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## *Snowmelt increase through albedo reduction*

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**PREFACE**

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# Snowmelt Increase Through Albedo Reduction

SAMUEL C. COLBECK

## INTRODUCTION

There has been interest in enhancing snowmelt for a long time (e.g. Woeikof 1885). Snow provides the main or the only source of water in many parts of the world, and snow sometimes gets in the way of man's activities and must be removed as quickly as possible. When snow is just in the way, methods that use a lot of expensive energy can be used only when small quantities of snow are involved and snow-free areas have a very high priority. When large quantities of snow must be melted, only natural energy sources can be considered and then relatively few things can be done to increase melt. Air temperature cannot be increased, winds cannot be generated, cloudiness cannot be reduced nor sunlight increased in any meaningful manner. However, snow albedo can be decreased simply and cheaply, and so snowmelt enhancement has concentrated on this one property of snow, its albedo. This has been done to enhance water supply (Bazhev 1975), extend the growing season (Nakamura and Onuma 1966), extend the navigational season in sea ice areas (Fedotov 1971), reduce avalanche hazards (Emel'ianov and Korolev 1969) and hasten river ice breakup (Cook and Wade 1968).

The subject of snow albedo has received a great deal of attention over the last 10 years because of widespread interest in the role of snow in climate. The high reflectivity of snow reduces the amount of solar energy absorbed by the earth, thus suggesting a positive feedback between snow cover and cold weather. This interest has led to a better theoretical understanding of the albedo of snow (e.g. Warren 1982). Snow albedo will be discussed after the properties of the snow surface are described. Since albedo is a sur-

face phenomenon, it is very important to understand the nature of the conditions at the surface.

## SNOW SURFACE CHARACTERISTICS

Many studies of crystal growth in snow have been done by the last two generations of snow scientists (e.g. Colbeck 1987), but relatively little study has been made of the snow surface itself. The reason is simply that the upper surface is subjected to rapidly changing conditions that make it quite complicated. By contrast, the base of a snow cover has a relatively stable temperature and temperature gradient. While the basal layers may remain frozen, the surface can experience brief melting even during periods of subfreezing weather. Large temperature gradients normally occur in response to diurnal changes in solar input and surface temperature.

I have recently looked into several aspects of snow surface conditions including surface hoar (Colbeck 1988), subsurface melt and windpumping, the movement of air through the pores of the snow by wind. Of these processes, the growth of surface hoar is clearly most relevant here. The formation of large crystals with flat surfaces that are strongly aligned (Fig. 1) clearly affects the directional dependence of albedo. The size of the crystals might also affect the albedo, although surface hoar of great size probably forms most often long before the onset of the melt season. Perhaps the critical phenomenon during the melt season is the formation of grain clusters and melt-freeze grains on the surface where their large size is important. While these considerations are all overshadowed by the effect of adding



*Figure 1. Surface hoar, which grows during clear, calm nights at high elevations. These large crystals are generally oriented and give the surface a directional reflectivity.*

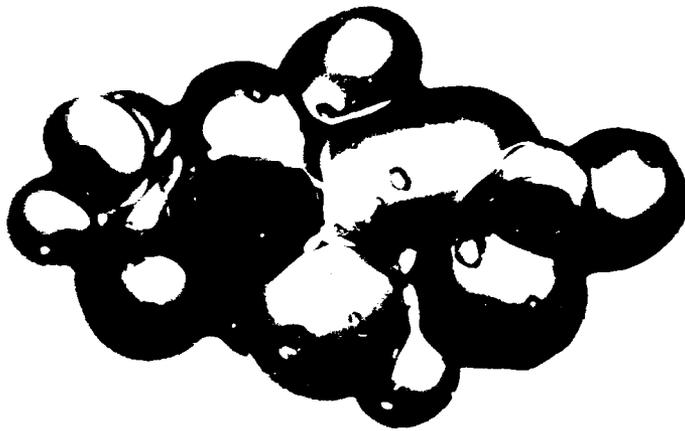
carbon black to a snow surface, the various naturally occurring surfaces are worth considering before looking at the unnatural ones. For one thing, the natural decrease of albedo through aging may be just as important as the increase due to the addition of darker materials.

There are basically two types of snow surfaces—wet and dry. Since only the wet surfaces are melting, I consider only wet surfaces here. Furthermore, I assume that the surface has been wet long enough that small rounded grains such as wind crust (Fig. 2) and special features like surface hoar are not present. The rate at which snow grains grow in the presence of liquid water is quite rapid even at low liquid contents (Wakahama 1968). In general, the albedo of fresh, dry snow is much higher than that of wet snow, even if the wet snow has refrozen. Accordingly, once dry snow starts to melt, its surface turns into large-grained, wet snow, possibly even before the entire snow cover rises to the melting temperature. Since albedo is a surface phenomenon, it is only the surface characteristics that matter here. Dry snow, with its high albedo, may delay the onset of melt, but most snow covers begin melting intensely once the melt season begins.

The surface of a melting snow cover is rarely slush but the surface is not completely free to



*Figure 2. Small, well-rounded crystals, which dominate wind crust. Their small size may affect their radiative properties.*



*Figure 3. Clusters of single crystals in freely draining, wet snow not subjected to melt-freeze cycles. Clustering results in large particles that affect albedo even when the snow is refrozen.*



*Figure 4. Amorphous particle from melt-freeze cycles of grain clusters as shown in Figure 3. These large particles also affect albedo.*

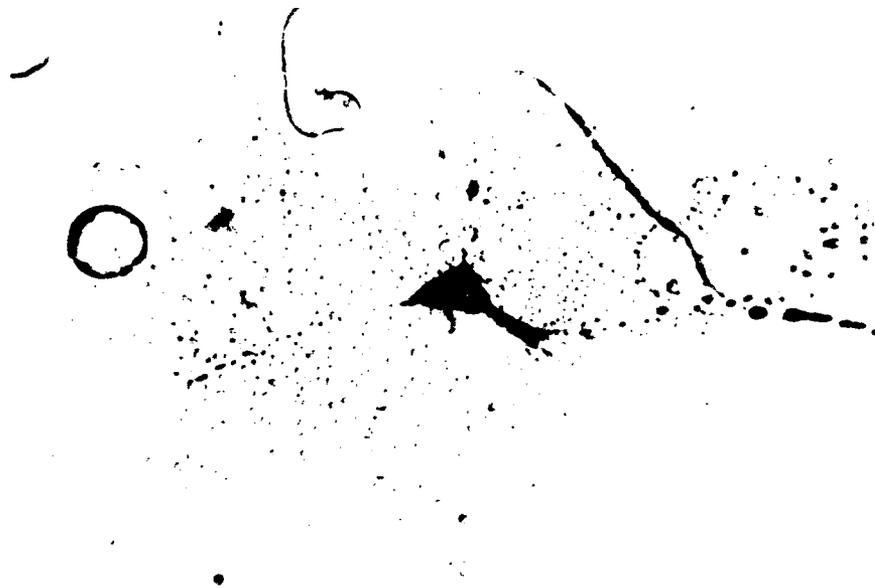
drain. Freely draining snow has a liquid content in the range of 3 to 6% by volume and has a very different grain geometry than slush or dry snow. In the absence of melt-freeze cycles, wet snow forms grain clusters like those shown in Figure 3. These form simply to minimize the surface free-energy and are stable as long as they do not freeze. If they do go through a few melt-freeze cy-

cles, they are altered to the form shown in Figure 4 where the result of repeated melting and re-freezing is apparent in the amorphous form of that large particle. The effect of grain size on the optical properties of snow is important, depending somewhat on the wavelength of the light. At optical wavelengths, the cluster or particle size can be considered the relevant size, although the single crystals of ice that constitute the cluster have liquid water among them and grain-boundary grooves and triple-grain junctions. The index of refraction for water and ice are nearly identical, and hence the fact that the interior of the cluster contains liquid is essentially irrelevant. In this sense there is no difference between the cluster of Figure 3 and the particle of Figure 4, when they are present.

Since these large particles tend to break down in strong radiational fields (Langham 1975), a snow surface subjected to intense sunlight is often loose and slush-like. The ice-to-ice grain bonds that form in grain clusters (Colbeck 1987) are strengthened by freezing. However, solar radiation absorption remelts the veins and even the crystal boundaries so that the clusters sometimes break down into individual crystals as intense melting proceeds. Because the upper surface is a free boundary of a porous medium, it tends to retain liquid by capillarity, as melt proceeds, and release it irregularly. Thus the upper few grains are often slush-like in the sense that the individual crystals are not ice-bonded to their

neighbors and the liquid content is relatively high. Of course this very wet layer occurs only during periods of intense radiation input; at night the surface often refreezes into a hard layer characterized by the grain clusters shown in Figure 3.

One other surface phenomenon in wet snow is worth mentioning, the occurrence of icy layers in



*Figure 5. Icy layer with visible triple-grain junctions. These can form on the surface after freezing-rain storms.*

snow. Icy layers form either by the capillary retention of meltwater by fine-grained layers where the water refreezes or by freezing rain on the snow surface (Fig. 5). While freezing rain may glaze the surface, icy layers are generally formed beneath the surface and reach the surface only after some melt has occurred. In either case, the surface has a markedly different character, even a glassy appearance.

### SNOW MELT PROCESSES

For glacial firn and deeper snow covers, the dominant energy fluxes are all at the upper surface. These fluxes are dominated by radiation and sensible and latent heat fluxes. For shallow snow covers, ground heating must also be considered and input from rain can be important locally. When snow covers are close to snow-free land or water surfaces, heat can be advected horizontally, but normally only the vertical components of heat flux are considered. When horizontal convection from warmer surfaces does occur, the melt rate is significantly increased, so that patchy snow covers might be expected to melt out quickly on their own and eliminate the need for albedo modification.

The turbulent fluxes of sensible and latent heat are less important than solar input on clear days of intense sunshine. Nevertheless, these inputs are often important and, during warm periods of strong, humid winds, the melt rate can be very high. Turbulent fluxes of sensible and latent heat to the snow surface are controlled by factors occurring in the first few meters above the surface and are increased by the intensity of the turbulence and the departures of temperature and humidity from their mean values (Male 1980). That is, the sensible heat transfer would be zero if the air were at  $0^{\circ}\text{C}$  and the latent heat transfer would be zero if the air were at the saturation vapor pressure over an ice surface at  $0^{\circ}\text{C}$ . The actual measurement of these parameters is difficult and is undertaken routinely at only a few places.

Male and Granger (1979) give an example of the sizes of the major fluxes to a melting snow cover (Fig. 6) and show that the flux of sensible heat can be a major factor in determining whether or not runoff occurs when the air is cool. This points out that the total energy balance is determined by a number of terms and that the balance often hangs rather delicately on the interplay of the radiation balance and energy losses due to evaporation, cool air, and conduction into sub-freezing snow. A slight adjustment to one of these

fluxes can easily change the delicate balance in very high alpine areas, where snowmelt is never intense, or in the early part of the melt season when solar input is large but the albedo is also high.

The useful portion of the incoming solar radiation is absorbed by the snow, with the remainder being reflected back toward space. Even on clear days the diffuse and secondary reflection is usually more than 10% of the downward component (Male 1980), but the major consideration is the strength of the solar radiation as it varies with altitude, latitude and season (Fig. 7). At any particular site the slope, orientation and cloudiness are also important. In mountain valleys shadows are important and the forest canopy often affects seasonal snow covers (Dozier 1979). With snow's high albedo, the amount of incoming radiation is often less critical than the amount lost to reflection, simply because the latter is one of the few things we can modify.

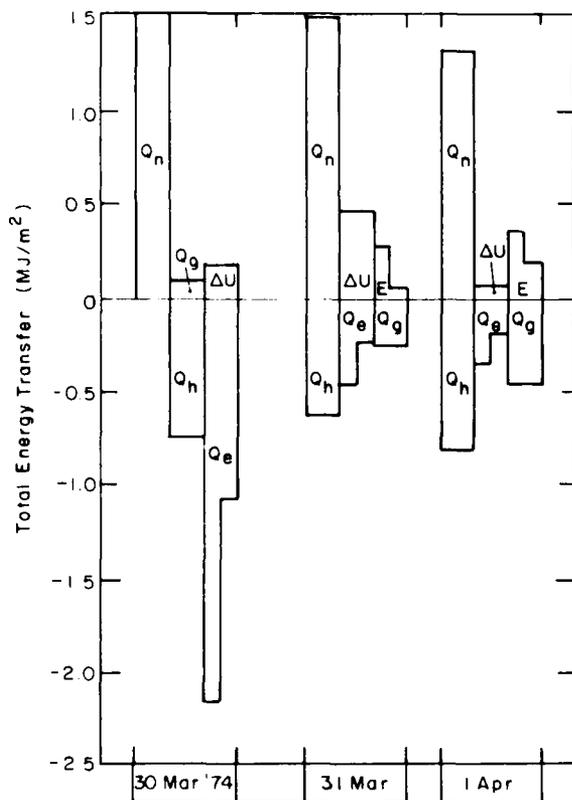


Figure 6. The relative sizes of major energy fluxes at one site (after Male and Granger 1979) with changes.  $Q_n$  is net radiation,  $Q_g$  is soil heat,  $Q_h$  is sensible heat flux,  $Q_e$  is latent heat flux, and  $\Delta U$  is change in internal energy.

## ALBEDO

Albedo is defined as the ratio of reflected to incident shortwave radiation, most of which is in the range of 0.3 to 3.0  $\mu\text{m}$ , the upper part of the ultraviolet through the visible and the near-infrared. The optical properties of snow have long been studied because of snow hydrology (e.g. Bohren and Barkstrom 1974), but more recent interest in climate has driven our theoretical understanding to a higher level (e.g. Warren 1982). Measurements of global albedo are still regularly taken because of climatic interests, as well as interest in such things as snow-covered sea ice (Grenfell and Maykut 1977). All of this work is directed toward understanding the albedo of naturally occurring snow but has application to the understanding of the artificial modification of snow albedo for two reasons. First, the albedo of snow covers decreases as the melt season progresses because of the accumulation of dust and

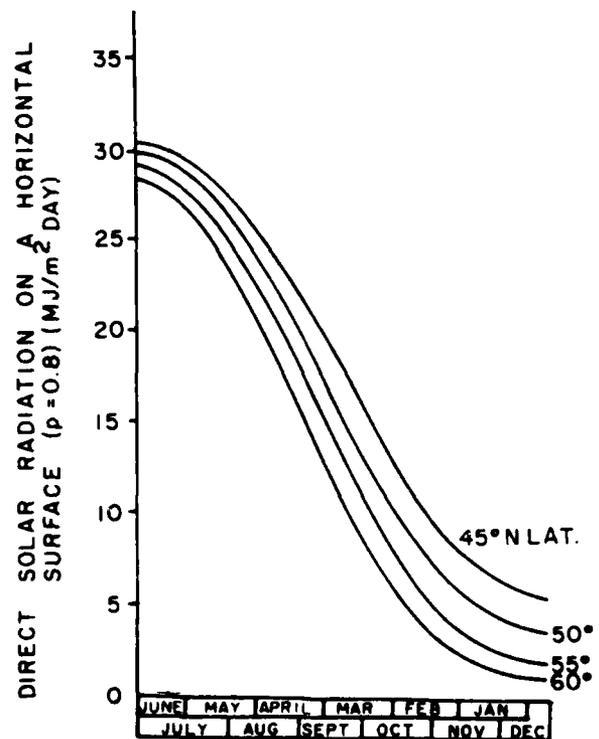


Figure 7. Solar radiation versus latitude and month for a transmissivity of 0.8 (after Male and Granger 1979). This would also vary with altitude and slope.

debris on the surface, compaction, thinning and the increase in grain size associated with metamorphism. Second, any man-made modification will decrease the albedo but not necessarily increase the melt rate since an insulating layer can be formed over the snow that would actually pro-

tect it. Thus anything added to the snow surface is done so at the risk that melting will be decreased below its naturally occurring value if too much material is added.

Berger (1979) found that the scattering coefficients for snow are inversely related to the grain

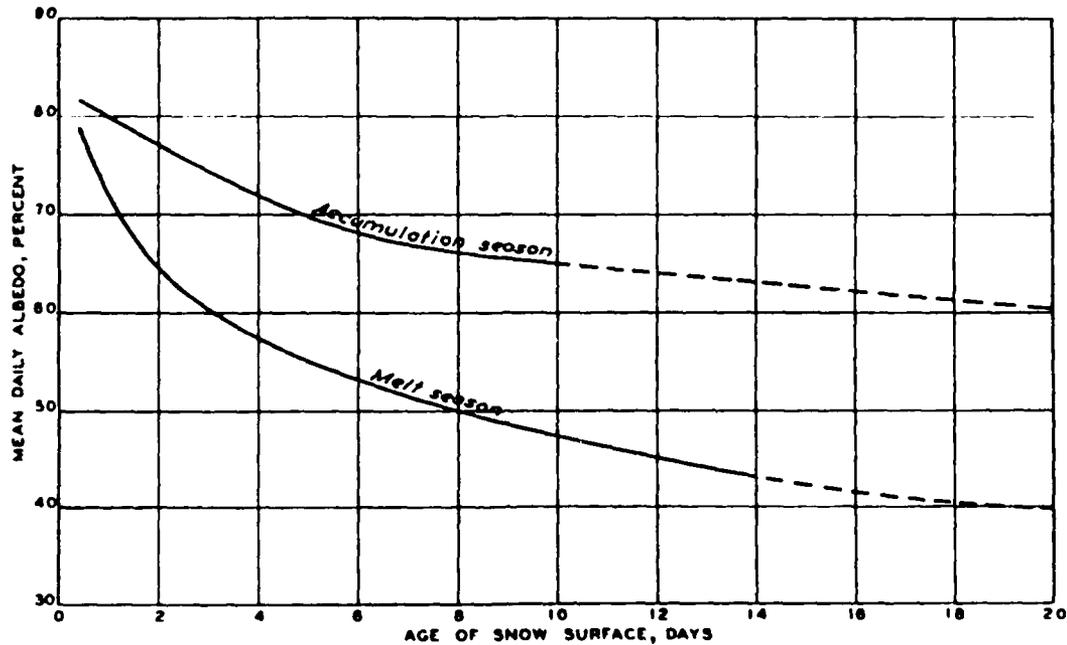


Figure 8. Albedo versus month and snow type (U.S. Army Corps of Engineers 1956). This relation would depend on local effects like dust and temperature.

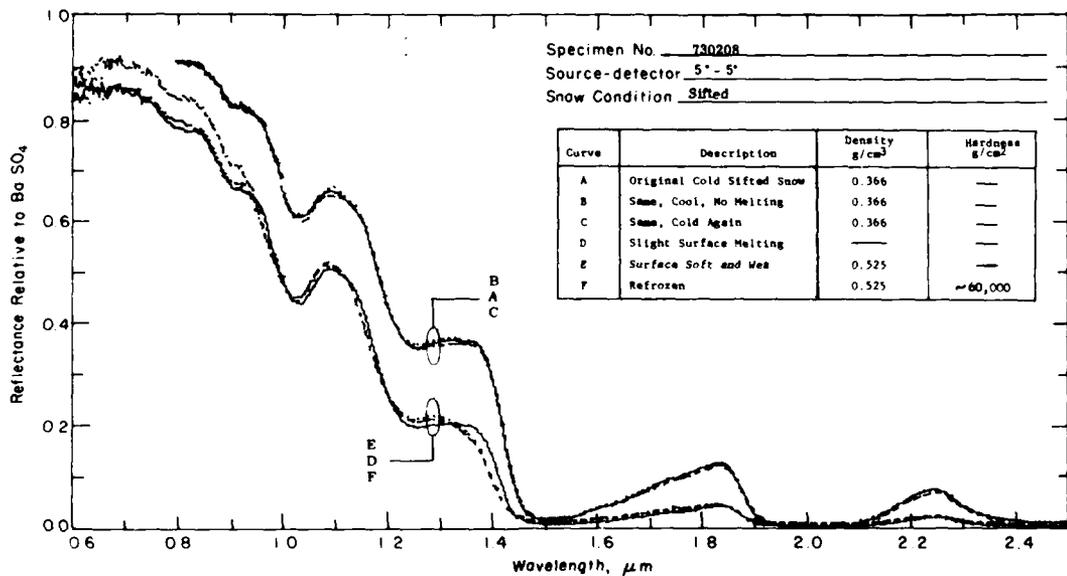


Figure 9. Reflectivity versus wavelength as measured by O'Brien and Munis (1975). The strong spectral dependence is clear.

size but proportional to density for the 2- to 20- $\mu\text{m}$  part of the electromagnetic spectrum. This effect is strong enough from 0.6 to 2.0  $\mu\text{m}$  that Hyvarinen and Lammasniemi (1987) suggested that the grain size and liquid-water content of snow could be measured with infrared instruments. In the visible region (0.4 to 0.75  $\mu\text{m}$ ), the albedo decreases less with age than it does in the near-infrared because the spectral absorption coefficients of ice increase at wavelengths beyond 0.8  $\mu\text{m}$  (O'Brien and Munis 1975). Thus albedo reduction with aging is partly due to increasing grain size, increasing density and increasing surface contamination. In shallow snow covers, the fact that vegetation begins to absorb sunlight also becomes a factor but, for snow over a flat surface, as little as 10 cm of snow has essentially the same albedo as an infinity thick cover (Grenfell and Maykut 1977). Nevertheless, O'Neill and Gray (1973) showed that the albedo of a shallow prairie snow cover decreases much more rapidly and to a lower value than that of a deep mountain snow cover. The U.S. Army Corps of Engineers (1956) found that albedo decreases during both the accumulation period and the melt season, but much more quickly during the melt season (Fig. 8).

Bohren and Barkstrom (1974) suggested that albedo is inversely proportional to the square root of the grain size and that it decreases with increasing wavelength. This effect is shown in the experimental results of O'Brien and Munis (1975) as shown in Figure 9. There are two distinct sets of curves in this figure, one set for snow that has never been wetted and one set for snow that has. Clearly once the snow is wetted and grain growth and/or clustering occur, the albedo decreases. The interesting thing is that it does not increase again when the snow is refrozen because the grain size and/or cluster size remain fixed. Thus it is not the presence of liquid water per se that affects the albedo, but rather the effect that liquid water has on grain growth or clustering as shown in Figure 3. These results were further quantified by Grenfell et al. (1981), as shown in Figure 10, where albedo is shown versus the square root of grain size for various wavelengths of interest here.

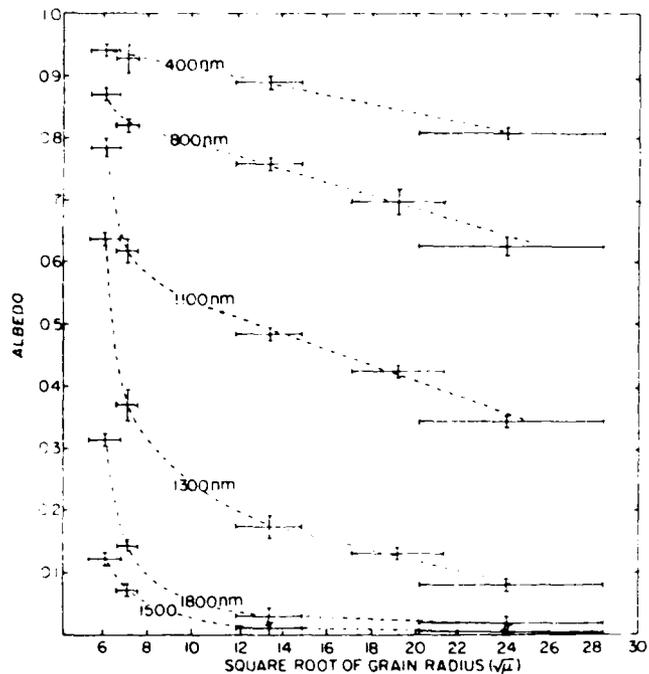


Figure 10. Albedo versus square root of grain size for various spectral wavelengths (after Grenfell et al. 1981).

## SURFACE DUSTING

The dusting of snow to increase the melt rate has long been practiced (e.g. Slaughter 1969); the Soviet literature seems to report this practice most extensively. Natural dusting also occurs and does not generally have the dangers of artificial dusting, including the possibility of forming an insulating layer by overuse. Also, natural dusting usually achieves an even distribution for the maximum effect. Local sources of wind-borne rock dust and soot are important in this regard (Woo and Dubreuil 1985), but dust transported over long distances seems to have the most dramatic effect. Dust from the Sahara Desert is often deposited in the Alps where de Quervain (1947) reported that one major dusting in 1947 caused the snow cover to disappear 1.25 months early. Similar experiences are reported in many parts of the world, depending greatly on prevailing winds and sources of large quantities of mi-

croscopic particles. Even the effect of dust contamination in the Antarctic has been investigated (Lettau 1977).

Small highly absorbent particles can reduce snow albedo from 5 to 15% even in quantities of 1 part per million (Warren and Wiscombe 1980). If desert dust and carbon soot are the most effective sources of these contaminants, the occurrence is highly variable geographically and annually. While volcanoes have episodic effects on snowmelt (Warren 1984), the limiting case of snow dusting is from the fires that would result in "nuclear winter." Warren and Wiscombe (1985) suggest that the effects of reduced albedo from these widespread events would persist for several years.

Slaughter (1969) reviewed the artificial darkening of snow surfaces; more recent work appears mostly in the Russian literature, with some work done elsewhere (e.g. Kaul et al. 1985). Savinov (1907) showed that soot could more than double radiation absorption but the optimum way to do this still is unclear. The dusting of snow is most effective when the snow is fresh and clean, but dusting then incurs the risk of recurring snowstorms that could temporarily bury the layers of dust. In fact very little fresh snow is necessary to eliminate the effect of dust and cumulative layers of dust might retard the melting. Also, dusting may increase the heat loss due to convection and evaporation so that the total effect may be limited by this negative feedback. This retarded melting arises in part because not all of the heat is transferred to the snow (Yosida 1948) and, in fact, the generation of useful heat probably becomes less efficient at lower temperatures where dusting is more highly desired. Unfortunately, dusting is not very effective when the minimum daily air temperature is below freezing, even at lower latitudes (Williams 1967).

If dusting is done, the optimum material depends on considerations like particle size, expense, color, density, spreadability, solubility, and environmental effects. While most of the underlying principles are well known, definitive experiments involving a complete energy balance and meaningful ablation measurements seem to be lacking. Furthermore, information could be transferred to another location only if the theory were well established and the local conditions well known. Nevertheless, it seems clear that finely dispersed, black, small particles are most effective. Hand and Lundquist (1942) found lampblack to be the most effective of the paint pigments, and such finely divided materi-

als seem more effective, in general (Arnold 1961, Skorik 1960, Williams and Gold 1963). In fact, Skorik (1960) found that small quantities of finely divided materials were more effective than large quantities of substances with low dispersivity. While this principle is clear and partly explains the very large reduction in albedo found by Warren and Wiscombe (1980) for very small quantities of soot, again there are no guidelines that one could use to design an application program.

Slaughter (1969) summarized the existing data including some comments on the persistence of the effect of albedo reduction. There seems to be general agreement that the effect decreases with time but the rate appears to vary significantly, depending perhaps on the size of the material used and the surface over which it is spread. Clear ice surfaces seem to either wash clean or concentrate the contaminants in less effective pockets while snow surfaces may retain the effect longer. It would appear that the migration of particles would decrease with increasing size so that the smaller particles, like carbon black, might be considerably more effective initially but the effect may decay more rapidly with time. Certainly experimental evidence is needed to assess the persistence of these materials on snow surfaces.

Persistence can be a problem when these materials accumulate on the surface of snow or ice due to repeated applications, or if debris falls onto the surface (Rogerson et al. 1986). Ashwell and Hannell (1966) found that ablation depended on whether the weather was predominantly overcast with winds or calm and sunny. This distinction makes sense from the viewpoint of the mechanisms involved, either dominantly convective heating or radiative heating, respectively. If solar input is restricted, the optimum thickness of the dark layer is very small, about 1 mm, according to Rhodes et al. (1987) depending on the exact nature of the controlling parameters including the type of material added. For the case of volcanic ash, Driedger (1981) found an optimum thickness of 3 mm under sunny conditions but found that ablation was enhanced as long as the ash layer was less than about 24 mm in thickness. Ostrem (1959) found that a layer of sand and gravel more than 5 mm thick decreased the melt rate of glacier ice. The general nature of this effect is shown in Figure 11, as suggested by Wilson (1958) and measured by Driedger (1981) for one case. The exact shape and size of the curve is dictated by meteorological parameters and im-

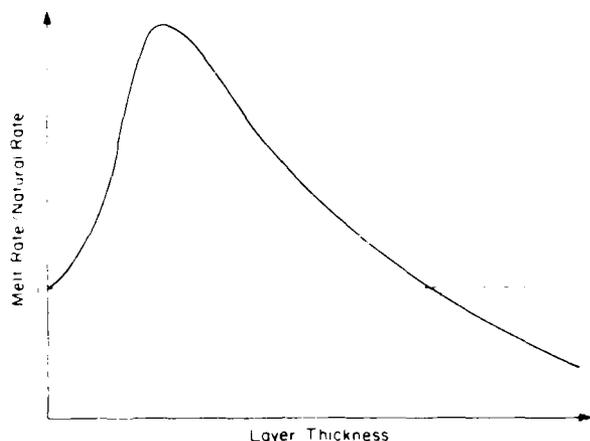


Figure 11. Generalized view of melt rate versus thickness of a dark layer. The optimum thickness increases absorption much more than it increases thermal insulation. Beyond a certain thickness the insulative effects are more important than the increased solar absorption and the melt rate decreases below its natural value.

purity characteristics, so that as the weather changes over a period of time along with the thickness of the dark layer, the integrated effect of the layer is difficult to predict. Finer materials appear to insulate better than coarse materials.

On the large scale, Tangborn and Lettenmaier (1981) looked for the effect on snow melt in moderate to large drainage basins of dusting from the 1981 eruption of Mount St. Helens. Although they reported that large changes were not observed, this negative result may have been due to the generally cloudy weather that prevailed over the region at the time. The results may have been different with sunny conditions and a more substantial snow cover.

Wilson (1953) also described the unstable nature of coated surfaces as roughness features develop along the surface and the heat reaching the ice begins to be affected by the variable thickness of the dark material. One notable effect is that the increased surface roughness might itself increase the turbulent transfer to or from the dark surface during windy periods. This emphasizes the complicated nature of the total energy balance of the system. There are other important considerations as well. If a glacier is dusted, the effect over a long period of time might be to lower the glacier's elevation enough to affect its microclimate and its overall mass bal-

ance. If the surface slope in the lower regions of a glacier is increased, the ice flow would be increased, which may affect the drainage of colder ice from higher elevations and have little effect on the overall glacier if the reservoir of ice is large enough. However, if the glacier is small, it might be sent into an irreversible decline by these processes. It is clear that dusting, like other anthropogenic influences on natural systems, provokes complicated responses that often have poorly understood consequences far beyond those originally intended.

## SUMMARY

Dark materials have been used for many years to enhance snow and ice melting for a variety of applications. There is a distinct lack of guidelines necessary to design such a program although much of the required information is already available. The general nature of the melt increase is shown in Figure 11, but different investigators have placed different values on the thickness of the layer that produces the maximum melt rate and the thickness of the layer at which the melt rate returns to the undisturbed value. The reasons for these differences are clearly due to the variability of weather conditions, physical properties of the dark material, and manner of application. The insulating value of a dark layer is counterproductive during cloudy and windy conditions but may have little effect during sunny conditions. Thus the prevailing weather conditions are very important but may be unpredictable. Fine-grained materials appear to be more effective because they can be finely dispersed, but they may be subject to erosion and early removal.

Since the albedo of snow decreases with age, especially once the snow is wetted, the melt rate increases naturally as the season progresses from winter into summer. This progression can be promoted by reducing the albedo before the air temperature warms to midsummer values but at some risk of having the darkening agent covered by subsequent snow. Repeated applications are both expensive and pose the threat of insulating the snow once they accumulate into one layer during heavy melt. Thus the design of an applications program must take a wide variety of factors into consideration, including weather, micrometeorology, and material characteristics. The optimization of a melt enhancement scheme would take some research under local conditions to ac-

count for insolation and other prevailing weather features. Even a well-designed program may fail in some years unless the local climate is quite predictable.

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