THE SPECTRAL ALBEDO OF SNOW:

670-80

EFFECTS OF AGE, DEPTH, CLOUD COVER, SUN ANGLE, AND IMPURITIES

by

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This is an abstract of a two-part paper we have recently submitted for publication (Wiscombe and Warren, 1980; Warren and Wiscombe, 1980). We present a method for calculating the spectral albedo of snow which can be used at any wavelength in the solar spectrum and which accounts for diffusely or directly incident radiation at any zenith angle. For deep snow, the model contains only one adjustable parameter, an effective grain size, which is close to observed grain sizes. A second parameter, the liquid-equivalent depth, is required only for relatively thin snow.

In order for the model to make realistic predictions, it must account for the extreme anistropy of scattering by snow particles. This is done by using the "delta-Eddington" approximation for multiple scattering, together with Mie theory for single scattering.

The peaks and valleys of the measured spectral albedo of snow coincide with the minima and maxima, respectively, of the spectral absorption coefficient of ice. Ice is very weakly absorptive in the visible but has strong absorption bands in the near-infrared (NIR), so snow albedo is much lower in the NIR. The NIR solar irradiance thus plays an important role in snowmelt and in the energy balance at a snow surface. Furthermore, while the visible albedo (for pure snow) is rather insensitive to variations in model parameters, the NIR albedo is very sensitive to snow grain size and moderately sensitive to solar zenith angle.

The spectral albedo from 0.3 to 5 μm wavelength is examined as a function of the effective grain size, the solar zenith angle, the snowpack thickness, and the ratio of diffuse to direct solar incidence. The decrease in albedo due to snow aging can be mimicked by reasonable increases in grain size (50-100 μm for new snow, growing to 1 mm for melting old snow).

The albedo is higher at all wavelengths for large zenith angles (low sun). Cloud cover affects albedo both by converting direct into diffuse radiation (and thereby changing the effective zenith angle), and also by altering the spectral distribution of the radiation.

The depth necessary to reach within 1% of the deep-snow albedo for visible light increases with grain size and decreases with impurity content. For old melting pure snow it can be as much as 20 cm liquid-equivalent (50 cm snow).

The model agrees well with observations of O'Brien and Munis (1975) for wavelengths above 0.8 μm (Fig. 1). In the visible and near-UV, on the other hand, the model for pure snow may predict albedos up to 15% higher than those which are actually observed (Fig. 2). The discrepancy is worst for old melting snow. Increased grain size alone cannot lower the model albedo sufficiently to match these observations. The two major effects which are neglected in the model, namely nonsphericity of snow grains and near-field scattering, are judged to be orders of magnitude too small to be responsible for the discrepancy. Insufficient snow depth and error in measured absorption coefficient are also ruled out as the explanation. The remaining hypothesis is that visible snow albedo is reduced by trace amounts of absorptive impurities.

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Small highly-absorbing particles, present in concentrations of only 1 part per million by weight (ppmw) or less, can lower snow albedo in the visible by 5-15% from the high values (96-99%) predicted for pure snow. These particles have, however, no effect on snow albedo beyond 0.9 μm wavelength where ice itself becomes a strong absorber.

Desert dust and carbon soot are the most likely contaminants. They enter snow as ice nuclei, as particles scavenged by falling snow crystals, and by fallout. Dust may be most important in regions near deserts. But careful measurements of spectral snow albedo in the Arctic and Antarctic point to a "grey" absorber, one whose imaginary refractive index is nearly constant across the visible. Thus carbon soot, rather than the red iron oxide normally present in desert dust, is strongly indicated at these sites. Soot is a normal component of the background atmospheric aerosol due to industrial sources as well as forest and brush fires.

Soot particles of radius 0.1 μm, in concentrations of only 0.3 ppmw, can explain the albedo measurements of Grenfell and Maykut (1977) on Arctic Ice Island T-3 (Fig. 2). This amount is consistent with some observations of soot in Arctic air masses. 1.5 ppmw of soot is required to explain the Antarctic observations of Kuhn and Siogas (1978), which seemed an unrealistically large amount for the earth's most unpolluted continent until we learned that burning of camp heating fuel and aircraft exhaust had indeed contaminated the measurement site with soot. The albedo of this site is unrepresentative of the rest of the Antarctic plateau, where measured total impurity concentrations (0.03 ppmw in ice cores) are far too low to affect the albedo. For a given soot concentration, the albedo is reduced more for large-grained snow than for fine-grained snow. This is because the radiation penetrates deeper in more coarsely-grained snow and encounters more absorbing material before it can re-emerge from the snowpack.

Midlatitude snowfields are likely to contain larger absolute amounts of soot and dust than their polar counterparts, but the snowfall is also much larger, so that the ppmw contamination does not differ drastically until melting begins. Nevertheless, the variations in absorbing particle concentration which <u>will</u> exist can help to explain the wide range of visible snow albedos reported in the literature.

In snowfields closer to desert areas and remote from population centers, such as the ice caps of the Tibetan Plateau, the principal absorptive component of the dust may be iron oxide instead of carbon. The snow would then exhibit a peak in the spectral albedo near $\lambda = 0.6$ μm (red-yellow), as has been seen by eye in Europe and New Zealand following dust storms in North Africa and Australia.

Greenland ice cores show higher dust concentrations during the ice age 18000 ybp, probably due to the expansion of world desert area at that time. The dust content of snow may have been high enough to reduce the albedo on southern parts of the North American and Scandinavian ice sheets.

Longwave emissivity of snow is unaltered by its soot and dust content. Thus the depression of snow albedo in the visible is a systematic effect and always results in more energy being absorbed at a snow-covered surface than would be the case for pure snow. Thus man-made carbon soot aerosol may continue to exert a significant warming effect on the earth's climate even after it is removed from the atmosphere.

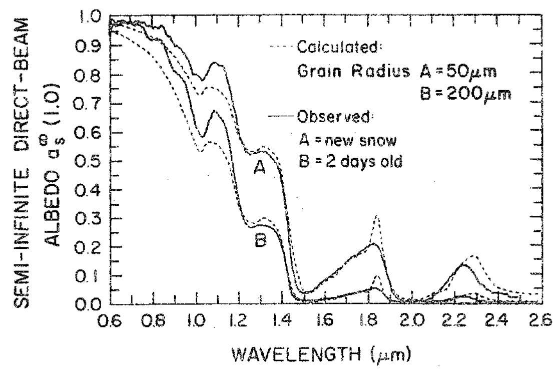

Figure 1. Comparison of calculated spectral albedo with laboratory observations of snow reflectance. Solid lines: Observations of O'Brien and Munis (1975), taken from their Fig. 5 and then corrected before plotting for the reflectance of the $BaSO_4$ standard they used. Source (zenith) angle 0°; detector (nadir) angle 30°. Both samples were from the same snowfall. (a) fresh snow, (b) snow aged naturally for two days with ambient temperatures hovering above and below freezing. Dashed lines: Direct-beam albedo for pure deep snow with overhead sun (zenith angle $\theta = 0°$), for grain radii: (a) r = 50 μm, (b) r = 200 μm.

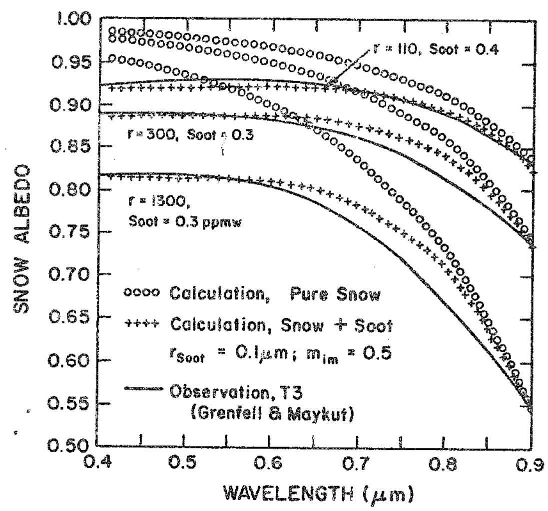

Figure 2. Comparison of calculated snow albedo with observations at visible wavelengths. Solid lines: Measurements of Grenfell and Maykut (1977, Fig. 1) made in summer 1974 at Ice Island T-3 in the Arctic Ocean. In order of decreasing albedo they are: (i) dry, cold snow, windpacked, deep drift $\rho = 0.4$ g/cm^3, (ii) 5 cm wet new snow over multi-year white ice, (iii) old melting snow, 28 cm thick. Circles: Calculated albedo of semi-infinite pure snow for diffuse illumination, with grain radii to match observations at $\lambda = 0.9$ μm. In order of decreasing albedo, they are: (i) r = 110 μm, (ii) r = 300 μm, and (iii) r = 1300 μm. Plus signs: Snow containing the specified concentrations of soot, using for the imaginary index of refraction m_{im} (soot) = 0.5 independent of wavelength. m_{re} = 1.8.

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