An Improved Method for Determining Snowmelt Onset Dates over Arctic Sea Ice Using Scanning Multichannel Microwave Radiometer and Special Sensor Microwave/Imager Data

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An improved method for determining snowmelt onset dates over Arctic sea ice using scanning multichannel microwave radiometer and Special Sensor Microwave/Imager data

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Abstract. Ablation of snow over sea ice is an important physical process affecting the Arctic surface energy balance. An improved understanding of the spatial and temporal variations in snowmelt onset could be used to improve climate simulations in the Arctic, as well as monitor the Arctic for signs of climate change. Utilizing an updated approach for monitoring snowmelt onset over Arctic sea ice, spatial variability in passive microwave derived snowmelt onset dates is examined from 1979 through 1998. The improved technique, termed the advanced horizontal range algorithm (AHRA), utilizes temporal variations in 18/19 GHz and 37 GHz passive microwave horizontal brightness temperatures obtained from the scanning multichannel microwave radiometer (SMMR) and the Special Sensor Microwave/Imager (SSM/I) to identify snowmelt onset. A qualitative assessment of spatial variability in snowmelt onset discusses the 1979 through 1998 mean snowmelt onset pattern, and it also illustrates that there are significant variations in snowmelt onset on an annual basis. Principal component analysis of the snowmelt onset dates suggests snowmelt onset variability is dominated by a zone of abnormally early (late) snowmelt onset near the Siberian coast and another zone of abnormally late (early) snowmelt onset near Baffin Bay. Statistical analysis between the first principal component and March-June monthly averaged Arctic Oscillation values implies that variations in snowmelt onset are related to alterations in the phase of the spring Arctic Oscillation.

1. Introduction

Snow plays an integral role in defining the Arctic's surface energy balance by influencing sensible and latent heat exchanges as well as the surface albedo. For instance, wet snow, with a relatively low albedo, absorbs approximately 45% more solar radiation than dry snow [Abdalati and Steffen, 1995]. The transition from a highly reflective dry snow cover to an absorptive wet snow cover also plays an important role in the temperature-albedo feedback mechanism [e.g., Curry et al., 1995]; melting snow decreases the surface albedo, leading to additional energy absorption, warmer air temperatures, and consequently additional snowmelt. Partly on the basis of this feedback mechanism, general circulation models typically predict climate warming in the Arctic region will be enhanced under doubled CO₂ scenarios [e.g., Gates et al., 1999; Maxwell et al., 1998]. An accurate method of monitoring snowmelt onset could therefore be used to validate and improve climate simulations during the Arctic snowmelt period. An improved understanding of the spatial and temporal variability in snowmelt onset could also be used as a proxy climate indicator.

The main objective of this paper is therefore to describe an improved technique for monitoring snowmelt onset dates over Arctic sea ice, based on the approach of Anderson [1997]. Satellite passive microwave data are used to derive the snowmelt onset dates because 1) microwave emission is sensitive to liquid water inclusions in the snowpack, 2) data collection is not significantly influenced by cloud cover and does not require solar illumination, and 3) a fairly continuous satellite passive microwave data set is available from 1979 through 1998 (from 1979 through August 1987, data were collected every second day, while from August 1987 through 1998, data were collected every day). For the purposes of this study, "snowmelt onset" is defined as the point when microwave brightness temperatures increase sharply due to the development of liquid water in the snowpack. It coincides with the "early melt" season discussed by Livingstone et al. [1987], and it is important to note that melt-freeze cycles may occur after the snowmelt onset date.

In addition to discussing improvements in the Anderson [1997] algorithm, this paper qualitatively discusses interannual variability in snowmelt onset by examining annual snowmelt onset distributions, as well as mean, range, and standard deviations maps of snowmelt onset. Furthermore, principal component analysis is used to quantitatively examine spatial variability in snowmelt onset, while correlation analysis is utilized to illustrate the Arctic Oscillation (AO) may exert a significant influence over snowmelt onset in some areas.

2. Background

The theoretical concepts underlying the use of microwave radiometry for snowmelt onset detection are well developed [e.g., Carney, 1992]. The acquired microwave Tₛ at a given frequency and polarization is a function of the surface emissivity at the given frequency and polarization as well as the physical temperature of the emitting medium [Ulaby et al., 1986]. During
winter, emissivities of dry first-year sea ice are approximately 0.888 (19H) and 0.913 (37H), while emissivities of dry multiyear sea ice are approximately 0.780 (19H) and 0.706 (37H) [Eppler et al., 1992]. Emissivities over first-year ice show remarkable consistency [e.g., Grenfell et al., 1988], with variations primarily due to scattering from the overlying snow cover [Eppler et al., 1992]. The emissivities over multiyear ice tend to have a wider range due to the varied histories of the floes. The overall lower emissivities in multiyear ice are related to the highly porous consistency [e.g., Grenfell et al., 1988], with variations primarily due to scattering from the overlying snow cover [Eppler et al., 1992].

As air temperatures increase in spring, liquid water forms between the snow grains [Colbeck, 1986], the snowpack becomes a source of microwave emission, and the resulting increase in surface emissivities (to 0.890 for 19H and 0.933 for 37H) and $T_B$ values can be used to detect snowmelt onset. For example, Campbell et al. [1984] demonstrated melt detection with single-channel Electrically Scanning Microwave Radiometer (ESMR) data, while Anderson et al. [1985], Anderson [1987], and Serreze et al. [1993] utilized single-channel scanning multichannel microwave radiometer (SSMR) data to detect snowmelt onset over Arctic sea ice surfaces.

The increase in $T_B$ values due to liquid water formation is, however, related to the frequency, such that the difference in $T_B$ values between 18 GHz and 37 GHz changes from positive in winter to zero or negative at snowmelt onset [e.g., Kunzi et al., 1982]. By monitoring daily variations in the difference between 18 GHz and 37 GHz horizontal polarization $T_B$ values, Anderson [1997] formulated an algorithm for detecting snowmelt onset over first-year and multiyear sea ice surfaces with SMMR and Special Sensor Microwave Imager (SSMI/I) data (note that SSM/I data are from 19 GHz and 37 GHz). Termed the horizontal range (HR) algorithm, snowmelt onset is calculated as the date when the difference between 18/19 GHz and 37 GHz horizontal polarization is less than or equal to 2 K. An examination of HR-derived snowmelt onset dates showed large spatial and temporal variability [Anderson, 1997; Anderson and Drobot, 1999], which is partly related to low-frequency atmospheric circulation [Drobot and Anderson, 1999].

Several other studies used passive microwave data to detect snowmelt onset over Arctic sea ice, including Comiso and Kwok [1996], who showed latitudinally averaged SSM/I data could be used to detect melt. However, they did not attempt a systematic mapping of the spatial and temporal variability with their data. Parkinson [1992] computed snowmelt onset as part of an effort to monitor the summer sea ice season, defined as the number of days in which the ice concentration was greater than a threshold value in a particular pixel. Although useful for marginal sea ice areas, this approach is not effective over areas where sea ice concentrations remain high throughout the year, such as the central Arctic basin. Snowmelt onset dates were also determined by Smith [1998], using SMMR and SSM/I 18/19 GHz and 37 GHz vertical polarization data. However, the Smith [1998] algorithm requires a pixel to have at least 40% multiyear ice, and it is therefore unable to determine melt onset over regions predominantly covered by first-year ice. The inability to monitor first-year ice regions has important ramifications for climate studies because first-year ice accounts for a significant portion of the Arctic Ocean ice cover [e.g., Rothrock and Thomas, 1990].

Active microwave data were used by Winebrenner et al. [1994] and Yackel et al. [2001] to detect snowmelt onset over Arctic sea ice surfaces. These approaches are complementary to the present study, because the high spatial resolution obtained by satellites such as RADARSAT could be valuable for validating and improving climate simulations during the Arctic spring. However, the relatively short time period of data collection (since 1995 for RADARSAT and since 1991 for the European Remote Sensing (ERS) series) limits their use as a proxy climate indicator.

3. Data

Snowmelt onset dates are derived with 25 km$^2$ daily-averaged brightness temperature data from the SMMR and three SSM/I radiometers (F8, F11, and F13), for the years 1979 through 1998. The data were obtained from the National Snow and Ice Data Center (www.nsidc.org) in Boulder, Colorado, on the polar stereographic projection. Since brightness temperature data originated from four different radiometers, all were converted to be consistent with the F8 SSM/I data using regression analysis obtained during overlap periods. Brightness temperatures at 18 and 37 GHz horizontal polarizations from SMMR were converted to F8 data using slope and intercept values provided by Jezek et al. [1991], while brightness temperatures at 19 and 37 GHz from the F11 SSM/I were converted to F8 with values from Abdalati et al. [1995]. Brightness temperatures from the F13 SSM/I were first converted to F11 using values from Stroeve et al. [1998], and then converted from F11 to F8 with values from Abdalati et al. [1995]. The conversion of brightness temperatures to a consistent data record is especially important in determining temporal trends in the snowmelt onset dates. If the data are not consistent, it is possible temporal trends in the snowmelt onset dates could be attributed to instrument differences, rather than real world conditions.

To eliminate land-to-ocean spillover (as discussed by Cavalieri et al. [1999]), a two-pixel buffer zone was applied to the land mask included with F13 data CDs. An ocean mask was also created to eliminate pixels affected by variations in the annual melt ice extent and weather effects. The ocean mask assigns melt to a pixel only if the ice concentration in a particular pixel was greater than 50% in winter for all years from 1979 through 1998.

Near-surface air temperatures from the International Arctic Buoy Programme/Polar Exchange at the Sea Surface (IABP/POLES) data set were used to assist in determining whether the improved algorithm is correctly determining the melt onset date. The IABP/POLES temperature data are a combination of Argos buoy data collected by the International Arctic Buoy Programme, North Pole manned drifting station data obtained from the Arctic and Antarctic Research Institute in St. Petersburg, Russia, and meteorological land station data obtained from the National Center for Atmospheric Research (NCAR). More information on the POLES data set was given by Rigor et al. [2000].

To examine whether variations in the atmosphere influence snowmelt onset dates, monthly averaged AO values were acquired from the Joint Institute for the Study of Atmosphere and Ocean (JISAO) at the University of Washington. Statistical associations between the AO and snowmelt onset were analyzed only for the months from March through June, since most snowmelt occurs during this period [e.g., Anderson, 1997; Smith, 1998].

4. Computation of Snowmelt Onset Dates Over Arctic Sea Ice

Although theoretical concepts suggest melt onset should be easily monitored by a large change in microwave brightness temperatures, this research indicates temporal information can improve derivation of snowmelt onset dates. Recall that with the HR algorithm [Anderson, 1987] melt onset is signaled as the first date where the difference between 18 GHz (19 GHz on SSM/I) and 37 GHz brightness temperatures are 2 K or less. Owing to
spatial and temporal averaging of the orbital data, sensor errors, and possibly mixed ice types, spurious snowmelt onset signatures are occasionally derived with the HR on days when the air temperature is clearly too low for snowmelt onset to occur (e.g., day 80 in Figure 1). In these cases the daily progression of differences in 19 GHz and 37 GHz brightness temperatures often returns to levels similar to those found prior to day 80 because there have not been any major physical changes in the snowpack. In contrast, when melt onset is more likely to occur (day 121), the time series of differences between 19 GHz and 37 GHz brightness temperatures changes rapidly from day to day after the melt date. At this point (day 121), the snowpack becomes a source of microwave emission, and continuous metamorphism in the snow grains likely leads to large changes in the observed microwave signatures from one day to the next. Therefore empirical evidence suggests that the temporal pattern in 18 GHz (19 GHz on SSM/I) and 37 GHz brightness temperatures contains information that is useful to melt detection.

By using the theoretical basis that the 18 GHz (19 GHz on SSM/I) and 37 GHz brightness temperatures should merge at melt onset, and the empirical findings that the temporal pattern in the differences between 18 GHz (19 GHz on SSM/I) and 37 GHz brightness temperatures changes after melt onset, an improved version of the HR algorithm, termed the advanced horizontal range algorithm (AHRA), is utilized to determine melt onset. The AHRA exploits changes in the difference between 18 GHz (19 GHz on SSM/I) and 37 GHz brightness temperatures over a 20-day period to derive snowmelt onset dates over Arctic sea ice from 1979 through 1998 (Figure 2). The AHRA tracks the 18/19 GHz horizontal polarization $T_B$ minus the 37 GHz horizontal polarization $T_B$ for each data point on each day. If the difference is greater than 4 K at a given point, winter conditions are assumed to exist, and the algorithm continues onto the next day for that point. Conversely, if the difference is -10 K or less (i.e., the 37 GHz horizontal polarization brightness temperature is 10 K or more greater than the 18/19 horizontal polarization brightness temperature), then liquid water is assumed to be present in the snowpack, and the AHRA classifies that day as the snowmelt onset date. In the intermediate phase, when the difference is less than 4 K and greater than -10 K, the AHRA determines if snowmelt onset occurred based upon a 20-day time series analysis of brightness temperatures. A difference is computed by subtracting the minimum $T_B(19H)-T_B(37H)$ from the maximum $T_B(19H)-T_B(37H)$ for the 10 days prior to the potential melt onset date, as well as for the period from the potential melt onset date to 9 days later. The former number is subtracted from the latter number, and if this difference is greater than 7.5 K, the algorithm assigns melt to that particular pixel. A large difference indicates the pattern in the time series of $T_B(19H)-T_B(37H)$ has changed, meaning much larger variability in the $T_B(19H)-T_B(37H)$ range is noticed after the potential melt onset date. As discussed above, the appearance of liquid water in the snowpack is a likely mechanism to cause this type of change. If the difference remains below 7.5 K, then it is unlikely that liquid water is present in the snowpack, and the algorithm moves on to the next day. The original 2 K limit of the AHRA was increased to 4 K for this work, and a bottom limit of -10 K was applied after studying examples from the case study sites displayed in Figure 3. The 7.5 K threshold was deemed superior to 2.5, 5.0, 10.0, 12.5, 15.0, and 20.0 on the basis of these case study sites as well.

Returning to the first example (Figure 1) to demonstrate melt derivation with the AHRA, the 19H-37H difference on day 80 is 1.2 K, so the AHRA will test over the 20-day window. From day 70 through day 79, the max-min $T_B$ difference is $6.6 - 4.4 = 2.2$ K, and the max-min $T_B$ difference from day 80 through day 89 is $6.0 - 1.8 = 4.2$ K. The after-before difference for day 80 is therefore $4.2 - 2.6 = 1.6$ K, which falls below the 7.5 K threshold, so melt onset is not derived. On day 121 the relevant differences are $6.0 - 4.0 = 2.0$ K for days 111 through 120 and $29.0 - (-7.6) = 36.6$ K for days 121 through 130. The after-
Figure 3. Case study sites used to develop the advanced horizontal range algorithm.

before difference for day 121 is therefore $36.6 - 2.0 = 34.6$ K, which is greater than the 7.5 K threshold, and melt onset is determined as day 121 with the AHRA.

To further verify melt derivation with the AHRA, 15 case study sites (Figure 3) of the 19 GHz-37 GHz $T_b$ values and IABP/POLES air temperatures are illustrated from 1997 (Figure 4) for first-year and multiyear sea ice. Over the multiyear ice sites, the 19 GHz-37 GHz differences generally remain consistently positive in the winter months, corresponding to 19 GHz emissivities being greater than 37 GHz emissivities. Volume scattering in the snowpack also reduces the microwave emission to a greater degree at 37 GHz than 19 GHz [e.g., Grenfell, 1986], further enhancing the difference between 19 and 37 GHz $T_b$ values. When the IABP/POLES air temperatures near 0°C, the 37 GHz $T_b$ approaches, and often surpasses, the 19 GHz $T_b$, resulting in the derivation of snowmelt onset. In some cases the AHRA-derived snowmelt onset date occurs slightly before the POLES air temperature reaches 0°C, consistent with the findings of Barber et al. [1995], who noted liquid water formation in Arctic snowpacks prior to 0°C air temperatures. Crane and Anderson [1994] also demonstrated snowmelt onset can occur at subzero temperatures with marked increases in solar radiation. Physically, the addition of sensible heat to the snowpack alters the geometric structure of the snow grains [Colbeck, 1986], and as liquid water may form on these snow grains, the resulting decrease in the emissivity difference between 19 and 37 GHz $T_b$ values leads to the derivation of snowmelt onset.

Over the first-year ice locations, the $T_b$(19H) - $T_b$(37H) differences remain much closer to the 4.0 K threshold during winter, because the emissivities of 19 GHz and 37 GHz are similar over first-year ice. Although the emissivity of 37 GHz is actually higher than 19 GHz, the 19 GHz $T_b$ is typically larger than the 37 GHz $T_b$ because volume scattering in the snowpack affects the 37 GHz channel more than the 19 GHz channel. In several cases the differences dip below the 4.0 K threshold of the AHRA, but quickly return to $T_b$(19H) - $T_b$(37H) values typical of premelt conditions. Similar to the multiyear ice locations though, the AHRA-derived snowmelt onset date occurs when air temperatures begin to remain around 0°C.

As an additional comparison, AHRA-derived snowmelt onset dates are compared with passive microwave-derived snowmelt onset dates from Smith [1998] and active microwave snowmelt onset dates from Winebrenner et al. [1994] over selected case sites in 1992 (Table 1). The melt dates for all three studies generally agree with each other within 4 days, except for the point at 74.7°N, -160.2°E, where the AHRA-derived melt onset date is 5 days later than Smith [1998] and 3 days earlier than
Winebrenner et al. [1994]. Smith [1998] states his algorithm was confused by an early warming at this point, which apparently did not affect the AHRA. Other differences between the AHRA and Smith [1998] are likely due to differences in the calculation of snowmelt onset, including preprocessing of the data. As previously mentioned, data for the AHRA algorithm were initially converted to a consistent time series record, which Smith [1998] did not perform. Overall though, there is good correspondence in the three methods. The mean absolute difference between the AHRA data and Smith's calculations is 2.59 days, while the mean absolute difference between the AHRA and the SAR data is 2.10 days, and the mean absolute difference between Smith's data and the SAR computations is 2.20 days (Table 1). The estimated error in the AHRA data is
Greenland ice sheet [e.g., Parkinson et al., 1987].

Table 1. Comparison of AHRA (A) snowmelt onset dates with Smith [1998] passive microwave (P) derived snowmelt onset dates and Winchbrenner et al. [1994] synthetic aperture radar (S) derived snowmelt onset dates in 1992. Relative differences (in absolute value) between the three methods are also highlighted.

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Mean Difference 2.59 2.10 2.20

Therefore roughly 2 days, similar to Smith [1998], with the advantage of the AHRA being that it computes melt onset over first-year sea ice.

5. Mean Snowmelt Onset Pattern

The 20-year mean snowmelt onset dates show snowmelt onset progresses roughly in a radial pattern, beginning with snowmelt along the southern ice edges and progressing northward (Figure 5). Melt onset typically starts prior to Julian day 70 in the Bering Sea, the Sea of Okhotsk, and along the Labrador coast. From Julian day 70 through Julian day 90, melt onset normally occurs in most of the Bering Sea and the Sea of Okhotsk, and some melt onset is noticeable in Hudson Bay, the south Greenland coastline, and the eastern Arctic Islands. The melt onset area expands to include the Hudson Bay coastline and southern portions of Davis Strait through Julian days 90 to 110, and melt also appears in the East Siberian Sea and along the Russian coastline during this time. From Julian day 110 through Julian day 130, snowmelt is typically detected along the edges of the Arctic Basin and throughout all of Hudson Bay. The next 20 days (Julian days 130 through 150) are marked by melt onset covering the Davis Strait and most of the southern Canadian Arctic Archipelago Islands. Snowmelt begins to rapidly expand northward at this point, and from Julian day 150 through Julian day 170 the total melt onset area nearly doubles. The latest melt onset dates are located in the Lincoln Sea north of Greenland, consistent with findings that minimum summer temperatures are located on the Greenland ice sheet [e.g., Parkinson et al., 1987].

6. Annual Snowmelt Onset Maps

In order to gain a better qualitative understanding of the variability in snowmelt onset, the distribution of the annual melt pattern (Figure 6), annual melt maps, and annual melt anomaly maps (Figure 7) are shown. It is clear that the occurrence of snowmelt in any given year rarely follows the mean snowmelt distribution discussed above. The melt distribution for 1981 best resembles the mean melt distribution, but it still exhibits areas where the melt occurs either earlier or later than average (Figure 6). Earlier melt is observed in 1981 for the Siberian and Chukchi Seas, and the west central Arctic Ocean, while later than average melt is observed in the Kara and Barents Seas (Figure 7).

In two of the years, 1979 and 1987, the melt distribution is greater than average in March and April, but then melt is delayed, and from May through July the spatial area of melt is less than average (Figure 6). In 1979, earlier melt occurs in the Bering Sea, along the Alaskan coast, and in portions of Hudson Bay, explaining the enhanced melting noted in March and April (Figure 7). Melt onset is delayed in much of the central Arctic however, explaining why the melt distribution is below average from May through July. In 1987, earlier melt is detected in Hudson Bay and the Sea of Okhotsk, while later melt is observed from the Siberian Sea northward into the Arctic Ocean (Figure 7).

Several years, including 1988, 1993, 1994, and 1998, show an opposite pattern to 1979 and 1987. The spatial melt area is below normal in March and April, but quickly increases such that from May through July the spatial area of melt onset is greater than average (Figure 6). In 1988 the vast majority of area that melts on average in March and April is delayed several weeks, with the exception of southern Hudson Bay (Figure 7). The enhanced melting seen from Julian day 160 through Julian day 190 is due to earlier melt in the Arctic Ocean north of the Laptev Sea. In comparison, during 1993 delayed melt is noted throughout all of Hudson Bay, and most of the other southern ice regions, with the exception of a small band of earlier melt along the northern edge of the East Siberian Sea and the Laptev Sea. The enhanced melt later in the season is due to earlier melt in the Arctic Ocean north of the Chukchi Sea. In 1994 the spatial melt pattern is similar to 1988, with earlier than average snowmelt seen in southern Hudson Bay, set against delayed melt in much of the remaining southern ice regions. Enhanced melt from Julian day 160 through Julian day 190 is a result of earlier snowmelt in the Arctic Ocean north of the Canadian Arctic Archipelago (Figure 7). In comparison with the other three years that have a similar melt distribution, 1998 displays a much more hemispherically based melt pattern (Figure 7). For instance, very early snowmelt is observed over much of the western hemisphere, especially in the Arctic Ocean. Later than average snowmelt is comparatively noticed in the Kara and Barents Seas, as well as the Arctic Ocean directly north of these Kara and Barents Seas (Figure 7).

Enhanced melting toward the latter stages of the melt period is also noted for 1989, 1991, and 1995 (Figure 6). In 1989, large sections of Hudson Bay experience later than average melt, while parts of the Sea of Okhotsk, Laptev Sea, Kara Sea, and central Arctic Ocean experience earlier than average melt (Figure 7). Similarly, later snowmelt is noted in Hudson Bay in 1991, while the Beaufort, East Siberian, and Laptev Seas all experience earlier melt. Although an average melt onset pattern is observed for most of Hudson Bay in 1995, earlier than average melt onset is noted in the Beaufort, Kara, and Barents Seas.

The melt pattern in 1990 stands out as a very abnormal melt season. Approximately 25% of the melt area is reached by Julian
mean snowmelt onset (1979 through 1998) as detected by the AHRA.

Figure 5. Mean snowmelt onset (1979 through 1998) as detected by the AHRA.

7. Variability in Snowmelt Onset Over Arctic Sea Ice

Spatial and temporal variability in the annual snowmelt onset dates is also illustrated with range and standard deviation maps, as well as with principal component analysis (PCA). PCA is a
technique that reduces a large data set into fewer new variables that still represent a large portion of the variance present in the original data set, and it is sometimes referred to as empirical orthogonal function (EOF) analysis. The result is a set of principal components (PCs) that are uncorrelated with one another. The primary advantage of using PCA in this study is to identify areas where the annual snowmelt onset dates vary in a similar fashion with one another.

The range in snowmelt onset at a given location is defined as the latest melt date minus the earliest melt date for that location. A range of 1-2 months is common in the central Arctic, while the range in the more southerly latitudes, such as Hudson Bay and along the continental coastlines, often exceeds 2 months (Figure 8). The large variability in the snowmelt onset dates suggests snowmelt onset is not solely defined by the seasonal transition in incoming solar radiation. It seems likely atmospheric circulation patterns may play a large role in determining melt onset, especially in the southerly regions. Utilizing the AHRA data, Drobot and Anderson [2001] stratify the Arctic into 13 subregions and suggest that earlier (later) than average snowmelt onset occurs where warm (cold) air advection and increased (decreased) cyclonic activity are present.

An examination of the standard deviation in melt onset (Figure 9) provides further evidence that atmospheric conditions

Figure 6. Shaded area represents annual snowmelt onset 2-day cumulative frequency distributions from 1979 through 1998. Solid line represents mean cumulative frequency distribution for the entire 1979 through 1998 data set. The 2-day window is necessary because scanning multichannel microwave radiometer data were only collected every second day.
play an important role in determining melt onset. While the range measures the absolute difference between the earliest and latest melt onset dates, the standard deviation provides the length in days between the 16.5 and 83.5 percentiles. In other words, it is the range in days between the third earliest and third latest melt onset dates at each pixel. Physically, the standard deviations indicate that the large ranges found in Figure 8 are not due to extreme values in a few years, but rather that large variability in the snowmelt onset dates exists in a wider range of years.

The principal component analysis is also useful for examining spatial variability in the snowmelt onset dates. The first PC, explaining 16% of the total variability in snowmelt onset, consists of one correlation center covering an area from the New Siberian Islands to Wrangel Island, north to the west central Arctic, and an opposite-signed center around Baffin Bay (Figure 10). Physically, the PCA analysis indicates that earlier (later) than average snowmelt onset in one center is correlated with later (earlier) than average snowmelt onset in the other. The two centers also closely resemble surface temperature differences related to the Arctic Oscillation [e.g., Thompson and Wallace, 1998]. The Thompson and Wallace [1998] study suggests positive (negative) phases of the AO are associated with abnormally warm (cool) surface air temperatures along the Siberian coast, and abnormally cool (warm) surface air temperatures near Baffin Bay. A reasonable hypothesis is that the first PC of the snowmelt onset data represents the snowmelt onset response to variations in the surface air temperature (associated with alterations in the AO phase).

In order to test whether annual variations in the AO are related to variations in snowmelt onset, the PC scores of the first PC are
Figure 7. Annual snowmelt onset and snowmelt anomaly maps from 1979 through 1998.
Figure 7. (continued)
Figure 7. (continued)
Figure 7. (continued)
correlated with monthly averaged March through June AO values (Figure 11). The statistically significant correlation ($r=-0.65$; $p$ value=0.02) suggests variations in the AO at least partly govern the date of snowmelt onset in portions of the Laptev Sea and Arctic Ocean and near Baffin Bay.

8. Conclusions

An improved snowmelt onset detection algorithm was presented for first-year and multiyear Arctic sea ice surfaces. The advanced horizontal range algorithm (AHRA) utilizes temporal information in the brightness temperature difference between 19 GHz (18 GHz for SMMR) and 37 GHz. From 1979 through 1998, snowmelt onset began on average in the Bering Sea and Sea of Okhotsk in the first week of March and progressed northward toward the central Arctic by the middle of July. The latest melt onset dates were observed in the Lincoln Sea, north of Greenland, in accordance with the minimum in air temperatures located over Greenland.

In comparison with the roughly radial northward melt progression of the annually averaged melt onset map, specific years showed a high degree of spatial variability. Most years typically have some regions of earlier than average melt and other regions of later than average melt. However, 1990 appeared to be an extraordinarily early melt onset year, with later than average snowmelt onset predominantly occurring only in the Beaufort Sea.

A principal component examination of snowmelt onset illustrated two dominant snowmelt onset centers. Earlier (later) than average snowmelt onset in an area from the New Siberian Islands to Wrangel Island, north to the west central Arctic, was related to later (earlier) than average snowmelt onset in an area around Baffin Bay. A statistical comparison between the PC scores suggested variability in the snowmelt pattern of the first
principal component closely resembled variations in the Arctic Oscillation.

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