ALBEDO OF BREAK-UP ICE COVERS

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INTRODUCTION

The break-up of river ice covers is usually preceded by a decay or ablation period during which the thickness and structural integrity of the ice sheet decreases. Although data is sparse, even higher rates of ice melt are thought to prevail during the ice clearance period which immediately follows the initial fracturing and fragmentation of the ice sheet. The changes in the overall mass and strength of ice during the pre- and post break-up periods can affect the timing and severity of the break-up process, particularly in regards to ice jamming and associated backwater flooding.

Although the hydrothermal heat transfer and atmospheric transfers of sensible, latent and radiative heat control the thermal ripening and melt of the snow and ice cover, it is radiation which causes the most significant reduction in the internal strength of an ice cover. Unlike the other heat flows, radiation can penetrate beneath the snow or ice surface and advance the warming and eventual melt of the internal ice structure (for example see Bulatov 1972; Ashton 1983). The depth to which solar radiation can penetrate a snow or ice cover is dependent on attenuation within the ice cover and on the surface reflection or albedo. A considerable amount of information exists about the changes in snow and ice albedos during the early and middle stages of melt but few albedo measurements have been conducted in the periods immediately preceding and following break-up. To be able to model the decay process leading to break-up and the overall mass balance of the ice clearing process, more must be known about the radiation budgets throughout these periods. This note defines the typical trends in albedo for the spring melt and break-up periods based on values from the literature and the results of field measurements conducted during the late stages of break-up. The data are being used as part of a larger study dealing with the energy balance of decaying river ice covers.

METHODS AND INSTRUMENTATION

Field measurements of river ice albedo were conducted on the Liard and Mackenzie Rivers within an approximate 20 km radius of Fort Simpson, N.W.T. (61° 53' N 121° 22' W) during the 1984 and 1985 break-up periods. A detailed site description and an analysis of break-up processes at this location are contained in Prowse (1986). Short-wave radiation was measured at 1.1 m above the ice surface with a Swissteco two-component pyranometer. Measurements were recorded on a Campbell Scientific Ltd. CR 21 data logger at 10 sec. intervals for periods ranging from 10-15 minutes (specific details concerning the brash ice measurements are described later). Table 1 gives the sky conditions and solar altitudes prevailing during the measurement periods. Solar altitudes are generally within the range (i.e. 30°) where differences in albedo for clear versus cloudy conditions are at a minimum. Albedo values are listed as a proportion (0.0 - 1.0), 1.0 representing complete reflectance.

Ice Type	α	γ (degrees)
Crystal pavement	.17	43
Candled black ice	.39	41
Granulated black ice	.55	42
Brash mean	.08	
max.	.15	30 - 38
min.	.04	

Table 1. Albedo (α) and solar altitude (γ).
All measurements were conducted under clear sky conditions.

REVIEW AND RESULTS

The following is a descriptive chronology of changes in albedo to be expected during the break-up period. Albedo values for snow and white ice are derived from the literature while those for decaying black ice are from the Liard field studies.

Snow and white ice

At the beginning of the melt season, river ice sheets are normally blanketed with a snow cover which can reflect a high proportion of the incoming short wave radiation. Observations of albedos in excess of 0.9 for newly fallen snow are common throughout the literature. As the snow ages, and metamorphic processes produce larger grains the albedo rapidly declines to a range of approximately 0.6 to 0.4 (see eg. Dirmhirn and Eaton 1975; USACE 1956) and with impurities to even lower levels. In cases where an ice sheet is not free-floating or where drainage is impeded, either by a lack of flow channels or deep snow accumulations, meltwater can pond on the surface. For example, the albedo of meltwater puddles on sea ice is approximately in the range 0.20 - 0.25 (Grenfell and Maykut 1977).

Once the snow cover is totally eliminated, reflective losses are dependent on the albedo of the underlying ice. Near-surface ice stratigraphy is usually determined by the freeze-up process, but two extremes in ice types are commonly identified: white and black (blue) ice. White ice normally has a small grained polycrystalline texture and is comprised either of dynamic ice forms produced at freeze-up, such as frazil ice (see Tsang 1982), or snow-ice which forms from the slushing and subsequent refreezing of the surface snow cover (see Adams 1981). Although Grenfell and Maykut (1977) report albedos for white sea ice ranging from 0.72 for "frozen multi-year white ice" to 0.47 for "melting first-year white ice", typical values of white ice forms on fresh water appear to be only slightly lower than that for well ripened snow. Examples from the literature include: 0.41 "slush ice"; 0.46 "snow-ice" (Bolsenga 1969); 0.4 "refrozen slush ice" and "pancake ice" (Bolsenga 1983); 0.33 "white granular ice" (Adams 1978). The most dramatic decrease in albedo, however, occurs when the surfaces become wetted with meltwater. Bolsenga (1977), for example, observed a reduction in the albedo of decaying refrozen snow-ice/slush from 48% when solidly frozen to 21% when intergranular spaces were occupied by liquid water, although part of this reduction can also be ascribed to changing solar altitude during the decay period.

With continued melt, the white ice forms also disappear usually revealing an ice stratum composed of relatively transparent black ice - a columnar ice formed by in situ freezing of the water column. The transparency of this type of ice is determined by the rate of freezing, which in turn determines the degree to which impurities and gases are expelled during freeze-out, and the prevalence of frazil ice, which may be incorporated into the sheet. In cases where the surface white ice stratum is especially thick, or where break-up occurs in the early stages of melt, the underlying black ice sheet may never become exposed.

Black Ice

Under cold conditions, clear black ice can have an albedo similar to that of open water. Bolsenga 1969, for example, reports a value of 0.1 for clear, snow-free lake ice. Once the surface of black ice begins to melt, however, the previously smooth surface becomes etched as the individual crystal perimeters begin to melt. Fig. 1 is an example of such a surface with a measured albedo of 0.17 (Table 1). This is the first stage in a "candling" process which can extend to considerable depths within the ice sheet. Fig. 2 illustrates a black ice layer which has fully candled to a depth of approximately 15 cm. This acicular melt process tends to further increase the albedo, in this case to 0.39, probably because of increased scattering at the crystal interfaces and the granulation of the crystal tops. As surface decay continues, the previous black ice surface converts to a completely granular form such as shown in Fig. 3. Without knowing the decay history of an ice sheet, differentiating granulated black ice from the various types of white ice is extremely difficult. In this study, large areas of the ice sheet surface appeared from a distance to be either melting snow or white ice but only on close inspection were revealed to be granulated black ice. Notably, this type of ice can have an albedo equal to, or even higher, than that of many other "white" ice surfaces. For example, the albedo of the surface in Fig. 3 was 0.55, as high as that for aged snow. A similar rise in the albedo of decaying black ice has been noted by Polyakova (1966).

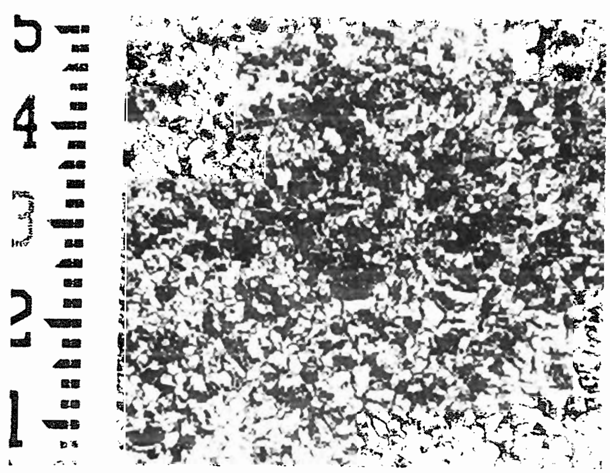

Figure 1: Crystal pavement. Early stages of black ice melt. Note etching of individual crystals.

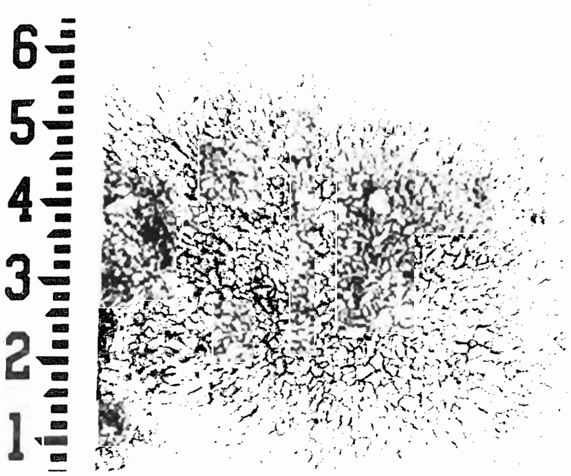

Figure 2: Surface of candled ice.

One possible reason for the relatively high albedo of granulated black ice is that the grains, having originated from black ice crystals, have a more angular surface and therefore higher reflectivity than the more rounded melt-freeze grains typical of other white ice forms. The onset of the granulation process is evident in the side-view of candling ice shown in Fig. 4.

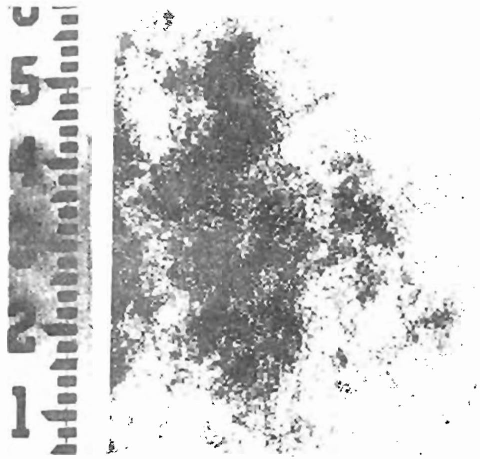

Figure 3: Granulated black ice. Angular grains indicate candling history.

Figure 4: Granulation at tops of black ice candles.

Figure 5: Swissteco albedometer suspended over brash ice in moving ice jam.

Brash Ice

Following the initial break-up of an ice cover, large portions of a river remain covered by fractured and fragmented ice either in the form of ice runs or ice jams. The duration, decay and release of these ice accumulations is partly dependent on melt produced by hydrothermal and atmospheric heat flows, and hence, in cases of radiative melt, the surface albedo. The surface composition of ice jams however, can vary significantly depending on the type of break-up process which leads to their formation (Prowse 1986). In thermal events (prolonged melt prior to break-up), extensive areas of the jams are comprised of large, relatively intact floes which can be assumed to have an albedo similar to that of the pre-break-up ice sheet. The remainder of thermal jams and the majority of the surface area of mechanically produced jams (limited pre-break-up decay, rapid run-off and break-up), are comprised of brash ice which can have a significantly lower albedo. In the case of the Liard River, there is often a brown tint to the brash ice, a product of extensive mixing in the turbid sediment-rich waters which accompany break-up. Because of the irregular reflective relief of brash ice and the variation in sediment and debris content, the albedo can be highly spatially variable. To obtain a representative albedo value for such a variable surface, common practice is to obtain radiation measurements from high elevations. During the 1985 break-up event, favourable conditions created by a releasing ice jam made it possible to obtain an average albedo of a brash ice surface by simply suspending the albedometer over the moving ice (Fig. 5). Over a period of approximately 1 hr. 20 min., 480 albedo measurements were recorded resulting in a mean albedo of .08 and a standard deviation of .02. A maximum of 0.15 was associated with the passage of a concentration of white ice floes and a minimum of 0.04 with a mixture of water and black ice.

SUMMARY

This note summarizes some of the changes in the albedo of snow and ice which are likely to prevail during the spring melt and break-up period. The rate of change in albedo is dependent on the type and thickness of individual strata and on prevailing weather conditions. The degree of albedo change is also dependent on how quickly river ice break-up follows the onset of melt. In the case of mechanical events, break-up may occur while the ice sheet remains largely snow covered. Only during thermal events would large areas of an intact ice sheet enter the latter stages of albedo change where the granulation of black ice causes a rise in reflectivity. The implications of the rise in albedo are twofold. Firstly, the higher albedo restricts radiative warming of flow beneath the ice cover and the associated melt of the ice bottom. It was believed previously that such warming could proceed relatively unhampered once the transparent layers of black ice were exposed. Second, the increase in reflectivity also slows the decrease in the mechanical strength of the ice sheet produced by internal radiative warming, especially in the case of thick ice sheets. For relatively thin ice sheets (e.g. 0.5m), however, the granulation process and resultant increase in albedo may not be complete before a significant proportion of the ice sheet candles. Under these conditions, the albedo can again decrease as the buoyancy of the ice sheet decreases and the near surface becomes saturated with water. The change in the albedo of black ice from low to high and back to low also offers an explanation for the adage of it being unsafe to venture onto an ice surface after the sheet has darkened for the second time.

The major implication of the low albedo of brash ice, is that the effect of radiation on total ice melt greatly increases during the break-up phase. Ice loss during break-up can be significant on large rivers such as the Liard and although much of this loss can be ascribed to hydrothermal melt (see e.g. Prowse and Marsh 1985), the importance of radiation melt has never been calculated, partly because of the lack of suitable albedo values. The magnitude and relative importance of the atmospheric and hydrothermal heat flows to brash ice covers is now the focus of further research.

REFERENCES

Adams, W.A. 1978. Effects of ice cover on the solar radiation regime in Canadian lakes. Verh. Internat. Verein. Limnol., 20, pp. 142-149.

Adams, W.P. 1981. Snow and ice on lakes. In: Handbook of Snow (D.M. Gray and D.H. Male, ed.) Pergamon Press, Toronto.

Ashton, G.D. 1983. Frist-generation model of ice deterioration. Proc. of Conference on Frontiers in Hydraulic Engineering, August 9-12, 1983, Cambridge, Massachusetts, pp. 273-278.

Bolsenga, S.J. 1969. Total albedo of Great Lakes ice. Water Resour. Res.5(5): 1132-1133.

Bolsenga, S.J. 1977. Preliminary observations on the daily variation of ice albedo. J. Glaciol. 18(80): 517-521.

Bolsenga, S.J. 1983. Radiation transmittance through lake ice in the 400-700 nm range. J. Glaciol. 27(95): 57-66.

Bulatov, S.N. 1972. Computation of the strength of the melting ice cover of rivers and reservoirs and forecasting of the time of its erosion. Proc. of Symposium on the Role of Snow and Ice in Hydrology, Banff, Alberta, September 1972, pp. 575-580.

Dirmhirn, I. and Eaton, F.D. 1975. Some characteristics of the albedo of snow. J. Appl. Meteor, 14: 375-379.

Grenfell, T.C. and G.A. Maykut. 1977. The optical properties of ice and snow in the Arctic Basin. J. Glaciol. 18(80): 445-463.

Polyakova, K.N. 1966. Characteristics of the melting of the ice cover and the opening of the middle Lena River. Sov. Hydrol., No. 3, pp. 276-292.

Prowse, T.D. 1986. Ice jam characteristics, Liard-Mackenzie rivers confluence. Can. J. Civ. Eng. 13(6): 653-665.

Prowse, T.D. and Marsh, P. 1985. Hydrothermal decay of ice jams. Proceedings of the 42nd Eastern Snow Conference, Montreal, Quebec, pp. 272-286.

Tsang, G. 1982. Frazil and anchor ice: A monograph. National Research Council of Canada, Subcommittee on Hydraulics of Ice-Covered Rivers, Ottawa.

USACE 1956. Snow hydrology. Summary of the snow investigations. U.S. Army Corps of Engineers, North Pacific Division, Portland, Oregon.