Cloud scattering optical depth and local surface albedo in the Antarctic: Simultaneous retrieval using ground-based radiometry

Paul Ricchiazzi and Catherine Gautier
Earth Space Research Group, Institute for Computational Earth System Science
University of California, Santa Barbara

Dan Lubin
California Space Institute, University of California, San Diego

Abstract. We have used solar irradiance measurements from a ground-based multichannel radiometer system deployed at Palmer Station, Antarctica (64°46'S, 64°04'W), during spring 1991 to simultaneously estimate cloud scattering optical depth and surface albedo. Irradiance measurements at 410 and 630 nm, in conjunction with a discrete ordinate radiative transfer (RT) model, enable this simultaneous retrieval by exploiting the wavelength dependence in Rayleigh scattering strength. The RT model is used in an inverse mode to find the values of surface albedo and cloud optical depth that match calculated and measured irradiances at both wavelengths. Under the homogeneous stratiform cloud cover for which the technique applies, surface albedo at 630 nm was consistently retrieved at above 0.9. For most homogeneous, overcast conditions, cloud optical depth (at 630 nm) is found to be in the range 20–50, with a most probable value of 25. This measurement and retrieval technique should be useful for compiling high-latitude cloud opacity and surface albedo climatologies of interest for global change and photobiology research.

1. Introduction

At high latitudes the surface radiation budget and the planetary albedo gradient are established by both the presence of cloud cover and the extent of sea and glacial ice. An accurate knowledge of both cloud and surface optical properties is fundamentally important for many high-latitude climate and global change studies, including the nature of the polar greenhouse effect [Tsay et al., 1989], the persistence of sea ice [Crane and Barry, 1984], the energy budget in the upper water column [Perovich and Maykut, 1990], and the ecological impact of springtime ozone depletion [Lubin et al., 1992; Smith et al., 1992]. Under cloudy skies, the high sensitivity in shortwave irradiance to large values of the surface albedo is well known and has been studied in detail by Gardiner [1987] for the Antarctic and by Leontyeva and Stammes [1993] for the Arctic. In satellite imagery of the polar regions, clouds are often difficult to distinguish from underlying high-albedo surfaces due to a combination of small shortwave radiance contrasts between the two, similar infrared brightness temperatures, and exaggerated bidirectional reflectance effects at low sun elevations. The task of detecting and classifying cloud and surface types in polar satellite imagery, without attempting to retrieve optical properties, requires multispectral techniques [Yamanouchi and Kawaguchi, 1992], or complex pattern recognition algorithms [Ebert, 1987; Welch et al., 1992]. Ground-based measurements can therefore play an important role in refining cloud and radiation budget studies at high latitudes.

The multiple reflection of photons between a cloud base and a snow or ice surface can vary the shortwave surface irradiance significantly, often doubling the downwelling irradiance relative to an open ocean surface [Gardiner, 1987], yet many radiative transfer analyses of Antarctic irradiance data use a fixed estimate of the surface albedo [e.g., Lubin and Frederick, 1991; Stammes et al., 1988]. Where the surface albedo is known a priori, the cloud scattering optical depth may be retrieved by comparing a detailed radiative transfer calculation to an irradiance measurement at a single visible wavelength. If the surface albedo is not precisely known (and many radiation monitoring programs in both the Arctic and Antarctic currently lack a specific measurement of albedo), then one must attempt to simultaneously retrieve this quantity along with the cloud scattering optical depth.

A technique to determine the regional albedo based on measurements of solar radiation intensity and polarization in a single spectral band has been described by Shiobara et al. [1987], for ground data obtained at Syowa Station, Antarctica. However, because it requires clear-sky measurements, their method may not be appropriate to provide the regional albedo under clouds for some ground site locations. For example, at a coastal site, where the surfaces which contribute to the regional albedo are heterogeneous, one would expect the regional albedo to depend on the cloud optical depth and cloud base height. Furthermore, the persistence of cloud cover in the maritime Antarctic may severely limit opportunities to make clear-sky observations.

The retrieval method presented here simultaneously retrieves both the cloud optical depth and regional surface albedo from measurements of downwelling surface irradiance. The scattering efficiency of cloud droplets has a negligible
spectral dependence for wavelengths below 1 μm [Goody and Yung, 1989], but there is a noticeable wavelength dependence in both Rayleigh scattering efficiency and the albedo of various snow and ice surfaces throughout the visible and ultraviolet. Our method takes advantage of these wavelength dependencies.

2. Measurements

This study is based on data obtained as part of an atmospheric science campaign carried out at Palmer Station, Antarctica (64°46'S, 64°04'W), during late 1991. The primary instrument for this study is a ground-based radiometer package designed for the validation of satellite radiation budget algorithms [Lubin et al., 1994]. One channel of this instrument measures surface irradiance in the channel 1 band of the advanced very high resolution radiometer (AVHRR) aboard the NOAA polar orbiters. This spectral interval is 580–680 nm (630 nm center). Adequate duplication of the AVHRR instrument response function [Kidwell, 1991] is accomplished by convolving the wavelength response of a Hamamatsu S-1336-8BK silicon photodiode with those of a Schott OG-570 filter and a Corion complete infrared suppressor. This filter/detector combination was assembled under a teflon irradiance cosine collector [Morrow et al., 1994] and packaged in a weatherproof and temperature-stabilized housing by Biospherical Instruments, Inc. (BSI), a San Diego company which engineers the National Science Foundation (NSF) Antarctic UV radiation monitoring network [Lubin et al., 1992; Booth et al., 1994].

This instrument package, fabricated by BSI as a "ground-truthing radiometer GTR 100," also includes BSI's standard filter/detector combination for a narrow-bandwidth irradiance measurement at 410 nm. From August through December 1991 the GTR 100 radiometer system was integrated into the NSF UV-monitor data acquisition system at Palmer Station [Lubin and Frederick, 1991], and irradiance measurements were recorded automatically several times per minute. Preseason absolute radiometric calibration of the GTR radiometer was performed by BSI. Postseason absolute radiometric calibration was performed by both BSI and us at the Santa Barbara Research Corporation (SBRC). Our postseason calibration agreed with BSI's preseason check to within 1% for the 630-nm channel. SBRC also provided a postseason measurement of the 630-nm channel ("AVHRR") spectral response, which showed excellent agreement with the design specifications. For this study, ancillary data include surface irradiance measurements made at 350 nm by the NSF scanning spectroradiometer at Palmer Station, and an all-sky record of cloud cover provided by both video and still cameras.

3. Radiative Transfer Modeling

Our goal is to use the radiation measurements mentioned above together with detailed radiative transfer calculations to deduce the atmospheric parameters which most strongly influence the shortwave surface irradiation budget. The analysis requires accurate radiative transfer predictions of the surface irradiance and depends on the following key ingredients: (1) good estimates of the solar flux incident at the top of the atmosphere within the 410- and 630-nm sensor passbands, (2) specification of parameters which determine the nature of atmospheric scattering and attenuation, due to aerosols, clouds, and atmospheric gases, (3) an accurate and robust numerical technique to solve the equations of radiative transfer in a nonconservative scattering atmosphere.

3.1. Solar Input

The radiative transfer (RT) calculations were performed at 1-nm intervals within the narrow 410-nm sensor pass band and at 5-nm intervals for the 630-nm sensor. We used the top-of-atmosphere solar spectra from LOWTRAN version 7 [Kneizys et al., 1983]. In our analysis we made corrections for the slow ±3.4% variation of the solar “constant” due to seasonal variations in the Earth-Sun distance. Satellite measurements show that fluctuations of solar output on shorter timescales are about a few hundreds of a percent in any day and about a few tenths of a percent over a given season [Frohlich, 1987]. These results imply highly stable solar input at the top of the atmosphere within the 630-nm sensor passband.

3.2. Rayleigh Scattering

In terms of the wavelength λ, the Rayleigh scattering coefficient σ is given by [Liou, 1980]

$$\sigma = \frac{8 \pi^2 (m^2 - 1)^2 (6 + 3 \delta)}{3 \lambda^4 N_0 (6 - 7 \delta)}$$

where m is the index of refraction of air, N is the number density of molecules, and δ is the depolarization factor. Since the index of refraction varies with wavelength, the wavelength dependence of the scattering coefficient is slightly different from the simple and familiar λ^-4 power law. Using results from the theory of dispersion of electromagnetic waves to relate m to N and using a depolarization factor of 0.0279, we write the Rayleigh optical depth as [Shevtze et al., 1980]

$$\tau(\lambda) = C_0 \int \frac{N}{N_0} dz$$

where

$$C_0 = \frac{1}{\lambda^4 (938.07 - 10.8426/\lambda^4)}$$

λ is the wavelength (μm), N_0 is the molecular density at the surface (cm^-3), and z is the altitude of the atmospheric layer (km). This equation for the Rayleigh scattering optical depth is also used in LOWTRAN 7. The total Rayleigh optical depth of the model atmosphere used in this study is 0.325 and 0.056 at 410 and 630 nm, respectively. Rayleigh scattering produces strong polarization effects which can significantly impact radiance predictions in the short wavelength region of the solar spectrum [Adams and Kattawar, 1970]. However, Kattawar [1990] has shown that irradiance predictions using a scalar radiative transfer equation are identical to those using a full vector treatment of the light polarization. Since our concern here is with the analysis of surface flux measurements, we adopt the simpler scalar approach.

3.3. Cloud Radiative Properties

The meteorology of the maritime Antarctic is characterized by frequently overcast conditions, usually due to low-level stratiform clouds [Lubin and Frederick, 1991]. The microphysical properties of Antarctic stratus clouds have been measured at a colder and more southerly location (near McMurdo Station, 75°S) by Saxena and Ruggiero [1990]. Their results indicate that cloud droplets consist mostly of liquid water and that
the drop size distribution is bimodal with a significant fraction of the droplets having very small radii, less than 2 μm. The effective radius $R_{\text{eff}}$ (the ratio of the third to second moments of the particle radius distribution function) of droplets in the second peak of the distribution is between 6 and 9 μm, which is closer to the typical values found for stratus clouds off the Australian coast (31°S–37°S) as reported by Pullenridge [1974].

With these considerations in mind and in order to narrow the scope of our presentation, our model clouds will consist of a single scattering layer at 1-km altitude with cloud droplets composed strictly of liquid water. Given the more temperate conditions at Anvers Island compared to the location of Saxena and Ruggiero's measurements, we will not consider the effects of a population of very small cloud droplets. Instead we use a lognormal drop size distribution with an effective radius of 8 μm and a sigma of 0.35 [Nakajima and King, 1990]. We have also performed a series of calculations with other values of cloud base height and cloud drop effective radius. These calculations, which will not be presented here, show that the retrieved values of surface albedo and cloud optical depth are insensitive to these parameters.

Mie scattering theory is used to derive the scattering properties of the cloud droplets. The Mie results are parameterized in terms of the scattering efficiency $Q_{\text{scat}}$, the single scattering albedo $\omega$, the probability that a extinction event scatters rather than absorbs a photon, and the asymmetry factor $g$, which indicates the strength of forward scattering. In the visible part of the solar spectrum the values of $Q_{\text{scat}}$, $\omega$, and $g$ are very weak functions of wavelength for $R_{\text{eff}} > 4 \mu m$. Hence the cloud radiative properties and, in particular, the cloud optical depth are essentially the same for both the 410- and 630-nm bands. The distribution of the scattered radiation is specified by the standard Henyey-Greenstein phase function [Henyey and Greenstein, 1941; Goody and Yung, 1989]. It has been shown that this phase function provides good accuracy when applied to radiative flux calculations [van de Hulst, 1968; Hansen, 1969].

3.4. Absorption Due to Atmospheric Gases

We base our estimates of gaseous absorption on the transmission models included in LOWTRAN 7. These models specify the transmission of all important atmospheric absorbers at a wavenumber interval of 5 cm$^{-1}$. They are based on fits to detailed line-by-line calculations and laboratory measurements that have been degraded to 20 cm$^{-1}$ resolution for assimilation into LOWTRAN.

In the present application, absorption by atmospheric gases is not a critical issue because the 410-nm band is essentially free of gaseous absorption, while the 630-nm band undergoes only weak absorption by ozone and water vapor. In our standard set of retrievals we fix the water vapor path (WVP) at 0.825 g/cm$^2$. This value corresponds to the average of a number of radiosonde profiles (TIGR data set [Chedin et al., 1985]) obtained between 60°S and 70°S, and also selected to match the surface temperature and relative humidity (RH) of local noon of the clearest day in the measurement period, day 262 ($T = 3^\circ$C and RH = 50%). Sensitivity to this assumption is explored in a later section. All radiative transfer calculations were performed with an ozone column depth of 200 Dobson units. During the time period under study the total ozone amount over Palmer Station, as measured by the total ozone mapping spectrometer (TOMS) on board Nimbus 7, actually changed by more than a factor of 2, varying between 145 and 307 Dobson units. However, since the ozone absorption occurs well above the scattering region (assumed to consist of low lying stratus clouds), the model calculations can be easily corrected for ozone variability by multiplying the model result by an ozone transmission factor. Based on LOWTRAN’s Chapman band absorption model, the corrected surface irradiance predictions in the 630-nm band was found to be

$$I_o = I_{200} \exp \left(-0.00007753 \frac{[O_3] - 200}{\cos(SZA)} \right)$$

where $[O_3]$ is the ozone column depth in Dobson units, SZA is the solar zenith angle, and $I_{200}$ is the predicted 630-nm surface irradiance assuming 200 Dobson units of ozone. The correction factor was computed once for each day based on the local noon TOMS measurement.

3.5. Radiative Transfer Code

The RT calculations were performed with the discrete ordinate (DISORT) radiative transfer code of Stamnes et al. [1988]. The gaseous absorptions by ozone and water vapor were incorporated into DISORT by assuming that the extinction optical depth is given by the negative natural log of the LOWTRAN 7 transmission function. This is not strictly correct for water vapor absorption in the 630-nm band because the transmission function does not follow the Lambert-Bouguer-Beer law. However, for a water vapor path of 1 g/cm$^2$ and a solar zenith angle of 60°, the downwelling irradiance in the 630-nm band is only attenuated 1.5% by water vapor absorption. Hence using the transmission functions in this way should introduce a very small uncertainty.

Because DISORT assumes a plane parallel geometry, we have limited considerations to solar zenith angles less than 75°, for which spherical geometry effects are negligible. Limiting the zenith angle conditions in this way also minimizes the uncertainties in our estimates of aerosol and gaseous absorption.

All computations were carried out with a standard subarctic-summer atmospheric model [McClatchey et al., 1972] for lack of atmospheric observations. The water vapor density of each layer of this model atmosphere was multiplied by the same constant factor to obtain a prescribed total water vapor path. The DISORT runs carried four radiation streams and used 33 vertical atmospheric layers. The surface reflectance was assumed to be Lambertian.

4. Analysis of Surface Irradiance Data

The first step in the analysis is to use DISORT to build a lookup table of predicted surface irradiance in the 410- and 630-nm bands. The lookup table includes entries for cloud optical depth, ranging from zero to 60, surface albedo, ranging from zero to 1, and solar zenith angle, ranging from 5° to 80°.

Figures 1a and 1b show the predicted surface irradiance at a solar zenith angle of 65° for the 410- and 630-nm sensors as a function of cloud optical depth. The greater spread in 410-nm irradiance at small optical depths, compared with the 630-nm band, is due to increased Rayleigh backscatter of the reflected radiation. At greater cloud optical depths the impact of Rayleigh scattering is minimized, and the shape of the two sets of curves becomes more similar.

Figure 2 displays the same information, but as a contour plot for the two channels (the solid contour lines are for the 630-nm
Figure 1. DISORT model calculations of downwelling surface irradiance in the (a) 630-nm band and (b) 410-nm band for a solar zenith angle of 65°. The RT calculations were carried out for cloud optical depths between zero and 60 and for surface albedoes between zero and 1.

Figure 2. DISORT model calculations of downwelling surface irradiance of the 630-nm (solid) and 410-nm (dashed) bands for a solar zenith angle of 65°.

Figure 3. Surface albedo in the 405–415 and 570–710 nm bands measured for a range of surface compositions characteristic of the maritime Arctic and Antarctic.
4.1. Normalization

Over most of the parameter space depicted in Figure 2 the two sets of contour curves intersect at fairly shallow angles, particularly for large cloud optical depths. This implies that the retrieved values of albedo and optical depth are sensitive to the measured surface irradiance and hence to the sensor calibration. Unfortunately, a radiometer calibration much better than about 5% is often very difficult to achieve. In addition, there are aspects of the radiative transfer calculation that are not treated sufficiently well to achieve good agreement with the measurements, for example, insufficient spatial or angular resolution. Since the retrieval method depends on consistency between the measured and computed values, it is necessary to normalize the DISORT calculations with the GTR. The normalization factors (one factor for each channel) were chosen to minimize, on a cloudless day, the deviation of the measured values from the DISORT clear-sky predictions. The model calculations were performed using maximal values of surface albedo from Figure 3, which are 0.96 and 0.95 for 410 and 630 nm, respectively. Hence, in making this normalization, we are assuming a value of the surface albedo on a given clear day. The values of retrieved optical depth and surface albedo for other days will depend on this assumed value. In a later section we assess the sensitivity of our results to this assumption.

Day 262 was the most cloud-free day in the time sequence, judged by both the radiometer traces and an all-sky video record. Figures 4a and 4b show the radiometer traces for the 630- and 410-nm sensors for day 262 compared with the DISORT clear-sky predictions for surface irradiance, assuming WVP = 0.825 g/cm². To achieve this level of agreement, the 410- and 630-nm calibration constants were multiplied by normalization factors of 0.967 and 0.957, respectively. Computing these factors, we excluded the 1700–1900 time interval because the radiometer traces indicate a slightly increased variability during this time span, possibly due to thin clouds. Based on how well the GTR measurements follow the model calculations, we estimate that these normalization constants have error bars of about 2%. The good agreement in the overall shape of both the 410- and 630-nm traces compared with the clear-sky calculated values can be taken as a measure of how well the GTR-100s teflon light diffuser simulates an "ideal cosine" angular response. This agrees with the angular response calibration performed by the manufacturer [Morrow et al., 1994], which indicates deviations from ideal cosine response no greater than a few percent.

Figures 5a and 5b show the same comparison for day 283, the second clearest day in the data set. On this day the computed normalization factors for the 410- and 630-nm channels were 1.017 and 0.984. The DISORT model parameters used in this comparison are the same as used for day 262. The slight increase of the 410- and 630-nm normalization factors implies a decrease in atmospheric transmission of 5.2% and 2.8% in the 410- and 630-nm channels compared with day 262. This is roughly consistent with the existence of a thin persistent cloud layer (τ = 0.4) on day 283. This contention is supported by the NOAA 11 AVHRR images for day 283, which show a band of scattered thin clouds extending across the middle of Anvers Island at UTC 1939.

Note that the normalization factors are within the ±5% tolerance expected for typically attained calibration accuracies. However, it comes at the price of assuming an average surface albedo that seems high for Palmer Station's coastal location. If one assumes a surface albedo substantially less than the pure snow value, the normalization factors differ from unity significantly. For example, if one assumed an average surface albedo of 0.50 in the vicinity of Palmer Station, the normalization factors for the 410- and 630-nm channels must be reduced to 0.859 and 0.937 for day 262. It would then be difficult to reconcile the measurements with the DISORT predictions and the predeployment and postdeployment calibrations, particularly for the 410-nm sensor. We have no on-site determinations of the aerosol optical depth during the period of our observations. However, the clear-sky measurements on day 262 can at least place bounds on the importance of aerosols on predicted surface irradiance. Figure 6 shows the theoretical versus measured irradiance comparison for day 262 assuming an aerosol visibility of 30 km. The spectral dependence of the aerosol model is based on the LOWTRAN 7 maritime model for a relative humidity of 50%. This comparison yields normalization factors of 0.905 and 0.885 for the 410- and 630-nm sensors, respectively. In this case our faith in the sensor calibrations forces us to assume a much larger aerosol visibility, so that the aerosol attenuation has little effect on the surface irradiance. Even if the aerosol optical depth were nonzero, the normalization factors should partially compensate for neglect of their
4.2. Optical Depth and Surface Albedo on Selected Days

In this section we present results for our optical depth and surface albedo retrievals. The results are for a set of days that are representative of the range of radiometer trace variability observed. Under the assumption that variability in time is symptomatic of spatial inhomogeneity, these days should be fairly representative of the range of cloud homogeneity during the period.

Figure 7 shows the radiometer traces and retrieved values of optical depth and surface albedo for several days in the observational run. The displayed values of surface albedo strictly apply to the spectral range of the 630-nm sensor. The albedo in the 410-nm band is related to it by the (piecewise) linear relation shown in Figure 3, which depends on surface composition mix, as discussed above. In the figure the retrievals for "white" and "blue" surface conditions are shown by solid and dashed lines, respectively. The gray areas in each plot indicate the excursions from the nominal "white" curve, which result from an estimated ±2% uncertainty in the sensor normalization factors. The diamond symbols in the surface albedo plots indicate the retrieved values of surface albedo using a separate radiometric measurement at 350 nm and assuming the retrieved optical depth is correct. The 350-nm values were obtained at hourly intervals from the NSF UV spectrometer [Lubin and Frederick, 1991]. As for the 410- and 630-nm channels, the 350-nm measurement was scaled by a normalization factor that was obtained by assuming a "pure snow" surface albedo of 0.96 on day 262. The error bars on each diamond symbol indicate the propagated effect of a 2% uncertainty in the 350-nm normalization factor.

On days that show smooth radiometer traces, for example, days 263, 284, and 286, the surface albedos are very nearly constant over the day and in the mid-90% range. These results are consistent with both the 350-nm measurements and the assumed surface albedo used to set the day 262 normalization factors. Over these 3 days, the optical depth varies between 5 and 50, if "white" conditions are assumed. The 2% uncertainty in the normalization factors has very little effect on the computed surface albedo, but a very large impact on optical depth. Compared with albedo, optical depth shows much greater sensitivity to whether white or blue surface conditions are assumed. However, that difference is small in comparison with the error bars that result from the 2% uncertainty in the normalization factors.

On days that show erratic radiometer traces, such as days 245, 253, and 279, the surface albedos vary more widely but still are in the 80-95% range. Evidently, the erratically varying radiometer traces are indications of inhomogeneous clouds for
Figure 7a. (Top) Measured downwelling irradiance at 630 nm (dot-dashed line) and 410 nm (dashed line) for day 263 compared with top-of-atmosphere (TOA) irradiance in the 410-nm (solid line) and 630-nm bands (dotted line). (Middle) Retrieved values of cloud optical depth assuming the "white" (solid line) and "blue" (dash-dotted line) surface conditions from Figure 3. The grey area indicates the propagated error that results from assuming a 2% uncertainty in the normalization factors discussed in section 4. (Bottom) Retrieved values of surface albedo in the 630-nm band assuming white (solid line) and "blue" (dash-dotted line) surface conditions from Figure 3. The grey area indicates the propagated error that results from assuming a 2% uncertainty in the normalization factors. Also shown is the computed surface albedo derived from a separate measurement at 350 nm (diamonds). The 350-nm estimate assumes the cloud optical depth for white conditions indicated in the middle panel.

which the DISORT plane parallel model fails. The technique also has difficulty retrieving surface albedo when the cloud attenuation is small, for example, days 262, 266, and 283. This is due to the insensitivity of the irradiance to the surface albedo at small optical depths and possibly due to the increased influence of a few small clouds in the radiometer field of view. In these cases, even though the surface albedos are unreliable, the retrieved values of optical depth may be fairly accurate as long as plane parallel conditions apply.

4.3. Sensitivity to Assumptions

The results presented so far are based on the assumption of a high surface albedo, on day 262. (Henceforth we will use the notation \( A_{\text{262}} \) to refer to the assumed surface albedo in the 630-nm channel on day 262.) To test how the retrieved optical depths and surface albedos depend on this hypothesis, the

Figure 7b. Same as Figure 7a but for day 284.

Figure 7c. Same as Figure 7a but for day 286.
Figure 7d. Same as Figure 7a but for day 245.

Figure 7f. Same as Figure 7a but for day 279.

Figure 7e. Same as Figure 7a but for day 253.

Figure 7g. Same as Figure 7a but for day 262.
Figure 7h. Same as Figure 7a but for day 266.

Figure 7i. Same as Figure 7a but for day 283.

Figure 8a. Same as Figure 7a but for day 284 and assuming a surface albedo of 0.95 for day 262.

analysis was repeated with new assumed values of $A_{262}$. These new values affect the results by changing the normalizations for the 350-, 410-, and 630-nm calibration factors. The retrieved values for day 284 are shown in Figure 8 for $A_{262}$ set to 0.95, 0.80, and 0.50. For these three frames the surface albedo in the 410-nm band is related to the 630-nm values by the white surface condition function of Figure 3. The computed value of albedo for the $A_{262} = 0.8$ case is very close to the nominal values. It is interesting to note that even in the case where $A_{262}$ is set to 0.5, the retrieved albedo values on day 284 still approach the 80% level, though in this case the computed values are not nearly as uniform in time as for the nominal case.

We have repeated this test for day 263 (which also shows smooth values of retrieved albedo) and have found very similar results. Assuming the surface albedo did not change appreciably between day 262 and 263, this gives further support for large values of $A_{262}$. Another piece of evidence is provided by 350-nm results, which show noticeably worse agreement as $A_{262}$ is decreased from 0.8 to 0.5.

To help illustrate the global properties of the retrieval algorithm, we created a two-dimensional grid of irradiance values for the 410- and 630-nm channels and computed the retrieved surface albedo at each point in the grid. Figure 9a shows a contour plot of computed surface albedo for a solar zenith angle of 65° and the nominal value $A_{262} (=0.95)$. The solid dots indicate the irradiance values (including normalization factors) measured on day 263 at times when the solar zenith was 65 ± 1°, which corresponds to local noon ±20 min. Notice that the irradiance measurements are aligned with the con-
tours of constant surface albedo, which is consistent with the results of Figure 7, that the computed albedo is nearly constant over the day.

The stability of the technique is illustrated in Figure 9b. This figure shows a contour plot of the rate of change of the retrieved surface albedo due to variations in $A_{263}$. The derivative was computed numerically by making small perturbations in $A_{263}$ about its nominal value of 0.95. The measured values cluster in a region where a reduction of $A_{263}$ by 0.1 results in a decrease of only 0.01 in the retrieved albedo on day 263. Hence, Figures 9a and 9b show that the nominal retrievals are both self-consistent and relatively insensitive to the conditions postulated for day 262.

A contour plot of the computed optical depth over the grid is shown in Figure 9c. In this case the contours of constant cloud optical depth are aligned normal to the cluster of measurements. Together, Figure 9a and 9c suggest that the scatter in the measured irradiance values is due only to variations in cloud optical depth.

For comparison, we show in Figures 10a–10c the same quantities as in Figure 9, but for an assumed day 262 albedo of 0.5. In this case the measured values of surface albedo cluster near 0.8 for day 263. This represents a dramatic increase in surface albedo between the two days and is difficult to explain because no snowfall was recorded at Palmer Station in this period. Furthermore, the cluster of measurements is not aligned with the contours of constant albedo. This variation in albedo is hard to explain given that it occurs within just 40 min and for a negligible change in solar zenith angle.

4.4. Retrievals Under Homogeneous Conditions

We would like to use our algorithm to obtain cloud and surface properties for all days in our data set. However, because our technique breaks down under conditions of fluctuating cloud cover, we are forced to limit the investigation to a

Figure 8b. Same as Figure 8a but assuming a surface albedo of 0.80 for day 284.

Figure 8c. Same as Figure 8a but assuming a surface albedo of 0.50 for day 284.

Figure 9a. Retrieved surface albedo as a function of the 630-nm irradiance and the ratio of the 410- and 630-nm irradiances for a solar zenith angle of 65° and assuming the nominal value of the day 262 surface albedo. The solid dots indicate the 410- and 630-nm irradiances measured for day 263 during the time interval when the solar zenith was 65° ± 1°.
subset of measurements taken during times of low irradiance variability. We will assume this subset represents times in which homogeneous cloud cover is present, though it is easy to think of situations where this would not be the case, e.g., orographic clouds. Measurements included in this set are those for which the instantaneous value of the 630-nm atmospheric transmission is within 5% of the hourly average and for which no more than 1% change is observed per 5-min interval.

Figures 11a and 11b show the optical depth and surface albedo for all measurements that satisfy this homogeneity criterion. It appears that the greater variability in the retrieved values of surface albedo before day 258 is probably due to the amplification of aerosol and WVP variations at low sun angles. Considering only the range of days with smaller fluctuations, it appears that the technique does not detect any significant change in surface albedo from day 263 through 287.

Some investigators have noted a tendency for the effective surface albedo in the visible to vary with the optical depth, presumably due to a varying mix of direct versus diffuse radiation [Wiscombe and Warren, 1980]. As Figure 12 shows, our analysis indicates a very slight reduction of effective surface albedo with optical depth. Unfortunately, the requirement of temporal stability seems to select against cases with optical depths less than 20. For optical depths greater than 20 the diffuseness of the radiation field is near its saturated value, making this effect hard to detect from our results. This probably also explains why there is no discernible dependence of the derived surface albedo on solar zenith angle (Figure 13).

This bias is illustrated in Figure 14, which shows a frequency distribution of optical depth derived from the complete set of measurements (dotted line) and for the “homogeneous” subset (solid line). In this figure, the distribution of optical depth for the full data set is computed assuming a surface albedo of 0.8 and using only the 630-nm channel measurements and calculations. Of course, this “full set” distribution is probably incorrect because it includes cases for which the plane parallel assumption is invalid. However, this comparison does point out that the subset of measurements for which our technique is useful may not be representative of the surface radiation bud-
Figure 10c. Same as Figure 9c but for an assumed surface albedo of 0.5 for day 262.

get climatology. It also underscores the need for modeling tools that do not rely on the plane parallel assumption.

5. Summary and Conclusions

In this paper we have determined cloud and surface radiative properties over the Antarctic from multispectral surface measurements. This has been achieved by means of a radiative transfer model, the discrete ordinate (DISORT) model of Stamnes et al. [1988]. Whereas the surface observations made with a set of custom-designed instruments were calibrated as well as possible (before and after deployment), reconciliation of model computations and surface observations in clear conditions required an additional adjustment ("normalization") of the surface observations by about 5%. While this value is larger than one would like, it is not totally unexpected, since it accounts for both remaining instrumental calibration uncertainty and model uncertainties.

The original goal of the study was to assess the radiative properties of antarctic clouds, but the largest difficulty encountered was the uncertainty in the surface albedo. Since surface albedo and cloud effects are strongly coupled over highly reflecting surfaces and because of multiple reflections between cloud base and the surface, we had to devise a scheme aimed at uncoupling the cloud and surface processes. Spectral transmission measurements in wavelength intervals insensitive to water vapor absorption (so that the only important physical variables that affect transmission are cloud and aerosol optical depth and effective surface albedo) have been used to that effect. The transmission difference (at small cloud optical depths) for two different wavelengths is due to variations in Rayleigh back-scattering of the reflected radiation. The transmission difference thus allows us to use these two irradiance measurements, together with model calculations, to solve for the cloud optical depth and surface albedo simultaneously. The scheme developed is based on measurements at 410 nm and over the wavelength interval of the AVHRR channel 1 sensor. The simultaneous solutions can be rather unstable because of the strong coupling between the surface and the clouds, particularly for large optical depth.

The albedo estimated in this study is higher than most reported albedo observations over Antarctica, and is consistent
with the surface albedo of fresh snow. Although we have shown that albedos ranging from 50% to 90% are possible solutions to our set of simultaneous set of equations, the use of additional observations in different wavelength regions suggests that the higher values are more consistent with these observations. Hence this raises the question of what is the regional spectral albedo of the surface in the area where we performed our study. Though we do not have independent measurements of the regional surface albedo, we did note that the surface condition in the vicinity of the station was essentially uniform throughout the period from day 230 to 290. Unbroken sea ice surrounded the station, and the coastal region (below the glacier) was almost completely blanketed with snow. To address this issue more quantitatively, we performed another set of spectral and broadband measurements at Palmer Station in fall 1994. The analysis of these measurements is under way.

Our uncertainty analysis indicates that the retrieval of cloud optical depth is highly sensitive to the "normalization" factor. Optical thickness differences of a factor of 3-4 were obtained on some days for an uncertainty of ±2% on the "normalization" factor. This high sensitivity of optical thickness to calibration (or normalization) error is rather disturbing, considering the fact that it is extremely difficult to calibrate surface instruments to better than ±5% uncertainty.

The ranges of optical thickness and surface albedo retrieved are reasonable, albeit a slightly larger than expected value for the surface albedo, but cover a large range of possible values, when accounting for observation uncertainty. The advantage of our retrieval method is that it relies on a number of spectral observations. While it only uses two spectral observations simultaneously, it verifies the results with other spectral observations, thus ensuring consistency with all available measurements.

This retrieval technique is probably not appropriate to make precise measurements of cloud optical depth, or to assess small variations of albedo such as might result from aging snow. However, even coarse sensitivity is useful. The regional albedo in the vicinity of our ground site is strongly affected by the distribution of sea ice off the southwest coast of Anvers Island. Due to the strong winds that are common to this region, the sea ice distribution around the station may change drastically from day to day and, consequently, may cause a factor of 2 variation of the regional albedo on the same short timescale. Our retrieval technique should be sensitive enough to detect such gross changes in regional albedo. This capability is particularly useful when continuous monitoring of the local distribution of sea ice is not available, for example, at an automated data collection site.

The measurements presented and discussed above are the first in a series of measurements to be made at Palmer Station to study cloud and surface effects on the surface radiation budget. This region of the globe has not yet been studied extensively despite its major importance on climate processes. While it is expected that satellite observations will be the primary source of radiation and cloud information over the Antarctic for climate studies, these data need to be supplemented...
by surface observations if they are to be properly interpreted. Deriving cloud properties over a highly reflective background is an extremely difficult task that can only be performed satisfactorily when we are able to characterize surface reflectance variations. This is an objective we hope to achieve in the near future.

Acknowledgments. This work was supported by the National Science Foundation, contract OPP9317120, and by the Department of Energy under grant DE-FC02-91ER61062. We are grateful to Santa Barbara Research Center for help with the postseason instrument calibration. Also, we are pleased to acknowledge Yang Shiren for his contributions to the development of our radiative transfer model.

References


Kattawar, G. W., Irradiance invariance for scattering according to a Rayleigh phase function compared to a Rayleigh phase matrix for a plane-parallel medium, Appl. Opt., 29, 2365–2367, 1990.


C. Gautier and P. Ricchiazzi, Earth Space Research Group, University of California, CRSEO Ellison Hall, Santa Barbara, CA 93106.

D. Lubin, California Space Institute, University of California, San Diego, MC A-021, La Jolla, CA 92039.

(Received May 31, 1994; revised April 11, 1995; accepted April 11, 1995.)