SOLAR AND INFRARED RADIATION CALCULATIONS FOR THE ANTARCTIC PLATEAU USING A SPECTRALLY-DETAILED SNOW REFLECTANCE MODEL

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The radiation balance at and above a snow-covered surface is affected not only by the high general level of the surface albedo, but also by the strong spectral variation of that albedo. Furthermore, the fact that clouds and snow have highly correlated optical properties means that the radiative problem of clouds over a snowy surface will be a particularly difficult one.

Both for its own intrinsic interest, and because it is in many ways as clean an example of snow-atmosphere radiative interactions as one is likely to find, we have chosen to make very spectrally detailed model calculations of solar and longwave radiation over Antarctica. These are done using an atmospheric radiation model of Wiscombe coupled to the recent snow reflectivity model of Wiscombe and Warren (1980). Typical clear and cloudy situations are studied. Radiative fluxes are examined particularly at the surface and the top of the atmosphere since these are the locations of most past and future measurements. The folly of deducing all-wave radiative properties from measurements in restricted spectral regions is made apparent by these calculations.

Because the snow of the Antarctic plateau is very clean, falls throughout the summer, and never melts, we can make accurate calculations of the surface albedo using our model for fine-grained snow. Fig. 1 shows the spectral albedo at the surface for snow grain size $r=100\mu m$ and solar zenith angle $\theta_0=84°$, typical of October at 90°S.

The calculations of radiation fluxes and heating rates in the atmosphere use Wiscombe's model with 37 vertical levels and 291 spectral intervals: 100 in the infrared ($5 \leq \lambda \leq 500\mu m$) and 191 in the

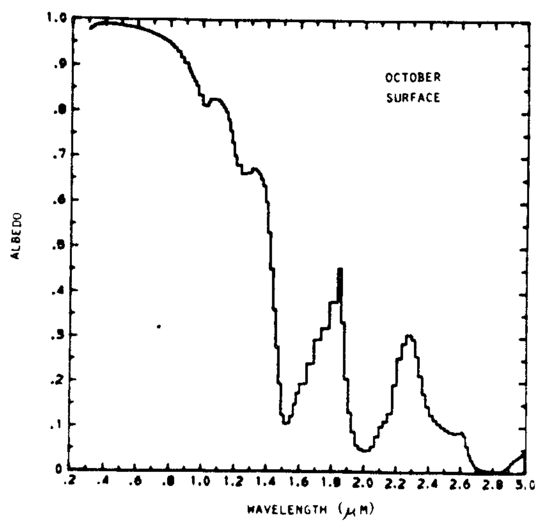

Fig. 1. Surface spectral albedo calculated for October at the South Pole. Snow grain radius $r=100\mu m$, clear sky, solar zenith angle $\theta_0=84.3°$.

solar spectrum ($0.2 \leq \lambda \leq 5\,\mu m$), designed to resolve all spectral variation in snow albedo, in solar flux, in terrestrial Planck function and in Rayleigh scattering. For the transmission of all molecular absorbers it uses the LOWTRAN3 program. The adding-doubling method is used in every spectral interval, executed as many times as there are terms in the exponential-sum fit to the transmission in that interval.

The clear-sky planetary albedo is shown in Fig. 2 for October (θ_o=84°) and January (θ_o=66°) at the South Pole, assuming the troposphere is saturated with respect to ice. The planetary albedo is reduced relative to the surface albedo due to absorption by atmospheric ozone (at $\lambda < 0.3\,\mu m$, $\lambda \approx 0.6\,\mu m$ and $\lambda \approx 0.75\,\mu m$) and water vapor (at $\lambda \approx 0.95\,\mu m$, $\lambda \approx 1.15\,\mu m$ and $\lambda \approx 1.4\,\mu m$). The planetary albedo is very slightly higher than the surface albedo only at $0.35 \leq \lambda \leq 0.43\,\mu m$, due to Rayleigh scattering. The reduction of albedo in the O_3 and H_2O absorption bands is more pronounced for October (low sun, long slant path) than for January (midsummer).

The surface radiation budget (spectral net flux) for January at the South Pole (clear sky, saturated troposphere) is shown in Fig. 3a. The visible radiation ($\lambda < 0.7\,\mu m$) is nearly all reflected, leaving a net flux of zero for these wavelengths. The absorbed solar radiation is almost entirely in the near-IR, where snow albedo is considerably lower than its visible values. The infrared radiation loss is centered on the terrestrial Planck function, narrowed by water-vapor vibration and rotation bands, and reduced at λ=15 μm by CO_2 absorption.

Fig. 3b shows the radiation budget at the top of the clear January atmosphere. The peaks of absorption in the solar spectrum are due to ozone at λ=0.3 μm and λ=0.6 μm, to oxygen at λ=0.75 μm, and to water vapor and snow at the longer wavelengths.

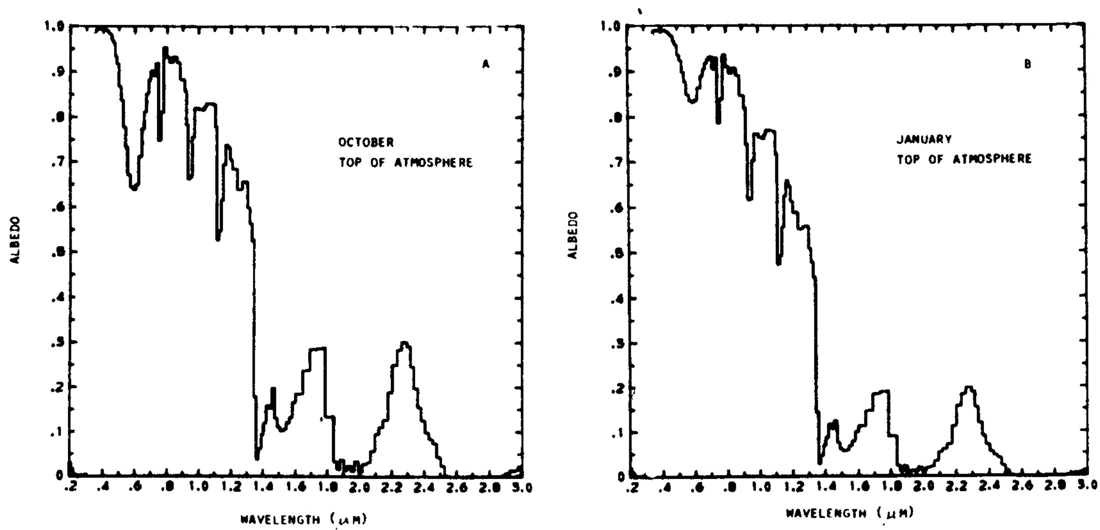

Fig. 2. Planetary spectral albedo calculated for South Pole. r=100 μm, clear sky, troposphere saturated with water vapor. (A) October temperature profile; zenith angle θ_o=84.3°. (B) Jan. temps.; θ_o=66.4°.

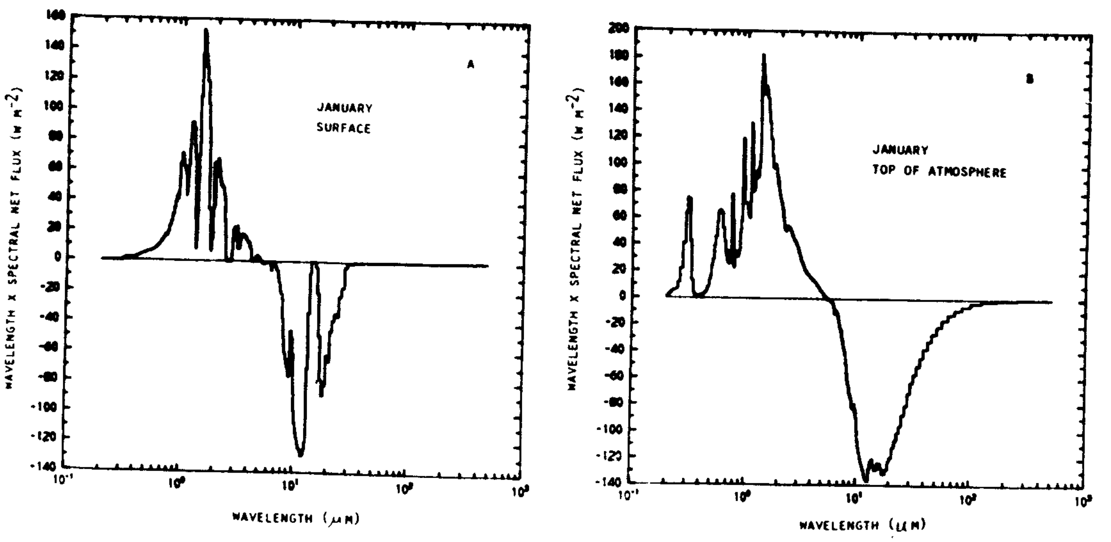

Fig. 3. Spectral net flux (multiplied by wavelength so that areas can be compared when plotted on this log-wavelength scale) for January at South Pole. Snow grain radius r=100 μm, clear sky, troposphere saturated, January temperature profile. (A) surface. (B) top of atmosphere.

Spectrally-integrated albedos are strongly affected by varying the snow grain size and weakly affected by varying the solar zenith angle or tropospheric humidity. For a January South-Pole temperature profile, with solstice sun angle and saturated troposphere, the total surface albedo is 85%, 82%, or 79% for snow grain radius r=50, 100, or 200 μm respectively. The planetary albedo is in each case 10% lower. Lowering the sun from zenith angle 66° to 84° enhances the spectrally-integrated surface albedo by 2% but reduces the planetary albedo by 2% (because of increased absorption in the longer slant path).

Two extreme tropospheric humidity conditions (saturated; or stratospheric mixing ratio) yielded net fluxes at the top of the atmosphere of -22 W m^{-2} or -44 W m^{-2}, respectively, for January clear sky at the South Pole. The spectrally-integrated surface albedo is higher by 2% under the saturated troposphere because water vapor absorbs in the same spectral regions as does snow.

The radiational effects of low, optically-thin cirrus clouds are investigated. Data reported from lidar backscatter, pyrheliometers, and ice particle replicators at the South Pole provide information about cloud height, optical thickness and particle size distribution. Our calculations can be compared with narrow-band and broad-band radiation fluxes measured at the surface and from satellites.

Reference

Wiscombe, W.J., and S.G. Warren, 1980: A model for the spectral albedo of snow, I. Pure snow. Submitted to J. Atmos. Sci.