Annals of Glaciology 5 1984
© International Glaciological Society

IMPURITIES IN SNOW: EFFECTS ON ALBEDO AND SNOWMELT

(Review)

by

Stephen G. Warren
(Geophysics Program and Department of Atmospheric Sciences, University of Washington, Seattle, Washington 98195, U.S.A.)

ABSTRACT

Very small (ppm) amounts of soil dust in snow can significantly reduce snow albedo and thereby affect the snow-surface energy budget. Ice cores from Greenland show enhanced dust concentrations in ice from the last glacial maximum, in amounts capable of causing measurable effects on snow albedo. This enhanced dust is probably due in part to the expanded desert areas at that time.

Volcanic ash layers visible in the Byrd station core reduced the snow albedo in West Antarctica when they were on the surface. The ash is unlikely to have had a long-term effect on albedo because of the episodic nature of volcanic eruptions.

Very large amounts of dust on snow can inhibit snow-melt by insulating the snow. A debris cover probably slowed the melting of parts of the North American ice sheet during its most recent decay phase.

Snow in the Arctic Ocean is presently suffering large-scale contamination by carbon soot from anthropogenic sources. Preliminary estimates indicate that soot concentrations in Arctic snow are sufficient to reduce snow albedo measurably.

1. THEORETICAL MODEL OF SNOW ALBEDO

In order to explain the reasons for the large observed variability of snow albedo, Wiscombe and Warren (1980) designed a theoretical model of radiative transfer for the scattering and absorption of sunlight in snow. This model is useful for any wavelength in the solar spectrum, 0.2 to 5.0 μm. Snow albedo is very high (over 90%) in the visible and near-ultraviolet, and is much lower in the near-infrared, so most of the light absorbed by snow is near-infrared sunlight.

The normal reduction of snow albedo with age is attributed to the increase in grain size (or decrease in surface-to-volume ratio) that normally occurs with snow metamorphism. Snow albedo is reduced at all wavelengths as the grain size increases. In order to match the observed course of spectral albedo in the near-infrared, effective grain radii in the range 50μm (new snow) to about 1 000 μm (old melting snow) were required in the model.

2. EFFECT OF DUST ON SNOW ALBEDO

Trace amounts of light-absorbing impurities can significantly reduce snow albedo in the visible wavelengths but have no effect on near-infrared albedo or thermal infrared emissivity.

Except in the Antarctic, most observations of visible snow albedo are lower than calculated for pure snow, and can be explained by assuming that the snow contained graphitic carbon "soot" in concentrations of 1 to 2 x 10^{-7} by weight (Warren and Wiscombe 1980). This soot was probably mostly the result of local (camp) pollution, although part of it may be due to widespread pollution from northern hemisphere industrial sources, as discussed below.

The impurities which are most likely to have widespread effects on snow albedo are carbon soot, continental dust, and volcanic ash. Soot is about 50 times more effective than dust, and about 200 times more effective than ash, at reducing snow albedo.

The theoretical effect of soot on snow albedo was confirmed by Grenfell and others (1981). They measured the spectral albedo of natural snow in the Cascade Mountains, Washington, USA, and also collected samples of the snow for soot analysis. The soot content (2 to 6 x 10^{-8} by weight) was roughly in accord with the albedo reduction, as discussed by Warren (1982[b]).

3. ICE-AGE-INDUCED DESERTIFICATION: A NEGATIVE FEEDBACK ON ICE-SHEET GROWTH?

Ice cores from Antarctica and northern Greenland show enhanced concentrations of insoluble continental dust in ice from the last glacial maximum (Thompson 1977, Lorius and others 1984). The enhancement is about a factor of 5 at Byrd station, 15 to 20 at Dome C, and 50 at Camp Century. The concentration of dust may have been 1 to 2 ppm by weight at Camp Century. This is enough to reduce visible snow albedo (at 0.4 μm wavelength) by 1% for new snow or 5% for old melting snow (Warren and Wiscombe 1980: fig.5(a),(c)). (This result was misquoted by Warren (1982[a]).) The effect on spectrally-integrated albedo would be about one-third of these values. We may expect larger quantities of dust at more southerly locations in Greenland and on the North American and Scandinavian ice sheets which were closer to the sources of dust.

The enhanced dust during the last glacial maximum may have been due to both expanded desert areas (Sarnthein 1978) and increased wind strength. The desertification may have been caused by the ice sheets themselves, by means of the associated cooling of ocean surface waters which suppressed the hydro-

logical cycle, as suggested by the general-circulation-modeling results of Manabe and Hahn (1977). This process leads to the well-known positive climatic feedback due to the albedo contrast between deserts and grasslands. However, the fallout of this desert dust in snow also leads to a potential negative feedback similar to one suggested by Kellogg (1975: fig.8). This speculation cannot be quantified until estimates of dust deposition rates on the Laurentide and Scandinavian ice sheets are available (Joussaume and others 1984).

4. VOLCANIC ASH IN SNOW

Volcanic ash layers visible in the Byrd station core (Gow and Williamson 1971) would have reduced the snow albedo in West Antarctica when they were on the surface (cf. Warren 1982[b]: fig.7). It is likely that these ash layers came from Antarctic volcanoes which may have owed their enhanced activity to the greater ice burden on the continent (Kyle and others 1981). This suggests another negative feedback on ice-sheet growth which was proposed by Bloch (1964). However, because of the episodic nature of volcanic eruptions, the ash would soon be covered with new snow, so its effect would be short-lived. Because the snow does not melt over most of Antarctica, the ash remains hidden from sunlight once it is covered by new snow. The effect of volcanic ash on snow albedo is thus unlikely to have long-term climatic significance.

5. VERY LARGE AMOUNTS OF DUST CAN INHIBIT ABLATION

When snow melts, the impurities often tend to collect at the surface rather than washing away with the meltwater. During the decay phase of the northern hemisphere ice sheets, the southernmost portions may thus have been covered with debris. If the debris layer is sufficiently thick, the insulation of the ice becomes more important than the albedo reduction, and ablation can be reduced. The terminus of the Malaspina Glacier in Alaska is one modern example (J Hall personal communication). The surface of such glaciers can even become vegetated. The critical thickness of debris, beyond which ablation is inhibited, has been measured as 5 to 10 mm for "dirt" (Ashwell and Hannell 1966), 20 mm for sand (Higuchi and Nagoshi 1977), and 20 to 30 mm for volcanic ash (Driedger 1981). The competing effects of debris are illustrated by contrasting cryoconite holes with dirt cones and glacier tables.

Layers of debris thicker than 20 mm may not have extended far from the ice front. For example, a 20 mm layer of dust could be formed by the ablation of 2 000 m of ice if the ice contained a dust fraction of about 10^{-5}, i.e. a factor of 10 higher than that found in ice from the last glacial maximum at Camp Century. It is thus doubtful that the delay of ice decay due to protection by dust would cause a noticeable lag in the oxygen isotope record in the ocean sediments. A quantitative assessment awaits estimates of dust deposition rates on the North American and Scandinavian ice sheets.

6. EFFECT OF SOOT POLLUTION ON SNOW ALBEDO

Carbon soot is produced by incomplete combustion. The major sources of soot at present are probably anthropogenic, both from the burning of coal and oil for industry, and from brush fires for agriculture. Natural forest fires can also contribute soot to the atmosphere. Soot is of current interest for its possible climatic effects as the absorptive component of the "Arctic haze" (Rosen and others 1981). This soot may be removed from the atmosphere in snowfall and thus be found in the snow on top of Arctic sea ice. Only one preliminary measurement of soot content in Arctic snow is now available (Rosen personal communication, cited by Warren 1982[b]). Rosen found weight fractions of 1 to 6 x 10^{-8}. A weight fraction of 5 x 10^{-8} is enough to reduce visible snow albedo by 2.5% (new snow) to 7% (old melting snow) and spectrally-integrated albedo by 1% (new) to 2% (old), according to the model of Warren and Wiscombe. As discussed below, the soot may actually be more effective than they calculated. Better estimates of the climatic importance of carbon particles in Arctic snow await the results of the large-scale snow-sampling and soot-measurement program now underway by Antony Clarke at the University of Washington.

7. IMPORTANCE OF THE LOCATION OF ABSORPTIVE IMPURITIES IN SNOW

Impurities can be more effective at reducing snow albedo if they are located inside the ice grains. Warren and Wiscombe (1980) modeled impure snow as an "external mixture", with separated ice and dust particles each surrounded by air. Ackerman and Toon (1981) have shown that the absorption of light is enhanced if the impurity is present as an "internal mixture". They did this for the case of soot particles in sulfuric acid droplets, which they modeled as concentric spheres, but the results also apply to soot in ice (cf. note added in proof to Warren 1982[b]). Chýlek and others (in press) have modeled soot in snow as an internal mixture, with small soot particles (mean radius about 0.1 μm) uniformly distributed through the much larger ice grains. They found that for a given effect on snow albedo, an external mixture requires about twice as much soot as an internal mixture requires.

These results will also apply qualitatively to continental dust: dust particles inside ice will be more absorptive than if they are surrounded by air. The enhanced absorption has not yet been calculated; it will probably not be as dramatic as for soot because of the larger size (radius ~ 1μm) of the dust particles.

ACKNOWLEDGEMENT

My research on optical properties of snow and ice is supported by National Science Foundation grant ATM-82-15337. This is Contribution No.680 from the Department of Atmospheric Sciences, University of Washington.

REFERENCES

Ackerman T P, Toon O B 1981 Absorption of visible radiation in atmosphere containing mixtures of absorbing and nonabsorbing particles. *Applied Optics* 20(20): 3661-3668

Ashwell I Y, Hannell F G 1966 Experiments on a snowpatch in the mountains of Sweden. *Journal of Glaciology* 6(43): 135-144

Bloch M R 1964 Dust-induced albedo changes of polar ice sheets and glacierization. *Journal of Glaciology* 5(38): 241-244

Chýlek P, Ramaswamy V, Srivastava V In press Albedo of soot-contaminated snow. *Journal of Geophysical Research*

Driedger C L 1981 Effect of ash thickness on snow ablation. The 1980 eruptions of Mount St. Helens, Washington. *US Geological Survey. Professional Paper* 1250: 757-760

Gow A J, Williamson T 1971 Volcanic ash in the Antarctic ice sheet and its possible climatic implications. *Earth and Planetary Science Letters* 13(1): 210-218

Grenfell T C, Perovich D K, Ogren J A 1981 Spectral albedos of an alpine snowpack. *Cold Regions Science and Technology* 4(2): 121-127

Higuchi K, Nagoshi A 1977 Effect of particulate matter in surface snow layers on the albedo of perennial snow patches. *International Association of Hydrological Sciences Publication* 118 *(General Assembly of Grenoble 1975 - Isotopes and Impurities in Snow and Ice)*: 95-97

Joussaume S, Rasool I, Sadourny R, Petit J-R 1984 Simulation of desert dust cycles in an atmospheric general circulation model (Abstract). *Annals of Glaciology* 5: 204-207

Kellogg W W 1975 Climatic feedback mechanisms involving the polar regions. *In* Weller G, Bowling S A (*eds*) *Climate of the Arctic*. Fairbanks, AK, University of Alaska. Geophysical Institute: 111-116

Kyle P R, Jezek P A, Mosley-Thompson E, Thompson L G 1981 Tephra layers in the Byrd station ice core and the Dome C ice core, Antarctica and their climatic importance. *Journal of Volcanology and Geothermal Research* 11(1): 29-39

Lorius C, Raynaud D, Petit J-R, Jouzel J, Merlivat L 1984 Late-glacial maximum-Holocene atmospheric and ice-thickness changes from Antarctic ice-core studies. *Annals of Glaciology* 5: 88-94

Manabe S, Hahn D G 1977 Simulation of the tropical climate of an ice age. *Journal of Geophysical Research* 82(27): 3889-3911

Rosen H, Novakov T, Bodhaine B A 1981 Soot in the Arctic. *Atmospheric Environment* 15(8): 1371-1374

Sarnthein M 1978 Sand deserts during glacial maximum and climatic optimum. *Nature* 272(5648): 43-46

Thompson L G 1977 Variations in microparticle concentration, size distribution and elemental composition found in Camp Century, Greenland, and Byrd station, Antarctica, deep ice cores. *International Association of Hydrological Sciences Publication* 118 (*General Assembly of Grenoble 1975 - Isotopes and Impurities in Snow and Ice*): 351-364

Warren S G 1982[a] Ice and climate modeling: an editorial essay. *Climatic Change* 4(4): 329-340

Warren S G 1982[b] Optical properties of snow. *Reviews of Geophysics and Space Physics* 20(1): 67-89

Warren S G, Wiscombe W J 1980 A model for the spectral albedo of snow. II. Snow containing atmospheric aerosols. *Journal of the Atmospheric Sciences* 37(12): 2734-2745

Wiscombe W J, Warren S G 1980 A model for the spectral albedo of snow. I. Pure snow. *Journal of the Atmospheric Sciences* 37(12): 2712-2733