Dependence of the energy balance of the Greenland ice sheet on climate change: Influence of katabatic wind and tundra

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SUMMARY

The summer-time atmospheric circulation at the margin of a large ice sheet, with an extended adjacent tundra, is investigated with a two-dimensional atmospheric model. The assumed topography corresponds to the section through west Greenland for which the Greenland Ice Margin Experiment (GIMEX) was performed. With this model, the local sensitivity of the ablation of the ice sheet to changes in the external parameters (which pertain to climate) is explored. The main focus is the sensitivity which is related to atmospheric dynamics. A comparison is made between the results with an exposed tundra and results with a snow-covered tundra. In the latter case, the katabatic wind is strongly reduced close to the ice margin, so that the downward sensible-heat flux for the ablation zone is more than halved. This shows that the influence of the exposed tundra on the energy balance of the ice sheet is important. Increasing the initial air temperatures leads to a stronger katabatic wind. As a consequence, the downward sensible-heat flux at the ablation zone is strongly enhanced. Neglecting the change of the katabatic wind would lead to a much smaller growth of this flux, which shows that the presence of an exposed tundra increases the sensitivity of the energy balance of the ice sheet to climate change.

For those runs the large-scale wind is zero, but in the presence of an inland-directed large-scale wind a strong surface wind parallel to the glacier front is found to develop above the ablation zone, by which the sensible-heat flux is increased. The advection of heated air from the tundra to the ice sheet plays a lesser role.

1. INTRODUCTION

This paper is concerned with the sensitivity of the ablation of the Greenland ice sheet to climate change. Ablation is defined here as the disappearance of glacier ice by either evaporation, or by melting followed by run-off. Its opposite is accumulation, the growth of a glacier by precipitation. The ablation zone is the area where the yearly sum of ablation exceeds the accumulation (note that if the glacier is in equilibrium, the ice underneath should flow upward with respect to the surface). In the accumulation area the converse holds. The frontier between the zones is called the equilibrium line.

The ablation rate of the ice sheet is determined mainly by the net radiation and by the turbulent fluxes of sensible and latent heat through the surface. The strong sensitivity of the ablation rate to changes in the incoming radiation (and hence climate change) is already well established. For a glacier surface that is close to the melting point, like the ablation zone of the Greenland ice sheet in summer, the net radiation is very sensitive

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to climate change, in consequence of the so-called albedo feedback (Oerlemans and Van der Veen 1984, Van de Wal and Oerlemans 1994).

This paper is devoted to the role of the turbulent fluxes. It is well known (Ambach 1963; Greuell 1992; Duykerke and Van den Broeke 1994; Henneken et al. 1994) that for the sloping ablation zone of the Greenland ice sheet, the daily-averaged flux of sensible heat attains values of several tens of watts per square metre (downward). However, especially in the neighbourhood of the equilibrium line, a considerable part of this is used for evaporation, which is a very energy-consuming way of ablation (the conversion of ice to water vapour requires 8.5 times as much energy as melting followed by run-off). Nevertheless, closer to the edge of the ice sheet, the largest part of this energy is left for ablation by melting (Duykerke and Van den Broeke 1994). It is recalled that an energy flux of $1 \text{ W m}^{-2}$ provides the energy to melt 0.26 mm water equivalent per day.

It might seem that the involved energy flux, some tens of W m$^{-2}$ as mentioned above, is of secondary importance for the energy balance and mass balance of the ice sheet. Indeed, during the ablation season, it is dominated by the net radiation. For instance, Duykerke and Van den Broeke (1994) report an observed daily average of 120 W m$^{-2}$, from 10 June to 31 July 1991, for a site where melting had created a low albedo. However, other sites in the ablation zone with higher albedo have much lower values of net radiation (Greuell 1992; E. A. C. Henneken, personal communication). Moreover, in other seasons, when the insolation is smaller or absent, the net radiation has the opposite sign. Averaged over the year, the radiation energy budget of the ice sheet is negative, whereas the contribution of the sensible-heat flux to the energy of the ice sheet is strongly positive (Greuell 1992; Van de Wal and Oerlemans 1994). Hence, the sensitivity of the turbulent fluxes to climate change is of considerable importance for determining the ablation.

Van de Wal and Oerlemans (1994) found by numerical experiments that a higher air temperature leads to a substantial increase in the ablative contribution of the turbulent fluxes. Moreover, the results appeared very sensitive to the choice of the coefficient of turbulent exchange for the surface. In their study the air temperature and the coefficient for turbulent exchange (which includes the wind speed) were prescribed externally. In reality there exists a mutual interaction between the temperature field and the wind field. Increasing the temperature difference between the cold ice surface and the warmer air will strengthen the katabatic wind, and this will in its turn considerably influence the air temperature and the energy budget of the ice sheet. This interaction is a subject of the present study.

Another feature which can make the mass budget of the ice sheet sensitive to climate is its being surrounded by large areas of exposed terrain (rocks and tundra). This was one of the reasons for setting up the Greenland Ice Margin Experiment (GIMEX), described by Oerlemans and Vughts (1993). Figure 1 shows southern Greenland, with the margins of the ice sheet and the transect where the GIMEX observations were made. At the GIMEX transect the width of the exposed area is about 150 km on average. This area can attain high temperatures in summer, a fact which possibly influences the energy balance, and hence the mass balance, of the ice sheet. Advection of air, which has been heated above the exposed surroundings, to the glacier is known to be very important for smaller glaciers (Oerlemans 1986). However, at the margin of the Greenland ice sheet, the surface wind is almost always directed from the ice sheet to the exposed terrain, so that the heat can be transported to the glacier only by upper-air currents. But there is another, less obvious, way in which the presence of exposed areas can influence the ablation of the ice sheet. The horizontal temperature gradients occurring at the ice sheet
margin for clear days in summer give rise to a horizontal pressure gradient, and in consequence of this the component of the surface wind perpendicular to the glacier front is accelerated. This thermal influence on the wind is well established by observation (Van den Broeke et al. 1994), but the consequences of this for the ablation of the ice sheet have hitherto not been investigated quantitatively.

The purpose of the present paper is to investigate by simulation the sensitivity of the turbulent fluxes to climate change, and to make clear the role played in this respect by the exposed area. This requires a numerical atmospheric model in which wind and turbulent exchange, and also long-wave radiation and its divergence, are calculated in terms of internal fields. The model that is used for this investigation is a two-dimensional mesoscale model (Meesters et al. (1994)). It has been tuned and tested by comparison with GIMEX observations on a clear summer day (12 July 1991).

2. SET-UP AND MOTIVATION

(a) Short description of the model

The model is a vertical section through the west coast of Greenland, corresponding to the observational GIMEX transect. The topography has heterogeneities on various scales that cannot be properly dealt with by the model. On the scale of the model the north–south variation of the topography of the ice sheet (Fig. 1) is important. Hence, a critical evaluation of the model results is necessary. On a smaller scale the exposed area is heterogeneous owing to the presence of hills (too small to be shown on a map with the same scale as the model). It is thought that this heterogeneity has a lesser effect on the calculations that concern the ice sheet. On the other hand the meteorological
parameters for the exposed area vary from site to site, whereas the model only yields grid-point averaged values. Nevertheless, if the data are properly interpreted, the agreement between the GIMEX observations and the model predictions appears satisfactory (Meesters et al. 1994).

The model is described in the appendix. A more thorough discussion of the assumptions is given in Meesters et al. (1994). The main features are as follows. The model topography is shown in Fig. 2; the left side corresponds to the west. The model has terrain-following coordinates. The horizontal grid distance is 20 km and the vertical grid distance is variable. The time base of the simulations is such that the highest solar elevation occurs at 1330 h (GMT minus 3 hours). Surface temperatures and surface fluxes are determined using energy-balance calculations. Exchange of energy and moisture at the surface is calculated according to the usual procedures for horizontally homogeneous turbulence. It is realized that this procedure may be inaccurate for the sloping ice surface, since in reality horizontal advection of heat is important there. The effects of this are still under study.

Figure 2. Model topography. 'abz' denotes ablation zone and 'accz' denotes accumulation zone. The tundra is exposed for runs A, C, D but for run B it is covered with snow. The dashed lines indicate the boundaries of the window for which upper-air results are shown.

(b) Standard run (A)

The point of departure is a simulation called 'run A' for which the conditions observed on 12 July 1991 are assumed. This run is almost identical to the run described in Meesters et al. (1994) with the sky clear and the large-scale wind zero. The temperature is initially prescribed according to the synoptic soundings at Egedesminde for this day. This field is assumed to be initially independent of the horizontal coordinate. This is not the case for the initial moisture field (contrary to what was assumed for simplicity in Meesters et al. (1994)), which is prescribed as follows. The initial mixing ratio, \( q \), is a linear function of the coordinates \( x \) and \( z \). At \( x = -150 \) km, \( q = 5 \) g kg\(^{-1}\) at sea level and 2 g kg\(^{-1}\) at a height of 4 km. Further, to incorporate the lower absolute humidity above the ice sheet due to the prevailing subsidence of air caused by the drainage of cold air at the surface, it is assumed that initially \( \frac{\partial q}{\partial x} = -2 \times 10^{-3} \) g kg\(^{-1}\) km\(^{-1}\) everywhere. Lastly, \( q \) is not allowed to drop below 0.5 g kg\(^{-1}\).

The ablation zone differs from the accumulation zone in albedo, density, and penetration of short-wave radiation. For the ablation zone, where the surface consists of ice, the albedo, \( \alpha \), is a function of the distance to the ice sheet: \( \alpha = 0.55 \) (10 km), 0.35
(30 km and 50 km), 0.50 (70 km) and 0.62 (90 km). For the accumulation zone, where snow covers the ice, $\alpha = 0.75$. These values are measured or interpolated results from GIMEX (Van de Wal and Oerlemans 1994). The initial snow/ice density, $\rho_s$, is 900 kg m$^{-3}$ (superimposed ice) at 10 km from the ice margin, and 500 kg m$^{-3}$ (old snow, Ohmura et al. 1991) at 90 km or more. Between 10 and 90 km linear interpolation is used for $\rho_s$. Penetration of short-wave radiation is treated in the appendix.

The presented results are for the second day of a two-day simulation, so that eventual switch-on effects are small.

Meesters et al. (1994) show that there is good agreement between the observed and simulated temperatures and wind vectors for the ablation zone for this date. However, for the present paper, it is the reliability of the calculated fluxes that is of the greatest importance. The turbulent energy fluxes are denoted by $H$ (sensible heat) and $L_e E$ (latent heat), in which $E$ denotes the vapour flux and $L_e$ the sublimation heat of ice. Throughout this paper $H$ and $L_e E$ are reckoned positive if downward. Note that $E$ is minus the evaporation rate. Unfortunately, there are few observations for comparison since direct (eddy correlation) turbulent-energy flux measurements were performed in GIMEX 1991 only for the Vrije Universiteit (VU) camp, 90 km from the ice-sheet margin.

The observed averages at this site for 12 July 1991 are $H = 11.4$ W m$^{-2}$ and $L_e E = -2.6$ W m$^{-2}$. The latter value is an estimate since the equipment to measure the moisture fluctuations was not operated during all hours of the day. The simulated values are $H = 11.4$ W m$^{-2}$ and $L_e E = -5.8$ W m$^{-2}$ respectively. The agreement between the sensible-heat fluxes is good, though probably somewhat fortuitous. Comparison of the latent-heat fluxes is difficult, since the observed value is uncertain and the simulated value depends strongly on the initial moisture content of the air in the model run. The latter has been found by pilot runs with different initial moisture contents. Nevertheless, the absolute error in the vapour flux is small.

Duynkerke and Van den Broeke (1994) tried to infer turbulent fluxes for other GIMEX stations in the ablation zone by application of standard flux-profile relations to the observed temperature and wind profiles. However, I hesitate to use these results for comparison, since their values for the VU camp are often several times higher than the values obtained there by direct measurement. The reason for this discrepancy is not yet clear, but the strong horizontal advection of heat can make the assumed flux-profile relations invalid.

(c) *Other runs (B, C, D)*

Three other runs were performed. The set-up of each of them differs from run A in only one respect. The investigations have the character of sensitivity tests, and are intended to highlight the dynamic relation between certain climatic elements and the turbulent-energy fluxes for the ablation zone of the ice sheet. It is not the purpose of the runs to mimic the effect of real climate change as such, which may vary strongly from place to place and from time to time, and which depends on many ill-known factors such as changes in albedo, humidity, clouds and precipitation. In the discussion at the end of this paper some consequences of the results for the dependence on climate of the Greenland ice sheet will be pointed out.

Run B differs from run A in that the whole tundra is covered with snow (for which the albedo is 0.75). By comparing results, the role of the heating of the tundra for the energy budget of the ice sheet is investigated. This requires a discussion of the dynamics of the glacier wind. Note that in Greenland large tundra areas are nowadays present in summer, but not in Antarctica, and that they only exist because of the relatively warm climate.
Run C differs from run A in that the initial profile of the air temperature is increased by 5 K over the whole atmosphere. This change is a slight exaggeration of the results that have been found (zonal average for 70°N, June–August) by simulations with general circulation models which were reviewed by Schlesinger and Mitchell (1987) (it concerns results of Hansen et al. (1984), Washington and Meehl (1984) and Wetherald and Manabe (1986)). The predicted change is rather independent of height for the summer months.

For simulation D an initial eastward wind with speed 2.5 m s⁻¹ is stipulated; and at a height of 5 km above the surface the wind is kept constant at this value. This set-up has been chosen to investigate the importance of the advection of heat from the tundra to the glacier.

It is realized that the two-dimensionality of the simulations has some artificial aspects, especially for simulation D. In reality the forcing of the eastward wind (simulation D) against Greenland leads to a deflection to the north and/or south (Putnins 1970). In other words the convergence of the eastward component, u, leads to divergence of the northward component, v. In the two-dimensional model the latter divergence is by definition zero. In consequence, u and also the vertical component, w, of the wind above the ice sheet will be overestimated. Though the calculated effects are expected to be exaggerated for an incoming wind speed of 2.5 m s⁻¹, they may be more realistic for a higher wind speed. Another potential problem with the two-dimensionality is that the advection at the northern and southern margin of the tundra is not described.

3. RESULTS

(a) Results for runs A and B

Figures 3(a) and 3(b) show the calculated potential temperature in a vertical section for runs A and B. The results for 1800 h are shown, which is about the time for which the differences between the two simulations are greatest. Figures 4(a) and 4(b) show the eastward component of the wind for the same time.

The differences between the results for both runs are large. For run A a convective layer has developed above the tundra. Its depth is about 1500 m, in accordance with the observational estimate by Duynkerke and Van den Broeke (1994). The temperature at reference height \( h = 10 \) m is 18 °C. As a consequence the temperature has a strong horizontal gradient at the ice margin. For run B, on the other hand, a stable boundary layer is present above the tundra, with a temperature of 1 °C at the reference height.

The wind fields above the ice sheet also differ conspicuously. For run A a relatively strong katabatic wind is present, which penetrates many tens of kilometres over the tundra. Between 1 and 2 km above the ice margin a weak return flow, compensating the stronger surface wind, is noted. Further, at the bottom left corner of Fig. 4(a), an opposing wind is seen. This is the sea breeze, which is caused by the sea–tundra temperature contrast. A fuller description of the sea breeze has been given by Meesters et al. (1994). The sea breeze has probably only little effect on the energy balance of the ice sheet.

For run B the katabatic wind is much weaker (but it is also more veered, so that Fig. 4(b), which only shows the eastward component, yields an exaggerated impression). (Values for the full wind speed will be shown in Figs. 7(a) and 7(b)). This is obviously related to the lack of a horizontal temperature gradient at the ice-sheet margin. Another difference is that the katabatic wind is not perceived over the snow-covered tundra. A closer analysis of the differences between runs A and B will now be given.

For both runs the air on the ice slope is cooled with respect to the air aloft, and flows downward. In run A the air ultimately reaches the tundra, where warmer air is
Figure 3. Potential temperature (°C) in the vertical plane at 1800 h, (a) run A and (b) run B.

present. There, a gravity current (Simpson et al. 1977) is expected to develop, i.e. the cold glacier air tends to flow underneath the heated air, and the latter is lifted. In reality the picture is more complicated than this, since in daytime the cold air itself is strongly heated by the surface (Van den Broeke et al. 1994). All this implies that air above the tundra is continuously removed from the surface by being heated and lifted, so that it makes place for fresh glacier air. This happens most rapidly in the afternoon, when the tundra temperatures are highest. Indeed, for this time, the highest wind speeds at the tundra are observed, though the wind speeds higher on the ice sheet are at their lowest then (Van den Broeke et al. 1994). In Fig. 3(a) both the convective boundary later above the tundra and the zone of rapid heating of the glacier air to the right of it are clearly visible.

For run B the situation is strongly different. The temperature of the snow-covered tundra surface cannot exceed the melting point. Hence, the surface temperatures there are comparable to the temperatures of the ablation zone of the ice sheet. Since the surface of the ice sheet is at a higher elevation, the potential temperatures are higher there (Fig. 3(b)). As the air flows down the slope its potential temperature diminishes continuously (through turbulent and radiative heat fluxes), but it does not essentially drop below the temperature of the tundra surface. Hence, contrary to run A, no gravity current can develop over the tundra: the katabatic wind stagnates near the ice margin.
This explains why moderate wind speeds are confined to the regions of the sloping ice sheet farther away from the ice margin.

It is interesting to make a comparison with the situation for Antarctica. It is well known (e.g. Tauber 1960) that the antarctic katabatic winds, when flowing from the slopes to the shelves, usually, but not always, dissipate within 10–20 km. Exceptions occur if the katabatic wind is strong (owing, for example, to orographic effects), or if the large-scale pressure fields favour higher wind speeds on the shelf (Bromwich 1989). The present run B is a case with weak flow and without large-scale pressure gradients, so that stagnation is expected also here.

Gallée and Schayes (1992) modelled the katabatic wind from the slopes of Antarctica onto an ice shelf during the polar night. The assumed slope angle for that run was 0.01 radians, which is comparable to the angles in the present run (Fig. 2). No large-scale pressure gradient was assumed. A pronounced dissipation at the ice-slope margin was found.

Gallée and Schayes (1992) analysed the stagnation (as seen for run B) by considering the dynamics in terrain-following coordinates. On the ice sheet a down-slope directed buoyancy force is present, since the boundary layer is colder than the outside air. At the slope margin the cold layer becomes thicker because of the presence of a stable-stratified boundary layer over the shelf. This causes an up-slope directed ‘thermal-wind term’,
using the terminology of Mahrt (1982). Hence, the drained cold air comes to rest there. For run A, on the contrary, the heating of the cold air implies that a ‘thermal-wind term’ develops which has the same direction as the buoyancy term.

The wind circulation, which has developed for run A, in its turn influences the temperature field. For the purpose of this study it is important to note that the temperatures in the boundary layer above the ablation zone are higher for run A than for run B. This is because of several closely related processes: the downflow of potential warm air in the katabatic flow, the subsidence caused by the horizontal divergence of the katabatic wind, and the advection of warm air from the tundra. It follows from the weakness of the eastward component of the return flow (Fig. 4(a)), and its limited extent, that advection of warm tundra air to the ice sheet is of little importance for run A. However, for a longer stimulation time this could be different.

Figure 5 shows $H$ and $L, E$ (positive if downward) for runs A and B. It is seen that $H$ dominates and that this flux is concentrated in the ablation zone. The latter feature is related to the sloping of the surface, and the consequent katabatic wind there. Higher on the ice sheet the turbulent fluxes are small, in consequence of the weak wind and of the prevailing strong stability.

![Figure 5: Downward sensible- $(H)$ and latent- $(LE)$ heat flux as a function of the place, averaged over the day, for runs A and B.](image)

First, a comparison of the calculated fluxes for run A (Fig. 5) with other results will be given. The increase in $H$ on approaching the ice margin was observed at the GIMEX stations, as far as can be gathered by calculations from profiles (Duynkerke and Van den Broeke 1994). This does not hold, however, for the stations at 8 km or less from the ice margin. Though these are beyond the resolution of the present model, the numerical results also hint at a decrease there.

The noted increase was also found numerically by Parish and Waight (1987), who simulated polar-night conditions for the margin of the antarctic. The horizontal resolution was 50 km, so that boundary effects (as seen in Fig. 5 for the similar run B) do not show up clearly. Parish and Waight also explain this strong dependence: a larger slope angle implies a larger katabatic wind, causing more turbulence and hence a larger downward sensible-heat flux at the surface, which in its turn enhances the negative buoyancy and consequently the katabatic wind strength. Hence, ‘the turbulent heat flux and katabatic wind are highly interactive’.
An interesting feature in Fig. 5 is that, higher on the ice sheet, $H$ is upwards on average. No observations of this are available, however, and the present numerical results as well as those of Van de Wal and Oerlemans (1994) indicate that the magnitude of the turbulent fluxes is small there.

In Fig. 5 $L_n E < 0$, implying that the evaporation is positive. This is in accordance with the GIMEX observations for the ablation zone (Duynkerke and Van den Broeke 1994). These observations also indicate that the evaporation within the ablation zone is larger at larger distances from the ice margin, but this is not confirmed by the present results. However, as has been indicated already in section 2(b), both the simulated evaporation and the evaporation calculated from the observed profiles can have a large relative uncertainty.

I now turn to the differences between runs A and B. In Table 1 the consequences for the energy fluxes (averaged over the 100 km adjacent to the ice edge, which area roughly corresponds to the ablation zone) are shown. Note that the net insolation is the same for all simulations, since cloudiness and albedo are the same for all runs. In reality one expects that the increase of the melting area is combined with a decrease in the albedo, but quantifying this effect is beyond the scope of this work.

<table>
<thead>
<tr>
<th>TABLE 1. SIMULATED ENERGY FLUXES, AVERAGED OVER THE DAY AND OVER THE ABLATION ZONE, AND CORRESPONDING MELTING RATE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Flux (W m$^{-2}$)</td>
</tr>
<tr>
<td>Run A</td>
</tr>
<tr>
<td>-------------------------</td>
</tr>
<tr>
<td>Net insolation</td>
</tr>
<tr>
<td>Incoming long-wave radiation</td>
</tr>
<tr>
<td>Outgoing long-wave radiation</td>
</tr>
<tr>
<td>Net radiation</td>
</tr>
<tr>
<td>Sensible heat (downward)</td>
</tr>
<tr>
<td>Latent heat (downward)</td>
</tr>
<tr>
<td>Total energy flux</td>
</tr>
<tr>
<td>Melting rate</td>
</tr>
</tbody>
</table>

The melting rates are expressed in mm of water equivalent per day.

In consequence of the lower air temperatures, the incoming (and, to a lesser extent, the outgoing) long-wave radiation for run B is diminished, so that the net radiation becomes smaller. However, Table 1 shows that the suppression is less than 2 W m$^{-2}$, averaged over the day and over the ablation zone.

For run B $H$ is conspicuously reduced (Table 1 and Fig. 5). Averaged over the ablation area, the downward sensible-heat flux changes from 19.3 W m$^{-2}$ to 9.2 W m$^{-2}$. It is also seen that the local maximum is reached at the ice margin for run A, but at some 60 km on the ice sheet for run B (Fig. 5). The explanation requires a look at the temperature and wind close to the ice surface.

$L_n E$ which has direction opposite to $H$, is also reduced, so that the change in the total turbulent-energy flux is from 14.0 W m$^{-2}$ to 6.7 W m$^{-2}$.

Figure 6 shows the temperatures for the surface ($T_s$) and for the reference height, 10 m ($T_a$), for 0000, 0600 and 1200 h. The results for 1800 h (omitted) resemble closely those for 1200 h. For the ablation zone of the ice sheet a slight suppression of the temperatures is seen for run B, as has already been discussed.
An interesting feature is that at 1200 h, $T_h < T_s$ for the higher part of the ice sheet. This implies that convection (with strong turbulent exchange in spite of low wind speeds) occurs there. The reason is that the (dry) ice sheet there has been heated rapidly in the morning, whereas the air temperature follows only slowly. As a consequence, the daily average of the sensible-heat flux is upward for that area.

The difference $\Delta T = T_h - T_s$ is similar for both runs, as can be inferred from Fig. 6. Hence, the strong reduction of $H$ in the ablation zone for run B cannot be explained by simply assuming that the flux is proportional to $\Delta T$.

$H$ depends on $\Delta T$ and on the wind speed at reference height, $U_{h}$, according to (neglecting the effect of pressure difference over the reference height)

$$H = \rho c_p C_H U_h \Delta T$$

in which $\rho$ is the air density and $c_p$ is the specific heat of air at constant pressure. $C_H$ is an exchange coefficient, for which it is assumed in the case of a stable surface layer (potential temperature $\theta_h > \theta_s$) that

$$C_H = \frac{1}{1 + 15 Ri_h (1 + 5 Ri_h)^{1/2}}$$
(Louis 1977), in which the bulk Richardson number, \( R_{ib} \), is

\[
R_{ib} = \frac{gh}{\theta} \frac{\theta_h - \theta_s}{U_h^2}
\]

with \( g \) the gravitational acceleration and \( h \) the reference height (\( h = 10 \text{ m} \)). In the case of \( \theta_h < \theta_s \), a more involved expression for \( C_H \) is assumed (Louis 1977), but this situation does not occur above the ablation area (though, as already seen, it occurs higher on the ice sheet at daytime).

Figure 7 shows the wind speed \( U_h \). For run A the maximum surface wind speed is reached close to the ice front, averaged over the day its value is \( 4.8 \text{ m s}^{-1} \). For run B the time-averaged wind is reduced to 0.7 times this value, the reduction being larger in the afternoon than in the morning, and the maximum is located at about 60 km from the glacier front.

In consequence of this, in the ablation zone the quantity \( C_H U_h \) is considerably suppressed for run B compared with run A. This appears to be the main cause of the difference in \( H \) for the ablation zone between both runs.

![Figure 7](image-url)  
Figure 7. Total wind speed (at 10 m), at various times, for (a) run A and (b) run B.
An interesting difference between Figs. 7(a) and 7(b) is the following. The highest wind speed is reached at about 1800 h for run A, but at about 0600 h for run B. The first behaviour is in accordance with the (summer-time) GIMEX observations (Van den Broeke et al. 1994), and is related to the dominating role of the thermal wind in the dynamics close to the ice margin. The second behaviour is typical for pure katabatic wind.

In summary it is seen that a large difference in $H$ occurs between runs A and B for the ablation zone. This difference is largely caused by the difference in the intensity of turbulent exchange at the surface, which is directly related to the difference in the strengths of the katabatic wind. Compared with this the reduction of $\Delta T$ is of minor direct importance.

(b) Results for run C

Figure 8 shows the potential temperature at 1800 h for run C. Of course in this figure the air temperatures are higher, both above the tundra and above the ice sheet, than for run A (Fig. 3(a)). However, the ice sheet can in general not follow the higher temperatures. Hence, both the air–ice temperature difference on the ice slope, and the horizontal temperature gradient at the ice margin, are larger for run C than for run A.

Figure 9 shows the eastward component of the wind for the same time. The results resemble those of run A (Fig. 4(a)), but the katabatic wind has become stronger. This

Figure 8. Potential temperature ($^\circ$C) in the vertical plane at 1800 h for run C.

Figure 9. Eastward component of the wind (m s$^{-1}$) in the vertical plane at 1800 h for run C.
is a consequence of both the larger air–ice temperature difference on the slope, and of
the larger horizontal temperature gradient at the ice margin.

In Fig. 10 the time-average turbulent fluxes for runs A and C are compared. Values
of surface energy fluxes averaged over the area of the ablation zone are shown in Table
1. The increase in $H$ for the ablation zone is $12.3 \text{ W m}^{-2}$ (Table 1), which is larger than
the increase in net radiation for the same zone ($10.1 \text{ W m}^{-2}$). The relative increase of $H$
is also large, 64%. $L_E$, which has the opposite sign but is less important than $H$ for the
investigated day, is also enhanced. As a net result the total turbulent-energy flux changes
from $14.0 \text{ W m}^{-2}$ to $23.6 \text{ W m}^{-2}$.

![Figure 10. Downward sensible- ($H$) and latent- ($L_E$) heat flux as a function of the place, averaged over the day, for runs A and C.](image)

Compared with simulation A, simulation C yields a higher evaporation over the
ablation zone, and a lower evaporation over much of the accumulation zone, as is shown
in Fig. 10. Over the ablation zone the turbulent exchange is stronger for run C, and this
promotes the evaporation. The converse occurs higher on the ice sheet where the wind
is very weak. There, the increased stability damps the evaporation for run C, even
though both the katabatic wind speed and the surface temperatures (and consequently
the surface vapour pressures) are higher for this simulation.

The incoming long-wave radiation is enhanced by $13.3 \text{ W m}^{-2}$ on average (Table 1).
In response, the surface temperature and the outgoing long-wave radiation increase. The
latter increase is high towards the centre of the ice sheet (not shown), but low for the
margin, since the surface temperature cannot exceed the melting temperature. Hence,
the increase in the net radiation is concentrated roughly in the ablation zone of the ice
sheet.

It is interesting to discuss the causes of the change of the turbulent fluxes. Figure 11
shows $T_s$ and $T_h$ for run C, for 0000, 0600 and 1200 h. (The result for 1800 h is not shown,
but it resembles closely the result for 1200 h.) On the whole, comparison of run C with
run A (Fig. 6(a)) shows a strong rise of both $T_s$ and $T_h$. However, the rise in $T_s$ is limited
by the reaching of the melting point, whereas $T_h$ may rise well above this.

This limit influences the difference $\Delta T = T_h - T_s$ in the ablation zone. If melting
occurs, the response of $\Delta T$ to an increase of the air temperatures aloft becomes stronger.
For most times of the day, the surface zone reaching the melting point is much broader
for run C than for run A. This stimulates the occurrence of large positive values of $\Delta T$
at daytime (1200 h in Figs. 6(a) and 11). At night the values of $\Delta T$ are more similar because the areas with melting are much smaller than at daytime.

It is clear that again, in the ablation zone, the change of $H$ is more than proportional to the change in $\Delta T$. For instance, for the grid point which is most close to the ice margin, the midday $\Delta T$ changes from 3.6 K to 4.5 K (an increase of 25%), whereas $H$ changes from 16.9 W m$^{-2}$ to 29.4 W m$^{-2}$ (an increase of 75%). The reason is that the coefficient for turbulent exchange, $C_H U_h$, strongly increases owing to the increase of the wind speed. Figure 12 shows the wind speed for run C. It is evident from a comparison with the results of run A (Fig. 7(a)) that $U_h$ is considerably enhanced. This causes an increase in $C_H U_h$.

The similarity between $\Delta T$ and the wind speed is to be expected. $\Delta T$ is roughly proportional to the difference $\Delta T^*$ in temperature between the katabatic layer and the layer aloft (this is confirmed by our model results, not shown). The latter is the primary driving force of the katabatic wind. According to analytical (Ball 1956) and numerical (Ye et al. 1990) models, the resulting equilibrium wind speed is, for a specific slope,
proportional to the square root of $\Delta T^*$ if the Coriolis force is neglected. But with the Coriolis force taken into account, as should be done for the present terrain (Meesters et al. 1994), the equilibrium wind speed becomes about proportional to $\Delta T^*$ (Ball 1956), and hence it is more sensitive to changes in $\Delta T^*$. Moreover, close to the ice margin, the wind speed is further enhanced by the larger thermal-wind term (following the terminology of Mahrt (1982)).

In summary it is seen that by increasing the air temperature (run C versus run A) the turbulent heat flow to the ice in the ablation zone is strongly enhanced, and that this is caused for a large part by the increase in the strength of the katabatic wind. The latter is a necessary consequence of the higher air temperatures above the ice and above the tundra. These cause higher (horizontal and vertical) temperature gradients, since the ice temperature cannot follow the air temperature once the melting point has been reached.

(c) Results for run D

First, the results for the wind components $u$, $v$ and $w$ for run D (for 1800 h) are presented in Figs. 13 to 15. Note that in Fig. 15 the unit cm s$^{-1}$ instead of m s$^{-1}$ is used. Comparison of Figs. 13 and 4(a) shows that the area with negative $u$ has decreased. Above the higher part of the ice sheet, where the wind was almost still for run A, an eastward large-scale wind is present for run D. Nevertheless a katabatic flow is still present up to over 100 km from the ice margin, and the westward component reaches to above 4 m s$^{-1}$. Above the tundra, the glacier wind is present up to some 40 km from the ice sheet.

Figure 14 shows that the wind has a strong northward component for run D. It exceeds 14 m s$^{-1}$ at some height above the ablation zone, which is an order of magnitude larger than the stipulated large-scale wind. For run A the $v$-component (not shown) has a maximum of 4 m s$^{-1}$, which value is reached only in a narrow zone above the ice margin. The existence of this $v$-component is obviously due to the Coriolis force. However, the present two-dimensional model may exaggerate the effect of this, as discussed in section 2(c).

The main elements of the dynamics leading to the strong wind for run D are probably as follows. By the counteraction of the warm eastward wind above the tundra and the cold glacier wind, a strong horizontal temperature gradient (and hence, pressure gradient) is created at the ice margin, which is detected up to a height of 2 km in Fig. 16. To this corresponds a strong thermal geostrophic wind difference in that zone, leading to a strong northward geostrophic wind. In the course of time the real wind will approach this geostrophic wind, at least in those regions where the effects of advection and turbulent mixing are small. The $v$-component can develop fully above the lowest part of the ice sheet, and it is advected further to the east by the eastward large-scale wind, experiencing only slow dissipation. Above the tundra, in the zone of the strongest temperature contrast, the $v$-component remains weaker. This can be explained from advective and convective effects.

The main conclusion concerning the horizontal wind is that the presence of a (warm) eastward large-scale wind, though directed opposite to the forcing of the katabatic wind, causes a strong increase in the surface wind over the ablation zone. This happens through the dynamic coupling to the northward component by the Coriolis force.

In the katabatic flow over the ice surface, the air is of course descending. Apart from this, and apart from sub-grid details, the air is rising everywhere, as is seen in Fig. 15. The rising is strongest on the left of the ice margin, where speeds up to 8 cm s$^{-1}$ are reached. This lifting is of course forced by the large-scale wind, and enhanced by the presence of the glacier flow.
Figure 13. Eastward component of the wind (m s\(^{-1}\)) in the vertical plane at 1800 h for run D.

Figure 14. As Fig. 13, but for the northward component.

Figure 15. As Fig. 13, but for the vertical component, in cm s\(^{-1}\).
The potential temperature at 1800 h is shown in Fig. 16. Comparison with the results for run A (Fig. 3(a)) shows that the fields are broadly similar. However, at the top of the picture, both the potential temperature and its vertical derivative are considerably diminished. This cooling of the middle troposphere is a consequence of the forced lifting of the air at the ice-sheet margin, by which the air is replaced by air which is potentially cooler. Close to the ice surface the westward component (continuous for run A) is diminished or reversed, so that the adiabatic warming of the surface flow is diminished or turned into cooling.

Another cause of the lower temperatures is the continuous inflow of sea-air. The temperature of the sea itself is −1.9°C. Hence, the air temperatures above the tundra are reduced. The reduction is stronger for run D than for run A, for which only a sea-breeze without large-scale wind was present.

Large amounts of sensible heat are produced at the tundra surface. Inspection of model results shows that the maximum spatial average is 250 W m⁻², which is reached at 1400 h. One would expect that with an eastward large-scale flow, heat from the tundra would be advected to the ice sheet, albeit not close to the surface. However, for the case of run D, this heating effect is apparently dominated by the (adiabatic) cooling effects mentioned above. It is unclear to what extent this result can be generalized.

Figure 17 shows the old (run A) and new (run D) daily-averaged fluxes of sensible and latent heat. In the ablation zone the downward flux of sensible heat is strongly increased owing to the much stronger wind there. Above the accumulation zone the upward flux increases by an order of magnitude. In run A the exchange was very small because of the still conditions. This shows that, for that zone, the results of runs A–C will be of limited value in practice. However, our primary interest is the energy budget of the ablation zone.

The evaporation is smaller for the new run, especially in the ablation zone. This is apparently because of the advection of moist air by the large-scale wind. The average total turbulent-energy flux for the ablation zone is 26.8 W m⁻² (14.0 for run A), as follows from Table 1.

The radiation components in the ablation zone are shown in Table 1. It appears that the net radiation diminishes by some 3.5 W m⁻², owing in part to the lower temperature aloft over the ablation zone and, in part, to the higher surface temperatures, and consequent higher radiation loss, which are caused by the increased exchange of sensible heat.
The temperatures at the reference height, and to a lesser extent the surface temperatures, are somewhat lower than for run A (not shown). This is mainly due to the reduction of the temperatures aloft, which has already been discussed.

Figure 18 shows the wind speed for run D. The wind is everywhere strongly increased, as has already been discussed. This clearly explains the larger energy exchange noted above.

As has been discussed in section 2(c), the quantitative results for this run should be interpreted with special care. It is found that incorporating an eastward large-scale wind leads to a stronger (and more northward) surface wind above the ablation zone, promoting the downward sensible-heat flux. The effect of advection of heat from the tundra by the large-scale wind is small for the present run. This is a consequence of the initial low temperature of the air as it is advected from the sea, and of the cooling by adiabatic expansion as the air ascends over the ice sheet.

Figure 17. Downward sensible- (H) and latent- (LE) heat flux as a function of place, averaged over the day, for runs A and D.

Figure 18. Total wind speed (at 10 m) for run D at various times.
Four simulations of the atmospheric circulation and the related surface fluxes have been performed for the western part of Greenland, with a broad (160 km) tundra zone separating the ice sheet from the coast. For the reference simulation (run A) the tundra is exposed and there is no large-scale wind. Insolation, initial temperature and moisture are prescribed according to observations on a clear summer day. For each of the other simulations (runs B, C, D) one feature in the set-up is changed. For run B the tundra is snow-covered; for run C initial air temperatures are increased by 5 K; and for run D an eastward geostrophic wind of 2.5 m s\(^{-1}\) is stipulated.

It has been shown (run A) that the turbulent flux of sensible heat attains appreciable values at the sloping edges of the ice sheet in summer, in accordance with measurements. These high values are reached owing to turbulence caused by a relatively strong surface wind on the ice edge, which occurs in consequence of two combined causes. The first is the presence of the ice slope, where the surface layer attains negative buoyancy and flows downward. The second is the temperature contrast between the ice sheet and the exposed tundra, causing a thermal forcing of the wind.

If instead the tundra is snow-covered (run B), the sensible-heat flux is strongly suppressed. The reason is the stagnation of the katabatic wind at the ice edge. Stagnation of katabatic winds at the edge of the Antarctica ice sheets, where no exposed areas are present, is well known from observations. It has been investigated theoretically by Gallée and Schayes (1992).

Of course if no slope was present the flux would also be suppressed, because heat loss of the boundary layer to the ice would not be compensated for by heating of the air by compression as the air flows downward (Manins 1992; Meesters et al. 1994). In that case the surface layer would become colder and more stable by loosing heat to the surface, which would inhibit the sensible-heat flux towards the ice. Such situations used to occur over flat terrain at night (Arya 1988).

Increase of the air temperatures (run C) causes a strong increase of the sensible-heat flux on the sloping edges of the ice sheet. This is because the surface temperature cannot exceed zero degrees, so that the vertical and horizontal temperature gradients grow. This stimulates both the negative buoyancy forcing and the thermal forcing of the wind at the ice margin, and hence the turbulent fluxes.

Van de Wal and Oerlemans (1994) noted the strong sensitivity of the heat flux to the choice of the turbulent-exchange coefficient. In their model this coefficient was prescribed externally. The present results show that the exchange coefficient should be treated as much as possible as an internal variable, to deal properly with the positive feedback related to the katabatic wind. Otherwise, the effect of temperature rise on the rise of the sensible-heat flux is underestimated. Moreover, the advective effect of the katabatic wind should also be treated internally.

It is found that the presence of a heated tundra is essential in bringing about the strong sensitivity of the sensible-heat fluxes to the air temperature, since without it the katabatic wind stagnates at the ice-sheet margin (run B). In this respect, the sensitivity for climate change is larger for the Greenland ice sheet than for Antarctica.

A related, though different, effect has been discussed by Oerlemans (1986). That concerned advection of air which was heated over ice-free terrain surrounding the glacier. Hence, it does not apply directly to large ice sheets, for which the surface wind is in general outward-directed. Of course advection of heat by upper-air currents is important for the climate of large ice sheets, as is shown by the strong climatic difference between Greenland (Putnins 1970) and Antarctica (Schwerdtfeger 1970). However, the advection effect is very small for simulations A to C because of the absence of a large-scale wind.
To investigate the role of advection a fourth simulation (D) was performed, for which an eastward large-scale wind was stipulated. It was found that the air temperatures above the ablation zone became lower instead of higher. This was due to the advection of cold sea-air, and to the adiabatic expansion of the air as it ascended above the ablation zone. It is difficult to draw a climatological conclusion from this. In the first place the duration of the present simulation is short and its spatial scale is small, and in the second place the two-dimensionality of the model can induce some spurious effects if a large-scale wind is present.

On the other hand, for simulation D the wind speed, and hence the turbulent exchange of energy for the ablation zone, were strongly enhanced. This occurred even though the large-scale wind is opposite in direction to the forcing of the katabatic wind. This can be understood from the formation of a front between the warm tundra air and the cold glacier air, and the deflection of the wind by the Coriolis force. The calculated wind speed may be exaggerated, however, since the divergence of the wind field in the y-direction was not incorporated in the model, and since one expects the length of the noted front to be as limited as the length of the tundra area, whereas in the two-dimensional model the front is effectively treated as infinitely long.

It is clear from the foregoing that the strong changes in the turbulent-energy fluxes cannot well be inferred directly from changes in the air temperatures at reference height above the ice sheet (which tend to adapt to the surface conditions), but are closely connected to changes in the surface-wind field. Understanding of the wind field for the ablation zone requires knowledge of the tundra–ice temperature contrast and its interaction with the wind field, in which the Coriolis force plays a crucial role. Incorporation of these elements will improve prognostic energy-budget models of the Greenland ice sheet.

Another point worth mentioning is the following. Glaciological models (e.g. Greuell 1992) commonly employ a simple parametrization of the incoming long-wave radiation, in which this quantity is assumed proportional to the fourth power of $T_h$ (in Kelvin). But in our simulations it is seen that the downward long-wave radiation is 5% higher for run C than for run A, whereas above the ablation zone, $T_h$ changes typically by 2 K, implying a change of only 3% in the fourth power of $T_h$. This implies that parametrizing the downward long-wave radiation for clear skies as a function of $T_h$ alone is likely to lead to errors over melting ice, namely to underestimation of the effect of the increase in the air temperature. The reason for this discrepancy is well understood: the air temperatures aloft that determine the radiation can be followed by $T_h$ only to a limited extent, since the surface remains at the melting point.

Part of the energy that the ice sheet gains from the sensible-heat flux is used for evaporation of the ice, and this energy should be subtracted from the energy available for run-off. Of course, evaporation has a direct ablatice effect, but this effect is small since conversion of ice to vapour requires 8.5 times as much energy as melting. For the present simulations the energy used for evaporation is rather small, especially for run D with an inland large-scale wind. However, higher evaporation values are often observed (for instance at the GIMEX VU camp). In that case the effect of climate change on the total turbulent-energy flux is diminished.

The influence of clouds has not been modelled. Their direct influence on ablation by modifying the net radiation on the ice sheet is obvious. For the subject of the present paper, the most important influence is probably the shielding of insolation on the tundra. This causes a decrease in the thermal forcing of the katabatic wind. Hence, the sensitivity of the ablation with respect to the presence of tundra, and with respect to climate change, is expected to decrease. Clouds also shield the insolation over the ice sheet, but this is
conjectured to be of lesser importance in this respect. Reasons for this are that (for the ablation zone in summer) the surface is always close to the melting point, much of the insolation is absorbed beneath the surface or reflected, and the incoming long-wave radiation increases if clouds are present.

It is too early to draw general conclusions concerning the change of the yearly energy balance and mass balance of the ice sheet in response to climate change. To do so the change of other parameters such as albedo, moisture content and precipitation should be taken into account, and the ice sheet should be considered in its entirety. Some general remarks pertaining to the ablation of the ice sheet can already be made, however. The present results indicate that adjacent exposed land areas promote ablation, and enhance the sensitivity of ablation to large-scale temperature changes. Of course a warm climate favours the presence of such areas. It is well known from other investigations (e.g. Van de Wal and Oerlemans 1994) that in such a climate, in consequence of melting and lower albedos on the ice sheet, the sensitivity of ablation on climate induced by albedo changes is also high. Hence the sensitivity mechanism discussed in the present paper cooperates with the major feedback mechanism that has been investigated so far. Hence, the sensitivity of ablation to climate change seems to be higher than what has previously been thought.

The magnitude of the local-wind-related sensitivity increase is difficult to assess on a yearly time-scale. The net radiation flux, though strongly positive in the ablation zone in summer, is on the whole negative when averaged over the year (Greuell 1992; Van de Wal and Oerlemans 1994). On the other hand the contribution of the sensible-heat flux is positive when averaged over the year, and concentrated in the ablation zone, where it is probably the dominating term in the yearly energy budget. Hence, though the strong sensitivity found by the present experiments only applies in summer, when the ablation zone is at the melting point and the tundra exposed, it is highly possible that it significantly enhances the sensitivity of the total yearly budgets. However, the evaluation of this claim requires the study of different weather conditions and of the other seasons.

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APPENDIX—DESCRIPTION OF THE MODEL

Details that depend on the conditions of particular simulations are treated in section 2. A description with more emphasis on the choice of empirical parameters and the involved dilemmas are given by Meesters et al. (1994).

(a) Coordinates and relief

The model uses a staggered grid (see Table A.1). Terrain-following coordinates are employed. The vertical coordinate, $\sigma$, is chosen in the simplest way that is possible:

$$\sigma = z - h(x) \quad (A.1)$$

in which $z$ is the altitude above sea-level and $h$ is the altitude of the surface above sea-level. The relief in the model is depicted in Fig. 2. Instead of the vertical wind speed $w$, the model uses $w_\sigma$ which is defined as

$$w_\sigma = \frac{\sigma}{dt} = w - \frac{dh}{dx} u. \quad (A.2)$$

The horizontal grid distance is $\Delta x = 20$ km. The highest model level is at $\sigma = 5000$ m.

TABLE A.1. ATTACHMENT OF QUANTITIES TO THE STAGGERED GRID OF THE MODEL

<table>
<thead>
<tr>
<th>$i$</th>
<th>$i + 1/2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$j + 1/2$</td>
<td>$u, v$</td>
</tr>
<tr>
<td></td>
<td>$\pi, H, E, F, F \downarrow, F \uparrow$</td>
</tr>
<tr>
<td>$j$</td>
<td>$r$</td>
</tr>
<tr>
<td></td>
<td>$w_\sigma, \theta$</td>
</tr>
</tbody>
</table>

Meaning of the symbols: $i$: number of the model column; $j$: number of the model level; $u, v$: horizontal velocity components; $w_\sigma = \sigma/\sigma$, in which $\sigma$ is the height above the surface; $\theta$: potential temperature; $\pi$: Exner function; $H$: turbulent-heat flux; $E$: vapour flux; $F, F \downarrow, F \uparrow$: downward and upward flux of long-wave radiation; $r$: horizontal momentum flux (two components).

(b) Dynamics of the atmosphere

The Exner function, $\pi$, is used instead of the pressure. It is determined by the hydrostatic equation

$$\frac{\partial \pi}{\partial \sigma} = -\frac{g}{\theta} \quad (A.3)$$

in which $\theta$ is the potential temperature and $g$ is the gravity acceleration. Between $\theta$ and the absolute temperature $T$, the following conversion relation exists:

$$\frac{T}{\theta} = \frac{\pi}{c_p} \quad (A.4)$$

with $c_p$ the specific heat of air.

Primitive equations are used, with the usual approximations for gentle slopes:

$$\frac{\partial u}{\partial t} = -u \frac{\partial u}{\partial x} - w_\sigma \frac{\partial u}{\partial \sigma} + f v - \theta \frac{\partial \pi}{\partial x} - g \frac{\partial h}{\partial x} + \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial u}{\partial \sigma} \right) \quad (A.5)$$

$$\frac{\partial v}{\partial t} = -u \frac{\partial v}{\partial x} - w_\sigma \frac{\partial v}{\partial \sigma} - f u - \theta \frac{\partial \pi}{\partial y} + \frac{\partial}{\partial \sigma} \left( K_m \frac{\partial v}{\partial \sigma} \right). \quad (A.6)$$
Herein, $u$ and $v$ are the horizontal components of the wind, $f$ is the Coriolis parameter, and $K_H$ is the coefficient for the vertical diffusion of momentum. In differentiating to $x$, the coordinate $\sigma$ (and not $z$) is kept constant. The term $-g d\theta/dx$ in Eq. (A.5) comes from the substitution

$$\theta \left( \frac{\partial \pi}{\partial x} \right)_z \rightarrow \theta \left( \frac{\partial \pi}{\partial x} \right)_\sigma + g \frac{d\theta}{dx} \quad (A.7)$$

which is found by application of

$$\left( \frac{\partial \pi}{\partial x} \right)_z = \left( \frac{\partial \pi}{\partial x} \right)_\sigma + \left( \frac{\partial \pi}{\partial \sigma} \right) \left( \frac{\partial \sigma}{\partial x} \right)_z \quad (A.8)$$

and Eq. (A.3). For the pressure-gradient term in (A.6), a constant value (adjusted to the geostrophic wind) is prescribed, since the model is two-dimensional. Advection (forward-upstream) and Coriolis force are calculated according to Pielke (1984).

The prognostic equation for the potential temperature, $\theta$, is

$$\frac{\partial \theta}{\partial t} = -u \frac{\partial \theta}{\partial x} - w_\sigma \frac{\partial \theta}{\partial \sigma} + \frac{\partial}{\partial \sigma} \left( K_H \frac{\partial \theta}{\partial \sigma} \right) + \frac{1}{\rho c} \frac{\partial}{\partial \sigma} (F_\downarrow - F_\uparrow) \quad (A.9)$$

in which $K_H$ is the coefficient for vertical diffusion of heat, $\rho$ is the air density, and $F_\downarrow$ and $F_\uparrow$ are the downward and upward long-wave radiation. The occurrence of $\pi$ in the denominator is a consequence of Eq. (A.4). Finally, the prognostic equation for the mixing ratio, $q$, is

$$\frac{\partial q}{\partial t} = -u \frac{\partial q}{\partial x} - w_\sigma \frac{\partial q}{\partial \sigma} + \frac{\partial}{\partial \sigma} \left( K_H \frac{\partial q}{\partial \sigma} \right). \quad (A.10)$$

Condensation and precipitation are not considered.

(c) Turbulence parametrization

The vertical turbulent exchange of momentum, heat and moisture is parametrized according to the European Centre for Medium-range Weather Forecasts model (Louis et al. 1981). For the wind speed maximum in the katabatic wind profile (where the gradient Richardson number, $Ri$, becomes infinite), zero exchange is predicted by the unmodified model. This is not realistic. Hence, the $Ri$-dependent correction factor in the turbulent-diffusion coefficient for the transition layer (Louis et al. 1981) is not allowed to drop below 0.05 in the model calculations.

(d) Radiation parametrization

The insolation, $S_\downarrow$, is prescribed as

$$S_\downarrow = S_0 \cos Z \exp(-\mu/\cos Z) \quad (A.11)$$

with $S_0 = 1380$ W m$^{-2}$, $Z =$ solar zenith angle, and

$$\mu = 0.2 - \beta h \quad (A.12)$$

where $h$ is the surface altitude and $\beta = 5 \times 10^{-5}$ m$^{-1}$. Equation (A.12) has been found by adjusting (A.11) to GIMEX results presented by Van de Wal and Oerlemans (1994).

The infrared radiative fluxes, $F_\downarrow$ and $F_\uparrow$, are calculated by using the integration parametrization as proposed by Savijärvi (1990), with effects of condensation neglected. Further details are described in Meesters et al. (1994).
ICE SHEET ENERGY BALANCE SENSITIVITY

(e) Surface parametrization

In Fig. 2 the prescribed surface types in the model are indicated. The temperature of the sea water is kept constant at its freezing point, \(-1.9 \, ^\circ\text{C}\) (Groen 1967). Its roughness length is \(10^{-2} \, \text{m}\), typical for water at moderate wind speeds (Arya 1988). The water vapour pressure at the surface is maintained at the saturation point.

For the tundra, the albedo is 0.2 (Putnins 1970), the roughness length is 0.01 m (Bowers and Bailey (1989) and Rott and Obleitner (1992) report slightly lower values), and the emissivity is 0.95. Vegetation is not explicitly incorporated. The heat diffusivity of the soil is \(2 \times 10^{-7} \, \text{m}^2\text{s}^{-1}\) (GIMEX result), and the conductivity 0.6 W (m K)\(^{-1}\). There are six model levels within the soil. The surface temperature, sensible-heat flux and latent-heat flux are calculated from the energy balance. A surface mixing ratio is not prescribed; the set of equations for exchange at the surface is closed by assuming a constant Bowen ratio, \(B\). Since Rott and Obleitner (1992) found that \(B\) is usually between 1 and 3 for a dry tundra, \(B = 2\) has been prescribed.

The albedo and density of the ice sheet are adjusted to the simulated case (see section 2). The emissivity is 0.97 (Kondo and Yamazawa 1986), and the roughness length is \(3 \times 10^{-3} \, \text{m}\) (based on measurements at the VU camp in the 1991 GIMEX campaign).

An algorithm has been set up to deal with the heat and mass transport within the upper layer of the ice. The most important details are as follows. A rigid grid is used with six levels within the ice, at depths of 2.5, 5, 9, 16, 25 and 50 cm (below the initial surface level). The evolution of temperature, density and meltwater content at each level involves radiation divergence, heat conduction and sinking of meltwater. Most of these processes are modelled in a straightforward fashion (along the lines of Greuell (1992)), but some details have to be discussed.

The extinction of short-wave radiation below the surface is prescribed, for the zone with superimposed ice, as

\[ S_{\text{net}}(z) = (1 - \alpha)S_{\downarrow 0} \left[ 0.4 \exp(-2\gamma_1 m) + 0.5 \exp(-\gamma_2 m) + 0.1 \exp(-\gamma_3 m) \right] . \]  

(A.13)

in which \((1 - \alpha)S_{\downarrow 0}\) is the net short-wave radiation, \(z\) is the depth below the surface, \(m\) is the mass per unit area between \(z\) and the surface, \(\gamma_1 = 0.0333 \, \text{m}^2\text{kg}^{-1}\) and \(\gamma_2 = 0.0015 \, \text{m}^2\text{kg}^{-1}\). Equation (A.13) has been constructed by fitting to values measured in glacier ice in the Alps (Ambach and Habicht 1961). For the snow zone the extinction is prescribed as (Greuell 1992)

\[ S_{\text{net}}(z) = (1 - \alpha)S_{\downarrow 0} \exp(-\gamma z) \]  

(A.14)

in which \(\gamma = 30 \, \text{m}^{-1}\).

The thermal conductivity is a function of the density according to (Anderson 1976; Male 1980):

\[ K(\rho) = a_0 + a_2 \rho^2 \]  

(A.15)

in which \(a_0 = 2.09 \times 10^{-2} \, \text{m}^2\text{s}^{-1}\) and \(a_2 = 2.5 \times 10^{-6} \, \text{m}^5\text{W} \, (\text{kg}^2\text{K})^{-1}\).

If the mass of meltwater at a level exceeds 0.1 times the mass of dry ice at the level, the superfluous meltwater drops to the level below. The threshold value 0.1 was reported by Ohmura et al. (1991).

(f) Initialization and time step

A time step of \(\Delta t = 60 \, \text{s}\) is used (but the diffusion calculation requires a splitting in smaller steps, and long-wave radiation is calculated only once every 12 time steps).
When initializing the model, the Exner function at the column on the left side of the model (above sea) is determined from a prescribed surface value ($\pi = \pi_s$, corresponding to a pressure of 1000 hPa), and the hydrostatic equation, (A.3). Subsequently, $\pi(x, \sigma_{top})$ is calculated along the model top by integration of Eq. (A.7), in which $(\partial \pi / \partial \sigma)_z$ is found from the equation for geostrophic equilibrium:

$$\frac{\partial \pi}{\partial \sigma} = \frac{f}{\theta} v_y$$  \hspace{1cm} (A.16)

in which $v_y$ is the $y$-component of the (prescribed) geostrophic velocity. Once initialized the top-values of $\pi$ are kept constant during the simulation.

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