large heat fluxes can exist in the cold surface layer of subpolar glaciers (Table 1). Close to the glacier front the mean annual upward heat flux at 10–15 m depth is estimated to be about 0.6 W/m<sup>2</sup>. In view of these large near-surface temperature gradients, heat loss of this order can freeze temperate ice at the base of the cold upper layer at the same rate as surface melting if the water content is maximum 3%. Close to the equilibrium line altitude and in the lower parts of the accumulation area, freezing at the base of the cold layer is probably less important.

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