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Longwave atmospheric radiation over Antarctica

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Abstract: The seasonal and spatial variation of downward longwave radiation ($L\downarrow$) at the surface of the Antarctic ice sheet is studied using data from four stations. $L\downarrow$ is found to depend on the fourth power of the surface temperature but on the sixth power of the temperature at the top of the surface inversion layer. The former result is shown to be a consequence of the dominance of longwave radiation in the surface energy balance while the latter raises some questions concerning the radiative properties of Antarctic clouds. Simple parametrizations of $L\downarrow$ as a function of temperature are suggested; these may find application in validating General Circulation Model radiation calculations over Antarctica.

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Introduction

The downward component of longwave radiation is one of the largest terms in the energy balance at the surface of the Antarctic ice sheets. It is also one of the most difficult terms to measure or to parametrize. The present study was motivated by a need to develop a longwave radiation climatology for Antarctica against which General Circulation Model (GCM) radiation calculations could be validated. Since only a limited number of radiation measurements over Antarctica are available and, in general, these will not be co-located with GCM grid points, the approach taken was to use the observations to develop regression relations between downward longwave radiation and more readily available climatological variables. These can then be compared with similar relationships derived from GCM data.

Two simple parametrizations are shown to give a reasonable description of the seasonal and geographical variation of longwave radiation over the East Antarctic Ice Sheet. The forms of the parametrizations also provide some interesting insights into radiative processes in the Antarctic atmosphere.

Data

Surface radiation measurements have been published for a number of Antarctic stations (e.g. Rusin 1961, Schwerdtfeger 1984) but are of rather variable quality. For this study, the following criteria were used to select data:

1) Instruments and measurement procedures must be well documented to ensure consistency of the measurements.
2) Measurements must have been made over permanent snow and ice to be representative of most of the surface of the Antarctic continent.
3) Measurements must be available for all months of the year.
4) Longwave radiation fluxes must be measured directly rather than computed from all-wave radiation balance.

Only measurements made at South Pole (SPO), Mizuho (MIZ), Syowa (SYO) and Neumayer (NEU) stations fulfilled all four criteria. Some details of these observations are given in Table I. In this study the published monthly mean radiation data for these stations were averaged if measurements from more than one year were available to produce a dataset containing twelve monthly averages for each station. The series of measurements from NEU is sufficiently long to provide an indication of the level of interannual variability. At this station, the interannual standard deviation of monthly means of downward longwave radiation is less than 10% of the long term mean. Measurements from a single year at other stations should thus be reasonably representative of climatological conditions.

The four stations considered span the entire range of climatological conditions encountered on the East Antarctic ice sheet. SPO is typical of the high, cold interior plateau, MIZ is situated in the peripheral region of high slope and is subject to strong katabatic winds, NEU and SYO are both typical of the coastal zone.

Analysis

Downward longwave radiation at the surface, $L\downarrow$, is, in general, a complicated function of the vertical distribution of temperature and water vapour in the atmosphere, of cloud cover and of the optical properties of any clouds present. Any parametrization of $L\downarrow$ in terms of temperature at a single level must be a gross simplification, but simple climatological relationships have been developed with moderate success. In the absence of cloud, $L\downarrow$ is dominated by emission from atmospheric water vapour and, to a lesser extent, carbon dioxide. Even in the relatively dry Antarctic atmosphere emissions from water vapour near the surface dominate $L\downarrow$.

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Table I. Details of stations used in this study.

Station	Location	Elevation (m)	Period of measurements	Reference
South Pole (SPO)	90°S	2835	1986–88	Dutton *et al.* 1989
Mizuho (MIZ)	71°S,44°E	2230	1979–80	Yamanouchi *et al.* 1981 Ishikawa *et al.* 1982
Syowa (SYO)	69°S,40°E	15	1987–88	Yamanouchi 1989
Neumayer (NEU)	71°S,8°W	35	1982–92	Schmidt & König-Langlo 1994

Thus, the effective radiative temperature of the atmosphere does not differ greatly from the surface temperature. Swinbank (1963) showed that, in cloud-free conditions and over a wide range of air temperatures, $L\downarrow$ was well described by the relation:

$$L\downarrow = 5{\cdot}31\times10^{-13}T_A^6 \quad (1)$$

where $L\downarrow$ is measured in W m^{-2} and T_A is the "screen" temperature (i.e. near-surface air temperature) in degrees K. This result is superficially surprising since, intuitively, one might expect a fourth-power relationship between radiative flux and temperature from the Stefan-Boltzmann law. An explanation of the sixth-power relationship was provided by Deacon (1970), who noted that there was a good climatological correlation between total column water vapour and surface temperature, arising from the strong (approximately exponential) variation of saturation vapour pressure with temperature. The optical depth (and hence the effective emissivity) of the atmosphere can be expressed as a function of the total column water vapour. When this is combined with the observed relationship between surface temperature and column water vapour it emerges that the effective emissivity is proportional to the square of the surface temperature and hence the radiative flux is related to the surface temperature by a sixth-power law over a wide temperature range.

Swinbank's empirical relationship was derived from observations made under clear skies. Under conditions of significant low cloud cover, $L\downarrow$ will be dominated by emission from cloud water droplets and ice particles. If the sky is totally overcast and the cloud is both optically thick (i.e. near "black") and low enough for absorption in the sub-cloud layer to be neglected, $L\downarrow$ will be close to the blackbody radiation at cloud base temperature, T_C, and hence proportional to T_C^4. For low-level clouds, T_C may be reasonably close to T_A so a fourth-power relationship between $L\downarrow$ and T_A under overcast conditions might be expected. During a given month, a station will experience a combination of clear, overcast and partly cloudy conditions so the relationship between climatological values of T_A and $L\downarrow$ is expected to be complex.

A strong surface-based temperature inversion is present over the Antarctic ice sheets for much of the year (Phillpot & Zillman 1970). The temperature at the surface may be up to 25 K colder than that at 200–300m above the surface and, under clear-sky conditions, the temperature, T_M, at the top of the inversion will be more representative of the radiating atmosphere than will the surface temperature since the warmer, moister air above the inversion is a more effective emitter than the cold, dry air within the inversion layer. Observations also indicate that, over the East Antarctic plateau, cloud bases generally coincide with the top of the inversion layer (Stone 1993). One therefore might expect to find a better climatological relationship between $L\downarrow$ and T_M than between $L\downarrow$ and T_A under both clear and overcast conditions.

Monthly mean values of T_M for SPO, SYO and NEU were calculated from the database of Antarctic temperature soundings described by Connolley & King (1993). For MIZ, only a limited number of temperature soundings are available (Kawaguchi *et al.* 1985). Monthly mean values of inversion strength, T_M–T_A, were calculated from these limited data (for some months as few as four soundings are available) and were then added to the respective monthly means of T_A, calculated from regular synoptic observations, to give the best possible estimate of T_M for MIZ.

Fig.1 shows monthly means of $L\downarrow$ plotted against a) T_A and b) T_M on log-log plots. Both temperatures are seen to be quite

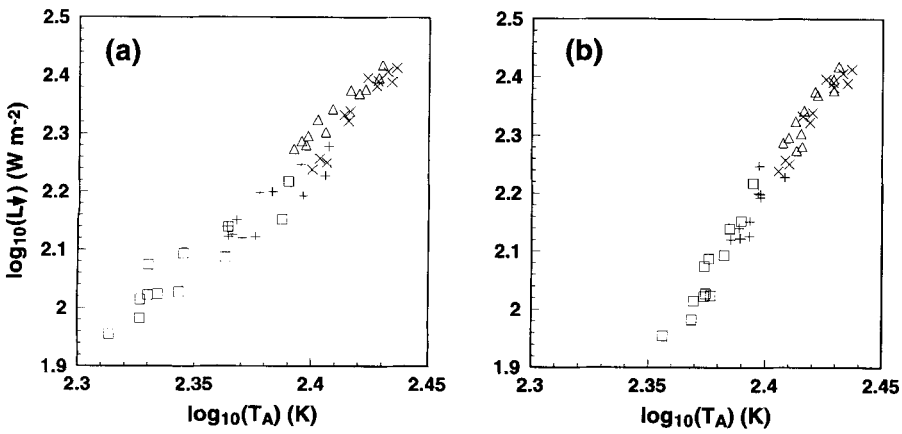


Fig. 1. Log-log plots of monthly mean downward longwave radiation as a function of **a.** surface air temperature and **b.** temperature at the top of the surface inversion. □- SPO, + - MIZ, Δ- NEU, × - SYO.

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good predictors of $L\downarrow$ and linear regression gives the following relationships:

$$L\downarrow = 1{\cdot}09\times10^{-7}T_A^{3{\cdot}85} \quad (2)$$

$$L\downarrow = 1{\cdot}31\times10^{-13}T_M^{6.29} \quad (3)$$

Thus when T_M is used as a predictor variable, a sixth-power relationship reminiscent of the Swinbank relationship emerges while using T_A leads to a fourth-power Stefan-Boltzmann law behaviour. Possible reasons for this behaviour are discussed below.

Regressing $L\downarrow$ on T_A^4 gives a best-fit line:

$$L\downarrow = 0.49 + 4.70\times10^{-8}T_A^4 \quad (4)$$

with a correlation coefficient of 0.97 and an RMS error of 12.2 W m^{-2}, while using T_M^6 as a predictor gives

$$L\downarrow = -5{\cdot}08 + 6{\cdot}68\times10^{-13}T_M^6 \quad (5)$$

with a slightly higher correlation coefficient of 0.98 and an RMS error of 9.8 W m^{-2}. There is thus little to chose between the two expressions as parametrizations of L$\downarrow$, although T_M may be preferred as a predictor on physical grounds. However, measurements of T_A are available from most Antarctic research stations and from an increasing number of automatic weather stations, while measurements of T_M are only available from the limited number of stations which operate radiosonde programmes.

It is important to recognize that eq. (4) and (5) are both empirical results derived from analysis of a limited dataset. While such relationships may provide some insight into radiative processes in the Antarctic atmosphere (as discussed below), the principal reason for performing the analysis was to develop simple empirical models for $L\downarrow$, using monthly mean temperature as the only predictor variable.

Discussion

The energy balance at the surface of the Antarctic ice sheets is given by:

$$S(1-\alpha) + L\downarrow - \varepsilon\sigma T_S^4 + H_S + H_L + G = 0 \quad (6)$$

where S is the global solar radiation, α is the surface albedo, T_S is the snow surface temperature, ε is the snow surface infrared emissivity, σ is the Stefan-Boltzmann constant, H_S and H_L are the atmospheric sensible and latent heat fluxes and G is the upward heat flux through the snowpack. During the Antarctic winter S is negligible over most of the continent. The highly stable stratification of the boundary layer restricts turbulent transport and monthly means of H_S are only 10–20 W m^{-2}, H_L is insignificantly small and G is also only a few W m^{-2} (Carroll 1982, King *et al.* 1995). By contrast, $L\downarrow$ is of order 100 W m^{-2}, so the dominant balance in eq. (6) must be between downward and upward longwave radiation, i.e. T_S^4 will be proportional to $L\downarrow$. The difference between monthly means of T_S and T_A rarely exceeds 3K even on the Antarctic plateau (Schwerdtfeger 1984, table 2.3), so a close relationship might be expected between $L\downarrow$ and T_A. Strictly, eq. (6) only holds for short-term (e.g. hourly) means and the monthly mean of (hourly) T_S^4 will, in general, differ from the fourth power of the monthly mean of T_S. However, examination of time series of hourly temperatures from Antarctic stations shows that the two quantities differ by less than 1% of their mean. Hence, eq. (6) will also hold to a very good approximation for monthly mean data.

During the summer, S is often >400 W m^{-2} on the Antarctic plateau but the snow surface albedo is 80–85% so the net shortwave radiation is generally <80 W m^{-2}. This is still significant in comparison with the longwave fluxes so one might not expect such a close relationship between $L\downarrow$ and T_S^4 during the summer months. In fact, if the regression analysis is repeated for the summer months (December–February) only, the goodness of fit is little changed but the best-fit regression line is now given by:

$$L\downarrow = -26{\cdot}99 + 5{\cdot}21\times10^{-8}T_A^4 \quad (7)$$

From eq. (6), it is obvious that one would expect a negative intercept in the $L\downarrow$–T_A^4 relationship if S is comparable with the longwave fluxes. However, longwave radiation measurements are subject to errors in the presence of solar radiation and these errors could contribute to the different regression relationships seen during summer and winter.

The reasons for the observed dependence of $L\downarrow$ on T_M^6 are not so obvious. Swinbank's formula (eq. 1) does predict such behaviour, but this was derived from measurements of $L\downarrow$ under clear-sky conditions with surface temperatures between 8 and 30°C. The monthly mean data used in the present analysis cover the range -46°C to 0°C and each monthly mean represents an aggregate of clear, overcast and partly cloudy conditions. The coefficient of T_M^6 in eq. (5) is larger than that in Swinbank's formula — this is not unexpected because the presence of clouds for part of any month will increase the average effective emissivity of the atmosphere. What is surprising is that a sixth-power relationship still holds for the monthly mean data.

Examination of cloud-cover reports from Antarctic stations shows that, on the majority of occasions, conditions are either overcast (7–8 octas cloud cover) or clear (0–1 octas cover). Overcast conditions occur more frequently in summer than in winter while, throughout the year, the interior of the continent is less cloudy than the coastal regions. In both regions, cloud radiation would be expected to make a significant contribution to the monthly mean $L\downarrow$. However, the magnitude of this contribution will depend not only on the frequency of cloud cover but also on the radiative properties of the clouds present.

Very few studies have been made of the infrared optical properties of Antarctic clouds. Stone (1993) used data from radiometersondes launched at the South Pole to deduce the effective emissivity of cloud layers. He concluded that the clouds were optically thin and non-black, with average

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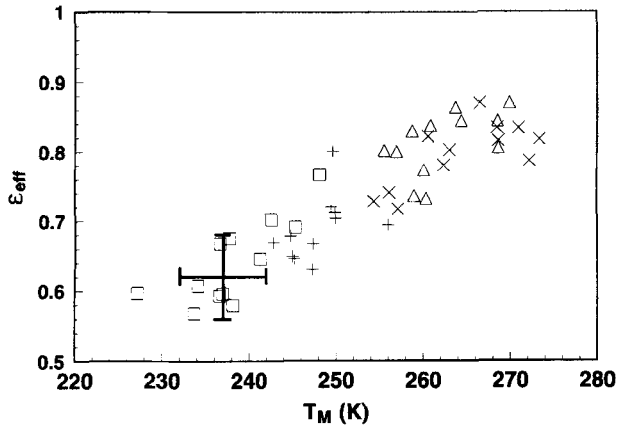


Fig. 2. Monthly mean atmospheric effective emissivity as a function of temperature at the top of the surface inversion. Station symbols are as for Fig. 1. The point with error bars indicates the range of cloud emissivities measured at SPO by Stone (1993).

effective emissivities of about 0.62. Their optical properties were found to be characteristic of ice clouds varying in ice water content from 3×10^{-7} to 6×10^{-6} kg m^{-3}. Because ice water content depends largely on cloud temperature (Heymsfield & Platt 1984), the effective emissivity will also increase with increasing temperature, leading to a higher power than four in the $L\downarrow$–T_M relationship under overcast conditions.

Fig. 2 shows the monthly mean effective emissivity of the atmosphere at each site, defined as

$$\varepsilon_{eff} = \frac{L\downarrow}{\sigma T_M^4} \tag{8}$$

as a function of temperature at the top of the inversion layer. The values of ε_{eff} for SPO are consistent with those determined by Stone (1993) at that station for overcast conditions. The increase in ε_{eff} with increasing temperature results from the tendency for cloudiness to increase with increasing temperature and also the temperature dependence of both cloudy and clear sky emissivities. Unfortunately, with only monthly mean data available, it is not possible to separate data into clear and cloudy conditions and hence determine the relative contribution of clouds to the observed temperature dependence of ε_{eff} on T_M.

If $L\downarrow$ is plotted against T_M on a log-log plot for each of the four stations individually, the best-fit power obtained is close to six for each station, indicating that the same ε_{eff} - temperature relationship holds both at the coast and over the interior. This similarity is somewhat surprising. Over the Antarctic interior, clouds have relatively low ice water contents and their emissivities may well be temperature dependent. However, at the Antarctic coast, temperatures are high enough throughout the year for most low clouds to contain significant amounts of liquid water and thus act like black bodies, with emissivities close to 1.0. König-Langlo & Augstein (1994) studied the dependence of $L\downarrow$ on cloud cover at NEU and showed that, to a very good approximation:

$$L\downarrow = (0.765 + 0.22c^3)\sigma T_A^4 \tag{9}$$

where c is the fractional cloud cover. Thus, under overcast skies (c=1) the effective emissivity is very nearly equal to 1.0 as would be expected of clouds having large liquid water contents. The apparent variation of monthly mean ε_{eff} with temperature seen at the coastal stations NEU and SYO (Fig. 2) is likely to result from variations in cloud cover, together with the variation of clear-sky emissivity with temperature, rather than a dependence of cloud emissivity on temperature. Any correlation between temperature and cloud cover would also introduce an *apparent* variation of effective emissivity with temperature but analysis of cloud cover data for NEU and SYO shows little evidence for such a correlation.

Conclusions

Analysis of monthly mean values of downward longwave radiation, $L\downarrow$, at four Antarctic stations has shown that $L\downarrow$ varies with the fourth power of the surface temperature and the sixth power of the temperature at the top of the surface inversion layer. Surface heat fluxes are small over the Antarctic ice sheets and the fourth power dependence on surface temperature results from a surface energy budget dominated by a near balance between upward and downward longwave radiation. Because of this, the fourth-power relationship (eq. 4) should not be used to predict $L\downarrow$ in regions where such a balance does not hold. For example, one would not expect eq. (4) to provide a good parametrization of $L\downarrow$ over coastal polynyas, where the surface heat fluxes in eq. (6) are of comparable magnitude to the radiative fluxes (Kurtz & Bromwich 1985).

The sixth power relationship between $L\downarrow$ and T_M (eq. 5) is reminiscent of Swinbank's (1963) formula for clear-sky radiation. However, because eq. (5) was derived from monthly mean data, including both clear, partly cloudy and overcast conditions, two further effects may contribute to this relationship. Firstly, over the interior of Antarctica, clouds are relatively tenuous and radiate like grey bodies, having effective emissivities which are strongly dependent on cloud ice water content, and hence on temperature. Secondly, variations in the amount and type of cloud cover will also lead to variations in effective emissivity. Evaluation of the relative contribution of each of these effects to the overall dependence of $L\downarrow$ on T_M will require the analysis of daily or even hourly radiation measurements in conjunction with correlative sky cover observations.

The parametrizations given by eq. (4) and (5) have both been shown to give a reasonable description of the geographical and seasonal variation of $L\downarrow$ and may prove of use in developing radiation climatologies for Antarctica. Measurements of T_A are available from an increasing number of automatic weather stations while surface temperatures can

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be mapped by satellite radiometers (e.g. Comiso 1994). Although T_M measurements are currently only available from radiosonde stations it may prove possible to use satellite sounder data to map T_M and thus provide an additional means of parametrizing $L\downarrow$. If cloud cover observations are available, a parametrization such as eq. (9), which explicitly accounts for the effects of clouds, may give better results than eq. (4) or (5). However, the coefficients in eq. (9) were derived from observations at a coastal station, where clouds are always "black". The present analysis confirms that clouds over the interior have emissivities significantly <1.0 and eq. (9) may not provide a good parametrization of $L\downarrow$ under these circumstances.

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References

Carroll, J.J. 1982. Long-term means and short-term variability of the surface energy balance components at the South Pole. *Journal of Geophysical Research*, **87**, 4277–4286.

Comiso, J.C. 1994. Surface temperatures in the polar regions from Nimbus 7 temperature humidity infrared radiometer. *Journal of Geophysical Research*, **99**, 5181–5200.

Connolley, W.M. & King, J.C. 1993. Atmospheric water vapour transport to Antarctica inferred from radiosonde data. *Quarterly Journal of the Royal Meteorological Society*, **119**, 325–342.

Deacon, E.L. 1970. The derivation of Swinbank's long-wave radiation formula. *Quarterly Journal of the Royal Meteorological Society*, **96**, 313–319.

Dutton, E.G., Stone, R.S. & DeLuisi, J.J. 1989. *South Pole surface radiation balance measurements April 1986 to February 1988. NOAA Data Report ERL ARL-17.* Silver Spring, Maryland: Air Resources Laboratory, 49 pp.

Heymsfield, A.J. & Platt, C.M.R. 1984. A parameterization of the particle size spectrum of ice clouds in terms of the ambient temperature and the ice water content. *Journal of the Atmospheric Sciences*, **41**, 846–855.

Ishikawa, N., Kobayashi, S., Ohata, T. & Kawaguchi, S. 1982. *Radiation data at Mizuho station in Antarctica in 1980. JARE Data Reports* No. 73. Tokyo: National Institute of Polar Research, 195 pp.

Kawaguchi, S., Kobayashi, S., Ishikawa, N., Ohata, T., Inoue, J., Satow, K. & Nishimura, H. 1985. *POLEX-South data, part 6. Aerological soundings of the lower atmospheric layer over Mizuho Plateau, East Antarctica. JARE Data Reports* No. 104. Tokyo: National Institute of Polar Research, 128 pp.

King, J.C., Anderson, P.S., Smith, M.C. & Mobbs, S.D. 1995. Surface energy and water balance over an Antarctic ice shelf in winter. *In Preprints, Fourth Conference on Polar Meteorology and Oceanography*. Boston: American Meteorological Society, 79–81.

König-Langlo, G. & Augstein, E. 1994. Parameterization of the downward long-wave radiation at the Earth's surface in polar regions. *Meteorologisches Zeitschrift*, **3**, 343–347.

Kurtz, D.D. & Bromwich, D.H. 1985. A recurring, atmospherically forced polynya in Terra Nova Bay. *Antarctic Research Series*, **43**, 177–201.

Phillpot, H.R. & Zillman, J.W. 1970. The surface temperature inversion over the Antarctic continent. *Journal of Geophysical Research*, **75**, 4161–4169.

Rusin, N.P. 1961. *Meteorological and radiational regime of Antarctica.* Jerusalem: Israel Program for Scientific Translations, 355 pp.

Schmidt, T. & König-Langlo, G. 1994. Radiation measurements at the German Antarctic Station Neumayer 1982–1992. *Berichte zur Polarforschung*, **146**. Bremerhaven: Alfred-Wegener-Institute fur Polar- und Meeresforschung, 1-66.

Schwerdtfeger, W. 1984. *Weather and climate of the Antarctic.* Amsterdam: Elsevier, 261 pp.

Stone, R.S. 1993. Properties of austral winter clouds derived from radiometric profiles at the South Pole. *Journal of Geophysical Research*, **98**, 12961–12972.

Swinbank, W.C. 1963. Long-wave radiation from clear skies. *Quarterly Journal of the Royal Meteorological Society*, **89**, 339–348.

Yamanouchi, T. 1989. *Antarctic climate research data, Part 1. Radiation data at Syowa Station, Antarctica from February 1987 to January 1988. JARE Data Reports* No. 144. Tokyo: National Institute of Polar Research, 193 pp.

Yamanouchi, T., Wada, M., Mae, S. & Kawaguchi, S. 1981. *POLEX-South data. Part 1 Radiation data at Mizuho station in 1979. JARE Data Reports* No. 61. Tokyo: National Institute of Polar Research, 350 pp.

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