Impact of Clouds on the Shortwave Radiation Budget of the Surface–Atmosphere System for Snow-Covered Surfaces

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ABSTRACT

Recent data from the Earth Radiation Budget Experiment (ERBE) have raised the question as to whether or not the addition of clouds to the atmospheric column can decrease the top-of-the-atmosphere (TOA) albedo over bright snow-covered surfaces. To address this issue, ERBE shortwave pixel measurements have been collocated with surface insolation measurements made at two snow-covered locations: the South Pole and Saskatoon, Saskatchewan. Both collocated datasets show a negative correlation (with solar zenith angle variability removed) between TOA albedo and surface insolation. Because increased cloudiness acts to reduce surface insolation, these negative correlations demonstrate that clouds increase the TOA albedo at both snow-covered locations.

1. Introduction

The availability of data from the Earth Radiation Budget Experiment (ERBE) has provided unique insights into many aspects of cloud–climate interactions (Ramanathan et al. 1989; Harrison et al. 1990). This was accomplished by separately averaging clear-sky top-of-the-atmosphere (TOA) radiative flux measurements so that the TOA radiation budget with clouds present could be compared to that without clouds. The ERBE results suggest that over bright snow surfaces the addition of clouds to the atmospheric column can act to reduce the TOA albedo (Ramanathan et al. 1989; Harrison et al. 1990), whereas theoretical model studies indicate the opposite (Shine et al. 1984; S. Warren 1992, personal communication). Harrison et al. (1990), however, caution that the ERBE clear-sky scene identification is considered highly unreliable over snow, a point that has been amplified in subsequent studies (Doelling 1991; Li and Leighton 1991).

Irrespective of the ERBE data, observational knowledge of how clouds impact the TOA albedo over snow would provide information that could be used to improve our knowledge of cloud radiative processes, particularly since this problem has never been addressed using radiometric data. Because they are highly reflective, clouds increase the TOA albedo over dark surfaces such as vegetation and oceans. But it is not necessarily obvious that this will occur over bright snow surfaces. If, for example, there is sufficient absorption of shortwave (SW) radiation by a cloud, then it could act to darken the atmospheric column relative to clear conditions. To obtain a more comprehensive understanding of this issue, ERBE SW pixel measurements have been collocated with near-surface SW measurements made at two locations: the South Pole and Saskatoon, Saskatchewan. These collocated datasets thus allow clarification as to whether clouds increase or decrease the TOA albedo over snow by using near-surface insolation as a measure of clear/overcast conditions.

2. Datasets

A detailed description of the South Pole instrumentation is given by Dutton et al. (1989), so only a brief discussion is given here. Near-surface insolation was measured by an upward-facing Eppley pyranometer located approximately 10 m above the ground on the roof of the Clean Air Facility (CAF) Building, which is part of the National Science Foundation's South Pole Station. Upward-reflected SW radiation from the surface was additionally measured by a downward-facing pyranometer mounted on a support structure 1 to 2 m (depending on snow depth) above the surface and

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located 80 m from the CAF Building. The pyranometer mount is designed to cast an insignificant shadow that does not affect the reflected SW measurements. The data are averaged over 24-h periods for purposes of this study. The spacially collocated ERBE data consist of net downward SW radiation at the TOA as determined from pixel measurements made by the SW cross-track scanner on *NOAA* 9. The collocation procedure is the same as that described by Cess et al. (1991). This satellite is in a sun-synchronous orbit and the scanner views the South Pole 14 times a day, thus providing realistic 24-h averages. A total of 131 collocated 24-h averages were obtained over five months: December 1985, January, November and December 1986, and January 1987.

The collocated measurements for Saskatoon (52.14°N, 107.07°W) are described by Li et al. (1993a). The near-surface instruments consist of upward- and downward-facing Eppley pyranometers that are mounted on a 2-m boom extending from the top of a 10-m tower. The data were recorded as 1-min averages. The surrounding terrain is flat agricultural land with some trees that cover less than 10% of the area within 20 km of the tower. Restriction was made to periods when the surface was covered with snow. The temporally and spatially collocated ERBE pixel measurements refer to the SW scanner on the *Earth Radiation Budget Satellite* (*ERBS*), whose orbit is inclined 57° relative to the equator and so provides sampling over the diurnal cycle. The measurements span the period from when the tower went into operation (October 1989) to the failure of the *ERBS* scanner (28 February 1990) and consist of 99 collocated measurements. Because of the restriction that snow be on the ground, these 99 measurements actually refer to a four-month period (November 1989 through February 1990).

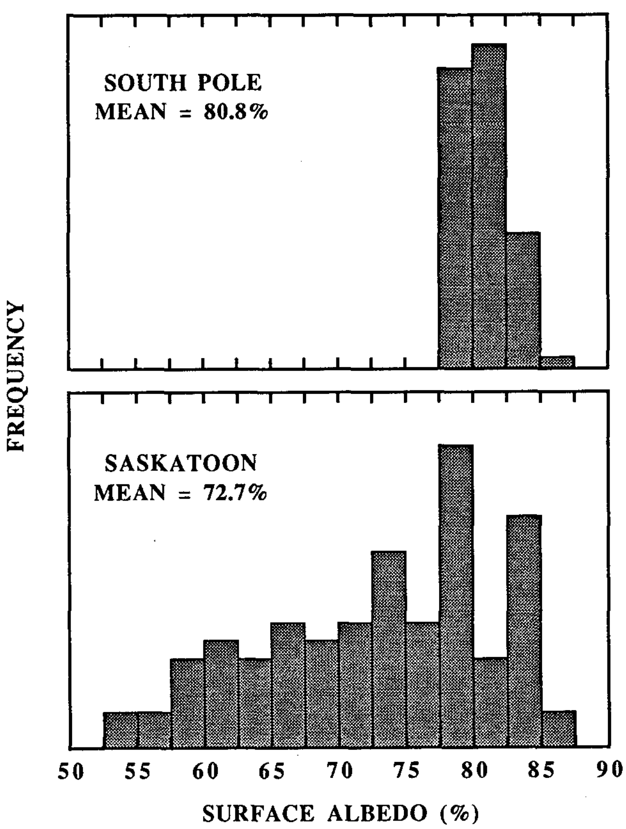

FIG. 1. Histograms of surface albedo measured at the South Pole and at Saskatoon.

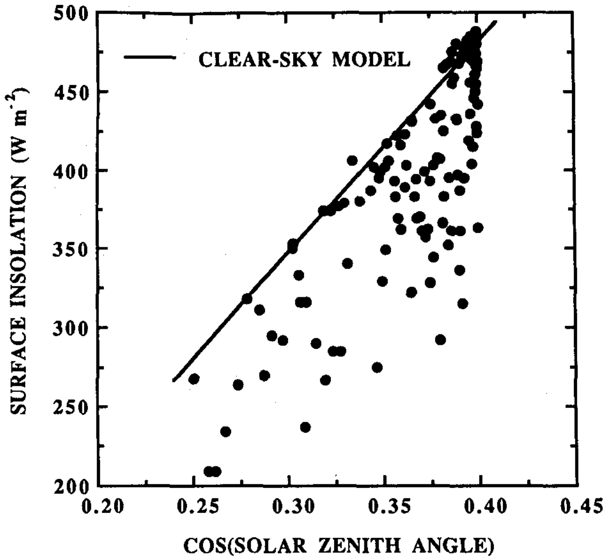

FIG. 2. Measurements of surface insolation at the South Pole as a function of the 24-h mean cosine of the solar zenith angle.

Uncertainty in the ERBE clear-sky identification over snow could impact the quality of the ERBE SW fluxes, because the angular-directional models that convert measured radiance to flux depend on whether the scene is clear or overcast. Nemesure (1991), however, has shown that any such error would be small.

The downward-facing pyranometers at both locations allow evaluation of the underlying snow albedos, which are presented as histograms in Fig. 1. That the snow surface albedo for Saskatoon exhibits greater variability is consistent with warmer temperatures causing coalescence (aging) of the snow particles and thus surface darkening, while the lowest albedo values are most likely indicative of thin or patchy snow.

3. South Pole results

Illustrated in Fig. 2 are surface insolation measurements, obtained from the upward-facing pyranometer at the South Pole and representing the same 24-h means as the collocated ERBE measurements, as a function of cos(solar zenith angle). Also shown are results from a clear-sky model described by Cess et al. (1991), but modified to incorporate a snow-surface albedo appropriate to the South Pole and utilizing a water vapor column abundance of 0.12 cm that represents an average over the measurement period. Water vapor abundances were estimated from temperature soundings by assuming the atmosphere was saturated relative to ice. The clear-sky model coincides with the upper envelope of the data, as is consistent with clear-sky insolations being larger than for overcast conditions, and the ample downward scatter in the data from this upper envelope shows that clouds are exerting a significant impact on SW radiation within the atmospheric column.

The collocated ERBE data allow a direct assessment of the question as to whether clouds increase or decrease the TOA albedo over the South Pole. Since clouds reduce the surface insolation, then if the TOA albedo, as measured by ERBE, were negatively correlated with surface insolation, this would mean that clouds enhance the TOA albedo. A complicating factor is that the TOA albedo and surface insolation also depend on solar zenith angle through its influence on atmospheric scattering, atmospheric slant path, and surface albedo. But the South Pole data offer a convenient means of circumventing this problem, because much of the data lie in a very narrow range of solar zenith angles (Fig. 2).

We first employ 66 collocated measurements that are restricted to cos(solar zenith angle) ranging from 0.38 to 0.40, together with the regression

$$\alpha = A_0 + A_1 \times \mathrm{INSO}/\mu, \qquad (1)$$

where α is the TOA albedo, INSO is the surface insolation, and μ is the cosine of the solar zenith angle. To minimize the effect of this small variation in μ, we adopt INSO/μ, rather than INSO, because the former incorporates that part of the μ dependence associated with the TOA insolation. Since clouds reduce INSO/μ, then clouds enhance the TOA albedo if $A_1 < 0$. A scatterplot of the 66 TOA albedo versus INSO/μ measurements is given in Fig. 3; the cluster of points on the right represents clear skies, while the positive upward trend in progressing to the left is caused by increasing cloudiness. This clearly demonstrates that clouds over the South Pole, for the time interval considered, increase the TOA albedo. Inspection of the monthly mean ERBE data (e.g., Harrison et al. 1990), for the same time interval, show just the reverse, and comparison of the ERBE clear-sky scene identification with the upward-facing pyranometer measurements suggests that the fault lies with the ERBE scene identification as was anticipated by Harrison et al. (1990). The ERBE scene identification tends to reverse the identification of clear and overcast pixels at this location.

There is a straightforward explanation for this reversal. The ERBE scene identification employs a maximum likelihood estimator (MLE) that makes use of a priori statistics of shortwave and longwave measurements for different surface and cloud conditions. Over snow/ice surfaces there are only two classifications: clear and overcast. Moreover the statistics used to assign a pixel as overcast were derived exclusively from measurements over ocean and snow/ice-free land. When applied over highly reflective snow, the MLE predicts a lower TOA albedo for overcast conditions relative to clear, contrary to the present South Pole findings, and so reverses the overcast and clear scene identifications.

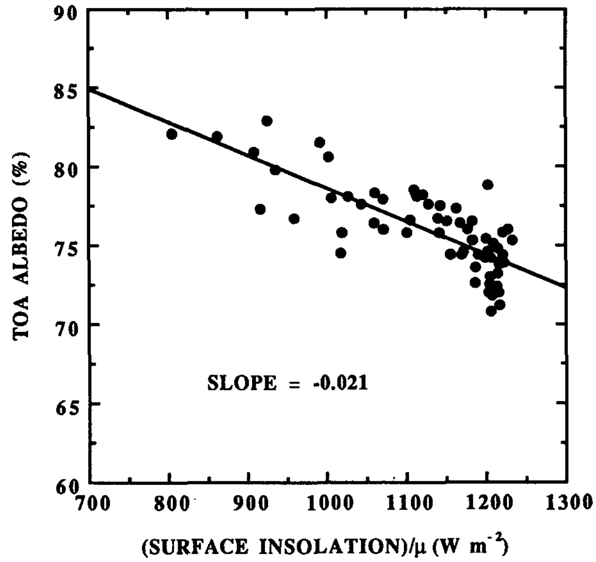

FIG. 3. South Pole TOA albedo as a function of surface insolation divided by μ. These refer to the 66 collocated measurements for which $0.38 < \mu < 0.40$.

The South Pole data serve another purpose. At different locations, such as Saskatoon, it is not possible to impose the restriction to a narrow range of μ values. An alternative approach is to remove the separate dependence on μ by adopting the two-variable regression

$$\alpha = B_0 + B_1(\mathrm{INSO}/\mu) + B_2\mu. \qquad (2)$$

To test the applicability of this, all 131 collocated measurements have been used to evaluate B_1 in (2), and as demonstrated in Fig. 4, this produces virtually the same result as did the 66 collocated measurements used with (1).

The collocated data provide additional information concerning the radiative impact of clouds on the atmospheric column. Measurements by the downward-facing pyranometer allow determination of the net downward SW flux at the surface. With SURF denoting this quantity, while TOP is the net flux at the TOA, then as discussed by Cess et al. (1993) the partial derivative

$$\left[\frac{\partial(\mathrm{SURF})}{\partial(\mathrm{TOP})}\right]_\mu$$

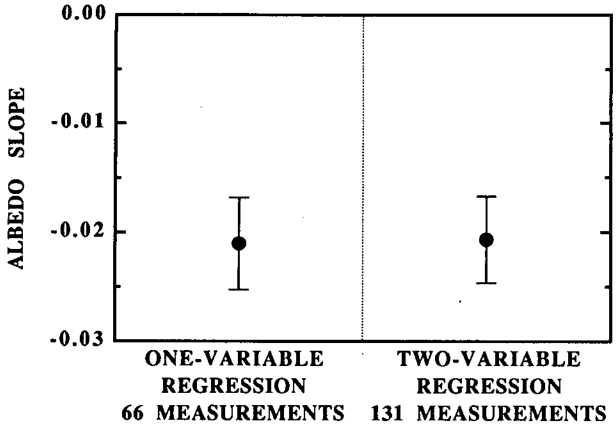

FIG. 4. Summary of albedo slopes for the South Pole as determined from one-variable and two-variable regressions. The vertical bar denotes the 95% confidence interval.

has a specific physical meaning. If the addition of clouds to the atmospheric column produces a greater decrease in TOP than in SURF, then adsorption of SW radiation by the column (TOP minus SURF) is decreased and the value of the above derivative is less than unity. Thus the derivative determines whether clouds radiatively (SW) heat (derivative > 1) or cool (derivative < 1) the column.

The 66 measurement pairs for which μ is essentially fixed provide an estimate of this derivative through the regression

$$\mathrm{SURF} = C_0 + C_1\mathrm{TOP}. \quad (3)$$

These measurement pairs are illustrated in Fig. 5 and demonstrate that the SW radiative impact of clouds acts to cool the atmospheric column at the South Pole. Again using a two-variable regression to remove the dependence on μ, then

$$\mathrm{SURF} = D_0 + D_1\mathrm{TOP} + D_2\mu \quad (4)$$

and the 131 measurement pairs produce a flux derivative (D_1) quite similar to that determined from the 66 collocated measurements using (3), as demonstrated in Fig. 6.

To summarize this section, use of data that are restricted to a small range of μ values in conjunction with one-variable regressions, or all of the data with the μ dependences removed through use of two-variable regressions, produce comparable conclusions. Over the South Pole the presence of clouds increases the TOA albedo and produces SW radiative cooling of the atmospheric column.

4. Saskatoon results

Illustrated in Fig. 7 are surface insolation measurements, as obtained from the upward-facing pyranometer at Saskatoon and for the times of the collocated

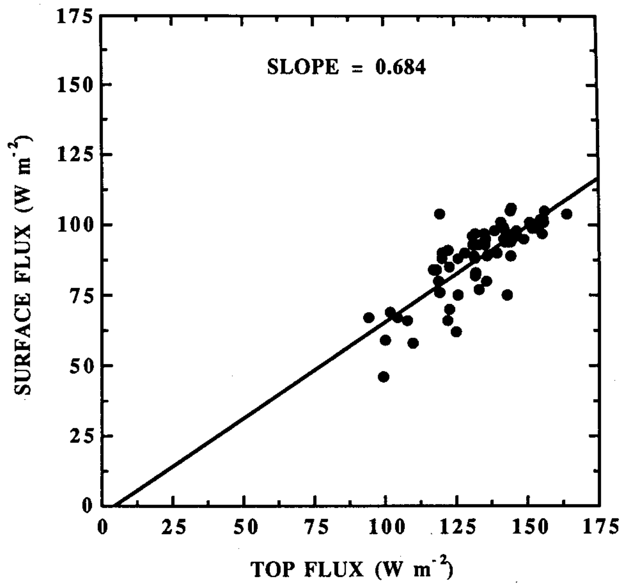

FIG. 5. South Pole surface flux as a function TOA flux for the 66 collocated measurements for which $0.38 < \mu < 0.40$.

ERBE measurements, as a function of cos(solar zenith angle). As for the South Pole (Fig. 2), these data demonstrate that clouds exert a substantial impact on atmospheric SW radiation at the Saskatoon location. Because of the large variation in surface albedo (Fig. 1), the clear-sky model has not been included in Fig. 7. The data show a continuous distribution of μ values from roughly 0.1 to 0.5, so that it is necessary to employ the two-variable regressions when interpreting the collocated data.

The albedo slope for Saskatoon, as given by B_1 in (2), is shown in Fig. 8 and compared to that for the South Pole. The Saskatoon slope is likewise negative

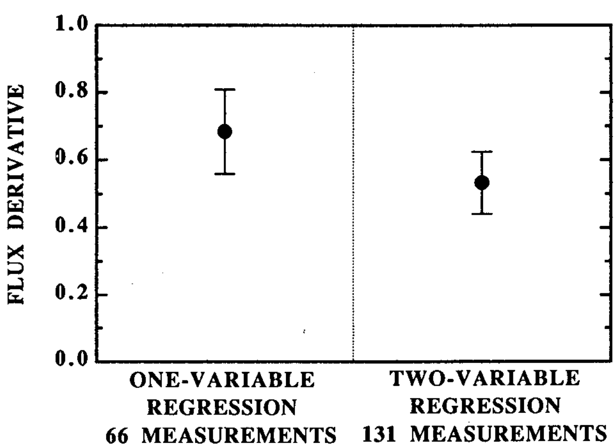

FIG. 6. Summary of flux derivatives for the South Pole as determined from one-variable and two-variable regressions. The vertical bar denotes the 95% confidence interval.

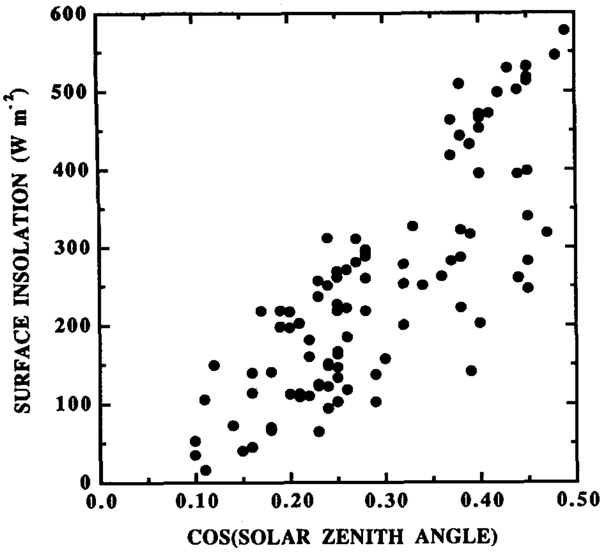

FIG. 7. Measurements of surface insolation at Saskatoon as a function of cosine of the solar zenith angle.

(Fig. 8) so that clouds again increase the TOA albedo over snow at this location. And again the processsed ERBE data show the reverse. Also shown is a similarly derived slope using collocated *ERBS* and Boulder Atmospheric Observatory tower data. These refer to a seven-month dataset (April 1986 through September 1986 and July 1987) as reported by Cess et al. (1993); they are for a snow-free surface for which the mean albedo is 18%. That the magnitude of the negative slope, and thus cloud-induced brightening, is greatest at Boulder is because of the darker surface. But this does not explain the difference between Saskatoon and the South Pole, since Saskatoon has the darker surface (Fig. 1). As will be discussed in the following section, this is probably attributable to differences in cloud type.

The Saskatoon flux derivative is likewise similar to that at the South Pole (Fig. 9), so here also the SW radiative impact of clouds acts to cool the atmospheric column. In contrast, the snow-free Boulder data in-

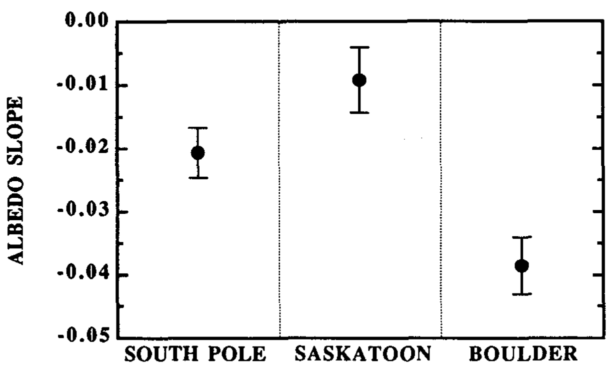

FIG. 8. Summary of albedo slopes for the South Pole, Saskatoon, and Boulder. The vertical bar denotes the 95% confidence interval.

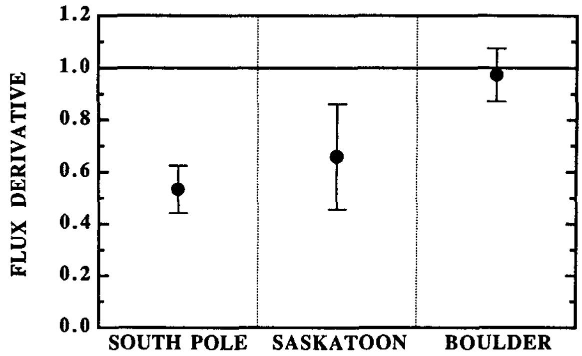

FIG. 9. Summary of flux derivatives for the South Pole, Saskatoon, and Boulder. The vertical bar denotes the 95% confidence interval.

dicate, in the context of a seven-month mean, that the SW radiative cloud impact is neutral.

5. Discussion of results

As previously discussed, theoretical model studies (Shine et al. 1984; S. Warren 1992, personal communication) indicate that the clouds increase the TOA albedo over snow-covered surfaces, as is consistent with the present observational study. To amplify this point, as well as to enhance our understanding of the observational results, we have utilized atmospheric SW radiation model results as described by Li et al. (1993b). The model incorporates 108 wavelength intervals from 0.285 to 2.5 μm and includes SW absorption by water vapor, ozone, and oxygen. We employ their results for a subarctic summer atmosphere with 1.6 cm of precipitable water, stratus, and stratocumulus cloud optical properties from Stephens (1978), and a surface albedo model for fresh snow that results in quite large surface albedo values (Fig. 10). Here the variability of the model's surface albedo is caused solely by its dependence on solar zenith angle. Because the issue of whether or not clouds might decrease the TOA albedo

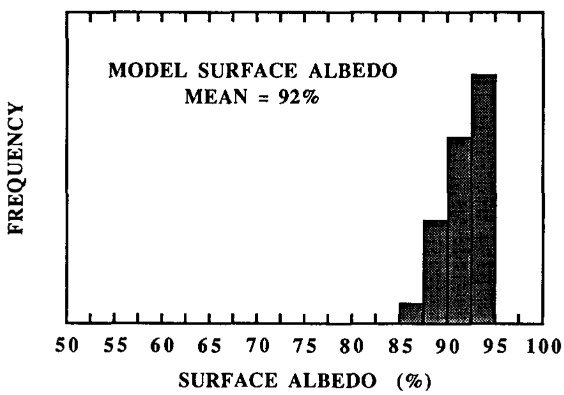

FIG. 10. Histogram of surface albedo from the atmospheric SW radiation model of Li et al. (1993b).

depends on the existence of a highly reflective surface, that shown in Fig. 10 provides a rather extreme model test.

The model predicts that clouds do in fact increase the TOA albedo as is demonstrated in Fig. 11. Here the progression from right to left corresponds to respective cloud optical depths of 0 (clear skies), 5, 10, 20, and 40. Note that these slopes are comparable to those for the South Pole and Saskatoon (Fig. 8). Note also that the model suggests a dependence on cloud type that probably is the cause of the previously discussed difference between the South Pole and Saskatoon slopes (Fig. 8). The stratus cloud model, relative to the stratocumulus model, is at a higher altitude and contains smaller drop sizes.

The μ = 0.303 and 0.500 model results shown in Fig. 11 reveal an interesting feature; the TOA albedo is positively correlated with μ (i.e., the albedo decreases with increasing solar zenith angle) irrespective of cloud optical depth. This is easily understood for clear skies. Because of the increase in atmospheric slant path with increasing solar zenith angle, absorption of shortwave radiation by atmospheric water vapor produces a decrease in albedo with increasing solar zenith angle (limb darkening), while atmospheric scattering produces limb brightening. The former is supressed over dark surfaces, such as ocean or vegetation, for which limb brightening prevails, while over bright snow the reverse

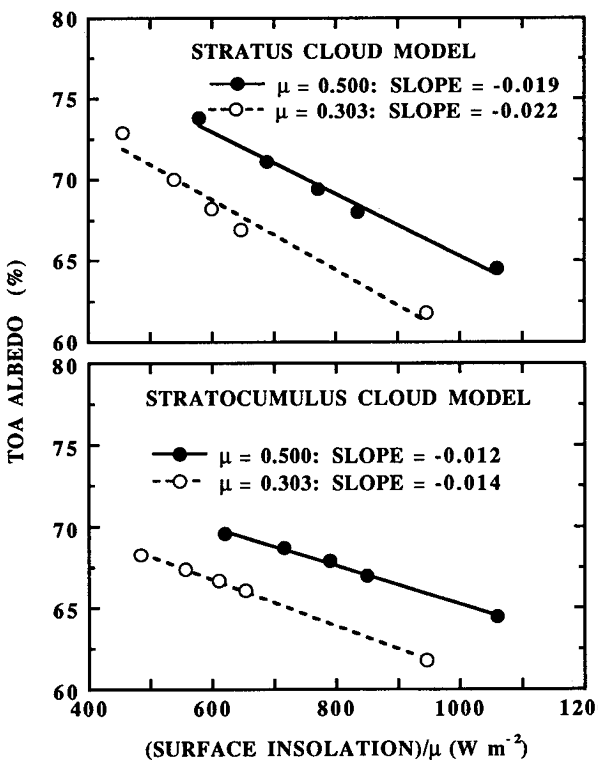

FIG. 11. TOA albedo as a function of surface insolation divided by μ from the atmospheric SW radiation model of Li et al. (1993b).

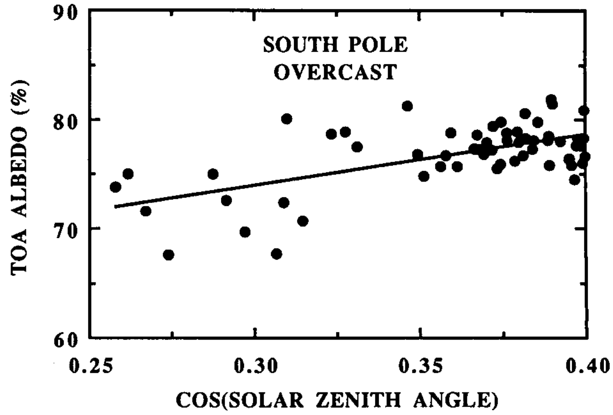

FIG. 12. Measurements of TOA albedo as a function of cos (solar zenith angle) for the South Pole under overcast conditions.

occurs so that there is limb darkening. This is consistent with the clear-sky results in Fig. 11 (the points on the right).

What is surprising is that limb darkening persists as cloud optical depth is increased to 40 (points on the left), because conventional wisdom dictates that scattering by a thick cloud will produce limb brightening irrespective of the underlying surface. The South Pole data, for which the surface albedo is reasonably representative of that used in the model (see Figs. 1 and 10), also indicate that limb darkening occurs for overcast conditions. To demonstrate this, the data have been screened to retain only those data for which clouds reduce the surface insolation by more than 30 W m^{-2} relative to the clear-sky upper envelope in Fig. 2. A positive correlation between overcast albedo and μ is clearly demonstrated in Fig. 12.

6. Concluding remarks

By collocating ERBE SW pixel measurements with near-surface SW measurements we have demonstrated that the addition of clouds to the atmospheric column increases the TOA albedo over two snow-covered locations: the South Pole and Saskatoon. That the ERBE data, by themselves, indicate otherwise appears to be the result of deficiencies in the ERBE scene identification procedure over snow-covered surfaces, with clear scenes being identified as overcast, and vice versa. The collocated data also demonstrate that the SW radiative impact of clouds on the atmospheric column is that of cooling.

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