Introduction

Antarctica has an important role in the global climate system; its ice sheet functions as one of the largest heat sinks in the world by extracting sensible heat from the atmosphere (Bromwich 1989). Yet relatively little is known about surface processes in the interior of the ice sheet due to its remoteness and inaccessibility. Experimental studies of the atmosphere over Antarctica and of the interaction between the ice sheet and the atmosphere can therefore greatly improve the understanding of the Antarctic climate and its role in the global atmospheric circulation.

The energy exchange between the Antarctic ice sheet and the atmosphere at a non-melting location can be described by the surface energy balance (SEB). For an infinitesimally thin surface layer without heat capacity, the SEB components are in balance:

\[ SR_{in} + SR_{out} + LR_{out} + H_S + H_L + G = 0, \] (1)

where \( SR_{in} \) and \( SR_{out} \) denote the incoming and outgoing components of the (solar) shortwave (SR) and (terrestrial) longwave radiation (LR). Net radiation (\( R_{net} \)) is the sum of net shortwave radiation (\( SR_{net} \)) and net longwave radiation (\( LR_{net} \)). The remaining non-radiative terms are the sensible heat flux (\( H_S \)), the latent heat flux (\( H_L \)) and the sub-surface heat flux (\( G \)). The fluxes are defined positive when directed towards the surface.

In the long Antarctic winter, when little or no shortwave radiation is present, the surface is cooled continuously through negative net longwave radiation (\( LR_{net} \)). To compensate this energy loss, the turbulent and sub-surface heat fluxes extract heat from both atmosphere and ice sheet, thereby cooling the near surface air and snow. If a surface slope is present, the cold and dense layer of air adjacent to the surface is forced down-slope by a horizontal pressure gradient (katabatic wind). This situation can be maintained throughout the winter, on occasion interrupted by the horizontal advection of warm air masses or strong large-scale winds. During the summer, the absorption of shortwave radiation introduces a diurnal cycle (except for locations near the pole). Since the temperature difference between the free atmosphere and the atmospheric surface layer (ASL) is reduced, the katabatic forcing is reduced as well and the pronounced katabatic wind maximum in the lower atmospheric boundary layer (ABL) disappears. The differences between day and night are illustrated by Fig. 1, which shows mean clear sky values of potential temperature and wind speed in the lower 300 m of the atmosphere, measured by tethered balloon at Kohnen station (75°00'S, 0°04'E, 2892 m above sea level, Fig. 2) in January and February 2002. The night-time temperature...
deficit in the lower 50 m of the atmosphere coincides with
the presence of a nocturnal jet ~ 20 m above the slightly
sloping flat snow surface. From the temperature profile it is
apparent that a katabatic forcing is acting in the lower 50 m
of the atmosphere, but whether this is the cause of the jet
remains inconclusive from the figure, as a nocturnal jet can
also be the representation of an inertial oscillation in a
stably stratified layer of air.

The Antarctic SEB can be studied in several different
ways. For example, general circulation models and
meteorological models are useful tools to study its spatial
and temporal variability (e.g. Genthon & Braun 1995, King
& Connolley 1997, Van den Broeke et al. 1997 and Van
Lipzig et al. 1999). Also, detailed meteorological
experiments can be conducted. SEB estimates from such
experiments have been presented for example by Carroll
(1982), Wendler et al. (1988), King & Anderson (1994),
Bintanja & Van den Broeke (1995), King et al. (1996),
Bintanja (2000) and Van As et al. (in press). These detailed
experiments can then serve to validate SEB calculations
from automatic weather stations (AWSs), which provide the
opportunity to calculate the year-round SEB on more
remote locations. SEB estimates from AWS data have been
presented, amongst others, by Stearns & Weidner (1993),

This paper presents the daily cycle of the SEB and ASL
structure using data of a measurement campaign which was
part of EPICA (European Project for Ice Coring in
Antarctica). ENABLE (EPICA-Netherlands Atmospheric
Boundary Layer Experiment) was performed at Kohnen
station, on the east Antarctic plateau (Fig. 2) between
8 January and 9 February 2002. The diurnal variation of the
SEB and/or the structure of the ASL at Antarctic locations
have also been presented by e.g. Wendler et al. (1988),
Kodama et al. (1989), Parish et al. (1993), Bintanja & Van

Fig. 1. Mean clear sky vertical profiles of potential temperature
and wind speed at 3 h and 15 h GMT, measured by tethered
balloon at Kohnen station in January and February 2002.

Fig. 2. Map of Dronning Maud
Land, Antarctica. Contour
lines are drawn for every
100 m of surface elevation.
Viola et al. (1999) and Bintanja (2000), but all of these experiments were performed at less elevated sites.

The paper is arranged as follows: firstly, the methods of calculating the SEB components will be described in the next section. Then we will discuss the general meteorological conditions at Kohnen, followed by a description of the mean clear sky daily cycle of SEB and ASL during ENABLE. Finally, a summary is presented.

Surface energy balance calculations

The methods of measurement and SEB calculation are discussed in detail by Van As et al. (in press). Here a summary is given.

Shortwave radiation

Broadband shortwave and longwave radiation fluxes were measured in both upward and downward direction by a Kipp & Zonen CNR1 combined pyranometer and pyrhiometer. In general, Antarctic radiation measurements can suffer from the effects of icing (SR and LR), a tilted sensor (SR_{in}) and a poor cosine response (SR_{in}). The deposition of ice on the sensors was generally small during ENABLE; this was checked every day. Also the horizontality of the sensor was checked daily; the tilt of the sensor did not exceed 0.5°. However, due to a poor cosine response of the SR_{in} sensor at zenith angles larger than ~80°, values of albedo (\( \alpha = |SR_{out}|/SR_{in} \)) were occasionally unrealistically high, exceeding unity, causing an underestimation of net shortwave radiation by up to ~20 W m\(^{-2}\).

This was overcome by calculating the ‘accumulated albedo’ (\( \alpha_{a} = \sum_{i=1}^{n} (|SR_{out}|/SR_{in}) \)) rather than the running daily mean \( \alpha = |SR_{out}|/SR_{in} \). The results of this method are not sensitive to the poor cosine response as insolation originating from large zenith angles contributes little to the daily mean SR_{in}; during ~30% of ENABLE the solar zenith angle was larger than 80°, but this yielded only 7% of the total shortwave radiation flux.

Since this procedure removes the daily cycle in albedo, we added a theoretical daily cycle for a dry, clean, semi-infinite snow pack, dependent on surface snow grain size (chosen as 10\(^{-4}\) m) and the diffuse fraction of SR_{in} (i.e. also on solar zenith angle) (Wiscombe & Warren 1980). This theoretical daily cycle of albedo contributed a maximum of 0.03 to the running mean albedo values. After this theoretical daily cycle of albedo contributed a maximum of 0.03 to the running mean albedo values. After this correction procedure albedo values ranged from 0.83 to 0.92 (Van As et al. in press). Hereafter SR_{net} was calculated from SR_{out}:

\[
SR_{net} = SR_{out} (1-1/\alpha).
\]  

The method is described in more detail by Van den Broeke et al. (2004a).

Turbulent heat fluxes

Turbulent heat fluxes were measured directly through eddy-correlation by Campbell CSAT3 sonic anemometers and KOH-3 Lyman-a hygrometers at \( \sim 2 \) m (H_{L} & H_{S}) and 10 m height (only H_{S}). Since the sonic anemometers produced several data gaps related to power supply and low temperatures, eddy-correlation data are available for 71% of the ENABLE period. To obtain a continuous record, Van As et al. (in press) showed that calculated bulk fluxes can reproduce the direct measurements of the turbulent heat fluxes with good accuracy: the mean difference between calculated and measured H_{S} is \( \sim 0.9 \) W m\(^{-2}\), the RMSD (root mean square difference) is 3.0 W m\(^{-2}\). The mean difference for H_{L} is \( \sim 0.7 \) W m\(^{-2}\), with a RMSD of 1.6 W m\(^{-2}\).

According to Monin-Obukhov similarity theory the sensible and latent heat flux can be expressed as

\[
H_{S} = \rho c_{p} u_{*} T_{*}, \quad H_{L} = \rho L_{v} u_{*} q_{*},
\]  

in which \( \rho \) denotes air density, \( c_{p} = 1005 \) J K\(^{-1}\) kg\(^{-1}\) is the specific heat of dry air at constant pressure and \( L_{v} = 2.83 \times 10^{6} \) J kg\(^{-1}\) is the latent heat of sublimation. The turbulent scales of wind speed, temperature and humidity can be approximated by the bulk method:

\[
u_{*} = \kappa u(z) / \ln(z \sigma_{T}) \quad T_{*} = (T(z) - T(0)) / \ln(z \sigma_{T}) \quad q_{*} = (q(z) - q(0)) / \ln(z \sigma_{q}),
\]  

where \( \kappa = 0.4 \) is the Von Kármán constant, \( u \) is wind speed (measured by cup anemometers), and \( T \) and \( q \) are temperature and specific humidity, respectively, (obtained with ventilated Vaisala HMP35C probes) at height \( z \) in the surface layer. \( z_{o}, z_{T}, \) and \( z_{q} \) are the surface roughness lengths associated with \( u, T \) and \( q \). The stability correction functions for momentum (\( \omega_{m} \)), heat (\( \omega_{T} \)) and humidity (\( \omega_{q} \)) depend solely on a non-dimensional stability parameter \( \xi = z/L_{s} \). Here \( L_{s} \) is the Obukhov length scale, defined as

\[
L_{s} = \frac{u_{*}^{2}}{g \kappa} \left( \frac{\theta_{v}}{\theta_{v,*}} \right) = \frac{u_{*}^{2}}{g \kappa} \left( \frac{T}{T_{*}} \right) \left( 1 + 0.61 q_{*} / \theta_{q,*} \right) \approx \frac{u_{*}^{2}}{g \kappa} \left( \frac{T}{T_{*}} \right) \left( 1 + 0.61 \right)
\]  

where \( g \) is the gravitational acceleration and \( \theta_{v} \) and \( \theta_{v,*} \) are the virtual potential temperature and its turbulent scale. The stability correction functions determined by Holtslag & De Bruin (1988) were used for stable stratifications (as recommended by Andreas 2002), and those by Paulson (1970) and Dyer (1974) for unstable stratification.

The roughness length was calculated from eddy-correlation data and was found to have a (constant) value of \( \sim 0.02 \) mm during ENABLE (Van As et al. in press). We used the polynomials suggested by Andreas (1987) to calculate \( z_{T} \) and \( z_{q} \) from \( z_{o} \) and \( u_{*} \).

Half-hourly means of \( T \) and \( u \) at height \( z = 2 \) m serve as input for Eqs (3–5). Specific humidity is calculated from

\[
q_{*} = \frac{q(z) - q(0)}{\ln(z \sigma_{q})}
\]
relative humidity (RH) measurements using the equation for saturation vapour pressure of Curry & Webster (1999). Prior to this, the RH measurements were corrected according to the method proposed by Anderson (1994), as the humidity sensors measure RH with respect to water instead of ice. This introduces potentially very large errors at very low temperatures. However, because of the small $H_L (< 1 \text{ W m}^{-2})$ in this temperature domain, the calculated SEB is not significantly affected.

Makkonen (1996) pointed out that these types of humidity sensors fail to record supersaturation with respect to ice, since the sensors remove supersaturation by nucleating ice crystals. Anderson (1996) however noted that supersaturation occurs infrequently at Halley, near the coast, and can be ignored. The occurrence of supersaturation on the Antarctic plateau needs further study with dewpoint sensors.

Surface temperature $T(0)$ is determined from outgoing longwave radiation applying Boltzmann’s law:

$$-L_R_{\text{out}} = (1 – \varepsilon) L_R_{\text{in}} = \varepsilon \sigma T(0)^4,$$

where $\sigma = 5.67 \times 10^{-8} \text{ W K}^{-4} \text{ m}^{-2}$ is the Stefan-Boltzmann constant. $\varepsilon$ is the broadband emissivity of the snow surface in the longwave part of the spectrum with a value close to unity. Due to the uncertainty of its value, here $\varepsilon$ is set to 1. Finally, surface humidity $q(0)$ is calculated from surface temperature assuming that the air at the snow surface is saturated with respect to ice.

### Sub-surface heat flux

Neglecting shortwave radiation penetration in snow (justified for fine-grained Antarctic snow according to Brandt & Warren (1993)) and ventilation of the upper snow layer, the thermodynamic equation of heat transport in snow simplifies considerably. Under the assumption of horizontal homogeneity the sub-surface heat flux is then defined as:

$$G = \left( k_e \frac{\partial T}{\partial z} \right)_{z=0},$$

where $k_e$ is the effective conductivity parameter, which embodies conduction in the ice lattice, diffusive transport in air between snow grains and the sub-surface equivalents of the sensible and latent heat flux resulting from vertical gradients in temperature and specific humidity in the (saturated) air in snow. We use the empirical relationship for $k_e$ suggested by Östin & Andersson (1991), which assumes a dependence on snow density only:

$$k_e(z) = -0.00871 + 0.439 \times 10^{-3} \rho_s(z) + 1.05 \times 10^{-6} \rho_s(z)^2.$$

Here $\rho_s$ is snow density in $\text{ kg m}^{-3}$, which was estimated from snow pit measurements during ENABLE, with values increasing from 323 $\text{ kg m}^{-3}$ to 383 $\text{ kg m}^{-3}$ between the surface and 1 m depth.

$G$ calculations were initialised by a temperature profile $T_k(0)$ which was determined from thermistor string measurements at AWS 9, which is at 1.8 km distance. Temperature profiles in all subsequent time steps were calculated using $T = T(0)$ and $\frac{\partial T}{\partial z} = 0$ as upper and lower boundary conditions, respectively. Accuracy and sensitivity of G calculations were checked by Van As et al. (in press). Calculated snow temperatures were found to be within 1 K of the measured snow temperatures.

### Accuracy of the calculated SEB

To assess the accuracy of the measurements and SEB calculations we have compared the measured surface temperatures (derived from $L_R_{\text{out}}$) with calculated surface temperatures for days selected in this study. In the latter case we assume $\text{SEB} = 0$ and solve Eq. (1) for $T(0)$. If the two values agree, it is a testimony of accuracy for

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<tr>
<td>$T(2 \text{ m})$</td>
<td>230.0 K</td>
<td>245.7 K</td>
<td>241.1 K</td>
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<tr>
<td>$q(2 \text{ m})$</td>
<td>0.17 g kg$^{-1}$</td>
<td>0.52 g kg$^{-1}$</td>
<td>0.31 g kg$^{-1}$</td>
</tr>
<tr>
<td>$u(10 \text{ m})$</td>
<td>4.8 m s$^{-1}$</td>
<td>5.3 m s$^{-1}$</td>
<td>4.8 m s$^{-1}$</td>
</tr>
<tr>
<td>$SR_{\text{net}}$</td>
<td>23 W m$^{-2}$</td>
<td>47 W m$^{-2}$</td>
<td>52 W m$^{-2}$</td>
</tr>
<tr>
<td>$L_R_{\text{net}}$</td>
<td>-30 W m$^{-2}$</td>
<td>-48 W m$^{-2}$</td>
<td>-60 W m$^{-2}$</td>
</tr>
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Figure 3 shows that there is good agreement between the 30 minute mean calculated and measured surface temperatures. The RMSD between calculations and measurements is only 0.9 K, which is equivalent to a surface flux of approximately 2.7 W m⁻². We feel that this is a good result for a SEB study under difficult conditions, and suffices especially when the focus is not on absolute values.

Furthermore, in several figures in this paper the standard deviation is indicated. As will become clear, the typical day to day variability in most measured parameters is much more significant than their random measurement error, but does still not dominate the daily cycle.

Results

General meteorological conditions

Kohnen station is situated on the eastern Antarctic plateau, approximately 500 km from the South Atlantic coastline.
AWS 9 has been in operation at this location since the austral summer of 1997/98. Figure 4 presents two hour means of AWS data of the four year period prior to and including ENABLE and Fig. 5 shows half-hourly means of \( T, q, u \) and wind direction (wd) during ENABLE. Basic climatological means and ENABLE means are given in Table I. The climate at Kohnen is dry and cold, with weak winds compared to Antarctic coastal sites or the escarpment region. During the experiment, mean 2 m temperature, specific humidity and 10 m wind speed were \( \sim 245.7 \) K, 0.52 g kg\(^{-1}\) and 5.3 m s\(^{-1}\), respectively, which is within the recorded variability of the same period in the four previous years as determined from the AWS data. Net shortwave radiation was slightly low during ENABLE (\( \sim 47 \) W m\(^{-2}\)).
Fig. 6. Power spectra of a. net shortwave radiation, b. 2 m temperature, and c. 2 m wind speed, measured by AWS 9 over a four year period. The dashed lines are the 99% significance level.

Fig. 7. a. The ratio of daily means of SR_{in} at the surface and at the top of the atmosphere, and b. the daily maximum of LR_{in} during ENABLE. Black bars denote the clear sky days used in this study, grey bars denote the remaining days.
and net longwave radiation somewhat high (~ –48 W m⁻²), which is at least partially caused by the passage of a low pressure system at the beginning of the campaign (Fig. 5).

To illustrate on which timescales the near-surface variability at Kohnen dominates, the power spectra of 2 h mean SRₙₑₙₑ, T and u at ~ 2 m height from the four year AWS dataset are shown in Fig. 6. In spite of the short summer period the figure shows a strong daily cycle for all three parameters, well exceeding the 99% significance level which is indicated by dashed lines. The pronounced daily cycle is clearly visible in Fig. 5 as well. Figure 6 also shows a significant yearly cycle in SRₙₑₙₑ and T, but not in the wind speed data. The higher harmonics of the daily and yearly cycles are a result of the limitations of the Fast Fourier Transform as they are caused by the non-sinus shaped cycles of the daily and annual signals. A dominant variability on synoptic timescales is not clearly identifiable. The 99% significance level is slightly exceeded by spectral peaks in temperature and wind speed at an oscillation period of approximately three days.

The direction and the high directional constancy (0.88, Reijmer 2001) of the near-surface winds at Kohnen suggest that katabatic winds occur frequently here, in spite of the small surface slope of 1.3 ± 0.3 m km⁻¹ in east/north-easterly direction (~ 63º) (derived from a 1 x 1 km digital elevation model (Bamber & Muller 1998)). Figure 5d shows that near-surface wind had the tendency to blow in the approximate downslope direction during ENABLE, with a mean direction of 49º. During the strong-wind event at the beginning of ENABLE the wind turned towards a more northerly direction.

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**Fig. 8.** The mean daily cycles of a. temperature, b. specific humidity, and c. wind speed at 10, 5 and 1 m height and at the surface over eleven clear sky days during ENABLE.

**Fig. 9.** The mean clear sky daily cycle of relative humidity (with respect to ice) at 2 m height during ENABLE. The bars indicate the standard deviation.
Mean clear sky daily cycle

Data selection
From ENABLE data, we selected days with a pronounced daily cycle. Since shortwave radiation forces the daily cycle, these days are characterized by clear skies. Fig. 7a shows the ratio of daily mean incoming shortwave radiation at the surface to its value at the top of the atmosphere for all ENABLE days. Partial cloud cover or overcast conditions during daytime result in lower ratios and are excluded from this study. A second method to identify cloudiness during a day is through LR
in, as high values of LR
in signify a high liquid water content of the atmosphere. The daily maximum values of LR
in are plotted in Fig. 7b. This enables us to eliminate days with night-time cloudiness, since these are not recognized in Fig. 7a. From here on, the results presented in this paper will describe the average daily cycle of the eleven clear sky days during ENABLE, indicated by black bars in Fig. 7. Icing occurred on multiple sensors on 28 January due to the presence of a fog layer; therefore this day is excluded as well. Table I shows that the mean temperature, specific humidity and wind speed of the selected days are lower than their mean ENABLE values. Although SR
net is higher in clear sky conditions, the surface is cooled more strongly through a more negative net radiation.

Temperature, humidity, wind speed and wind direction
Figure 8 shows the mean clear sky daily cycle in temperature, specific humidity and wind speed at the approximate heights of 10, 5 and 1 m and at the surface. During the first part of the day, T and q strongly increase with height and a large wind shear is present. From ~ 2 h GMT onwards, the surface starts to warm. The air temperatures at the three presented heights start to increase ~ 0.5 to 2 h later. Specific humidity shows similar behaviour because of the strong dependence of saturation vapour pressure on temperature and the near saturation of the air. Wind shear decreases significantly during these morning hours, as stability decreases.

At approximately 9 h GMT convection starts as surface temperature exceeds the temperatures in the ASL. Mean surface specific humidity exceeds the atmospheric humidity values slightly earlier, indicating the start of sublimation at ~ 8 h. From this moment, wind speed in the ASL increases uniformly until it reaches a maximum around noon. The maximum surface temperature and specific humidity occur at ~ 15 h, again peaking ~ 1 h before the air temperatures and specific humidity values do. Striking is the large variation of surface temperatures in excess of 14 K during the daily cycle. After approximately 16 h, temperature and specific humidity decrease. During this collapse of the convective system, the wind speed drops rapidly. A clear minimum is recorded between 17 h and 18 h at all heights. In the evening, a large wind shear is again recorded.

The daily cycle of relative humidity (with respect to ice) at 2 m height is plotted in Fig. 9. Only one level is shown,
Fig. 12. The mean clear sky daily cycle of a. stability parameter $z/L_*$ and b. its effect on the magnitude of the turbulent heat fluxes (see Eq. 4) during ENABLE. The bars indicate the standard deviation.

Fig. 13. (opposite column) The mean daily cycles of the turbulent scales of a. temperature, b. specific humidity, and c. wind speed over eleven clear sky days during ENABLE. The bars indicate the standard deviation.
because there is not much variation with height. At the end of the evening and beginning of the night the air is saturated. RH decreases from \(\sim 4 \text{ h GMT} \) onwards, which is caused by rising temperatures, until a minimum value of \(\sim 89\% \) is reached at the end of the afternoon. From then on RH increases again. The figure does not show a large daily cycle; the air of the ASL is near saturation the entire day. We note that values indicating supersaturation in Fig. 9 are spurious readings as the instruments are not capable of measuring RH > 100\%. Supersaturation is likely to occur and needs attention in future experiments.

Wind directional constancy at 2 m during the eleven clear sky days of ENABLE is very high with a value of 0.95. Figure 10 shows that the wind blows persistently from an east/north-easterly direction during these days. This direction is close to the slope direction of \(\sim 63^\circ \). Figure 11 shows the mean clear sky daily cycle of wind direction during ENABLE. The down-slope wind direction at night is consistent with the behaviour of a katabatic wind, but a rotation to the left due to the Coriolis force is not observed. During the day the wind direction in the mixed layer is most likely determined by large scale winds and is also east/north-easterly. We can identify a slight anti-clockwise rotation of the wind vector between \(\sim 9 \text{ h} \) and \(16 \text{ h GMT} \). The standard deviation shows that the daytime variability in wind direction is smaller than at night.

Stability and turbulent scales
Figure 12a shows the mean clear sky daily cycle of the stability parameter \( x = z/L_\ast \), which is positive for stable
stratification and negative for an unstably stratified layer. The figure also indicates slightly unstable conditions between approximately 9 h and 17 h GMT. The remainder of the day the stratification is stable. Stability peaks between 19 h and 22 h, after an increase early in the evening. This happens rapidly in comparison with the relatively gradual decrease in stability in the first part of the day. Standard deviation is large but can be understood from the non-linear dependence of $z/L_*$ on $u_*$ and the large day to day variability in $u_*$ as will be shown below. However, the daily cycle in stability is similar for all eleven clear sky days.

The behaviour of $z/L_*$ can be understood from the diurnal variability in the turbulent scales $u_*$ and $T_*$ (Eq. 5 shows that $z/L_*$ is fairly independent of $q_*$). The sign of stability parameter $z/L_*$ is fully determined by $T_*$. A high friction velocity $u_*$ acts to increase the absolute value of $L_*$ through enhanced vertical mixing. From this, we expect maximum stability to occur when $u_*$ is small and $T_*$ is large and positive. This is confirmed by Fig. 13, in which $T_*$, $q_*$ and $u_*$ are plotted. The sudden increase in stability in the early evening is caused by low near-surface wind speeds. The noontime maximum in $u_*$ coincides with negative values of $T_*$ which results in slightly unstable conditions. Friction velocity peaks ~ 2 and 3 h before $T_*$ and $q_*$ do, respectively. Furthermore, $T_*$ and $q_*$ obtain their minimum values when gradients in $T$ and $q$ are largest, which occurs prior to the maximum in air temperature and specific humidity (Fig. 8). Standard deviation is relatively large for $u_*$ due to the day to day differences in daily mean wind speed. However, the shape of the daily cycle of $u_*$ is similar for all eleven clear sky days.

To investigate how the stability parameter influences the turbulent heat fluxes through Eq. (4), we have plotted the contribution of the stability correction parameter $y$ to the absolute value of the turbulent heat fluxes in Fig. 12b. Clearly the influence is large. During daytime the correction increases the fluxes only by ~ 6%, but at night the stability correction reduces the turbulent heat fluxes by ~ 50%.

Surface energy balance

Figure 14a shows that the mean incoming solar radiation over the eleven clear sky days during ENABLE contributes a maximum of approximately 680 W m$^{-2}$ at noon to the surface energy balance. Most of this is reflected at the surface due to the high surface albedo (mean value of ~ 0.86 during ENABLE), leaving a maximum of 107 W m$^{-2}$ for net shortwave radiation. Incoming and outgoing longwave radiation are relatively small due to the low temperatures in the atmosphere and at the surface. $LR_{\text{net}}$ displays a strong daily cycle due to the daily temperature variation of the surface (Fig. 14b). $LR_{\text{net}}$ has an amplitude of ~ 6 W m$^{-2}$ over a day; the maximum value is attained in the afternoon, when the ABL temperature peaks. The surface looses energy through $LR_{\text{net}}$. Since the surface continues to warm even after the maximum in $SR_{\text{net}}$ (Fig. 8a), $LR_{\text{net}}$ lags $SR_{\text{net}}$ by approximately three hours. Figure 14c shows that this phase difference results in a maximum in net radiation approximately one hour before noon with a value of ~ 30 W m$^{-2}$.

In spite of its small value, $R_{\text{net}}$ dominates the SEB, as shown in Fig. 14d. The other components, of which the $G$ has the greatest magnitude, all tend to counteract $R_{\text{net}}$. From Fig. 13 we expect a negative $H_q$ between approximately 9 h and 16 h GMT and a negative $H_L$ between 8 h and 18 h GMT, which is confirmed by Fig. 14d. Due to the small values of the surface roughness lengths and the limited amount of solar radiation, $H_q$ and $H_L$ are small, contributing less than 9 W m$^{-2}$ and 2 W m$^{-2}$ to the mean SEB, respectively. The latent heat flux is especially small due to the small gradients in specific humidity, owing to the low temperatures.

The SEB components have all been determined independently. According to Eq. (1) the components should be balanced for a non-melting skin surface layer without heat capacity. The residual energy plotted in Fig. 14d is the sum of all components and shows that this condition has not been met entirely. This residual energy can be attributed to sensor and calculation inaccuracies.

Summary

In this paper we presented the summertime clear sky daily cycle of near-surface meteorological variables and the surface energy balance near Kohnen base, located on the high Antarctic plateau. Absorbed solar radiation drives the diurnal cycle in the other components of the SEB, even though the high albedo (~ 0.86) results in only a relatively small $SR_{\text{net}}$. Since $LR_{\text{net}}$ reaches a minimum value at ~ 14–15 h GMT, when surface temperature is high, $R_{\text{net}}$ peaks just before noon. The turbulent heat fluxes and the sub-surface heat flux distribute the surplus radiative energy over the ASL and the upper snow layer. In spite of large near-surface temperature gradients, the mean turbulent heat fluxes are small due to the low surface roughness and generally weak winds. Between ~ 9 and 16 h surface temperatures exceed air temperatures, which results in weak convection. A short-lived, shallow mixed layer is formed in the otherwise stably stratified ASL. $T_*$ and $q_*$ reach their maximum values before $T(0)$ and $q(0)$ do around 14–15 h, while $u_*$ and $u$ peak simultaneously around noon. The directional constancy at daytime is remarkably high, which is proof of a stationary large-scale setting. After ~ 16 h stability rapidly increases due to an increase in $T_*$ and a sudden drop in $u_*$. Maximum stability is reached between ~ 19 h and 22 h, when $T_*$ and $u_*$ attain their maximum and minimum values, respectively. In this stable layer a nocturnal jet develops. Although the jet is located in the layer of temperature deficit and it is directed in the downslope direction, further analysis is necessary to determine
whether this jet is entirely katabatically driven, or partially is a manifestation of an inertial oscillation.

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References


