Journal of Glaciology, Vol. 29, No. 102, 1983

SPECTRAL REFLECTANCES OF SNOW AND FRESH-WATER ICE FROM 340 THROUGH 1 100 nm

By S. J. BOLSENGA

(Great Lakes Environmental Research Laboratory, National Oceanic and Atmospheric Administration, 2300 Washtenaw Avenue, Ann Arbor, Michigan 48104, U.S.A.)

ABSTRACT. Measured spectral reflectances of new and moderately metamorphosed snow were generally >80% from 340–950 nm. From 950–1 100 nm a characteristic dip and rise of spectral reflectances occurred. One spectroradiometer scan over a deteriorated snow patch showed much lower spectral reflectances than fresh snow, but the shape of the curve remained similar to that of fresher snow. Spectral reflectances for clear ice contrasted sharply with those for snow. In general, values were <10% and the curves lacked distinctive shape. Higher spectral reflectances, due to "lighter"-appearing ice in the measurement area, were measured at some sites. Refrozen slush, pancake, brash, and slush curd ice revealed spectral reflectance curves similar in form to each other, but which varied significantly in the range of spectral reflectances for each ice type. Generally, reflectances rose slowly from 340 nm to a peak near 550 nm. From 550–775 nm reflectances decreased slowly but significantly. A slight dip and rise in reflectances occured from 775–850 nm after which values again dipped significantly (850–900 nm). From 950–1 100 nm, a dip and rise in reflectances similar to that for snow was observed. The amount of slush included seems to control the reflectances of these ice types. All measurements were acquired with a pair of scanning spectroradiometers having picowatt accuracy, adapted to obtain, automatically, simultaneous readings of incident and reflected radiation from 340–1 100 nm. The spectroradiometers were field-calibrated using Sun-plus-sky radiation as a calibration source.

RÉSUMÉ. *Reflectances spectrale de la neige et de la glace d'eau douce entre 340 et 1 100 nm.* Les reflectances mesurées de la neige fraîche ou peu transformée étaient, en général, supérieures à 80% dans la bande 340–950 nm. Entre 950 et 1 100 nm interviennent un creux et un pic de reflectance. Un parcours au spectroradiomètre sur une neige transformée a montré des reflectances spectrales beaucoup plus faibles que pour la neige fraîche, mais la forme de la courbe est restée semblable à celle de la neige plus récente. Les reflectances spectrales pour de la glace claire contrastent fortement avec celles pour la neige. En général les valeurs étaient de moins de 10% et les courbes montraient des lacunes nettes. En certains points de fortes reflectances spectrales, dues à de la glace apparaissant comme plus «brillante» dans la zone observée, ont été mesurées. La neige fondante regelée, les crêpes, le «brash», et de la glace de «slush» figée ont donné des courbes de reflectance spectrale se ressemblant l'une l'autre, mais significativement décalées l'une de l'autre dans la gamme des reflectances spectrales. Généralement, la reflectance augmente faiblement à partir de 340 nm jusqu'à un maximum près de 550 nm. De 550 à 775 nm les reflectances décroissent lentement mais significativement. Une petite oscillation par creux et pic se place entre 775 et 850 nm après quoi les valeurs plongent à nouveau de manière sensible (850–900 nm). Entre 950 et 1 100 nm une oscillation par creux et pic semblable à celle de la neige s'observe. La quantité de «slush» qu'il inclut semble régler la reflectance de chaque type de glaces. Toutes les mesures ont été faites avec une paire de spectroradiomètres à balayage, précis au picowatt, construits pour donner automatiquement une lecture simultanée des radiations incidentes et réfléchies entre 340 et 1 100 nm. Les spectroradiomètres ont été étalonnés sur le terrain sur les radiations naturelles solaires et célestes.

ZUSAMMENFASSUNG. *Spektrales Reflexionsvermögen von Schnee und Süsswassereis im Wellenlängenbereich 340–1 100 nm.* Im Wellenlängenbereich von 340–950 nm betrug das gemessene spektrale Reflexionsvermögen von frischem und mässig umgebildetem Schnee im allgemeinen mehr als 80%. Zwischen 950 und 1 100 nm trat eine charakteristische Ab- und Zunahme des spektralen reflexionsvermögens auf. Die Abtastung eines älteren Schneeflecks mit einem Radiometer ergab weit geringere Werte als bei frischem Schnee, doch verlief die Reflexionskurve ähnlich wie dort. Die spektrale Reflexion bei klarem Eis unterscheidet sich wesentlich von der für Schnee. Im allgemeinen lagen die Werte unter 10% und die Kurven zeigten keinen unterschiedlichen Verlauf. An einigen Stellen mit „heller" erscheinendem Eis wurden höhere Werte der spektralen Reflexion gemessen. Wiedergefrorener Schneeschlamm, Pfannkucheneis und Trümmereis sowie Eismatsch zeigten Reflexionskurven, die ihrer Form nach ähnlich waren, sich jedoch im Bereich der spektralen Reflexion für jeden Eistyp wesentlich unterschieden. Im allgemeinen stieg das Reflexionsvermögen allmählich von 340 nm zu einem

https://doi.org/10.3189/S0022143000008352 Published online by Cambridge University Press

Gipfel bei 550 nm. Zwischen 550 und 775 nm nahm das Reflexionsvermögen langsam, aber deutlich ab. Eine leichte Ab- und Zunahme trate zwischen 775 und 850 nm auf, worauf die Werte wieder deutlich sanken (850–900 nm). Zwischen 950 und 1 100 nm wurde eine Ab- und Zunahme des Reflexionsvermögens ähnlich wie bei Schnee beobachtet. Der Anteil an Schneeschlamm scheint für das Reflexionsvermögen dieser Eistypen bestimmend zu sein. Alle Messungen wurden mit einem Paar von Abtastradiometern mit Picowatt-Genauigkeit gewonnen, die zur automatischen Aufzeichnung der einfallenden und reflektierten Strahlung zwischen 340 und 1 100 nm eingerichtet waren. Die Geräte wurden im Feld mit Hilfe der Sonnen- und Himmelsstrahlung geeicht.

Introduction

Winter in the North American Great Lakes is of sufficient length and severity that a partial or total ice cover is produced. Each Great Lake possesses its own set of characteristics both physical (volume, surface area, shore length, etc.) and hydrometeorological (temperature, precipitation, evaporation patterns, etc.). During a normal winter the aerial ice extent varies from 15% on Lake Ontario to 95% on Lake Erie (Rondy, 1976). During a severe winter, the aerial ice extent on all of the Great Lakes can approach 100%. Each of the ice types common to the Great Lakes has a unique appearance, mode of formation, and physical characteristics. The physical characteristics (strengths, reflectances, crystallographies, etc.) are studied to provide information useful for solving problems in winter navigation, shore-line engineering, hydropower generation, water supply, and water quality. One of those physical properties, the spectral reflectances of the various ice types, is examined here.

A pilot study was conducted in early 1967 to obtain the total (wavelength-integrated) albedo of various types of ice common to the Great Lakes (i.e. clear, slush, ball, pancake, etc.) at only one solar altitude and cloud condition. Measurements ranged from 10% for clear ice to 46% for snow-ice (Bolsenga, 1969). The program indicated that many gaps existed in our knowledge of the albedo of fresh-water ice types. A new program was subsequently initiated to develop a catalog of information that would provide measurements of the total albedo of fresh-water ice within a wide range of solar altitudes (10–45°), cloud types, and cloud amounts. A measurement system was fabricated consisting of upward- and downward-facing pyranometers mounted on a tripod and boom assembly. Reports from that study (Bolsenga, 1977, 1979, 1980) commented on the diurnal variation of, and on the influence of changing cloud conditions on, the total albedo of ice.

The need for ice and snow reflectance measurements over various spectral ranges was emphasized by these and other earlier studies. With data at reasonably small wavelength intervals, reflectances could be determined corresponding to the current or future sensitivity of detectors designed for many applications. In order to satisfy this need, two scanning spectroradiometers manufactured by the EG&G Corporation, were used to measure radiation in the 280–1 100 nm range* (10 nm bandpass slits). Two instruments were required to measure incident and reflected radiation at each wavelength simultaneously (one instrument directed to the zenith and the other to the nadir, both synchronized at equal scanning speeds). As elaborated elsewhere (Bolsenga and Kistler, 1982), optimum instrument configuration requires that incoming radiative flux be recorded simultaneously with reflected flux since reflectances calculated from non-simultaneous spectroradiometer scans can be subject to significant errors due to possible changes in the atmosphere or solar altitude during the period of the scan. At the

* It should be noted that even though the instruments are capable of measuring flux at wavelengths as low as 280 nm, spectral reflectance graphs shown here artificially suppress data from 280–330 nm due to a presumed lack of sufficient flux in the natural environment to produce accurate spectral reflectance calculations through most of that range.

measurement site the spectroradiometers were mounted on a boom 4.6 m long supported at each end by tripods, placing the reflected unit about 1.5 m above the ice surface.

Few studies have been conducted on the spectral reflectances of ice. One of the most recent (Grenfell and Maykut, 1977) deals with reflectance measurements of sea ice and refrozen melt ponds. In addition to the fact that sea ice rather than fresh-water ice was studied, Grenfell and Maykut used only one "spectrophotometer" requiring time to complete an incident and subsequently a reflected scan. It is emphasized that the instruments or data in this study are not adjusted to the human eye response (CIE curve). Perovich and Grenfell (1981) report on the spectral reflectance of laboratory-grown young sea ice.

Duggin (1980) independently developed a technique similar to that used here but with different equipment. Both studies were apparently conducted concurrently and arrived at similar conclusions with respect to the technique involved. Duggin calibrated in the field using a $BaSO_4$ coated plate, whereas in this program both instruments were pointed to the zenith using the Sun plus sky as a calibration source. The instrumentation used in this study, including wiring to control the two spectroradiometers as well as software to process the data in the field immediately, is described in detail in Bolsenga and Kistler (1982).

In an early, but classic study, Krinov (1947) determined the spectral reflectances of a great variety of natural surfaces, including a limited number of measurements of snow crusted with ice. Krinov recognized the problems associated with changing atmospheric conditions and solar altitude:

> "As stated above, the illumination of the object and the standard surface must be the same. Thus the spectrograms were usually obtained when the sky was clear and seldom when the sky was clouded and only when the clouds were low (cumulus, strato-cumulus, broken cumulus, etc.) and the total amount of cloud in the sky did not exceed 0.3; in this case exposures were made only when the sun broke through the clouds. In a few cases spectrograms were obtained when there was a continuous even cloud cover extending across the entire sky. Usually the spectrograms were taken near midday beginning two hours before noon and ending two hours after noon."

In order to achieve practical results comparable with those reported in the literature, the standard 8° field-of-view of the spectroradiometers was modified by the use of hemispherical cosine diffusers. Initial cosine response tests at 550 nm showed agreement with ideal cosine response within a few per cent depending on the angle of the diffuser from a calibration light source. Further tests showed that diffuser cosine response differed increasingly from true cosine response with increasing wavelength.

Due to these difficulties, all of the measurements reported here were collected with the plastic diffusers but under overcast skies unless specifically noted. Observers classified skies as heavily overcast only when entirely sure of the lack of direct radiation as indicated by the unaided eye. In some cases described later, matte paper, which produced much improved cosine response, was taped over the hemispherical diffusers.

Measurements of snow and ice reflectances were acquired at several sites including the Straits of Mackinac (lat. 45° 42′ N., long. 84° 30′ W.) located between Lakes Michigan and Huron (Fig. 1) and North Lake, a small inland lake located near Ann Arbor, Michigan (lat. 42° 25′ N., long. 85° 57′ W.).

Ice at the Straits of Mackinac sites was relatively young at the time of the measurements. First ice of the season, as detected by side-looking airborne radar and reported by the U.S. National Weather Service, was located in a near-shore area south of Bois Blanc Island on 18

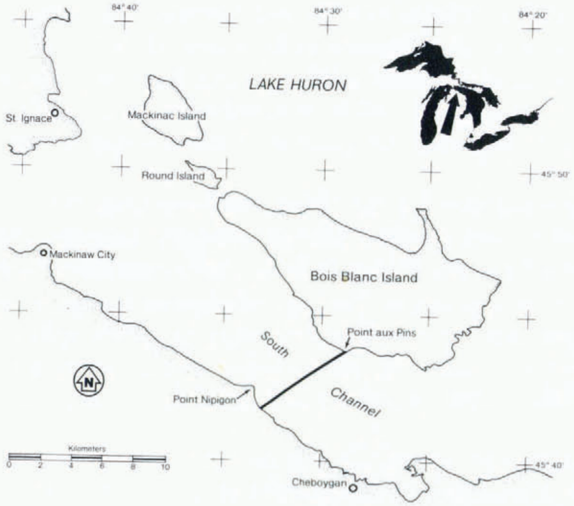

Fig. 1. Location of Straits of Mackinac measurement sites. Spectral reflectance data were collected on traverses along the solid dark line from the mainland to Point aux Pins.

January 1980. Spectral reflectance measurements spanned the period 19–22 February 1980. On 25 January 1980, thin ice (10–20 cm) was reported to cover fully the area of the measurements from Bois Blanc Island to the mainland on the south-west. On 28 January 1980, satellite imagery and U.S. Coast Guard reports verified that condition. On 30 January 1980 the ice was 23 cm thick at one location in the area of the spectral reflectance measurements. The thickness increased to 38 cm by 13 February 1980. The ice continued to cover the area fully from 25 January through 2 April 1980.

Ice growth on North Lake proceeded as one might expect for a small inland lake unaffected by high winds or significant water currents. Snow-ice formed first due to snow-fall into near-freezing water and subsequent congelation. A warm spell caused break-up of the thin ice layer. The "white-ice" fragments were subsequently frozen with new clear ice in certain near-shore areas. A moderate amount of snow accumulated on that ice and subsequent mild temperatures melted the snow to a slush layer which refroze later due to lower temperatures. Snow loads were never sufficient to fracture the ice and cause water to percolate upward into the snow layer. Melting and refreezing of overlying snow and additional growth of clear ice did, however, continue throughout the season.

Analysis

A descriptive classification of Great Lakes ice types was made prior to this study by Marshall (1966, 1977 [a], [b]). Spectral reflectance measurement sites were selected on the basis of his classifications. Measurements were made on snow, clear ice, refrozen slush ice, brash ice, pancake ice, and slush curd ice.

An extensive amount of research on the reflectance of snow has been conducted with pyranometers, thus confining the data to integrated values in the 300–2 000 nm range. Experimental work on the spectral reflectances of snow in small wavelength increments by

O'Brien and Munis (1975), O'Brien (1977), and Grenfell and others (1981) seems to agree reasonably well with these results although O'Brien and Munis' data, it should be noted, were collected under artificial conditions and with the source beam and detector at various angles to the surface. Warren and Wiscombe (1980), Wiscombe and Warren (1980), Choudhury (1981), and Warren (1982) have addressed the question of the modeling of the spectral reflectance of snow.

Several scans were made over 3–4 cm of new snow overlying 15 cm of mostly clear lake ice. Skies were totally overcast with sporadic light snow flurries. Incoming radiation for all runs was totally diffuse as observed visually. Figure 2 shows results from two runs during a period without snow flurries. Spurious data points in this and subsequent figures are due to instrumentation noise. It is emphasized that these points, evidenced by sharp spikes in the graph, are not considered to represent real data. The curves are characterized by a narrow range of values (81–89%) from 340–900 nm after which a pronounced dip to values of about 70% occurs near 1 050 nm.

After these measurements the surface was liberally trampled by the observers. Care was taken to expose only a minimum of the underlying ice surface. The measurements over the trampled surface compared to those over undisturbed snow show small (2–3%), consistently lower spectral reflectances of the trampled surface from 340–900 nm. From 900–1 100 nm little differences in spectral reflectance was noted. Contrary to the very small differences in measured spectral reflectances, the trampled surface appeared quite different to the eye from the undisturbed surface. Perhaps the "look angle" of the spectroradiometers as opposed to the "look angle" of the eye of the observer accounted for the discrepancy. Measurements were made again on snow at the same location three days later. The surface had obviously metamorphosed during the intervening period, being hard packed and etched by the wind. The shape and numerical values of the curves are very similar (differences, generally <5%) to those in Figure 2. Scans over other hard-packed snow surfaces showed slightly lower spectral reflectances than those for the new snow shown in Figure 2.

One run was made over a snow surface that had reached an advanced stage of deterioration. The mass was dark in appearance and nearly at the stage of complete melting. Sky cover was totally overcast but thin in spots. Weak beam radiation was observed sporadically through thin portions of the overcast, but only during about 10% of the period for the run. The shape of the curve is similar to that of fresher snow but with greatly reduced values (Fig. 2), more appropriately ascribed to poorly reflecting refrozen slush.

Depending on bubble content, fracture patterns, and thicknesses, clear ice can demonstrate a remarkable ability to transmit solar radiation at certain wavelengths. In a study of the transmittance of photosynthetically active (400–700 nm) radiation through ice, Bolsenga (1981[a]) found that nearly 90% of the incoming radiation was transmitted through 28 cm of clear ice. (The transmittance dropped to 10% for the same ice surface with a 3 cm thick layer of snow.) This property is important in considering the spectral reflectances of ice, since the underlying water or the bottom of the lake is often clearly visible through clear ice and can obviously influence reflectances in certain spectral ranges.

Three dual spectroradiometer runs over approximately 30 cm of clear ice were made at the Straits of Mackinac location (Fig. 3). The clear ice probably formed in a zone of open water between large areas of ice of different types. Sky cover was dense overcast (radiation totally diffuse). Neither the Sun nor shadows were visible throughout the period of the runs. The ice surface appeared moist when the 5 m diameter area was cleared of snow for the measurements.

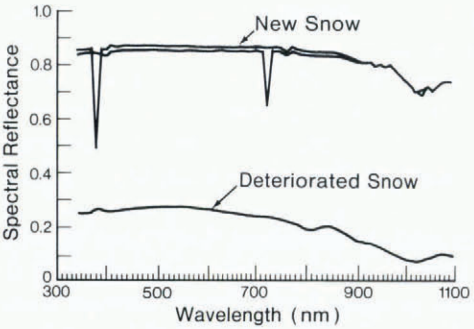

Fig. 2. Spectral reflectances of new snow (upper trace) as compared to spectral reflectances of the same surface trampled by the observers. Also shown are spectral reflectances of a much deteriorated patch of snow.

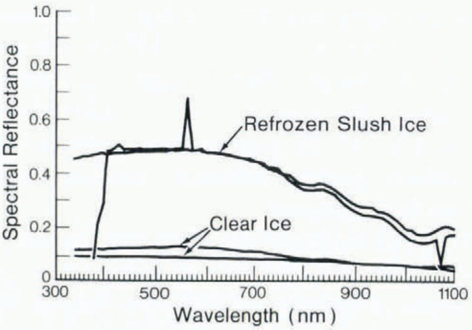

Fig. 3. Spectral reflectances of Straits of Mackinac clear ice compared to North Lake clear ice (upper clear-ice trace). The lower trace marked refrozen slush ice is for data collected with plastic diffusers. Lack of transmittance of the matte paper causes erroneous spectral reflectances from 340 through 410 nm as shown by the upper refrozen slush-ice trace.

Temperatures were near freezing and snow at the site compacted easily. It is postulated that mild weather caused water to percolate through the snow layer to the snow–ice interface. The surface appeared very black, no doubt partly due to the 22 m water depth at the site. The depth of the water precluded any influence of the bottom on scan values. The curves lack any abrupt changes in spectral reflectance with recorded reflectances remaining in a narrow (5–10%) range for three spectral scans. A general decrease in reflectance throughout the spectrum is evident with a slight, almost imperceptible rise in values from 340 to roughly 500 nm, where the highest values occur.

In another case, the spectral reflectances of 15 cm of clear ice on North Lake with a water depth of only 1 m was measured (Fig. 3). Although the surface was predominantly clear ice, several fragments of "white ice" (snow-ice or refrozen slush) from a previously melted ice cover were frozen in the clear ice matrix and numerous cracks were noted. Higher spectral reflectances from 340–850 nm for this ice surface as compared with the Straits of Mackinac surface can be ascribed to cracks and the "white-ice" inclusions.

Refrozen slush ice is one of the most common ice surfaces on the Great Lakes and on smaller inland lakes. High snow-falls with intervening high temperatures or rain often reduce an overlying snow cover to slush. Subsequent refreezing often occurs due to lowered temperatures. Three scans with matte paper over the plastic diffusers and one scan with plastic diffusers alone were made over a refrozen slush-ice surface (Fig. 3) at the Straits of Mackinac about 150 m offshore. Since the ice was located in the near-shore zone, formation of the surface probably began in mid-January as noted earlier in this paper. A thick overcast was present and no shadows were observed during the measurements. Since data from the three matte-paper runs were nearly equal, data from only one such run are shown in Figure 3. Data from the run with plastic diffusers are nearly equal to spectral reflectances for the run with paper diffusers from 340–750 nm. Even though skies were totally overcast, above 750 nm a disparity of 2–4% is evident. Differences between the two runs at higher wavelengths are most likely to be due to uneven radiance (intrinsic radiant intensity emitted by a radiator in a given direction) distribution over the sky hemisphere (Petherbridge, 1955; Kasten and Möller, 1960; Gordon and Church,

1966) combined with the fact that (1) refrozen slush is not a diffuse reflector, and (2) cosine response errors of the plastic diffuser are severe at higher wavelengths. The errors are not considered to be of sufficient magnitude seriously to hamper the usefulness of the data for most applications. The type of ice surface and the associated likelihood of specular reflection influence the possibility of this type of error. Specular reflection does not occur with Lambertian surfaces, and without specular reflection, spectral reflectance errors attributed to uneven sky radiance have not been detected.

Two runs taken 30 min apart over another refrozen slush-ice surface on North Lake (Fig. 4) show a significant difference of about 10% from the spectral reflectances of refrozen slush ice at the Straits of Mackinac from 340–700 nm. Differences in spectral reflectances between these refrozen slush-ice surfaces can be attributed to variations in surface texture and bubble content.

Since pancake ice is composed of fragments of a previously formed ice cover, the individual pancakes consist of any type of ice (clear ice, snow-ice, etc.) with raised rims of congealed small ice fragments, ranging in size from sand to gravel, caused by grinding of one fragment against another. The result of this abrasive process appears as a "whitish" rim enclosing an inner portion varying in appearance from white (snow-ice, slush ice) to black (clear ice). Spectral reflectances of such ice could be expected to vary widely, depending on the make-up of the inner portion of the individual pancakes and the matrix in which the individual pancakes are frozen.

Only one type of pancake ice was measured in this study (Straits of Mackinac, approximately 1.5 km offshore) in which the individual pancakes were formed from refrozen slush-ice or snow-ice fragments. The pancakes were not frozen together horizontally, but in a mildly chaotic fashion. Since the ice was located a fair distance from shore, it is speculated that a weak ice cover formed, only to be fragmented by winds and water currents. The fragments subsequently resided in a zone of mild, but prolonged turbulence where continued abrasion of the edge of one piece against another caused the characteristic raised rims and circular shape. Higher winds then thrust one pancake partially over another in the "mildly chaotic fashion". A significant amount of snow was swept from the measurement site with a broom, including the areas under and between the slightly upthrusted pancakes. Three runs were made over the surface under a heavy overcast. Figure 4 shows spectral reflectances from one run lacking spurious data points. Spectral reflectances are generally about 50% from 340 to 700 nm. After 700 nm the spectral reflectances gradually decrease to nearly 20% at about 1 050 nm. The dip and subsequent rise in spectral reflectances from 950–1 100 nm is similar to that found in snow.

The spectral reflectances shown in Figure 4 are not likely to be characteristic of the spectral reflectances of all types of pancake ice. "White" ice (refrozen slush or snow-ice) formed most of the surface. Darker ice such as clear ice could as easily have been involved. The spectral reflectances of the pancake ice measured here can be perceived as similar to the spectral reflectances of refrozen slush ice. The spectral reflectances of the other form of pancake ice described can be predicted to be a combination of the spectral reflectances of refrozen sluch ice (rims only) and clear ice. Similarities between the measured spectral reflectances for pancake ice and those shown for the refrozen slush ice in Figure 3 are noteworthy. Using pyranometers, Bolsenga (1969) measured the albedo of snow-free pancake ice composed of "white" ice pancakes in a clear ice matrix as 31% under 4/10 altocumulus at a solar altitude of 32°. The average spectral reflectance of the data shown in Figure 4 is 40%. Clear ice comprised a greater proportion of the total area for the ice in the earlier study, which accounts for the lower albedo value.

Two distinct varieties of slush curd ice were measured at the Straits of Mackinac. Slush curd

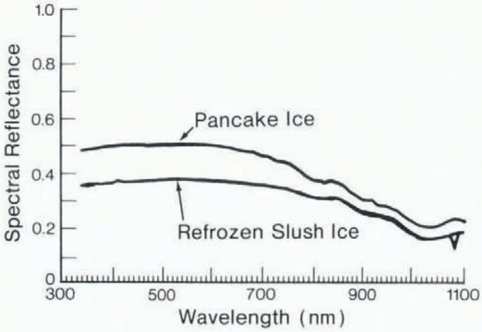

Fig. 4. Spectral reflectances of refrozen slush ice on North Lake (two runs) and spectral reflectances of pancake ice on the Straits of Mackinac.

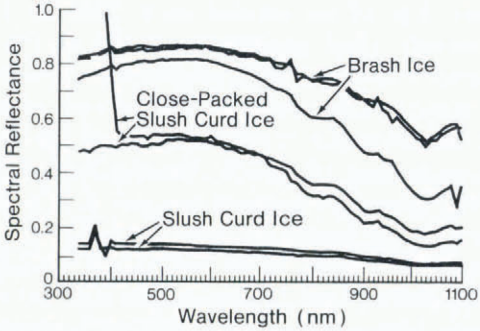

Fig. 5. Spectral reflectances of slush curd ice and brash ice. In the lowest two traces, the upper trace is from a scan over a "lighter" portion of the ice while the lower trace is from a "darker" portion. In the middle two traces, lower spectral reflectances are from a run with plastic diffusers and upper spectral reflectances are from paper over plastic diffusers. In the upper three traces, spectral reflectances of brash ice with upthrusted blocks (upper traces) and the same ice surface with appreciable snow removed are shown.

ice forms when a thick slush layer freezes into a coagulated or thickened pattern with intervening clear-ice areas. The different appearance of the different types of slush ice was probably caused by differing slush contents of the water combined with varying degrees of water turbulence. At the first site the slush curds were frozen in a matrix of clear ice. The visual appearance of the ice was highly variable due to a variety in size and concentration of the individual curds. The ice was 30 cm thick over water 24 m deep. Multiple runs were made at two points to observe an anticipated difference in spectral reflectances over portions of the ice which appeared different. A medium-thick to thick overcast was present during the runs.

As shown in Figure 5 a small, but easily distinguishable, difference of 2–3% from 340 to 1 000 nm was measured between runs over the "darker" and "lighter" surfaces. The similarity of the slope of the curves to that of clear ice curves at the Straits of Mackinac on the same day (Fig. 3) is striking. Spectral reflectances of the slush curd ice are slightly higher than those of the clear ice in all cases, but with the large number of whitish-appearing slush curds, it is surprising that the spectral reflectances were not much higher.

At the second site the surface of the slush curd ice appeared to the casual observer as snow-ice or refrozen slush. Closer scrutiny revealed definite slush curds about 10–30 cm in diameter and closely packed (no clear ice). Figure 5 shows the spectral reflectances under a thick overcast using plastic diffusers and reflectances from the same ice surface with matte paper over the plastic diffusers. The differences between the two runs at longer wavelengths is explained by sky radiance differences interacting with the type of ice and with diffuser transmittances.

Brash ice could be described as one of the most visually distinguishable types of ice. It is perhaps a paradox that the same ice type has little chance of being ascribed a "characteristic" set of spectral reflectance values. The individual blocks of ice might have been composed originally of any type of ice ranging from clear ice to snow ice with spectral reflectances varying accordingly. Final arrangement of the ice fragments after congelation is usually chaotic, but preferred orientation of the fragments is possible. A wide variation of spectral reflectances would thus occur for most brash-ice sites on clear or partly cloudy days due to changing solar altitudes over the period of a day interacting with ice-block orientation. Spectral reflectances of brash ice are thus not only site specific, but also diurnally dependent.

The brash ice measured here was located at the Straits of Mackinac nearly 5 km offshore. Individual ice fragments were composed of clear ice 4–5 cm thick. Light was observed to penetrate the blocks easily imparting a "bluish" color to some areas of the mass. Skies were heavily overcast during all runs. The first two runs were completed with a considerable amount of snow both covering and deposited between the individual ice fragments. After those runs, much snow was removed from the measurement area by hand and with a broom. The ice with an undisturbed snow cover indicated high (>80%) spectral reflectances over most of the visual range (Fig. 5). Reflectances lowered fairly smoothly after 700 nm. Near 1 000 nm a dip and rise was found similar to that observed in snow. For the scan after the snow had been partially removed, reflectances were lower than for the undisturbed surface at all wavelengths with the disparity increasing with increasing wavelengths.

Summary

Results of measurements with a state-of-the-art, dual, scanning spectroradiometer system have provided spectral reflectance data under over-cast skies for several types of fresh-water ice commonly found in the Great Lakes as well as many smaller bodies of water. It is hoped that the scans will provide information useful for remote sensing, energy-budget analysis, winter aquatic biology, and for spectral reflectance modeling of snow and ice. Fuller details of this work are available in Bolsenga (1981[b]).

Acknowledgements

D. C. Norton (Great Lakes Environmental Research Laboratory) assisted in the field work and development of the apparatus for collecting field data. S. I. Outcalt and G. H. Suits (University of Michigan) provided invaluable technical advice. R. D. Kistler (Great Lakes Environmental Research Laboratory) and R. Watson (EG&G Inc.) assisted in development of the data acquisition and processing software.

MS. received 19 January 1982 and in revised form 25 June 1982

REFERENCES

Bolsenga, S. J. 1969. Total albedo of Great Lakes ice. *Water Resources Research*, Vol. 5, No. 5, p. 1132–33.

Bolsenga, S. J. 1977. Preliminary observations on the daily variation of ice albedo. *Journal of Glaciology*, Vol. 18, No. 80, p. 517–21.

Bolsenga, S. J. 1979. Solar altitude effects on ice albedo. *NOAA Technical Memorandum* ERL GLERL-25.

Bolsenga, S. J. 1980. New observations on daily variations of ice albedo. *NOAA Technical Memorandum* ERL GLERL-27.

Bolsenga, S. J. 1981[a]. Radiation transmittance through lake ice in the 400–700 nm range. *Journal of Glaciology*, Vol. 27, No. 95, p. 57–66.

Bolsenga, S. J. 1981[b]. Spectral reflectances of freshwater ice and snow from 340 through 1 100 nm. *University Microfilms International. Publication* No. 81–25,073.

Bolsenga, S. J., *and* Kistler, R. D. 1982. A dual spectroradiometer system for measuring spectral reflectances. *Journal of Applied Meteorology*, Vol. 21, No. 5, p. 642–47.

Choudhury, B. 1981. Radiative properties of snow for clear sky solar radiation. *Cold Regions Science and Technology*, Vol. 4, No. 2, p. 103–20.

Duggin, M. J. 1980. The field measurement of reflectance factors. *Photogrammetric Engineering and Remote Sensing*, Vol. 46, No. 5, p. 643–47.

Gordon, J. I., *and* Church, P. V. 1966. Overcast sky luminances and directional luminous reflectances of objects and backgrounds under overcast skies. *Applied Optics*, Vol. 5, No. 6, p. 919–23.

Grenfell, T. C., *and* Maykut, G. A. 1977. The optical properties of ice and snow in the Arctic basin. *Journal of Glaciology*, Vol. 18, No. 80, p. 445–63.

Grenfell, T. C., *and others*. 1981. Spectral albedos of an alpine snowpack, by T. C. Grenfell, D. K. Perovich, and J. A. Ogren. *Cold Regions Science and Technology*, Vol. 4, No. 2, p. 121–27.

Kasten, F., *and* Möller, F. 1960. *Advanced investigations of the visual range under overcast sky (visual range and albedo especially in polar regions)*. Mainz, Meteorologisch-Geophysikalisches Institut der Johannes Gutenberg-Universität. (Final Technical Report, U.S. Dept. of Army Contract DA-91-591-EUC-1300.)

Krinov, E. L. 1947. *Spektral'naya otrazhatel'naya sposobnost' prirodnykh obrazovanii*. Moscow, Laboratoriya Aerometodov, Akademiya Nauk SSSR. [English translation: Spectral reflectance properties of natural formations. *Canada. National Research Council. Technical Translation* 439, 1953.]

Marshall, E. W. 1966. Air photo interpretation of Great Lakes ice features. *University of Michigan. Great Lakes Research Division, Institute of Science and Technology. Special Report* No. 25.

Marshall, E. W. 1977[a]. The geology of the Great Lakes ice cover. 1. *University Microfilms International. Publication* No. 77–26,298.

Marshall, E. W. 1977[b]. The geology of the Great Lakes ice cover. 2. *University Microfilms International. Publication* No. 77–26,298.

O'Brien, H. W. 1977. Observations of the ultraviolet spectral reflectance of snow. *U.S. Cold Regions Research and Engineering Laboratory. Report* 77–27.

O'Brien, H. W., *and* Munis, R. H. 1975. Red and near-infrared spectral reflectance of snow. *U.S. Cold Regions Research and Engineering Laboratory. Report* 332.

Perovich, D. K., *and* Grenfell, T. C. 1981. Laboratory studies of the optical properties of young sea ice. *Journal of Glaciology*, Vol. 27, No. 96, p. 331–46.

Petherbridge, P. 1955. The brightness distribution of the overcast sky when the ground is snow covered. *Quarterly Journal of the Royal Meteorological Society*, Vol. 81, No. 345, p. 476–78.

Rondy, D. R. 1976. Great Lakes ice cover. (*In* Limnology of lakes and embayments. *Great Lakes Basin framework study*. Appendix 4. Ann Arbor, Great Lakes Basin Commission, p. 105–17.)

Warren, S. G. 1982. Optical properties of snow. *Reviews of Geophysics and Space Physics*, Vol. 20, No. 1, p. 67–89.

Warren, S. G., *and* Wiscombe, W. J. 1980. A model for the spectral albedo of snow. II. Snow containing atmospheric aerosols. *Journal of the Atmospheric Sciences*, Vol. 37, No. 12, p. 2734–45.

Wiscombe, W. J., *and* Warren, S. G. 1980. A model for the spectral albedo of snow. I. Pure snow. *Journal of the Atmospheric Sciences*, Vol. 37, No. 12, p. 2712–33.