Energy- and mass-balance observations of the land–ice–ocean–atmosphere system near Barrow, Alaska, USA,
November 1999–July 2002

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ABSTRACT. We present results from a comprehensive field study carried out near Barrow, Alaska, USA,
designed to characterize local- to intermediate-scale sea-ice processes in the Arctic coastal zone of
central importance to the annual cycle and evolution of the coastal sea ice. Included in this are the
behavior of the snow cover of the ice and adjacent tundra and lake system; concurrent studies of mass
balance of the sea ice and lake ice; interaction of shortwave radiation with the shore-fast ice and the
adjacent land surfaces; evolution of the area coverage and distribution of the various surface types; and
the resulting regional albedo values. Maximum snow depths decreased during 2000–02 from 0.38 m to
0.26 m. Ice-melt rates in 2001 were 0.05 and 0.028 m d–1 at the top and bottom of the sea ice
respectively, two to three times larger than observations from the central Arctic. Detailed surface results
combined with aircraft photography were used to calculate regional albedos for the late spring and early
summer of 2001. Values ranged from 0.8 for all cold snow-covered surfaces to approximately 0.4 for
melting sea ice and lake ice vs 0.18 for bare tundra. Regional and surface-based values of cumulative
shortwave radiation entering the ice were consistent, indicating that albedo sampling on a scale of
200 m can provide a useful representation for regional sea-ice albedo.

INTRODUCTION

There is a continuing accumulation of evidence that the Arctic climate is warming; the sea-ice cover has showed
diminished extent, seasonal duration and thickness particularly for multi-year ice (Rothrock and others, 1999; Cavaliere
and others, 2003). This has been accompanied by large fluctuations in the first-year (FY) ice extent characterized by
rapid retreat of the ice from the coastal zone early in the melt season. The retreat is accelerated by feedback
interactions between solar radiation and the ice cover that act to modulate changes in input of solar radiation to the ice
and ocean, synoptic-scale atmospheric variability, the length of the melt season and the rate of fall freeze-up.

Determination of the physics of these interactions, development of suitable parameterizations, and proper
inclusion of them in regional and large-scale energy-balance models requires a detailed understanding of a wide range of
small-scale processes (Ebert and others, 1995; Arbetter and others, 1997; Holland and others, 1997), particularly during
the melt season when radiative forcing is strongest. Local-scale processes associated with the coastal sea ice (primarily
FY ice), bare tundra and coastal lakes are critical components of the system and apply over an extensive zone.
FY sea ice makes up more than one-third of the maximum sea-ice cover, and lakes cover up to 50% of the coastal
tundra in Siberian and North American Arctic lowlands.

The evolution and interaction of the mass and energy balance of the different coastal surface types is largely
unexplored, however, and comparison with corresponding processes in the central pack is needed to provide an
accurate model of the system as a whole. The Arctic coastal zone is a critical area for several reasons. Radiation-driven
melting processes act early and strongly at the coastal boundaries of the ice pack where spatial albedo contrasts
are very large. Land-ice boundaries are particularly sensitive areas. Snowmelting over land surfaces proceeds faster and
occurs earlier than over sea ice, and near-surface summer temperatures can rise well above freezing. This provides an
advective heat source for the ice, and the summer ice melt begins and propagates out from the coasts. Changes in
albedo and ice thickness during the melt season follow the same pattern as in much of the central pack, but proceed at
significantly greater rates.

EXPERIMENTAL PROGRAM

The coastal zone in the vicinity of Barrow, Alaska, USA, is particularly well situated for studies of these kinds of
interactions. This area provides site stability and easy accessibility so that detailed temporal coverage is practical
for a modest investment of logistics effort. We carried out a comprehensive study in this area over a 3 year period,
designed to characterize the annual cycle and evolution of sea-ice properties on local to intermediate scales in the
Arctic coastal zone. The program also included monitoring of the snow cover and the temperature regime of the
adjacent tundra and lakes.

This involved concurrent studies of the mass balance of the snow and ice; the interaction of shortwave radiation with
the shore-fast ice and the adjacent land surfaces; the
The program was carried out from November 1999 through June 2002 at the four principal sites shown in Figure 1a. For continuous monitoring of the seasonal evolution, automated stations were deployed during the late fall and visited periodically to retrieve the data. In April and from late May through the end of June (just before ice break-up), intensive on-site observations were carried out to characterize cold-season conditions and monitor the transition to and evolution of the melt season when ice conditions evolved rapidly and were highly non-uniform.

The automated measurements included (i) vertical temperature profiles using thermistor strings extending from the air through the snow and ice into the upper part of the water column (or 0.8 m into the tundra), (ii) snow depth using an array of 16 depth gauges designed to be read without disturbing the local snow cover, and (iii) ice thickness using heated wire/crossbar gauges. Data were collected at each site at 2–4 week intervals during the cold season and every 2–3 days during the melt season. The instrumented site on Elson Lagoon is shown in Figure 1b. During the on-site intensive measurement periods we observed spectral albedos, spectral transmissivity profiles through ice, ice salinity profiles, melt-pond areal coverage and water depth, and surface topography of undeformed ice.

The fifth site, in the Beaufort Sea just to the east of Point Barrow, was equipped with an automatic monitoring station during 2001 and 2002 to record the thickness and temperature profiles. The array of sites covered three dynamically distinct sea-ice zones, the tundra and a freshwater lake.

To extrapolate the surface-based observations to larger spatial scales, aerial-photographic survey flights were carried out during the melt season over an area of about 100 km² including the surface sites and covering the coastal sea ice and adjacent land.

**RESULTS**

**Snow cover**

Snow thickness vs time is shown in Figure 2 for the three sea-ice sites. The Chukchi site was established first each year and spanned most of the season. It showed a steady build-up of snow throughout the winter followed by a marked increase in April and May to a maximum depth of 0.40 m. The maximum snow depth on Elson Lagoon was the same. The total snow accumulation decreased in 2001 and again in 2002. Snow-depth measurements at the end of the ice-growth season along a 200 m profile suggest that average snow depths over undisturbed ice are lower, indicating that despite their small cross-section (20 x 50 mm) the snow gauges (as well as the surrounding installations) may have contributed to catchment of drifting snow at these sites. The onset of melt in both 2000 and 2001 occurred on about 1 June and the snow melted away in 7–10 days. In 2002, the melt began 2 weeks earlier and the undrifted snow was almost completely melted by 1 June.

Snow density evolution was similar in all three years. Shortly after deposition, the density in the windpack reached a value of about 0.33 (±0.01) Mg m⁻³ during the cold season, typical of winter windpacked conditions. At the base of the snow was a layer of depth hoar with a density of approximately 0.15 Mg m⁻³. For the deeper snowpacks, this layer reached about 0.05–0.1 m in thickness in May, while for the thinner cases (Imikpuk Lake (IM) in 2000 and Chukchi Sea (CS) in 2001 and 2002) there was only a trace of depth hoar. At the onset of melt, melting at the surface and percolation of the water produced a layer of superposed ice at the snow–ice interface and destroyed the depth-hoar layer. During the melt, the snow density increased to near 0.4 Mg m⁻³ due to compaction and recrystallization under warm conditions. This picture is consistent with snowpack evolution described in a previous experiment (Perovich and others, 1998).

A time series was not obtained for the tundra since automatic stations were installed only at the ice sites, but observations in mid- to late April or early May showed that
the maximum mean snow depth was 0.31, 0.30 and 0.28 m in 2000 through 2002 respectively. The surface of the tundra was very uneven while the surface of the drifted snow was relatively flat in late spring. As a result, the standard deviation for snow depth was approximately 0.1 m a–1 and the maximum depth was essentially the same for all three years. As at the three ice sites, the bulk of the snow on the tundra melted away in about 7–10 days, although the thickest snow remained for an additional 3–4 days.

It is evident from Figure 2 that the maximum snow depth for the ice sites generally decreased and that there were significant site-to-site differences each year. This is probably due in part to differences in wind strength and drift patterns combined with ice surface conditions during initial snow deposition, which can influence the roughness of the ice surface and the tendency of blowing snow to adhere to it. Unfortunately, we do not have direct observations to test this hypothesis.

**Mass balance**

Figure 3 shows the full mass balance at the ice sites for the 2000 season. Establishment of the ice cover occurred in early October 1999, and ice growth proceeded into the beginning of the following June to a maximum of 1.54, 1.49 and 1.92 m of growth at the CS, Elson Lagoon (EL) and IM sites respectively. The greater ice thickness at the IM site was primarily due to the much thinner snowpack and to the higher thermal conductivity of the freshwater ice. Although differing slightly in magnitude, the temporal pattern shown is also representative of the two following years. Maximum ice thicknesses for those cases were 1.6 m at CS on 1 June in both 2001 and 2002. For the EL site the corresponding values were 1.54 and 1.6 m.

As in 2000, the ice surface ablation in 2001 began in early June as soon as the snow cover had melted away. Ice-melt rates and the corresponding heat fluxes are shown in Figure 4. The axes on the right side give the net heat flux associated with a given melt rate of ice assuming a density $\rho_{\text{ice}} = 0.9 \text{ Mg m}^{-3}$ and a heat of fusion $L_f = 335 \text{ J g}^{-1}$. The heat flux for 1–10 June is actually underestimated by a factor of 2.3 since the snow rather than ice was melting during that interval. The actual heat fluxes are indicated by the open diamonds. For comparison with a central Arctic case during a vigorous melt season, we consider the results of Perovich and others (2003, fig. 9) for the SHEBA (Surface Heat Budget of the Arctic Ocean program) melt season.

![Fig. 3. Mass balance at the ice sites for the 2000 season. The curves at the top of each panel give the air temperature at 2 m. The water temperature is shown at the bottom. The central panel shows snow thickness (gray) and ice thickness (interface between red and dark-blue contours). The contours show the temperature structure within the ice. The temperature scale is shown at the right.](image-url)
During the peak of the melt season, surface and bottom ablation rates at Barrow were as high as 0.05 and 0.028 m·d⁻¹ respectively in mid-June, corresponding to heat fluxes of 175 and 90 W·m⁻². The corresponding maximum melt rates for SHEBA were slightly less than half as large: 0.024 and 0.012 m·d⁻¹. It is clear that melt rates in the coastal zone were significantly greater than at the site in the central pack. After break-up of the coastal pack, the remaining ice melted away completely, but we were not able to monitor the progress during that period. Melt rates at Barrow during the other years were of the same order of magnitude, although the details differed in response to interannual variations in the synoptic weather patterns.

**Surface albedo**

Observations of total shortwave albedo, \( \alpha \), were carried out using a Kipp & Zonen albedometer as an integral part of the intensive observation program. Measurements were made each year in April to obtain the stable winter values and were resumed in late May through late June to cover the variations during the onset and development of the melt season until the ice became unstable. Observations were taken at 2.5 m intervals along 200 m lines at each site and the results averaged to obtain a representative value for the various sites. A compilation and extended discussion of the full albedo dataset has been presented by Grenfell and Perovich (2004). For the present discussion, we make use of results from 2001 for the sea ice. The seasonal evolution of total albedo for that year is shown in Figure 5. The corresponding results from SHEBA (Perovich and others, 2002) are also plotted to provide a comparison with the central Arctic.

Considerable temporal variability is present due to synoptic weather events as well as hydrographic effects involved in melt-pond equilibrium (Eicken and others, 2004). These show significant inter- and intra-annual fluctuations. In general, with the onset of melt, the albedo at the ice sites decreased steadily as the ice decayed except for two episodes of snowfall as indicated. These events deposited only a few cm of new snow, and the layers remained on the ice for 2–4 days. The final albedo values after the ice had melted are taken to be 0.07 following Pegau and Paulson (2001).

When melt ponds were present, the ice surface was usually inhomogeneous on a small enough scale that the instrument was recording a mixture of ponded and bare ice. In order to determine the albedo of individual areas of bare and ponded ice, we recorded a visual estimate with each observation of the percentage of ponded ice in the field of view. A linear regression of albedo vs this percentage was then extrapolated to 0 and 100% to determine \( \alpha_{bare} \) and \( \alpha_{MP} \), the melt-pond albedo. A description and illustration of the analysis is presented by Grenfell and Perovich (2004, fig. 7b).

Using this method, we found that the albedo for bare coastal ice showed qualitative behavior consistent with previous observations at Barrow (Grenfell and Perovich, 1984) and comparable to those reported for multi-year (MY) ice (Perovich and others, 2002) consistent with the different melt rates as mentioned above. Of particular significance is that the albedo for drained ice was quite stable throughout the melt season. We found \( \alpha_{MY} \sim 0.6 \) for bare FY ice as compared with \( \alpha_{MY} \sim 0.64 \) from the SHEBA data.

**Regional albedo**

The 200 m spatial averages of albedo described above cover a relatively small distance scale but provide well-defined directly measured albedos for the individual surface types. Of interest for large-scale modeling, however, is the short-wave input over regional areas on the order of 100 km² or more. To address this, we extended the present set of surface-based observations, combining the site albedos with areal distributions of surface types determined from high-resolution aircraft photography.

A digital camera was mounted looking vertically downward in a specially outfitted Cessna 185 aircraft operated by Terra-Terpret Inc. We carried out photographic survey flights in May and June at altitudes of 700–1700 m. Strip mosaics were generated from each series of images and analyzed to monitor the relative areas of different surface types. Two representative mosaics are shown in Figure 6. These show the contrast between conditions just prior to the onset of melting and after melt ponds were well established.

In 2001, the low cloud conditions were such that we were able to complete eight flights spaced over the melt season, the data from which were analyzed to provide areal distributions of surface types. These results are shown in Figure 6. The patterns are similar to those reported previously for the central Arctic (Perovich and others, 2002) with the coastal zone showing greater spatial variation than for the central pack. A correlation analysis of the areal distributions of surface type and melt rate is presented by Grenfell and others (1998) in relation to regional weather patterns.
season from late May through 23 June. Each flight covered
the same grid pattern perpendicular to the coast from 3
to 5 km inland out to the edge of the flaw lead, a distance
of 3–10 km depending on the extent of the coastal ice
cover at the time. The flaw lead is a persistent feature of the
local ice pack generated by the differential motion of the
offshore ice relative to the shore-fast ice providing a natural
boundary for the coastal zone ice cover.

The fractional area of each surface type, $A_s$, was deter-
ymined using the image-processing software package ENVI
with the results shown in Figure 7. Individual surface types
were selected objectively in each mosaic from histograms of
the pixel brightness and red–green–blue (RGB) color
distributions of the digital images. The isolation of the
various classes selected was verified by examining three-
dimensional scatter plots to make sure there was no
significant overlap among the clouds of points assigned to
the various surface types.

We isolated and identified the following classes: snow
cover, bare ice, dark bare ice, light melt ponds, dark melt
ponds, light bare lake ice, dark bare or ponded lake ice,
open water and bare tundra. Not all classes were present in
a given mosaic. For the analysis we separated out areas of
sea ice, lake ice and tundra to minimize ambiguity in
classification. For clarity of presentation, we have combined
the light and dark sea-ice and light and dark pond categories
into bare sea ice and melt ponds respectively. The classes of
dark lake ice and ponded lake ice have also been combined
because the occurrence of ponds on the lake ice was rare
and of short duration, and the dark areas typically replaced
the ponds when the surface water percolated into the ice.

Individual values of total albedos, $\alpha_R$, were assigned by
combining the end-point values for each surface type using
the method described above. Assignment of light vs dark
types involved selection of sub-intervals of certain traverses
where the surface was noted as dark, dirty or clean in the
field record. The values used here are given in Table 1. The
uncertainties shown in the bottom row are based on
standard deviations from the 200 m transects. Regional
albedos were then calculated using the equation:

$$\alpha_R = \sum_s A_s \alpha_s,$$

where $s$ spans the number of surface types observed at a
given time for each region considered. The results are
presented in Figure 8 and represent a spatial average over
distance scales on the order of 10 km. All three surfaces had
the same value of $\alpha_R = 0.8$ in winter because they were
covered with an optically thick overlying snow cover. As the
melt season developed, $\alpha_R$ then decreased to individual
lower summer values. The final values for early July assume
that the sea ice and lake ice have disappeared, leaving open
water. We have also assumed that the tundra albedo
remained the same as the last measured value. The results
in Figure 8 are generally consistent with the site averages
shown in Figure 5. Unfortunately, flying conditions were
limited by the persistent low-level cloud conditions,
precluding a more detailed time series.

![Fig. 6](image)

**Fig. 6.** Strip mosaics from the 2001 season before melt onset and during the interval of well-developed melt pond.

![Fig. 7](image)

**Fig. 7.** Surface albedo evolution from aircraft mosaics for sea ice (a), lake ice (b) and tundra (c) during the 2001 season.
DISCUSSION

Model calculations by Maykut and Untersteiner (1971; hereafter MU) indicate that if the summer albedo of the sea ice itself decreases below 0.47 the ice pack will disappear after 2–3 years. While the snow depths we observed decreased each year, they were still close to the climatological value of 0.3 m used in the MU model, so we expect the model to provide a reasonable approximation to the Barrow case. Our value of about 0.4 is significantly below the 0.47 limit, indicating that the decay should be correspondingly more rapid, consistent with the annual break-up of the coastal ice near Point Barrow. While the model does not take into account a variety of dynamical effects, the decay of the ice and the decrease in local ice concentration magnifying the dynamical behavior and associated feedbacks are certainly accelerated by the lower albedo.

Figure 9 shows a comparison of the cumulative amount of shortwave (SW) radiation transmitted into the sea ice at the sea-ice sites with the corresponding regional values. These were calculated using the above-observed albedo sequences in conjunction with total incident SW energy fluxes provided by the US National Oceanic and Atmospheric Administration (NOAA) Climate Monitoring and Diagnostics Laboratory (CMDL) (personal communication from R. Stone, 2002) from the integral \( \int \frac{F_{\text{inc}}}{C_0} \frac{1}{C_1} ds \). Albedo sets were interpolated onto a daily grid, and the above integral was evaluated numerically through 23 June, the date of the last regional albedo analysis. The series for the surface sites was continued through 11 August, and the details are presented in Grenfell and Perovich (2004).

The regional total SW energy input reached 524 MJ m\(^{-2}\) by 23 June vs 561 MJ m\(^{-2}\) for the surface site analysis. The regional totals are slightly below the surface site totals, probably because the albedos were slightly higher away from the coastline; however, the two curves fall within the error bounds of each other. In fact, the cumulative total for the EL site alone was 518 MJ m\(^{-2}\). We conclude that the analysis of the surface sites with a 200 m sampling distance gave a good representation of the regional values. These results do not include the effects of open leads, which were scarce in the present study. Although the influence of leads would need to be taken into account in a more general setting, the albedo is well known for open water (e.g. Pegau and Paulson, 2001), so this would only require data on the fractional coverage of leads in the areas in question.

By 11 August the total for the surface sites reached 1279 MJ m\(^{-2}\). Since 3 MJ m\(^{-2}\) are needed to melt 0.01 m of ice, assuming 100% efficiency, enough energy was absorbed to melt the full thickness of the coastal ice cover by about mid-June and 4.2 m of ice by 11 August, more than 2.5 times the maximum ice thickness of about 1.6 m. This is consistent with the total disappearance of the coastal ice in July. The error bars shown are based on variations among individual sites and represent estimated values of standard deviation. These are not representative of interannual variations, though. In general, inter-site differences depend on the dates of melt onset and ice break-up as well as snowpack accumulation and decay, but in a given year the albedo

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Table 1. Total albedos, \( \alpha_{\text{tot}} \), of the various surface types included in the calculation of regional albedo during 2001

<table>
<thead>
<tr>
<th>Date</th>
<th>Snow cover</th>
<th>Light bare ice</th>
<th>Dark bare ice</th>
<th>Light melt pond</th>
<th>Dark melt pond</th>
<th>Lake light</th>
<th>Lake dark/melt pond</th>
<th>Bare tundra</th>
<th>Open water</th>
</tr>
</thead>
<tbody>
<tr>
<td>29 May</td>
<td>0.80</td>
<td>0.60</td>
<td>0.58</td>
<td>-</td>
<td>-</td>
<td>0.60</td>
<td>-</td>
<td>0.14</td>
<td>0.067</td>
</tr>
<tr>
<td>1 June</td>
<td>0.80</td>
<td>0.60</td>
<td>0.58</td>
<td>-</td>
<td>-</td>
<td>0.60</td>
<td>-</td>
<td>0.14</td>
<td>0.067</td>
</tr>
<tr>
<td>11 June</td>
<td>0.80</td>
<td>0.60</td>
<td>0.58</td>
<td>0.25</td>
<td>0.20</td>
<td>0.60</td>
<td>0.53</td>
<td>0.18</td>
<td>0.067</td>
</tr>
<tr>
<td>13 June</td>
<td>0.80</td>
<td>0.60</td>
<td>0.48</td>
<td>0.20</td>
<td>0.16</td>
<td>0.55</td>
<td>0.46</td>
<td>0.14</td>
<td>0.067</td>
</tr>
<tr>
<td>14 June</td>
<td>0.80</td>
<td>0.59</td>
<td>0.56</td>
<td>0.20</td>
<td>0.16</td>
<td>0.56</td>
<td>0.39</td>
<td>0.15</td>
<td>0.067</td>
</tr>
<tr>
<td>15 June</td>
<td>0.80</td>
<td>0.50</td>
<td>0.50</td>
<td>0.20</td>
<td>0.16</td>
<td>0.56</td>
<td>0.34</td>
<td>0.15</td>
<td>0.067</td>
</tr>
<tr>
<td>22 June</td>
<td>0.80</td>
<td>0.61</td>
<td>0.52</td>
<td>0.27</td>
<td>0.10</td>
<td>0.45</td>
<td>0.32</td>
<td>0.18</td>
<td>0.067</td>
</tr>
<tr>
<td>23 June</td>
<td>0.80</td>
<td>0.58</td>
<td>0.52</td>
<td>0.15</td>
<td>0.09</td>
<td>0.45</td>
<td>0.28</td>
<td>0.18</td>
<td>0.067</td>
</tr>
<tr>
<td>Error estimates</td>
<td>0.01</td>
<td>0.01</td>
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<td>0.02</td>
<td>0.02</td>
<td>0.01</td>
<td>0.02</td>
<td>0.01</td>
<td>0.003</td>
</tr>
</tbody>
</table>
evolutions at the sites considered here are correlated in the sense that they respond to the same synoptic weather conditions. Intennarkal variations can be significantly larger (Grenfell and Perovich, 2004).

We note that a physics-based model for the evolution of melt ponds taking into account the melt rate, surface geometry, meltwater percolation and runoff has been developed incorporating data from the present field experiment (Eicken and others, 2004). While a number of questions remain, this work provides the formulation of the problem and the basis for future development.

An effect of probable climatological importance raised by Grenfell and others (2003) with regard to melt-pond modeling in a warming environment is the spatial redistribution of melt ponds following early melt onset followed by a significant freeze-up associated with increasing instability in weather conditions. This effect is likely to accompany a warming trend, as typified by the 2002 melt season that began 2 weeks earlier than expected. In this case, the melt ponds refroze completely, essentially becoming lake ice, and the spatial distribution of the ponds was significantly modified when melting resumed.

The lower albedo of bare FY ice compared to MY ice is due in large part to the presence of suspended particulate material (SPM) in the FY ice blown out from the shore and/or suspended in the water at the time of freeze-up. Pigmented organic material can also produce a significant influence. We expect that this will be an important consideration throughout the entire Arctic coastal zone.

The complete dataset for this experiment has been compiled on DVD disks. It is available from the first author upon request.

CONCLUSIONS

A comprehensive field study was carried out near Barrow to characterize local-to intermediate-scale sea-ice processes in the Arctic coastal zone of central importance to the annual cycle and evolution of the coastal sea ice. This period spans an interval of significant warming and increasing retreat of the ice pack. Maximum snow depths decreased from 2000 to 2002 from 0.38 m to 0.26 m but they are still within the range of the local climatology. Maximum ice thickness was in the interval of significant warming and increasing retreat of the Arctic coastal zone of central importance to the annual characterizations of melt-pond processes in the Arctic coastal zone relative to the central pack show that the

early melt onset of 2002 (14 May) presented a modified situation that is not described by a simple temporal extrapolation of the melt-season length. This involved an instability in the weather associated with warming that modified the spatial distribution of the ponds in an unexpected way. Observations of the type presented here are critical for understanding sub-gridscale processes associated with satellite observations and large-scale heat and mass-balance modeling.

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