

# The observed katabatic flow at the edge of the Greenland ice sheet during GIMEX-91

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## Abstract

Observations performed in the melting zone of the Greenland ice sheet and over the adjacent tundra in the summer of 1991 are described. The experimental area is the region near Søndre Strømfjord (67°N, 54°W), which is relatively dry and sunny, resulting in the highest mean temperature in Greenland in July. The katabatic wind is dominantly present over the ice; it influences the energy balance near the surface through the sensible and latent heat flux. With the aid of a tethered balloon it was observed that the thickness of the katabatic layer is typically 100 to 200 m. An interesting aspect of the katabatic wind appears to be the acceleration of the flow in the late afternoon due to the large temperature gradient at the border between tundra and ice. Further on the ice, this effect is no longer important for the dynamics of the katabatic flow. The net radiation is the main driving force there. An attempt is made to estimate the importance of these thermal wind effects compared to the buoyancy forcing. It is concluded that near the edge of the ice surface winds are driven by the horizontal pressure gradient, imposed by the thermal contrast between tundra and ice. A comparison is made between the observed katabatic wind and those in the Antarctic.

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## 1. Introduction

The Greenland Ice Margin EXperiment (GIMEX) was conducted to study the relation between ablation and meteorological conditions in the atmospheric boundary layer over a melting snow or ice surface. This article will concentrate on the observed katabatic flow that shapes the boundary layer over the ice.

Katabatic flow is a widespread phenomenon occurring over sloping terrain. It originates from the cooling of air over a sloping surface, causing the air to become negatively buoyant and consequently to flow downslope. The mechanism drives the nocturnal katabatic flow in mountainous ar-

eas, which is of interest for the transport of pollutants in the stable nocturnal boundary layer. Observational studies in this field have been performed by, among others, Manins and Sawford (1979b) and Doran and Horst (1983). They found that the thickness of the katabatic layer is typically several metres to several tens of metres, depending on the steepness and length of the slope, local topography, surface roughness and ambient winds and stability (Petkovsek and Hocevar, 1971; Aritt and Pielke, 1986). An extreme example of katabatic flow is observed over the slopes of the Antarctic continent. Due to the high reflectivity of the snow and its radiative properties the Antarctic continent acts as a strong heat

sink, which results in an almost ever present surface-based inversion. A considerable number of observational studies have been carried out in Antarctica (Parish, 1981; Schwerdtfeger, 1984; Heinemann, 1988; Pettré and André, 1991). Lately satellites have also been used to study Antarctic katabatic winds (Bromwich 1989). The flow is especially strong on the steep slopes of the coastal zone. The vertical extent of the katabatic layer at Antarctica is typically several hundreds of metres.

On mountains glaciers and in the ablation zone of ice caps melting of ice and snow keeps the surface temperature at 0°C. The sensible heat flux is therefore directed from the air, which has a temperature higher than 0°C, towards the surface. This cooling of the air also causes katabatic flow. Observations of katabatic flow over melting glaciers and snow fields have been made by, for instance, Stretten et al. (1974), Munro and Davies (1978) and Ohata (1989). The development of these so-called glacier winds depends on factors such as the size of the ice or snow mass, the ambient wind and the structure of the surrounding surface. Ohata (1989, 1991) pointed out that the glacier wind itself is of importance for the mass balance of a glacier or ice cap, because the turbulent structure of the surface layer largely depends on this wind. Energy balance models

designed to simulate mass balance (e.g. Oerlemans, 1991) should deal with this aspect.

The purpose of this article is to describe the observed katabatic wind regime over the south-western part of the Greenland ice cap during the GIMEX-91 expedition, and compare it with the Antarctic situation. The observational area was near Søndre Strømfjord, 67°N 51°W. The climatological situation in this part of Greenland differs dramatically from that in Antarctica where the inland slopes generally show homogeneous surface conditions, no ablation occurs and where there is no adjacent land mass or tundra. Katabatic flow on the Greenland ice sheet has not been studied very often or for long periods. Ice-crossing expeditions performed the first measurements and erected several temporary climatological stations, on top as well as near the edge of the ice sheet (Loewe, 1935). Putnins (1970) provides a more detailed climatological survey of Greenland, but almost no wind data are available for the areas between coast and ice edge. Hedegaard (1982) statistically analysed wind data collected over several years by the different coastal stations. He states that in many cases the drainage of cold air from the ice sheet determines the wind climate in the coastal zone, some 170 km away from the ice edge. Surface wind data from the

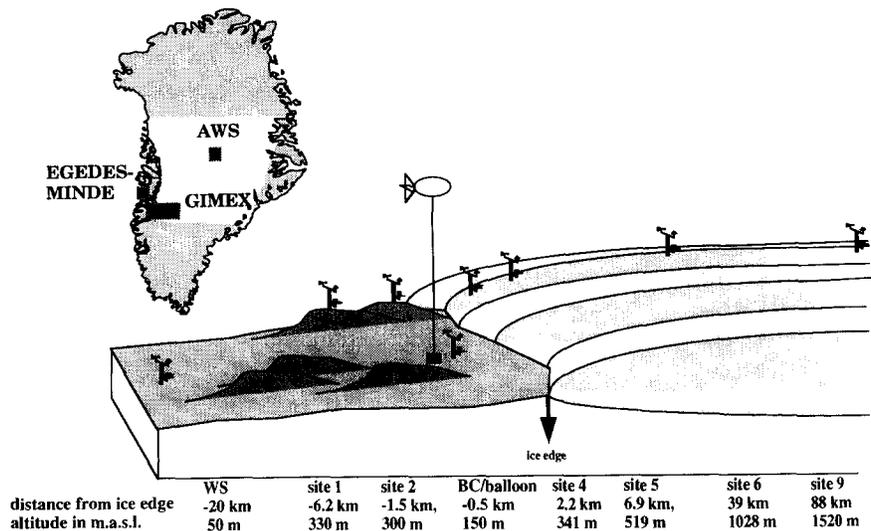


Fig. 1. Locations of the eight different sites used in the present study on a transect perpendicular to the ice edge.

only inland station in his analysis (Søndre Strømfjord, situated 150 km from the coast and 20 km from the inland ice, see Fig. 1) show a predominant northeasterly wind component. This is probably due to the channelling effect of the fjord. It should be noted, however, that in this part of Greenland the large-scale flow is from the east most of the time.

## 2. Katabatic forcing mechanisms

Mahrt (1982) analysed the momentum balance of gravity flows using the momentum Eq. (1), assuming the flow to be hydrostatic, shallow and (in our case) two-dimensional ( $\partial/\partial y = 0$ ):

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = g \frac{\theta_d}{\theta_0} \sin \alpha + \cos \alpha \frac{g}{\theta_0} \frac{\partial(\theta_d h)}{\partial x} + fv - \frac{\partial \overline{u'w'}}{\partial z} \quad (1)$$

where  $u$  is the velocity along the slope,  $w$  the velocity perpendicular to the slope,  $\alpha$  the slope angle,  $\theta_d$  the temperature difference of the layer compared to a background reference state  $\theta_0(z)$ ,  $g$  the acceleration of gravity,  $h$  the layer thickness and  $f$  the Coriolis parameter. The first term on the right-hand side, the buoyancy force, by definition drives the katabatic flow. The second term is referred to as the thermal wind term. The thermal wind term drives sea-breeze circulations if the terrain is flat. The third and fourth terms on the right-hand side represent the Coriolis-force and the vertical divergence of momentum flux, respectively. We do not consider the influence of the synoptic forcing, although it can be of great importance for the onset and final strength of the katabatic flow (Aritt and Pielke, 1986).

According to the buoyancy term in Eq. (1) the strength of the temperature inversion near the sloping surface and the angle of inclination of the surface are the parameters that locally determine the katabatic forcing over the ice. As the air flows downslope the temperature will increase due to adiabatic heating. At the same time the katabatic layer cools by turbulent heat transfer towards the colder surface. Stable ambient conditions will tend

to decrease the temperature deficit of the layer following the motion. As a result the temperature difference between the katabatic layer and the ambient air depends on the turbulent cooling rate near the surface and the lapse rate of the ambient air. Other potentially important factors are entrainment of warmer air at the top of the katabatic layer due to turbulence (Manins and Sawford, 1979a) and mass divergence or convergence in the katabatic layer (Parish and Waight, 1987), causing the katabatic layer to become thinner or thicker, respectively.

## 3. The GIMEX-91 experimental set-up

In the present study wind-, temperature- and humidity-data from eight different sites located along a transect perpendicular to the ice edge were used (Fig. 1). The most westerly station on this transect is the Søndre Strømfjord weather station (site WS, 20 km west of the ice edge, altitude 50 m.a.s.l.). The most easterly one is the boundary layer research unit of the Free University of Amsterdam (88 km from the ice edge, altitude 1520 m.a.s.l., from now on referred to as site 9). Between these two sites six meteorological masts were erected by the University of Utrecht, the sites being numbered according to Fig. 1. Besides these eight stations we used observations collected by an automatic weather station (site AWS) at the top of the ice (72°N, 40°W) set up by the University of Wisconsin.

Characteristic for the expedition area—and more generally for the west-coast of Greenland below 72°N—is the wide strip of tundra between ice cap and sea; this strip has a maximum width of 150 km. During the summer the tundra is completely free of snow, except for the highest parts of the hills. In July this area is the warmest place of Greenland with a daily mean temperature exceeding +10°C (Ohmura, 1987).

Shaped by past ice-front movements, the topography of the tundra is E–W oriented; this affects the stable air streaming off the ice. For this reason measurements at sites 1 and 2 were performed on top of hills at altitudes of 300 and 330 m.a.s.l., respectively. Data from these sites

are not representative for the lower part of the katabatic layer streaming off the ice. In order to make a direct comparison, a reference mast was erected in the valley (site BC), see Fig. 1.

Wind measurements at sites 1–6 and BC were performed at a height of 6 m, while wind data measured at site 9 were logarithmically interpolated to this level (using data from 4 and 8 m). Standard wind measurements at WS were performed at 10 m. Temperature and relative humidity were measured at 2 m. Unless stated otherwise, we use hourly mean values of these parameters, based on 2-minute sampling. Data from WS and AWS are 3-hourly synoptic observations (01, 04, 07, 10, 13, 16, 19 and 22h00 LT); these are therefore statistically less relevant than the mast-data. Nevertheless, they provide very valuable information about the conditions at places far away from the ice edge.

Vertical soundings were performed with a cable balloon at site BC up to 1000 m above ground level every 01, 07, 10, 13, 16, 19 and 22h LT, sampling every 10 seconds temperature, relative humidity, wind speed, wind direction and air pressure. Afterwards height (using the hydrostatic approximation), specific humidity and potential temperature were calculated and interpolated using a least-squares fit at intervals of 40 m. The mast at BC also served to cover the lowest 5 m of the balloon soundings.

The total measuring period for the GIMEX-91 expedition was 52 days (10/6/91–31/7/91), but 30 days of data were also available from the GIMEX-90 expedition (18/7/90–15/8/90). An analysis of the katabatic flow for a five-day period selected from the 1990 data set has been carried out by Van den Broeke (1991).

#### 4. Mean observations near the surface 5–24 July 1991

A period of 20 days from the 1991 data set was selected for a more detailed study. This period, running from 5–24 July 1991, was characterised by weak large scale flow, clear skies (mean cloud amount at Søndre Strømfjord: 18%) and small interdiurnal variations. In Fig. 2 standard synop-

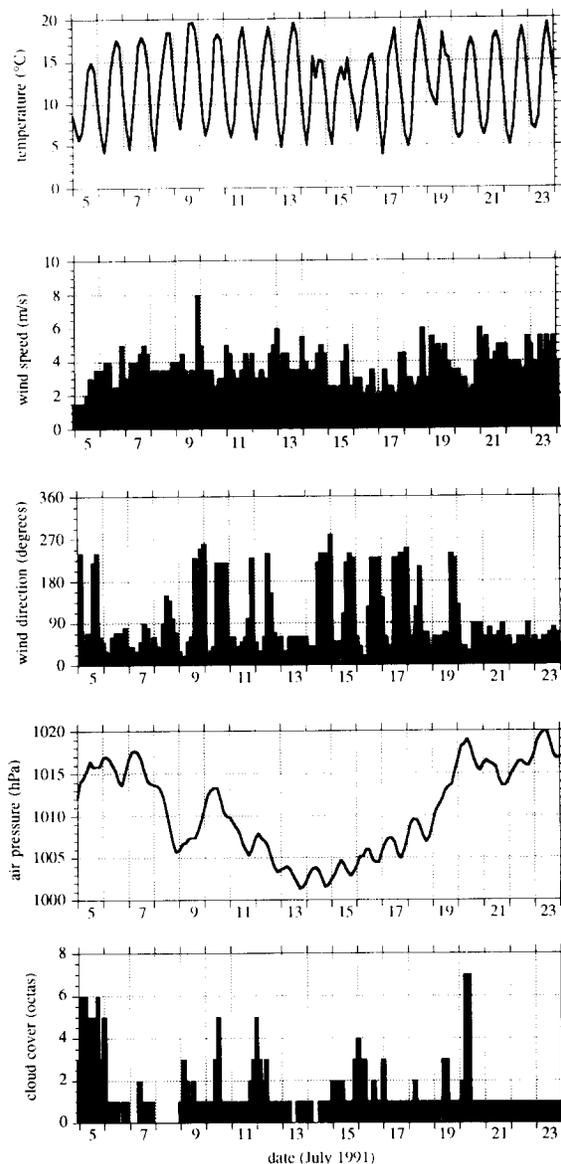


Fig. 2. Standard synoptic observations at WS during 5–24 July 1991.

tic observations at WS during this period are presented. The mean maximum temperature during the observed period was 19.2°C and the mean minimum 4.7°C. Long periods with relatively low cloud cover and high temperatures are quite common in this dry area.

As mentioned in the preceding section, the measurements of wind direction at Søndre

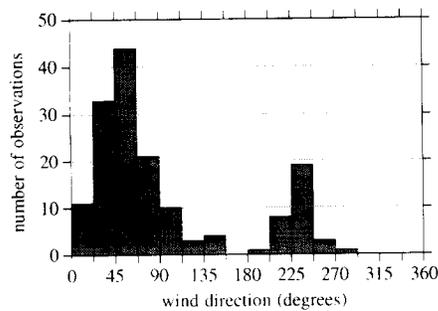


Fig. 3. Wind directions in the period 5–24 July 1991.

Strømfjord are believed to be strongly influenced by the local topography. Fig. 3 shows that the winds observed in this period are predominantly from the NE and SW, following the direction of the fjord. Statistical analysis of wind data by Hedegaard (1982) covering a period of three years (1973–1976) also reveals a remarkable maximum of winds from the NE for Søndre Strømfjord; this is obviously caused by the fjord which channels the cold air as it drains from the ice.

Table 1 shows the mean observed quantities at the different sites as observed in the period 5–24 July 1992. Balloon soundings at Egedesminde (69°N 53°W) were used to determine a mean background potential temperature profile. This profile was assumed to be constant in time. It can

be seen from this table that the air in the ablation zone 2 m above the ice has a negative temperature difference of 4°C compared to this background profile. This temperature deficit is removed when the air flows over the comparatively warm tundra. The downslope wind component is largest on the ice sheet near the edge. The directional constancy is defined as the ratio of averaged absolute windspeed and vector averaged windspeed (Schwerdtfeger, 1984). A directional constancy of 1 indicates that the wind blows from one direction all the time, whereas a value of 0 means that the wind blows from random directions. The directional constancy was calculated using hourly mean values of the period 5–24 July and is also shown in Table 1. The values, although calculated for a relatively short period, are generally high, especially in the ablation zone.

The hourly mean resultant wind vectors are plotted in Fig. 4. In Figs. 5, 6, 7 and 8 the mean daily cycles of wind speed, absolute temperature, potential temperature (corrected to BC altitude = 149 m.a.s.l.), and specific humidity over the ice (a) and the tundra (b) are presented, all for the period 5–24 July 1991.

Fig. 4 clearly shows that over the ice where the slope is significant (sites 4–9) the vector of the observed wind has an important component along

Table 1  
Mean observed quantities at the different sites in the period 5–24 July 1992

Site	WS	1	2	BC	4	5	6	9	AWS
Altitude (m.a.s.l.)	50	330	300	150	341	519	1028	1520	3105
Mean temperature at 2 m (°C)	11.7	11.2	9.4	9.2	5.0	4.0	1.7	-0.3	-11.5
Mean specific humidity at 2 m (g/kg)	5.0	-	-	4.3	4.1	-	-	3.6	1.0
Mean potential temperature at 2 m (°C)	10.7	13.0	11.2	9.2	6.9	7.6	10.3	13.1	17.7
Mean background potential temperature (°C) (at Egedesminde)	9.8	11.2	11.1	10.3	11.3	12.2	14.7	17.2	-
Mean potential temperature deficit (°C)	+0.9	+1.8	+0.1	-1.1	-4.4	-4.6	-4.4	-4.1	-
Slope angle (idealised parabolic profile) (degrees)	-	-	-	-	3.4	2.3	1.1	0.7	-
Mean wind speed at 6 m (m/s)	3.5	3.7	3.9	3.6	5.4	5.8	5.6	5.4	2.8
Directional constancy	0.51	0.77	0.86	0.88	0.97	0.98	0.98	0.94	0.61
Mean downslope component at 6 m (m/s)	-1.5	-2.8	-3.3	-2.7	-5.1	-5.1	-4.5	-2.7	-1.0

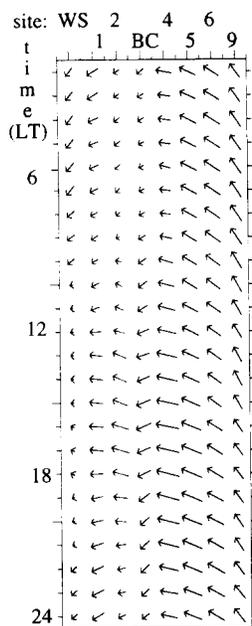


Fig. 4. Hourly mean resultant wind vectors.

the fall line of the topography, which is directed E–W. The deflection of the surface wind from the fall-line is largest at site 9 ( $55^\circ$ ) where the katabatic forcing is small (slope angle only  $0.7^\circ$ ) and is balanced by friction and the Coriolis force. The downslope transport of air is especially large near the ice margin, where the slopes are steeper (see Table 1) and the Coriolis force of less importance. Over the tundra close to the ice margin (site 1, 2 and BC) the wind field is dominated by the katabatic wind streaming off the ice. On top

of the ice at AWS (no slope) as well as at site WS no significant katabatic influences are seen. The wind direction in the valley at BC is clearly influenced by the local topography of the tundra, forcing the wind in a nearly NE direction. The increased friction over the tundra, as a result of the larger roughness length compared to the ice, plays a major role in deflecting the easterly surface winds towards more northeasterly directions at sites 1 and 2. The directional constancy decreases over the tundra and on top of the ice, where no katabatic forcing occurs.

The mean daily cycles in absolute wind speed over the ice as presented in Fig. 5a, are remarkably different from those observed over the Antarctic continent. The observed wind speeds along the GIMEX transect reflect the interaction of several different mechanisms that drive the surface winds. The implications of Table 1 and Figs. 4–8 will now be discussed and an attempt will be made to relate the observed wind-, temperature- and humidity fields at each location.

The AWS is situated in the dry snow zone, of which the height contour approximately coincides with the July  $-10^\circ\text{C}$  isotherm (Ohmura, 1987). Even in summer, the absolute air temperature at screen height remains negative (Fig. 6). The snow pack is not heated sufficiently to reach melting point, partly due to the high surface reflectivity of approximately 85%. Since snow conducts heat very poorly, conditions often become unstable over the dry snow when insolation is large; the release of sensible heat can go up to  $50\text{ W/m}^2$  (Ohmura, 1982; Wendler et al., 1988). This causes

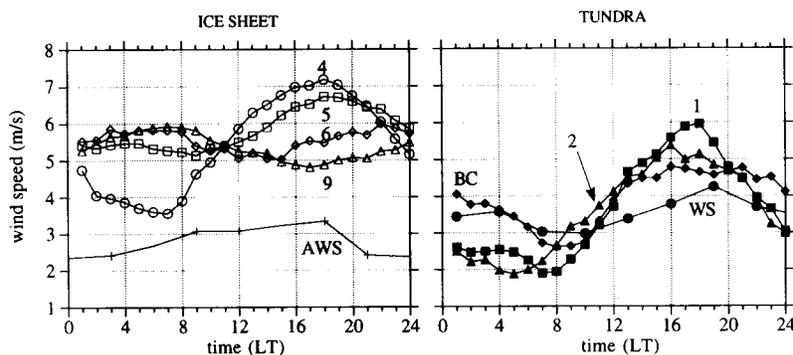


Fig. 5. Mean daily cycles of wind speed over ice sheet and tundra in the period 5–24 July 1991.

large diurnal cycles in temperature at screen height, up to 10–15°C in the interior of the Greenland ice cap (Loewe, 1935). This is con-

firmed by the measurements at AWS. The katabatic forcing is weak at this site since surface slopes are gentle, typically < 0.1°. In combination

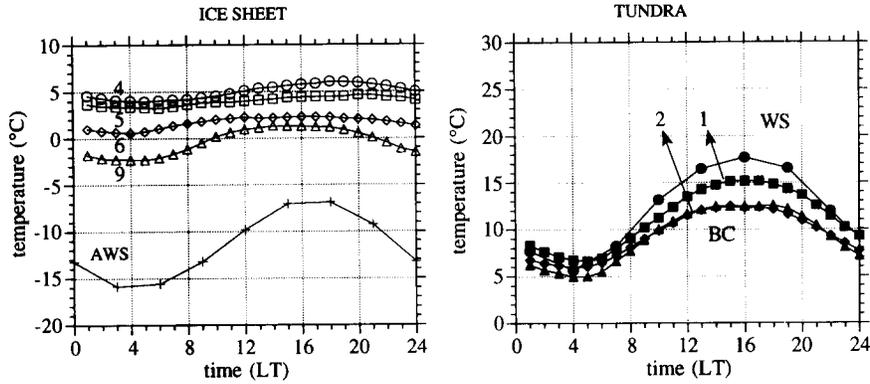


Fig. 6. Mean daily cycles of absolute temperature over ice sheet and tundra in the period 5–24 July 1991.

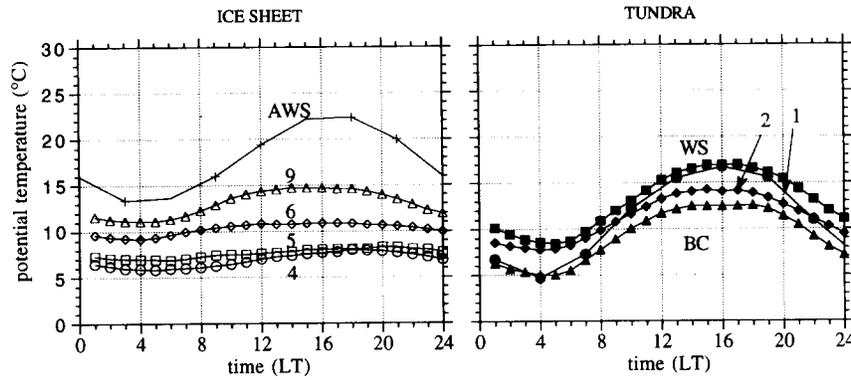


Fig. 7. Mean daily cycles of potential temperature (corrected to BC altitude = 149 m.a.s.l.) over ice sheet and tundra in the period 5–24 July 1991.

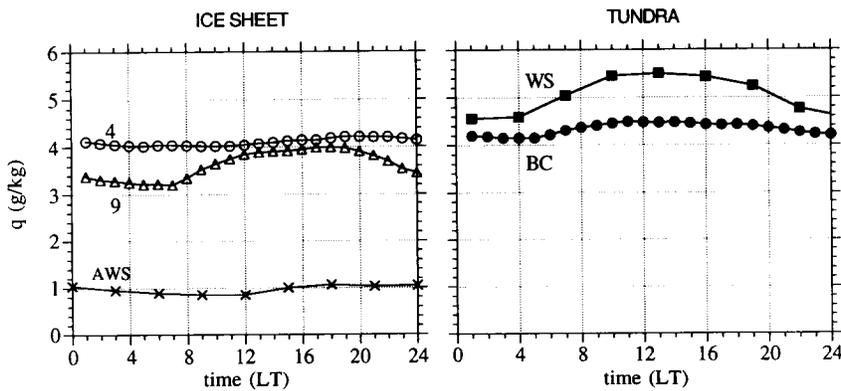


Fig. 8. Mean daily cycles of specific humidity over ice sheet and tundra in the period 5–24 July 1991.

with convective conditions, exchange of momentum with the free atmosphere will be enhanced, resulting in maximum wind speeds in the afternoon and low directional constancy (Fig. 5, Table 1). The sparse inland observations (Putnins, 1970) confirm this diurnal variation. The humidity content is very low (Fig. 8).

The absolute temperature at site 9 is negative at night and becomes positive at 10h00 LT (Fig. 6). This site is situated in the wet-snow zone near the equilibrium line, where on a yearly basis mass is neither gained nor lost. A large part of the available energy is used for heating and melting the snow pack, which refreezes at night, resulting in a strong damping of the daily temperature cycle at screen height compared to the dry snow zone (Greuell, 1992). However, the presence of a small slope is enough to generate a katabatic wind at this site with a maximum of 5 m/s at 07h00 LT, which lags the minimum temperature by 3 hours. This reflects the inertial response of the wind field to the thermal forcing. A similar observation was made by a Swiss team, working near Jacobshavn at the location 69°N 49°W at 1175 m.a.s.l. (Ohmura et al., 1991), and by Wandler et al. (1988) on an Antarctic slope. For this reason we conclude that the buoyancy forcing at site 9 is primarily responsible for the observed daily cycle in wind speed.

In the Søndre Strømfjord area the bare ice zone can be as wide as 100 km at the end of the melt season. Sites 6, 5 and 4 are situated in the ablation zone where all the snow has melted and the surface consists of ice. The albedo of an ice surface is substantially lower than that of dry snow, approximately 50% (Oerlemans and Vugts, 1992; Van de Wal and Russell, 1994). Absolute air temperatures at screen height are positive all day at these sites, whereas the surface temperature of the melting ice by definition is 0°C. The result is a sensible heat flux that is continually directed towards the surface, which indicates that the air is cooled when advected downslope (Van den Broeke et al., 1992; Duynkerke and Van den Broeke, 1994). This cooling is confirmed by the decreasing potential temperature when the air is followed down the slope (Fig. 7). The air maintains an almost constant temperature deficit over

the melting ice (Table 1). This deficit, according to Eq. (1) drives the katabatic flow.

In contrast to site 9 the wind field at sites 6, 5 and 4 shows a maximum in wind speed during the daytime. This acceleration of the air is strongest near the edge, and can be attributed to the strong heating of the tundra. The heat balance of the adjacent tundra differs dramatically from that of the melting glacier surface. During daytime a large upward sensible heat flux of 400–500 W/m<sup>2</sup> was observed, resulting in strong heating of the boundary layer over the tundra up to 1500 m. Over the ice, the flux of sensible heat is directed downwards. This results in a potential temperature at screen height at site BC which exceeds that over the adjacent ice (site 4) by typically 7–10° C. The heat source from the tundra influences the katabatic flow in several different ways:

First, during daytime, the air column up to ±1500 m is several degrees warmer than the air over the adjacent ice, so a pressure gradient is imposed on the air over the ice (thermal wind effect, see section 2), enhancing the downslope katabatic flow. Second, at greater altitude we expect advection of warm tundra air over the ice cap as a compensating effect. The transport of the warmer air towards the ice will increase the temperature deficit of the katabatic layer over the ice compared to its surroundings, increasing the buoyancy force. Third, turbulent entrainment of this warm air at the top of the katabatic layer will tend to act as an interfacial drag as well as to decrease the temperature deficit. To complicate matters even more, the slope of the surface increases rapidly towards the ice margin. The buoyancy force, as shown in (1), increases almost linearly with slope angle. This effect will again enhance the katabatic flow locally.

In Fig. 5 a weak nocturnal maximum in wind speed can be discerned at site 5, but at site 4 a pronounced minimum in wind speed occurs. This is probably due to the formation of a stable boundary layer over the tundra at night. The potential temperature over the tundra at location BC falls below that over the ice (Fig. 7). This is analogous to a situation described by Bromwich (1989) for Antarctica, where colder maritime air over the ice shelf, which is at rest, stops the

katabatic flow at the surface, but allows it to persist at some altitude above the ice shelf, where buoyancy is neutral.

The cold katabatic flow will influence the climate over the tundra close to the ice margin. When convective conditions prevail over the tundra the katabatic air will tend to flow close to the surface; this is reflected in the potential temperature depression at BC in the valley compared to site 2, situated at the top of a hill. The potential temperature at WS shows an undisturbed daily cycle. The relatively high nocturnal temperatures

at sites 1 and 2 compared to BC are probably due to enhanced mixing as a result of the exposed character of these sites.

The water vapour content  $q$  increases in a downstream direction (Fig. 8) both over the ice and over the tundra, indicating evaporation from both these surfaces.

In conclusion, the most important forcing parameters (surface energy balance, ambient air stability and slope angle) vary substantially along a transect dry snow zone-ablation zone-tundra, and closely interact with each other. The large

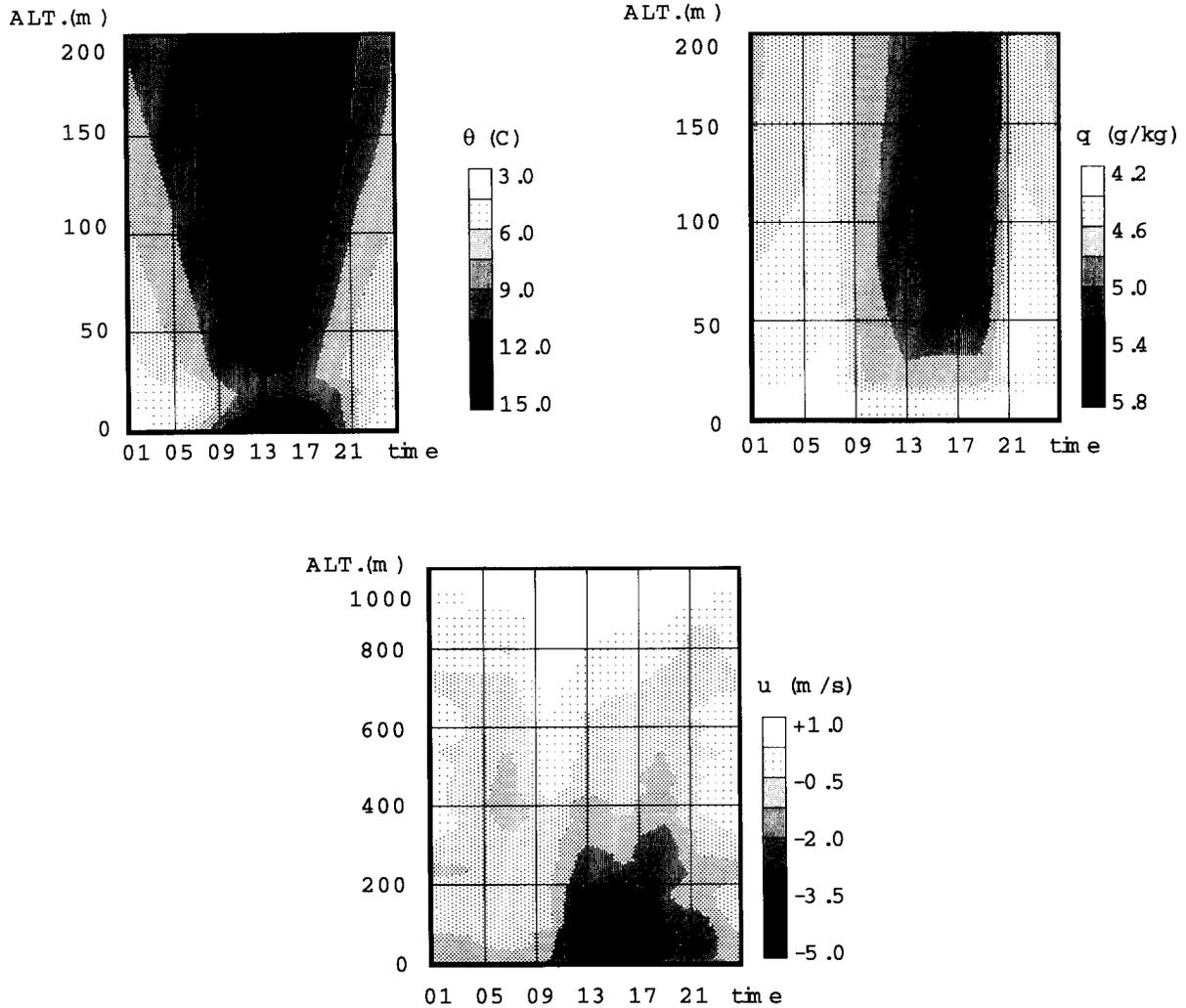


Fig. 9. Averaged daily cycles of potential temperature  $\theta$ , specific humidity  $q$  and downslope (E-W) wind component  $u$  in the period 5-24 July 1991. Note different scale in  $u$ -plot.

thermal differences at the ice-tundra border will accelerate the flow. Moreover it is very probable that warm tundra air recirculates towards the ice, modulating the katabatic flow upstream. These processes might provide a substantial feedback mechanism enhancing the ablation near the edge and influencing the mass balance of the ice sheet.

### 5. Data obtained with a tethered balloon

To gain more insight into the vertical structure of the katabatic flow, we performed tethered balloon soundings up to about 1000 m above ground level. These soundings were made at 3 hourly intervals at site BC, close to the ice edge. For the period of 5–24 July data from more than 100 ascents are available. The averaged daily cycles of potential temperature  $\theta$ , specific humidity  $q$  and downslope (E–W) wind component  $u$  for this period were calculated and are presented in Fig. 9 (note the difference in scaling of vertical axis). The cross-slope component revealed no regular diurnal variation and is not shown here.

The formation of a convective boundary layer over the tundra can be seen in Fig. 9a, b. A notable phenomenon is the ever present inversion of potential temperature and specific humidity in the lowest 100 m, reflecting the katabatic wind advecting cold and dry air from the ice. At night this inversion reaches the surface, due to radiative cooling near the ground and condensation of water vapour. Typical values of inversion strength range from 4 to 6 K for potential temperature and 0.3 to 0.5 g/kg for specific humidity. The daytime inversion starts at approximately 15 m above ground level. Strong negative gradients can be detected below this level, reflecting the usual unstable conditions over a low-reflective surface during large insolation.

A maximum in wind speed is observed at approximately 100 m above ground level. A reversal in wind direction is observed at a maximum height of 800 m (see Fig. 9c) in the late afternoon, suggesting a return flow above this altitude. In the early morning the altitude at which the downslope wind component changes sign goes down to a minimum value of 500 m.

These data show that at daytime a deep convective boundary layer forms over the tundra and at night a well developed stable nocturnal boundary layer.

### 6. Preliminary analysis of the momentum balance

From the foregoing it is concluded that the circulation near the ice margin of central West Greenland is dominated by a katabatic wind system [forced by the buoyancy term in Eq. (1)] disturbed by a local thermal circulation [forced by the thermal wind term in Eq. (1)]. For a calculation of the absolute magnitude of the buoyancy forcing an estimate is needed of the background temperature, the potential temperature deficit compared to the background near the ice surface, and the slope angle at the different sites [see Eq. (1)].

Using 12-hourly radio-soundings at Egedesminde (some 200 km further to the north, Fig. 1) over the period 5–24 July 1991, a time-averaged background potential temperature profile  $\theta_0(z)$  was obtained showing a lapse rate of 5 K/km, indicating stable background conditions. The potential temperature deficit of the katabatic layer can now be defined as the difference between the potential temperature of the well-mixed katabatic layer (Fig. 7a) and the background potential temperature  $\theta_0(z)$ , which is considered to be independent of time.

It is possible to fit a parabolic profile to the ice-topography covering the transect of the measurements (a detailed map is not available), using certain assumptions described in Paterson (1981) and Oerlemans and van der Veen (1984). The surface elevation  $h_s$  is then given by:

$$h_s = \left( \frac{2\tau_0 x}{\rho g} \right)^{1/2} \quad (2)$$

where  $\tau_0$  represents the yield stress for ice in the approximation of perfect plasticity (assumed to be 100 kPa) and  $\rho$  the ice density (900 kg/m<sup>3</sup>). The two-dimensional position of the measuring sites (see Fig. 1) can be described reasonably well by this idealised profile. From Eq. (2) the slopes

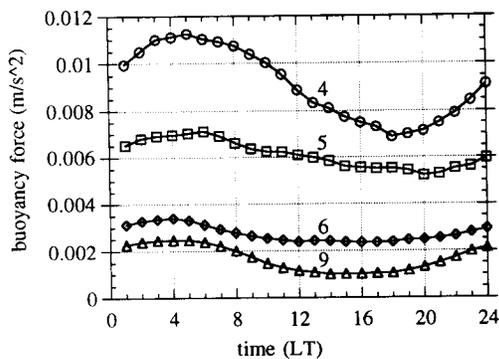


Fig. 10. Buoyancy force near the surface calculated using Eq. (1).

at the different locations of the sites are calculated, which are given in Table 1. Now the buoyancy force near the surface can be calculated with Eq. (1). The result is presented in Fig. 10.

The daily temperature cycle at the different sites and the strong slope dependency of the buoyancy force (stronger forcing for steeper slopes) are clearly reflected in this figure, as could be expected in view of all the assumed simplifications. This result is capable of explaining the observed local wind field at site 9 (maximum during the night), but not at the other sites; The shift of the maximum in wind speed (Fig. 5) towards noon going from site 6 to site 4 can thus only be explained by the increasing influence of the thermal wind term in Eq. (1). Unfortunately it is not possible to make an accurate calculation of this thermal wind term, since not enough data are available concerning the height of the katabatic layer. However, an order-of-magnitude estimate can be made from the balloon soundings near the ice margin. Assuming that at 16h00 LT the katabatic layer at site 4 is 200 m thick with an average negative temperature deficit of  $3^\circ$ , while at the same time a convective boundary layer over site BC has developed with an average positive temperature deficit (compared to the same background) of  $3^\circ$  over 1500 m. Since the distance between the two sites is only some 3 km, the resultant thermal wind force [from Eq. (1)] equals  $0.06 \text{ m/s}^2$ , an order of magnitude larger than the buoyancy forcing for mast 4 at that time (Fig. 10). Consequently, the thermal wind force very proba-

bly accounts for the strong deviation of the wind field from the “buoyancy force-balance”. Future model simulations are expected to confirm and further quantify the exact influence of this local thermal forcing.

## 7. Greenland vs Antarctic katabatic winds

It is interesting to compare the summertime katabatic winds of Greenland and Antarctica. The area of Antarctica where the snow slopes have comparable magnitude to those on Greenland is called intermediate plateau. The violent katabatic wind storms as observed at some coastal stations (Parish and Bromwich, 1989) are believed not to be forced locally.

Observations of the vertical structure of katabatic winds on the intermediate plateau are sparse and mostly performed during the Antarctic summer, when circumstances allow balloon soundings. Vertical profiles of katabatic wind in Adelie Land, East Antarctica, have been analysed by Sorbjan et al. (1985) and Kodama et al. (1989). Ohata et al. (1985) studied the structure of katabatic winds in Queen Maud Land, East Antarctica.

Beside some spots near the coast and on the peninsula, no melting occurs on Antarctica. Unstable conditions prevail over the dry snow surface at daytime (Wendler et al., 1988), which leads to the mixing of the boundary layer air with the free atmosphere. In spite of this, the directional constancy of Antarctic summertime katabatic winds is surprisingly high. The reason for this is a mesoscale/synoptic pressure gradient that is also directed downslope. Kodama et al. (1989) suggested that the warmer coastal environment is responsible for this mesoscale thermal wind. Their calculations confirmed this.

The boundary layer over the melting zone of the Greenland ice sheet will, on the contrary, always be stably stratified in summer, as long as air temperatures are positive (analogous to glacier winds). This results in a very high directional constancy of the surface wind. Superimposed on this forcing is the thermal wind circulation near the border between ice edge and tundra. So in

Greenland the tundra acts in the same way as the warm coastal environment in Antarctica, but only as a secondary forcing on a smaller spatial scale.

During the winter, when the tundra in Greenland is covered with snow and the Antarctic coastal seas are frozen, katabatic winds are forced by the negative surface long wave radiation budget, and the daily cycle of the katabatic winds will be comparable in both places. Especially above Greenland, though, the observations are too sparse to confirm this.

## 8. Concluding remarks

From the observations presented above it can be concluded that in summer the katabatic regime is dominantly present over the Greenland ice sheet. The large diversity in surface conditions during the ablation season makes the katabatic forcing complex. In the study area the adjacent tundra has a large influence on the katabatic regime, and tends to accelerate the flow near the ice margin during periods of large insolation. This thermal wind effect influences the regular katabatic regime over a substantial part of the ablation zone. At night, when the air over the tundra cools more than over the ice, it tends to oppose the katabatic wind over the ice, but over a much smaller horizontal area.

Calculations by Duynkerke and Van den Broeke (1994) have shown that the increased wind speed in daytime significantly enhances the sensible heat flux towards the ice. This means that part of the heat produced over the tundra is used indirectly to increase ablation in the melting zone near the edge. The latent heat flux will also increase and partly compensate this effect, but is probably less important. Entrainment at the top of the katabatic layer directly increases the temperature of the katabatic layer and, consequently, increases ablation since the entrained air will be effectively mixed in due to the high wind speeds.

Model simulations must be performed in order to quantify the effect of enhanced ablation in places where the tundra enhances the katabatic wind. Model studies are also necessary to study the spatial extent of the thermal wind-effect. If

this wind-effect, integrated over the whole ice sheet, does indeed significantly influence the energy balance at the ice- or snow-surface, future mass balance models dealing with the Greenland ice sheet should certainly take this effect into account.

A comparison between the summertime Greenland and Antarctic katabatic winds reveals many interesting points. While different forcing mechanisms are involved, the nature of the surface winds and the directional constancy are highly comparable.

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