LARGE–SCALE PATTERNS OF SNOW MELT ON ARCTIC SEA ICE
MAPPED FROM METEOROLOGICAL SATELLITE IMAGERY

by

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ABSTRACT

The seasonal progression of snow melt on the Arctic pack ice is mapped from satellite shortwave imagery (0.4-1.1 micrometers) for four spring/summer seasons (1977, 1979, 1984 and 1985). This provides the first detailed information on the temporal change of the ice surface albedo in summer and of its year-to-year variability. The average surface albedo of the Arctic Basin for the years investigated falls from between 0.75 and 0.80 in early May to between 0.35 and 0.45 in late July and early August. In the central Arctic, where ice concentration remains high and ponding on the ice is limited, the July albedo ranges from 0.50 to 0.60. Overall, melt progresses poleward from the Kara and Barents Seas and from the southern Beaufort and Chukchi Seas, with the melt fronts meeting on the American side of the Pole. There are substantial year-to-year differences in the timing, duration and extent of the melt interval. The progression of melt in May and June of the earliest melt year (1977) was about 3 weeks ahead of the latest year (1979). By late July, the central Arctic was essentially snow free in 1977 and 1979, but more than 50% snow covered in 1984. Although limited in extent, our data base suggests relationships between snow melt and Arctic surface air temperatures in spring, spring cloudiness and the extent of late summer ice.

INTRODUCTION

The timing, duration and extent of snow melt on the pack ice has long been recognized as a critical variable influencing the summer climatic regime in the Arctic Basin, with potential impacts on other parts of the Northern Hemisphere (Fletcher 1966). The snow melt has implications for the long-term mass balance and stability of the sea ice and may serve as an indicator of CO_2-induced climatic change (Barry 1985).

Up to now, direct information on melt and the resultant changes in surface reflectivity (albedo) in the basin has been limited to measurements taken at drifting stations or on fast ice and to a few aircraft programs (e.g. Laktionov 1953, Untersteiner 1961; Langleben 1971; Bryazgin and Koptev 1970; Weaver and others 1976, Holt and Digby 1985). Others have used these data to estimate regionally averaged summer albedo (e.g. Larsson and Orvig 1962; Marshunova and Chernigovskiy 1966; Posey and Clapp 1964; Hummel and Reck 1979, Robock 1980; Kukla and Robinson 1980). Estimates of July surface albedo in the central Arctic range from 0.40 to 0.65.   More recently, microwave satellite data have been used to identify the earliest phase of snow melt onset (Anderson and others, 1985) and also to estimate sea ice concentration and, indirectly, Arctic surface albedo (Carsey 1985). In the latter study, the areal coverage of bare ice was deduced from the data and the proportion of melt ponds and leads was estimated from published reports. Parameterized albedos were then assigned to each surface type and weighted according to their coverage. Central Arctic albedo in mid-July 1974 was estimated to be 0.58 with values about 0.10 lower around the margins of the basin.

In the present study, we have used operational daily meteorological satellite imagery to map the changes of surface brightness and texture associated with four spring/ summer seasons in the Arctic Basin (Figure 1). Parameterized albedos have been assigned to different brightness classes by analyzing satellite data on an image processor.

DATA AND METHODS

Imagery from the Defense Meteorological Satellite Program (DMSP) near-polar orbiter served as the primary data source for assessing surface conditions. Shortwave (0.4- 1.1 micrometers) images with resolutions of 0.6 km for direct read-out products and 2.7 km for orbital swath
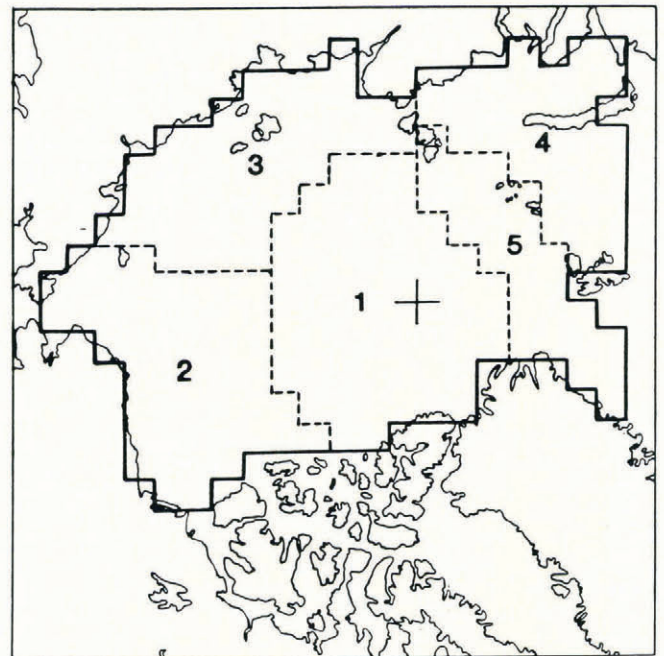

Fig.1.   Arctic Basin study zone (heavy line) divided into five regions (dashed lines): 1) Central Arctic, 2) Beaufort/Chukchi Seas, 3) East Siberian/Laptev Seas, 4) Kara/Barents Seas and 5) Northwest North Atlantic.

format images were used. Supplemental data included NOAA Very High Resolution Radiometer (VHRR) and Advanced VHRR 1.1 km resolution imagery.

Snow melt on the ice is recognized in the shortwave images by a characteristic decrease in brightness and a change in surface texture. As the snow dissipates, melt ponds form and bare ice is exposed (Barry 1983) (Figure 2). Landfast ice and first-year pack ice brighten somewhat when melt ponds drain (Zubov 1945; Lapp 1982; Holt and Digby 1985). Comparisons of DMSP imagery with synchronous 80 m resolution Landsat imagery confirm the recognition of surface features in the lower resolution products. The comparisons also show that, in most cases, brightness and textural changes due to melt processes on the ice may be successfully distinguished from variations in ice concentration. This distinction was most difficult where U.S. Navy/NOAA ice charts (Godin 1981) depict ice concentrations of less than 75%; less than 10% of the basin during most of the melt season (Figure 3).

Basin-wide maps of surface brightness and texture were constructed manually in three-day increments from May through August 1977, 1979, 1984, and 1985. Repetitive coverage and characteristic textures permitted the differentiation of moving clouds from the surface (Robinson and others 1985). Interactive image processor analyses of selected scenes showed good agreement with visual classification of surface brightness. Maps were digitized using the National Meteorological Center standard grid which divides the basin into 212 cells. At least one cloud-free scene per 3-day interval was available over more than 80% of the basin from May to mid-August. Missing cells were either: (1) assigned observed brightness values from an immediately preceding or subsequent chart, (2) considered to be open water, if shown as such on the Navy/NOAA weekly ice chart closest to the analyzed interval, or (3) handled as missing data (less than 10%, on average).

Four ice-surface classes are identified. The literature on ground and aerial observations, referred to earlier, shows that these classes represent: class (1) fresh snow cover over 95% of the ice; class (2) snow covers between 50-95% of the surface, with the remainder being bare or ponded ice (in spring this is considered the initial stage of active snow melt); class (3) the final stage of active snow melt, with between 10-50% of the ice surface snow covered and with numerous melt ponds, or, following pond drainage, predominantly bare ice, with snow patches and scattered ponds; and class (4) heavily ponded or flooded ice with less than about 10% snow cover, or exposed bare ice.

Local observations indicate that when class 3 is first identified the surface is in the final stage of active melt and remains so for approximately the next 2-3 weeks. At later dates, when class 3 is charted, the surface is predominantly drained bare ice (Kuznetsov and Timerev 1973; Hanson 1980). Classes 2 and 3 are illustrated for the Beaufort Sea in Figure 2. Class 4 is most commonly found in regions of fast ice. The absence of the numerous flaws and leads in the fast ice and its proximity to surface run-off from land results in more extensive flooding of this ice than over pack ice.

Large-scale surface albedo values adopted for the charted classes are: class (1) 0.80; class (2) 0.64; class (3) 0.49; and class (4) 0.29, with standard deviations between 0.08 (class 3) and 0.05 (class 4). These values were adjusted for average summer cloudiness (Robinson and others 1985) by ±0.05 for brightness classes 1-3 and ±0.02 for class 4, based on data from Buzuyev and others (1965), Kuznetsov and Timerev (1973) and Grenfell and Perovich (1984). Albedo is also decreased to account for the presence of open water within the pack (Cogley, 1979). For class 1, the correction ranges from -0.17 when the ice concentration is 75%, to -0.01 when the ice concentration is 99%. Ice concentrations were based on U.S. Navy/NOAA weekly ice charts.

Class albedos were calculated by measuring the brightness of selected clear-sky scenes on a digital image processor. Linear interpolation was made between homogeneous bright snow on multiyear ice and dark open-water. Clear-sky albedos of these targets were estimated from measured ground and aerial data (e.g. Hanson 1961, Nazintsev 1964, Langleben 1971, Bryazgin and Koptev 1970,
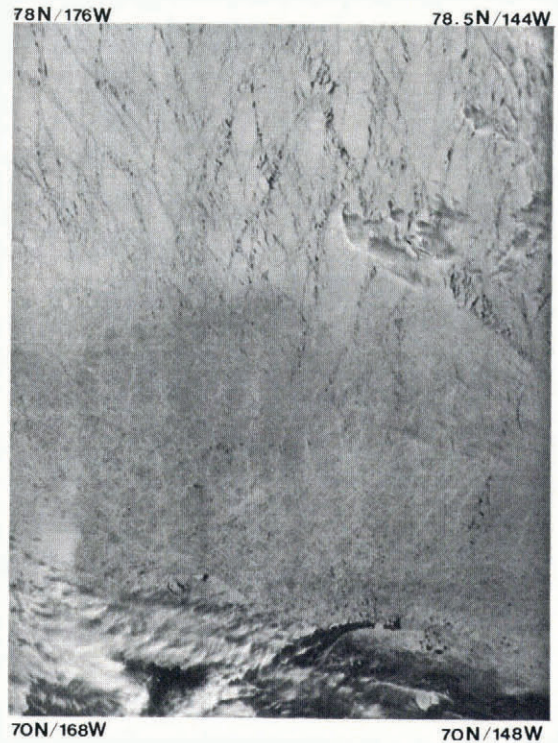

78N/176W — 78.5N/144W

70N/168W — 70N/148W

Fig.2. High-resolution (0.6 km) DMSP shortwave image of the Beaufort Sea on June 17, 1978 showing stages of snow melt (classes 2 & 3) on the sea ice.

Payne 1972, Grenfell and Maykut 1977, Pautzke and Hornof 1978, Cogley 1979). The snow-covered targets were assigned an albedo of 0.80 until late June, after which a value of 0.70 was used. The open-water albedo was taken as 0.12. Specular reflectance is minimal over these surfaces at relatively high solar zenith angles during the Arctic summer and at the satellite viewing time and angle (Taylor and Stowe 1984). This procedure has been used in other studies (e.g. Preuss and Geleyn 1980; Robinson and Kukla 1985) and found to be particularly well suited for use with the broad-band DMSP imagery (Shine and Henderson-Sellers 1984).

RESULTS

Our maps show that in all four years melt began in the Barents and Kara Seas and the southern Beaufort and Chukchi Seas, then progressed along the Arctic coast of Asia and towards the American side of the Pole. In 1977, active melt (classes 2 and 3) covered 50% of the basin by the end of May (Figure 3A), while over half of the basin was categorized as class 3 or 4 by the end of June. Melt began almost 3 weeks later in 1979 (Figure 3B), yet, as in 1977, more than half of the basin was classified as class 3 or 4 by the end of June. In 1984, melt began prior to May 1 in the southern Beaufort and Chukchi Seas (Figure 3C). However, it was not as extensive in May as it was in 1977. In 1984, areas with concentrated narrow leads in seas bordering the Asian continent resulted in a large-scale surface brightness equivalent to unbroken ice undergoing partial melt (class 2). This gives the appearance in Figure 3C of quite extensive basin snow melt early in May. However, an examination of high-resolution DMSP and AVHRR images indicates that while melt did begin quite early in the southern Beaufort and Chukchi Seas and portions of the Kara and Barents Seas in 1984, it did not begin over more than 10-15% of the basin until late May. Over half of the basin was classified as classes 3 and 4 by the end of June 1984. Melt began at about the same time in 1985 (Figure 3D) as in 1979, with about 25% of the basin categorized as class 2 at the end of May. Approximately 50% of the basin was classified as classes 3 and 4 by the end of June.
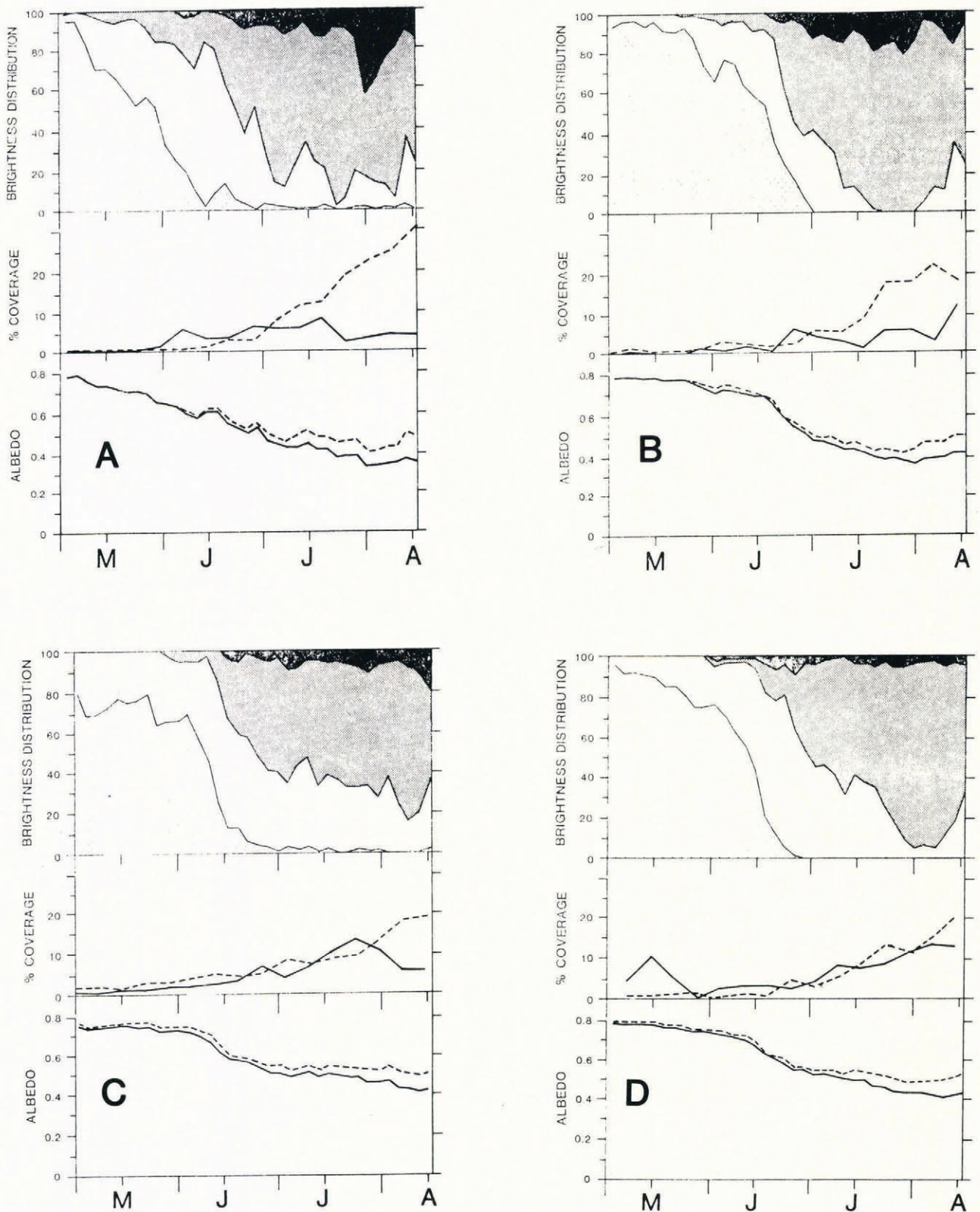
Fig.3. A. top: Progression of snow melt and subsequent ponding and drainage on Arctic sea ice from May to mid-August 1977, as shown by the changing distribution of brightness classes (cf. text for class descriptions). Classes shaded from light grey (class 1) to dark grey (class 4). Areas with less than 12% ice concentration or open water omitted.

middle: Percentage of the basin with open water or with less than a 12% (1/8) concentration of ice (dashed line) and percentage of the basin with a 12%-75% (1-6/8) ice concentration (solid), according to 1977 weekly Navy/NOAA ice charts.

bottom: Basin-wide albedo (including sea ice and open water) from May-August 1977 (solid line). Albedo of ice (sea water excluded) in areas with at least 75% ice concentration (dashed).

B. Same as 3A, except for 1979.

C. Same as 3A, except for 1984 and ice concentrations of 10% (1/10) and 80% (8/10) where marked as 12% and 65% in A.

D. Same as 3A, except for 1985 and ice concentration of 10% and 80%.

All sectors of the basin showed signs of melt by the beginning of July, although there were considerable year-to-year variations in melt intensity during this month. In 1977, a mid-month snowfall in the central Arctic temporarily brightened the surface and increased the albedo. Melt advanced steadily in July 1979, leaving the area essentially snowfree by the last week of the month. The central Arctic continued to have a considerable snow cover throughout July 1984, as was the case until the last week of the month in 1985.

Soon after the minimum albedo was reached in late July or early August of each year, fresh snow began once again to cover the central part of the basin. The newly snow-covered areas did not appear as bright on the imagery as full spring cover, probably due to the presence of open leads and undrained ponds. Out of the four years, snow cover was most extensive by mid-August 1985. Heavy cloudiness allowed only sporadic views of the basin surface in the latter half of all Augusts, when it appeared that fresh snow increasingly covered the basin.

Due to the earlier melt in 1977 and the combination of melt and lower ice concentration in 1984, basin albedo averaged 0.73 in May 1977 and 1984, compared with 0.77 in 1979 and 0.76 in 1985 (Table I). Basin albedo was highest in June 1979 (0.66), and lowest in June 1977 (0.58).

TABLE I. AVERAGE SURFACE ALBEDO OVER THE ARCTIC BASIN (FOR EACH MONTH IN THE FOUR STUDY YEARS)

	1977	1979	1984	1985
May	0.73	0.77	0.73	0.76
June	0.58	0.66	0.61	0.65
July	0.43	0.44	0.48	0.48
August*	0.36	0.40	0.42	0.42

* for the period August 1-17

By July, basin albedo dropped to 0.43 in 1977, 0.44 in 1979, and 0.48 in 1984 and 1985. The first half of August averaged 0.36 in 1977, and between 0.40 and 0.42 in the other years, the major difference in 1977 being the relatively large amount of open water in the basin (Figure 3). The late July and early August albedo of the ice surface (sea water excluded) in those parts of the basin where ice concentration exceeded 75% was approximately 0.45 in 1977 and 1979 and 0.50 in 1984 and 1985 (Figure 3).

A regional breakdown (Figure 1) of surface albedo shows that the early advance of extensive snow melt in 1977 compared with the other three years was most pronounced in the Central Arctic (Table II). Later in the summer the albedo differences between years in the coastal seas were a result of differences in ice extent. This is particularly evident in the East Siberian and Laptev Seas, where the early August albedo ranged between 0.25 (1977) and 0.47 (1984).
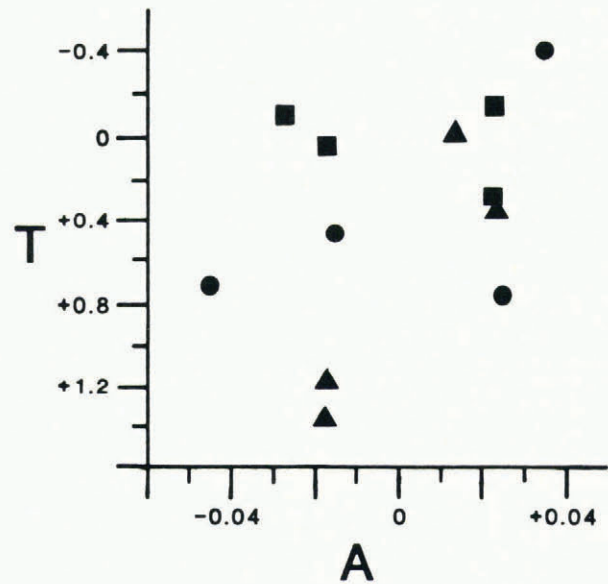

Fig.4. Comparison of monthly anomalies of surface albedo over the Arctic Basin and surface air temperature from 65-85°N for May (triangles), June (circles) and July (squares) of the four study years. Albedo anomalies from the average of the four years. Temperature anomalies based on 1951-1970 means (Jones 1985; Climate Monitor 1985a, 1985b).

TABLE II. AVERAGE SURFACE ALBEDO OVER REGIONS OF THE ARCTIC BASIN (FOR EACH MONTH IN THE FOUR STUDY YEARS)

		Central Arctic	Beaufort/ Chukchi	E. Siberian/ Laptev	Kara/ Barents	NW North Atlantic
May	1977	0.78	0.74	0.73	0.66	0.72
	1979	0.79	0.76	0.79	0.77	0.73
	1984	0.79	0.75	0.74	0.63	0.70
	1985	0.79	0.77	0.79	0.69	0.74
June	1977	0.66	0.58	0.55	0.48	0.60
	1979	0.76	0.62	0.67	0.58	0.62
	1984	0.69	0.59	0.63	0.51	0.59
	1985	0.73	0.65	0.62	0.54	0.62
July	1977	0.52	0.42	0.40	0.27	0.46
	1979	0.53	0.41	0.42	0.32	0.45
	1984	0.58	0.45	0.49	0.30	0.50
	1985	0.59	0.47	0.48	0.31	0.47
August*	1977	0.50	0.33	0.25	0.19	0.44
	1979	0.53	0.37	0.39	0.22	0.43
	1984	0.54	0.42	0.47	0.15	0.43
	1985	0.54	0.41	0.45	0.17	0.44

* for the period August 1-17

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In the central Arctic, where ice concentration generally remains above 95% throughout the summer, the July albedo was 0.52 in 1977, 0.53 in 1979, 0.58 in 1984 and 0.59 in 1985. These satellite-derived albedos are within the range of published estimates (e.g. Hanson 1961, Larsson and Orvig 1962; Posey and Clapp 1964; Marshunova and Chernigovskiy 1966; Schutz and Gates 1972, Hummel and Reck 1979; Robock 1980; Kukla and Robinson 1980; Barry 1983; Carsey 1985), which vary from 0.40 to 0.65. The large range of the estimates reflects the high year-to-year variability in the extent of snow melt, degree of ponding and frequency of summer snowfalls. Differing approaches and limited data bases available to earlier researchers may also be a factor.

Towards the middle of August, when the extent of fresh snow cover in the central Arctic began to increase, albedo rose by 0.05 to 0.10 in all years; however, our satellite-derived albedo in the first half of August is lower than that reported earlier from drifting stations (Nazintsev 1964, Pautzke and Hornof 1978). This may be the result of our large-scale averaging, as opposed to the ground observations made mostly on thick multi-year ice surrounding drifting stations. Alternatively, the August snowfalls may have come relatively late or been less extensive in the four years of our study.

DISCUSSION

The geographic progression of seasonal melt identified in our maps was compared with earlier studies. The climatological summary of Marshunova and Chernigovskiy (1978) shows a concentric pattern of melt over the Arctic pack ice progressing towards the Pole by early July. From a study of sequential ESMR data for 1974, Campbell and others (1980) suggested that melt began along the Siberian coast in May and moved as a roughly linear front across the Pole, reaching the Canadian and Greenland coasts about a month later. Re-examination of their data by Crane and others (1982), indicated that the color coding of the microwave imagery created a partially spurious impression of a progressing melt front.

Climate models suggest that an early loss of snow cover may have an impact on ice extent later in the summer (Maykut and Untersteiner 1971; Semtner 1976). In 1977, when the snow melt and ponding came early, the subsequent late summer ice extent was considerably reduced compared to ice cover in the other years. This relationship was most evident in the Beaufort and Chukchi Seas and the East Siberian and Laptev Seas regions. Whether such a dependence was coincidental or not must remain speculative until more years are examined.

Cloud cover, mapped at approximately three-day intervals in the summers of 1977 and 1979, showed a late May–early June maximum in extent (averaging approximately 90%) and thickness over the basin, followed by a period of less extensive (approximately 75%) and somewhat thinner cover extending into early August (Robinson and others 1985). This suggests that early stages of surface melt may be related to the early season cloudiness, which appears to be associated with the poleward retreat of the Arctic front and advection of air from lower latitudes by synoptic disturbances (Barry and others 1986). Early melt may also be enhanced by the increase in infrared radiation at the surface due to the cloud cover. This results in an increase in surface net radiation over high–albedo Arctic surfaces (e.g. Ambach 1974).

The relationship between snow melt onset and large-scale Arctic surface air temperatures was examined using monthly temperature and albedo anomalies (Figure 4). Temperature anomalies are based on 1951-1970 means for 65-85°N (Jones 1985; Climate Monitor 1985[a] and 1985[b]). Albedo anomalies are calculated from the average of the four study years (cf. Table I). A relationship between positive anomalies of temperature and negative anomalies of basin-wide surface albedo is apparent in May and three of the Junes studied, but is not evident in July. This suggests that the temperature data are not representative of the inner Arctic Basin, but are related to melt in the coastal seas located near the reporting stations.

CONCLUSION

This work provides the first direct evidence of the fluctuations of snow cover and surface albedo across the entire Arctic Ocean in spring and summer. The timing of snow melt on the Arctic sea ice is shown to vary greatly from region to region as well as from year to year. This variation will have a significant impact on the heat and mass balance of the Arctic, since the surface albedo and the thermodynamics of the ice are strongly related to the presence of snow. Improved knowledge of the spring and summer surface albedo in the Arctic is important in climate models and may also help in recognizing the initial signs of any climatic changes induced by CO_2 and other trace gases.

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