Spatiotemporal variations of snowmelt in Antarctica derived from satellite scanning multichannel microwave radiometer and Special Sensor Microwave Imager data (1978–2004)

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[1] We derived the extent, onset date, end date, and duration of snowmelt in Antarctica from 1978 to 2004 using satellite passive microwave scanning multichannel microwave radiometer (SMMR) and Special Sensor Microwave Imager (SSM/I) data. A wavelet-transform-based method was developed to determine and characterize melt occurrences. About 9–12% of the Antarctic surface experiences melt annually. This is more than twice the surface melt extent measured in Greenland. Seasonally, surface melt primarily takes place in December, January, and February and peaks in early January. Regression analysis over the 25 year period of study reveals a negative interannual trend in surface melt. Nevertheless, the trend inference is not statistically significant. Large year-to-year fluctuations characterize the interannual variability. Extremely high melt occurred in the 1982/1983 and 1991/1992 summers, while extremely weak melt occurred in the 1999/2000 summer. A strong correlation with air temperature suggests that the melt index can serve as a diagnostic indicator for regional temperature variations. Periodic melting has been observed over Ross Ice Shelf, Ronne-Filchner Ice Shelf, the West Antarctic ice streams, and outlet glaciers in the Transantarctic Mountains. The Antarctic Peninsula, West Ice Shelf, Shackleton Ice Shelf, Amery Ice Shelf, and the ice shelf along the Princess Ragnhild Coast experienced the most persistent and intensive melt and should be closely monitored for their stability in the future, given the recent disintegration of the Larsen Ice Shelf A and B.


1. Introduction

[2] Portions of the Antarctic Ice Sheet annually experience surface melting. Wet snow has a low visible and near-infrared albedo and absorbs as much as 3 times more incoming solar radiation than dry snow with a high albedo [Steffen, 1995]. The variation in snowmelt extent and duration affects the Earth’s radiation budget and hence is a factor in global climate changes. Furthermore, the Antarctic Ice Sheet and its surrounding ice shelves are extremely sensitive to both atmospheric and seawater temperature changes. Mercer [1978] postulated that the Antarctic ice shelves would be the component of the Antarctic glacier system most responsive to “greenhouse” warming. He anticipated a southerly retreat of the ice shelf margins starting with the ice shelves of the Antarctic Peninsula. His anticipation seems especially prescient given the rapid disintegration of the Wordie Ice Shelf during the 1980s [Doake and Vaughan, 1991] and the catastrophic collapse of Larsen Ice Shelf A in 1995 [Vaughan and Doake, 1996] and Larsen Ice Shelf B in 2002 [MacAyeal et al., 2003]. These ice shelves survived thousands of years of climate variations before their recent disintegrations [Gilbert and Domack, 2003]. On the basis of the observations from sequential satellite images, Scambos et al. [2000] concluded that meltwater ponding on the surface during the summer seasons fills and magnifies the ice crevasses and provides the conditions for the ice shelf disintegration. Their findings suggest that snowmelt amount could serve as an indicator of the stability of the ice shelves and perhaps as a proxy indication of climate change.

[3] Antarctica is the coldest, windiest and, on average, highest continent on the Earth. Harsh climate, inaccessibility, long dark polar nights in winters, and logistical difficulties impose serious difficulties and challenges for traditional field observations. With the ability to peer through clouds and observe day and night, satellite passive
microwave sensors provide an ideal instrument for mapping snowmelt in often cloudy polar regions. The scanning multichannel microwave radiometer (SMMR) onboard the Nimbus 7 satellite recorded microwave radiation every other day from October 1978 to August 1987 [Knowles et al., 2002]. The Special Sensor Microwave/Imager (SSM/I) from the Defense Meteorological Satellite Program (DMSP) [Armstrong et al., 2003] recorded daily radiation from July 1987 to the present. The availability of spatially coherent and temporally continuous passive microwave data provides a viable means for investigating the intra-annual and inter-annual variability of surface melt. Using SMMR and SSM/I data, surface melt over the Greenland Ice Sheet has been thoroughly examined by various researchers [Mote et al., 1993; Mote and Anderson, 1995; Abdalati and Steffen, 1995, 1997, 2001; Joshi et al., 2001]. However, only a limited number of studies have reported on Antarctic surface melt. Zwally and Fiegles [1994] mapped the length of the melt season over the Antarctic continent using SMMR data spanning the period 1978–1987. Ridley [1993] examined SMMR and SSM/I data spanning 1978–1991 over the Antarctic Peninsula. Fahnestock et al. [2002] analyzed snowmelt on ice shelves of the Antarctic Peninsula during the period 1978–2000. Torinesi et al. [2003] examined the spatial variability of surface melt duration in Antarctica during 1980–1999.

[4] Our research updates the previous investigations with an extended time period of passive microwave records using a more sensitive and more accurate melt detection and tracking method. The surface melt analysis is extended over the period 1978–2004 across the entire Antarctic continent and all the surrounding ice shelves. A wavelet based method is employed in our melt analysis. With this new method, we are not only able to derive melt extent and melt duration with an improved accuracy, but are also able to create time series maps for melt onset date and melt end date during 1978–2004. On the basis of 25 years of surface melt information, we analyze the spatial and temporal trend and variability of snowmelt. We also characterize the seasonal melting cycle and identify the most extensive and intensive melt zones in Antarctica. The timing and extent of the extreme melt events in the past two and a half decades are also determined and discussed. Finally, we examine the relationship of surface melt extent and duration with near-surface air temperature at a regional scale.

2. Methodology

2.1. Melt-Induced Edges on Time Series Curve of Brightness Temperature

[5] A passive microwave sensor detects the naturally emitted microwave energy from snowpack within its field of view. Passive microwave data are brightness temperatures calibrated from recorded microwave energy. The relationship between the microwave brightness temperature \( T_B \) and near-surface physical temperature \( T_s \) of snowpack can be represented by the first-order Rayleigh-Jeans approximation [Zwally and Gloersen, 1977]:

\[
T_B = \varepsilon T_s \tag{1}
\]

where \( \varepsilon \) is the emissivity of the near-surface snowpack. The physical basis for snowmelt algorithms is that the microwave emissivity of a snowpack increases rapidly in response to the introduction of a small amount of moisture. The increase in the liquid water content makes the wet snow act as a black body and emit more energy, resulting in a distinct rise in brightness temperature [Zwally and Gloersen, 1977; Ulaby et al., 1986]. The microwave emission drops either as the liquid fraction increases or upon refreezing, because the ability of the ice particles to scatter radiation emitted from deeper in the snow is reestablished. The brightness temperature increase at the melt onset and decrease at the time of freeze-up are detectable by microwave sensors at frequencies in excess of 10 GHz [Ulaby et al., 1986; Abdalati and Steffen, 1997].

[6] When the daily brightness temperature \( T_B \) is plotted as a one-dimensional time series curve, strong edges (discontinuities) emerge on the curve at the onset of melt and at the time of freeze-up. Curve A in Figure 1 shows the brightness temperature variation for a wet pixel that experienced melting during the austral summer. The most prominent feature of this curve is that the rapid increase in \( T_B \) at the onset of snowmelt in early summer stands in sharp contrast to the lower \( T_B \) observed during nonmelt conditions of spring and winter, forming a strong upward edge on the time series curve. An apparent prolonged period (plateau) of elevated brightness temperatures is observed during the summer. During late summer, a dramatic decrease in brightness temperatures corresponding to snow refreezing creates an obvious downward edge on the curve. Curve B in Figure 1 shows the brightness temperature variation for a typical dry pixel, where the physical temperature of snow remains below the melting point all the time. Abrupt changes (strong edges) are absent for the dry pixel, and its brightness temperature varies gently and smoothly.

2.2. Channel Selection

[7] The SMMR sensor has ten microwave channels, including both vertical and horizontal polarizations for 6.6 GHz, 10.7 GHz, 18 GHz, 21 GHz and 37 GHz. The SSM/I sensor has seven channels, the vertical polarization for 22 GHz and both polarizations for 19 GHz, 37 GHz and 85 GHz. Although the 85 GHz channel has the best spatial resolution (14 km), it tends to be contaminated by water vapor and clouds in the atmosphere [Mätzler, 2000].

[8] We compared the brightness temperatures of all the channels for the SSM/I sensor at various locations. As demonstrated in Figure 2, the low-frequency channels are more responsive to melt onset than high-frequency channels. This is primarily due to the fact that absorption and hence emission are greater at the higher frequencies and there is less difference between wet and dry snow conditions for the higher-frequency channels. Although the emissivities for wet snow at horizontal and vertical polarizations are nearly equal, the emissivity of dry firm at horizontal polarization is significantly lower than the vertical polarization [Zwally and Fiegles, 1994]. Therefore the transition from dry snow to wet snow causes a greater increase in horizontal brightness temperatures than vertical brightness temperatures at the same frequency (Figure 2). Overall, we observed that the brightness temperature of the 19 GHz horizontally polarized channel exhibits the sharpest transitions at the melt onset and at the time of refreezing.
Since our melt detection algorithms rely on the edge strength of temperature transitions induced by melt and refreeze events, we chose to use the 19 GHz horizontally polarized channel of the SSM/I data (18 GHz for SMMR) in our analysis.

2.3. Multiscale Wavelet-Transform-Based Method

In the past decade, various algorithms have been proposed for extracting snowmelt information from multi-channel satellite passive microwave data. To detect snowmelt occurrence, many researchers have employed a single channel brightness temperature and a threshold empirically determined by field observations or physical experiments [e.g., Mote et al., 1993; Mote and Anderson, 1995; Ridley, 1993; Zwally and Fiegles, 1994; Torinesi et al., 2003]. Some researchers have also used a composite index derived from two channels of passive microwave data, including the normalized gradient ratio (GR) [Steffen et al., 1993] and the cross polarization gradient ratio (XPGR) [e.g., Abdalati and Steffen, 1997, 2001; Fahnestock et al., 2002]. Ramage and Isacks [2003] introduced diurnal amplitude variations (DAVs) in snowmelt analysis, which are the running differences between early morning and late afternoon brightness temperature observations. Joshi et al. [2001] applied a derivative of Gaussian (DOG) edge detector to time series curves of daily brightness temperature and determined the timing of melt and refreeze occurrences.

On the basis of the wavelet transform of the daily brightness temperature time series, we performed a multi-scale edge detection for deriving snowmelt onset date, melt end date, melt duration, and melt extent [Liu et al., 2005]. In the same vein as the work of Joshi et al. [2001], our method is based on the observation that strong and significant edges in the brightness temperature \((T_b)\) time series curve signify snow melting and refreezing events. We first decompose the time series brightness temperature values into multiscale components through a wavelet transform. Then, the edges are tracked and analyzed across scales. To differentiate significant edges corresponding to melting events from weak edges associated with random signal perturbation and noise, an optimal edge threshold is statistically determined by variance analysis and a bimodal Gaussian curve fitting technique [Liu et al., 2005]. On the basis of the principle of spatial autocorrelation, we also developed a spatial neighborhood operator for detecting and correcting the possible errors brought about by strong noise or heterogeneity of data pixels [Liu et al., 2005]. Through detecting and tracking strong and significant edges in the brightness temperature time series curve, our algorithm is able to determine whether and when a pixel experienced melt.

Figure 1. Daily brightness temperature variations of the 19 GHz horizontal channel of the SSM/I for typical wet and dry pixels. The wet pixel is located on the Antarctic Peninsula, and the dry pixel is located in the interior Antarctic Ice Sheet. The horizontal axis is the sequential number of days from 1 July 2001 to 30 June 2002.

Figure 2. Comparison of brightness temperatures for channels of different frequencies and polarizations.
The melt onset date is identified by locating the position of the first significant upward edge along the time axis. The melt end date is indicated by the last significant downward edge along the time axis. The melt duration, namely, the time interval between the melt onset date and the melt end date, is estimated by tracking and accumulating the time intervals between significant upward edge and downward edge pairs. It excludes the refreezing period between the melt onset date and the melt end date. [11] Our wavelet transform based edge detection method has multiple advantages over conventional snowmelt algorithms. First, our method explicitly searches and tracks strong edges on brightness temperature curves across scales and therefore provides a more precise estimate for the melt onset date, melt end date and melt duration. This is in contrast to some previous melt studies that use a melt index with an absolute brightness threshold to detect melt occurrence. Because of the variation in emissivity of the snowpack at different locations, the use of a fixed absolute brightness temperature as a melt threshold often causes errors in the timing of melt occurrences. As shown by Liu et al. [2005], the conventional melt index threshold method tends to produce too late a melt onset date and too early a melt end date for each melt event and hence too short a melt duration. Second, our statistical approach produces a more objective and reliable edge threshold compared with the trial-and-error process used by Joshi et al. [2001] for edge threshold selection. It also circumvents the problems and frequent unavailability of field observations for threshold determination. In most previous studies, the melt threshold value was empirically determined from field observations of the volumetric water content of a snowpack or near-surface air temperature at one or more specific sites. Since one pixel from the satellite passive microwave data covers a relatively large ground area, e.g., 25 km by 25 km for SSM/I data, the in situ point observations of the water content of snowpack or air temperature may not be representative for the entire ground area covered by the pixel of passive microwave data, resulting in a biased threshold value. Because of the hostile environment in polar regions, acquisition of field observations is extremely difficult and costly. Furthermore, the explicit incorporation of spatial contextual reasoning through a neighborhood operation enables our method to check and remove potential errors from purely temporal analysis of brightness temperature time series. Therefore our melt computation results are spatially more consistent and less noisy, compared with previous similar studies [Liu et al., 2005].

[12] Our method has been implemented using C++ programming language. A detailed description and evaluation of the wavelet transform based method has been presented by Liu et al. [2005].

### 3. Melt Information Extraction From Satellite Passive Microwave Data

[13] The satellite passive microwave data used in our analysis include the SMMR data collected by the NASA Nimbus 7 satellite and the SSM/I data collected by the DMSP-F8, DMSP-F11, and DMSP-F13 satellites. We processed the data spanning the period from 26 October 1978 to 30 June 2004. The temporal coverage of each sensor is listed in Table 1. We employed the 18 GHz horizontal polarization channel from the SMMR data and the 19 GHz horizontal polarization channel from the SSM/I data.

[14] The observations from both ascending and descending orbits are available. We chose the observations that occur closer in time to the strongest melt period for most sensitive detection of snowmelt. Those include the observations from the Nimbus 7 SMMR ascending orbits, DMSP-F8 descending orbits, DMSP-F11 and DMSP-F13 ascending orbits. As shown in Table 1, observations over Antarctica were made near local noon by the SMMR ascending orbits and at dusk (about 17:30 local time) by the DMSP-F8 descending orbits and the DMSP-F11 and DMSP-F13 ascending orbits.

[15] The continuous passive microwave data over 1978–2004 were acquired by four sensors. These sensors are different in terms of the incidence angle, over-flight time, or effective field of view. In addition, the 18 GHz horizontally polarized channel of the SMMR sensor is used to match the 19 GHz channel of the SSM/I sensors. For the sake of consistent analysis, the microwave brightness temperatures from these sensors are recalibrated. On the basis of the overlapping observations, Jezek et al. [1991] computed linear regression equations for all channels between the SMMR and SSM/I F8 sensors to remove systematic differences. Abdalati et al. [1995] conducted the similar work between the SSMI F8 and F11 sensors. Using the SSM/I F8 data as a baseline, we converted the SMMR,
SSM/I F11 and SSM/I F13 data into SSM/I F8 equivalent values using the regression coefficients derived by Jezeck et al. [1991] and Abdalati et al. [1995], as shown in Table 2. The brightness temperature differences between SSM/I F11 and SSM/I F13 are insignificant, due to very similar orbital characteristics (Table 1). Therefore the same set of regression coefficients between SSM/I F11 and SSM/I F8 are applied to the conversion between SSM/I F13 and SSM/I F8.

[16] The SMMR and SSM/I data have been processed into Equal Area Scalable Earth (EASE)-Grid Temperature Brightness ($T_b$) products at the National Snow and Ice Data Center (NSIDC). A grid cell for the 18/19 GHz channels covers a 25 km by 25 km ground area. A grid with 177 x 221 cells retrieved from the EASE-GRID Southern Hemisphere data fully covers the Antarctic continent and a small portion of the surrounding ocean. To eliminate ocean contamination, we applied an Antarctic coastline extracted from a Radarsat SAR image mosaic [Liu and Jezek, 2004] to mask the EASE-Grid data. The melt computation was carried out only for pixels that lie on the ice shelves and/or the ice sheet.

[17] The SMMR data are available every other day. Brightness temperatures for the alternate missing days are calculated through linear interpolation from adjacent days. Also, there are instances of missing data for some days in the SSM/I EASE Grid data, which are indicated by a value of 0. To remove the false edges induced by 0 values on the time series curve, we replace the 0 values with the linearly interpolated brightness temperatures. For each pixel, we extract the corresponding brightness temperature values from daily SMMR and SSM/I grids to form 365 days of $T_b$ time series. The time series starts with 1 July and ends on 30 June of the next year in order to center the Antarctic austral summer season in the middle of the $T_b$ curve. The brightness temperature data of the 18/19 GHz channels are sequentially processed pixel by pixel for every year using the wavelet transform based method. Three output grids are produced for each year: melt onset date grid, melt end date grid, and melt duration grid. Because the SSM/I data were missing for 28 days in December, 1987, the melt calculation for the 1987/1988 austral summer is incomplete. Therefore this year is not included in the following analysis of spatiotemporal variability.

### Table 2. Regression Coefficients for Data Adjustments Between Different Sensors

<table>
<thead>
<tr>
<th>Conversion</th>
<th>Slope</th>
<th>Intercept</th>
<th>Correlation Coefficient</th>
</tr>
</thead>
<tbody>
<tr>
<td>SMMR 18 GHz horizontal to SSM/I F-8 19 GHz horizontal</td>
<td>1.06</td>
<td>−2.79</td>
<td>$R &gt; 0.99$</td>
</tr>
<tr>
<td>SSM/I F-11 and F-13 19 GHz horizontal to SSM/I F-8 19 GHz horizontal</td>
<td>1.008</td>
<td>−1.17</td>
<td>$R &gt; 0.99$</td>
</tr>
</tbody>
</table>

4. Spatiotemporal Variation of Snowmelt in Antarctica

4.1. Snowmelt Extent and Duration

[18] By processing the SMMR and SSM/I data, we are able to map the spatial extent of snowmelt across the continent on a daily, monthly or seasonal basis. In order to gain an insight into the spatial variability in melt occurrence, we compiled an average melt duration map for the period 1978–2004 (Figure 3). The shaded area shows the total melt extent, which consists of grid cells that were detected as melting for at least one day on average during summers from 1978 to 2004. The total melt extent is 2,359,375 km², covering 17.27% of the continent. The grey level of each grid cell represents the average melt duration (melt days) per year during 1978–2004.

[19] Relatively extensive and continuous melt areas include the Larsen Ice Shelf on the Antarctic Peninsula, George VI Ice Shelf and Wilkins Ice Shelf around Alexander Island, Amery Ice Shelf, West Ice Shelf, Shackleton Ice Shelf, ice shelves along Queen Maud Land, and Ross Ice Shelf (Figure 3). Melt areas are also scattered along the coasts of Wilkes Land and Marie Byrd Land, on the Ronne-Filchner Ice Shelf, and in glacial valleys in the Transantarctic Mountains. There exist strong regional variations in melt duration. On average, 25% of the melt areas only experienced melting of shorter than 1 week each year, and about 71% of the melt zones experienced melting of shorter than 30 days (Figure 4). Only about 9% of the melt zones experienced more than 60 days of melting. Ice shelves on the Antarctic Peninsula, Amery Ice Shelf, Shackleton Ice Shelf, West Ice Shelf, and the ice shelf along the Princess Ragnhild Coast experienced the longest periods of melting on average, due to their relatively low latitude and low elevation (Figure 3). Significant parts of the Ross Ice Shelf and the Ronne-Filchner Ice Shelf experienced melting, although average annual melt duration was less than one week per year.

[20] In a Geographical Information System (GIS) environment, we calculated the distance of each melt pixel to the coastline (ice sheet or ice shelf margin) [Liu and Jezek, 2004] and the elevation value of each melt pixel from the Antarctic digital elevation model [Liu et al., 1999]. The relationships of the melt duration and melt occurrence rate with the topography and proximity to the ocean were analyzed based on all melt pixels. As shown in Figure 5, the melt duration and melt occurrence rate decrease with the increasing distance from the coast. For the continental ice sheet, surface melting can extend inland up to 600 km from the coastline. However, 60% of the melt areas are concentrated within 100 km wide coastal strip, and 90% of the melt areas are located within 300 km of the coastal ice margins. Seventy-one percent of the area that falls within the 100 km wide coastal zone experienced surface melting. The average duration is about 20 days for the melt zones that are located less than 100 km from the coastline. Snowmelt was observed on the ice sheet to an elevation as high as 1500 m above the mean sea level on the Antarctic Peninsula and in the Transantarctic Mountains, but 60% of the melt areas are located below 200 m in elevation (Figure 6). About 95% of the snow surface with an elevation below 50 m experienced
melting, having an average melt duration of 29 days. In general, the melt duration and frequency in Antarctica decreases with the rise in elevation as might be expected based on the atmospheric temperature lapse rate (Figure 6). However, this elevation dependence pattern is complicated by the latitudinal effect. The melt zones with an elevation between 50 m and 100 m are mainly distributed in the Ross Ice Shelf and Ronne-Filchner Ice Shelf. Because of the high latitude, the melt zones in this elevation range have relatively lower melt duration (about 10 days) (Figure 6). The melt zones with an elevation above 1000 m are mainly located in the Antarctic Peninsula. The melt zones in this high elevation range have relatively high melt duration (about 15 days) due to their low latitude.

4.2. Seasonal Melt Cycle and Spatial Variation of Snowmelt Onset and End Dates

We delineated the daily melt extents for every year from 1978 to 2004, except for the year 1987/1988. To understand the typical seasonal cycle of surface melt, the melt extent values of corresponding days of each year are averaged across the past 25 years to produce mean daily melt extents. Days are sequentially labeled with 1 July as the first day and 30 June of the next year as the last day.

![Figure 3. Average annual melt duration during austral summers from 1978 to 2004.](image)

![Figure 4. Frequency distribution of surface melt duration in Antarctica.](image)
The daily variation of 25-year mean melt extent is displayed in Figure 7 to show the typical seasonal melt distribution cycle. Slight melt occurred in late October and November. In early December, the melt begins to spread and the melt extent expands rapidly. Melt reaches its maximum coverage in early January. From early February, refreezing begins to occur and spatial coverage of melt drops off steadily. Melt typically terminates in late February or early March for most melt zones.

Most surface melting in Antarctica occurs during December, January and February (Figure 7). The 25 year mean melt extent and mean frequency of melt occurrence

![Figure 5](http://example.com/figure5.png)

**Figure 5.** Relationship of surface melt with distance to coastal margin. The melt duration curve shows average melt days for melt pixels that are located at different distance ranges from the coast. The melting distribution rate curve indicates the percentage of melt pixels that are located at a specified range in relation to the total number of melt pixels in Antarctica. The melting occurrence rate curve shows the percentage of the melt pixels in relation to the total number of melt and nonmelt pixels within a specified range of distance to the coast.

![Figure 6](http://example.com/figure6.png)

**Figure 6.** Relationship of surface melt with surface elevation. The average melt duration curve shows average melt days for melt pixels within a specified range of elevation values. The melting distribution rate curve indicates the percentage of melt pixels within a specified range of elevation values in relation to the total number of melt pixels in Antarctica. The melting occurrence rate curve shows the percentage of the melt pixels in relation to the total number of melt and nonmelt pixels within a specified range of elevation.
within each of these three summer months are shown in Figures 8a–8c. The mean melt frequency indicates the percentage of melt days in each summer month averaged over 1978–2004. The mean melt frequency can be interpreted as the melt occurrence probability on a day within an austral summer month. The melt extent in January is similar to that in December but is over 2 times as large as in February. The melt occurrence frequency is considerably higher in January than in December. By February, only the ice shelves on the Antarctic Peninsula still have sustained high melt frequency, and other melt zones have much reduced melt occurrences (Figures 8a–8c).

[24] We have produced melt onset date and melt end date maps for each year from 1978 to 2004 (Figures 9 and 10).

Figure 7. Seasonal variation of mean melt extent averaged over 1978–2004. The vertical axis is the average area of melt, and the horizontal axis is the sequential number of days starting with 1 July.

Figure 8a. Melt extent and occurrence frequency in major summer months: December.
Figure 8b. Melt extent and occurrence frequency in major summer months: January.

Figure 8c. Melt extent and occurrence frequency in major summer months: February. The gray level shows the percentage of melt days in each month.
Figure 9. Melt onset date in Antarctica during 1978-2004.
Figure 10. Melt end date in Antarctica during 1978-2004.
Figure 11. Melt duration in Antarctica during 1978–2004.
These maps demonstrate respectively the expansion and contraction of melt zones in space and time. As expected, melt zones expand generally from the coast toward the interior, from lower latitude to higher latitude, and from lower elevation to higher elevation. Melt zones contract in an opposite direction. This is because the air temperature increases and decreases approximately according to this pattern as the summer season progresses [Comiso, 2000].

4.3. Interannual Variability and Extreme Melt Years

[25] The spatiotemporal variations in melt duration for the Antarctic continent are shown in Figure 11. The melt extent and melt index are used to quantify the interannual variability of the snowmelt. The areal extent of melt for each year is calculated by summing the areas of all pixels that experienced at least one day of melting in the year. The concept of melt index introduced by Zwally and Fiegles [1994] is used to measure the total melt amount. The melt index MI for a year is defined in equation (2) as

$$MI = \sum_{i=1}^{N} A \cdot md_i$$

where $A$ is the areal size of one grid cell, $md_i$ is the number of melt days (duration) within a year for the pixel $i$, and $N$ is the total number of melt pixels for the year. As the melt index incorporates both the melt extent and duration information, it provides an overall measurement of melt amount for a region.

[26] We computed the melt extent and melt index on a yearly basis from 1978 to 2004 (Figure 12). Interannual changes in the snowmelt are investigated by examining the time series of the melt extent and melt index. To uncover possible interannual trends, we conducted linear regression analysis on the time series of melt extent and melt index using the least squares method. Over the 25 year period, the melt index decreases by about 274,820 day km² (0.66% of the mean) annually, and the melt extent shows a slight decrease of 8,600 km² (0.63% of the mean) annually. Breaking annual data into separate summer months, we found out that both the melt index and melt extent exhibited a negative trend in December and January and a positive trend in February. However, caution must be exercised in interpreting the trend analysis results. As indicated by small correlation coefficients ($R$) and 95% confidence intervals of the fitted slope coefficients (Figure 12), the fitted trend lines are not statistically significant.

[27] Lack of statistical significance in the trend study reflects large interannual variability in melt occurrence. We calculated the median and median absolute deviation of the melt extent and melt index during 1978–2004 to quantify the interannual variability. Because of the presence of anomalous melt years, the median is a more robust estimator for the central tendency of annual melt magnitude than the mean, and the median absolute deviation (MAD) is a more robust measure of variability (dispersion) than the standard deviation. The median absolute deviation is defined as the
median of the absolute deviations about the median [Barnett and Lewis, 1994]. On the basis of the median and MAD, we developed an anomaly indicator to identify anomalous melt years:

\[
\text{MAD} = \text{median} \left\{ \left| ME_j - \text{median} \{ ME_j \} \right| \right\}
\]

where \( ME_j \) is the melt extent for the year \( j \), and \( ME_j \) is the anomaly indicator for the year \( j \). Similarly, we can calculate the anomaly indicator in terms of melt index \( MI_j \), which replaces the melt extent \( ME_j \) in equations (3) and (4).

![Figure 13](image)

**Figure 13.** Seasonal variation of melt extent for different regions averaged over 1978–2004. The vertical axis is the average area of melt, and the horizontal axis is the sequential number of days.

We label a year as an extremely high melt year if the anomaly indicator \( AI_j \) for that year is greater than 3.0. Namely, the melt extent \( ME_j \) or the melt index \( MI_j \) for that year is greater than the median plus 3 times the median absolute deviation. We label a year as an extremely low melt year if the anomaly indicator \( AI_j \) for that year is less than \(-3.5\). This year has the least melt extent and duration during the past 25 years.

### 4.4. Regional Differences in Surface Melt

[29] We divided the Antarctic continent into eight regions according to spatial continuity and similarity in seasonal melt cycle (Figure 13). The seasonal melt cycle for each region is described by the mean daily melt extent averaged over 1978–2004. As shown in Figure 13, the Antarctic Peninsula region has the longest melt season, spanning from late October until early March of the next year. This region is the first to show melting, the last to show refreezing. In contrast, the Ronne-Filchner Ice Shelf region and the Ross Ice Shelf region have the shortest melt season, mainly occurring in late December and January. The melt season of other regions is primarily December, January, and the early part of February.

[30] For each region, we calculate the annual melt extent and melt index between 1978 and 2004, and trend lines are also fitted (Figure 14). Most regions, including the Antarctic Peninsula region, the Amery Ice Shelf region, the Queen Maud Land region, the Marie Byrd Land region, and the Ross Ice Shelf region, exhibited a negative interannual trend in both melt extent and melt index. Only Ronne-Filchner Ice Shelf region had a positive trend in both melt extent and melt index. For the Shackleton Ice shelf region (including the West Ice Shelf) and the Wilkes Land region, the melt index has a negative trend, while the melt extent has a slight positive trend. Again, the trends identified from the regression analysis are not statistically significant (Figure 14).

[31] We calculated the median, MAD, and coefficient of variation (CV) in terms of melt extent, melt index, and melt...
duration over the past 25 years for each region (Table 3). The CV is defined as the ratio of the MAD to the median. It indicates the relative interannual variability. Using the anomaly indicator defined earlier, we also identified extremely high and extremely low melt years for each region (Table 3).

The most extensive, intensive and consistent melting in Antarctica occurred in the Antarctic Peninsula region. Its
Table 3. Snowmelt Variability in Different Regions in During 1978–2004

<table>
<thead>
<tr>
<th>Statistics</th>
<th>Peninsula</th>
<th>Ronne-Filchner</th>
<th>Queen Maud Land</th>
<th>Amery</th>
<th>Shackleton</th>
<th>Wilkes Land</th>
<th>Ross</th>
<th>Marie Byrd Land</th>
<th>Total</th>
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</thead>
<tbody>
<tr>
<td>Melt extent, km²</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Median</td>
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<td>87500</td>
<td>317188</td>
<td>91563</td>
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melt extent and duration are the highest among the regions and its interannual variability of melt extent, melt index and melt duration is the smallest. The sustained melting is attributable to the gentle surface slope on Peninsula ice shelves and the greater solar radiation intensity at the lower latitudes and, possibly, the advection of the warm humid air masses from northerly directions. Scambos et al. [2000] suggested that recent breakups of ice shelves on the Antarctic Peninsula [Vaughan and Doake, 1996; MacAyeal et al., 2003] were correlated with the ponded water brought about by long and extensive melt on the ice shelves. Three other regions with steady and extensive melt are the Amery Ice Shelf region, the Queen Maud Land region, the Shackleton Ice Shelf region. For these regions, melt extent and duration are also quite large and consistent from year to year. In comparison, the Marie Byrd Land region and the Wilkes Land region have a short melt duration. In these two regions, surface melting is scattered along a narrow band of the coast. Of the eight regions, the Ronne-Filchner Ice Shelf region and especially the Ross Ice Shelf region have the shortest melt duration and the greatest interannual variability in melt extent. In the Ronne-Filchner Ice Shelf region, the extensive melting occurred in the 1990/1991, 1991/1992 and 1996/1997 austral summers. In these years, melt zones extended along the Ronne Ice Shelf inland up to the Evans Ice Stream and along the Filchner Ice Shelf up to the Slessor Glacier valley. In contrast, no melting or only a very small extent of melting occurred during 1993/1994, 1999/2000, 2000/2001, and 2003/2004 austral summers. In the Ross Ice Shelf region, the interannual variability in melt extent is even greater. In the extremely high melt years, such as 1982/1983 and 1991/1992 austral summers, spatially continuous melt spread over most of the Ross Ice Shelf. Short-duration melting periodically covered the West Antarctic Ice Streams, including the Echelmeyer Ice Stream, MacAyeal Ice Stream, the Binschadler Ice Stream, Kamb Ice Stream, and Whillans Ice Stream. However, in some years the entire Ross Ice Shelf region remained frozen and no melt was detected (Table 3 and Figure 14). Sporadic melting was also detected in the glacial valleys in the Transantarctic Mountains up to 85°S, despite the high latitude and high elevation. Melt patches and their spatial

**Figure 15.** Relationship of surface melt index with monthly mean temperature of December and January.
growth and decay in the Transantarctic Mountains appear linked to the locations of storm tracks. The adiabatic warming of Katabatic winds flowing downslope and the relatively low albedo of the ice are conducive to local temperature rise and to intensive solar insolation, which are capable of supplying sufficient heat to produce temporary surface melting.

5. Correlation Between Surface Melt and Regional Climate

[33] The relationship between surface melt occurrence and regional climate was analyzed by examining the correlation of surface melt extent and melt index with concordant historical near-surface air temperatures recorded by weather stations. In our analysis, four manned stations were used, and their geographical locations are marked in Figure 13. The meteorological data for these stations were obtained from the Web site of the Antarctic Cooperative Research Center (http://aadc-maps.aad.gov.au/aadc/portal/) and the Web site of the British Antarctic Survey (BAS) (http://www.antarctica.ac.uk/met/data.html). The monthly mean station temperature was used to represent the regional climate condition. Because of very limited temperature records and strong fluctuations in surface melt, regression analysis was not conducted for the Ross Ice Shelf region, the Ronne-Filchner Ice Shelf Region and the Marie Byrd Land region. For other five regions, we first performed a regression analysis between regional melt index and the selected station air temperatures, respectively, for December and January. A statistically significant linear relationship between melt index and the monthly mean temperature exists for both December and January. The correlation strength for December is considerably higher than that for January for all the five regions.

[34] To reveal the overall relationship between the summer surface melt and near-surface air temperature, a linear regression equation is fitted for each of the five regions using the average December and January melt index as the response variable and the average of the monthly mean temperatures of December and January as the predictor (independent) variable. The regression analysis results are shown in Figure 15 and Table 4. A clear positive correlation between surface melt index and air temperatures is indicated by high correlation coefficients. The statistical significance of the linear relationship is confirmed at over 99% confidence level by Student’s t tests (Table 4). Alternating extremely warm and cold summer temperatures correspond to extremely high and low melt years. As shown in Figure 15, the surface melt index generally increases as summer temperature increases. The slope of the regression line indicates the quantity of surface melt increase by 1°C summer temperature rise. The melt associated with a warming of 1°C is close to one standard deviation of the natural interannual variability of the melt index (Table 4). The 95% confidence interval is also computed for the slope coefficient for each region. For the five primary melt regions listed in Table 4, we infer at the 95% confidence that a 1°C summer temperature increase is accompanied by a melt index increase between 2,580,000 to 6,130,000 day km⁻².

[35] The intercept of the regression equations can be used to estimate the average monthly temperatures (at the selected station) required for the occurrence of surface melt in the corresponding region. Surface melting occurs for the Antarctic Peninsula region when the monthly mean temperature during December and January at Rothera station is above −4.8°C, for the Amery Ice Shelf region when the monthly average temperature at Mawson station is above −2.6°C, and for the Queen Maud Land region when the monthly average temperature at Syowa station is above −3.3°C. The surface melting is negligible for most of regions when the monthly mean temperature is lower than −5°C. This finding is similar to that of Zwally and Fiegles [1994].

[36] Positive correlation is also found between the melt extent and the monthly mean temperature for all regions. However, the strength of correlation is not strong. The correlation coefficient R is below 0.5 for most of regions, except for the Amery Ice Shelf region (0.6) and the Queen Maud Land region (0.53). The weaker linear correlation between the melt extent and the monthly mean temperature...
is probably the result of topographic effects. With increasingly higher temperature, the spread of melt further inland would occur at a slower rate than estimated by linear regression models due to the relatively high surface slope at the coastal zone of the continental ice sheet.

6. Discussion and Conclusions

[37] In this research, we derived the melt extent, melt onset date, melt end date and melt duration during 1978–2004 by applying our wavelet transform based method. The quantitative information about snowmelt extent and duration will be useful input for modeling and predicting climate and sea level changes.

[38] Our analysis shows that surface melt on the Antarctic ice sheet and its surrounding ice shelves are linked to regional climate changes. The strong positive correlation with the near-surface air temperature suggests that the melt index can serve as a diagnostic indicator of changing climate as recommended by Zwally and Fiegles [1994]. Although the observed melting is correlated with regional air temperatures overall, some scattered short-duration melt patches on the Ross Ice Shelf and in the Transantarctic Mountains cannot be explained by synoptic scale climate conditions. Melting associated with warm synoptic-scale meteorological conditions is commonly more uniform over larger areas at the same surface elevation and progresses in terms of surface elevation and distance to the coast. The scattered and spatially discontinuous melt pattern in these regions is likely affected by sporadic katabatic wind heating of the surface.

[39] Examination of melt extent and melt index time series reveals a negative interannual trend over the past 25 years in Antarctica. This finding compares favorably to the slight cooling trend derived by Comiso [2000] with the AVHRR infrared data from 1979 to 1998. At a regional scale, our data analysis indicates that only the Ronne-Filchner Ice Shelf region had a positive trend, and all other regions exhibited a negative trend in surface melt. These regional trends are in agreement to the findings of Zwally and Fiegles [1994] (9 years of SMMR data) and of Torinesi et al. [2003] (18 years of SMMR and SSM/I data). Through examining passive microwave data from 1978 to 1991, Ridley [1993] concluded a positive trend in melt duration for three small ice shelves on the Antarctic Peninsula with a confidence level of lower than 85%. His result is likely biased by a short period of the surface melt increase during 1985–1990. For the Antarctic Peninsula the slight negative trend over the past 25 years derived from our study agrees with that of Comiso [2000], who revealed a similar trend during 1979–1998 for most of the Peninsula at the pixel level. Like previous research [Zwally and Fiegles, 1994; Torinesi et al., 2003] the negative melting trends inferred from our least squares regressions are not statistically significant.

[40] The most prominent feature in the surface melt time series over the past 25 years is a large degree of year-to-year fluctuations. Alternating high and low melting years rule out reliable statistical inference of melt trends. With a sensible anomaly indicator, we identified the anomalous melting years in Antarctica. The overall interannual variability in melt extent in Antarctica is largely dictated by the melt situations in the Ross Ice Shelf region, the Ronne-Filchner Ice Shelf region, and the Wilkes Land region. These three regions have the greatest interannual variability in melt extent. Other regions, especially the Antarctic Peninsula region, have relatively stable and consistent melt occurrence from year to year.

[41] Spatially, snowmelt mainly occurred on the Antarctic peripheral ice shelves, and most of the grounded ice sheet remains frozen throughout all the seasons. In most years, melt occurred only over 9–12% of the continent. In the past 25 years, about 17% of the Antarctic snow surface in total has experienced surface melting. This is in strong contrast to Greenland where melt normally occurred over 40% of the ice sheet each year, and cumulatively about 66% of the Greenland ice sheet has experienced surface melt during 1979–1991 [Mote et al., 1993]. The limited melt extent on the Antarctic ice sheet is mainly due to its high latitude and relatively high surface slope in the coastal zone, which limits the expansion of melt zones toward the interior of the ice sheet. Nevertheless, it should be noted that, despite the small fraction being melt zone, the absolute melt extent in Antarctica is over two times as large as in Greenland.

[42] Our numerical analysis demonstrated that the latitude, surface elevation, and proximity to the ocean are three primary factors shaping the spatial pattern of the melt occurrence and duration in Antarctica. With increasing latitude, elevation and/or distance to the coast, the melt onset date is generally later, melt end date is generally earlier, and duration is generally longer. Seasonally, surface melt in Antarctic continent takes place primarily during December, January, and February and peaks in the early January.

[43] Surface melting is closely related to the stability of the Antarctic glacial system and global sea level changes. The Larsen Ice Shelf, the George VI Ice Shelf and the Wilkins Ice Shelf on the Antarctic Peninsula have experienced the most intensive surface melting in the past 25 years. Scambos et al. [2000] and MacAyeal et al. [2003] show that ponding water resulting from intensive melting during summers provides the conditions for ice shelves to break up. Our melt analysis indicates that melting on the Shackleton Ice Shelf, West Ice Shelf, and lower part of the Amery Ice Shelf, the ice shelf along the Princess Ragnhild Coast in Queen Maud Land was also quite intensive and persistent in the past 25 years. If the surface melting is further intensified, these ice shelves may be susceptible to similar disintegration and therefore should be closely monitored in the future. In addition, we observed that extensive surface melting periodically occurred over the Ross Ice Shelf, the Ronne-Filchner Ice Shelf, the West Antarctic ice streams, and outlet glaciers in the Transantarctic Mountains, despite the high latitude. Because of the very gentle surface slope, the Ross Ice Shelf and the Ronne-Filchner Ice Shelf are most responsive to extremely warm summers. Additional increases in melt extent on these two giant ice shelves associated with warming summers would be much greater than other regions. The West Antarctic ice streams and valley glaciers in the Transantarctic Mountains are the most dynamic and fast moving in the Antarctic continent. The Ross Ice Shelf is the main outlet for these fast moving ice streams. The impacts of periodic surface melting on the dynamic nature of these ice streams and on the stability of
the Ross Ice Shelf are largely unknown at present. Given the ongoing debate concerning global climate and uncertain trend inference of surface melting at present, continued monitoring of surface melt of the Antarctic continent is of particular value for a reliable evaluation of the magnitude and direction of long-term melting and climate change trends.

[44] Acknowledgments. This work was supported by NASA grant NAG5-10112 and NSF grant 0126149. The authors want to thank the National Snow and Ice Data Center (NSIDC) in Boulder, Colorado, for providing the SMMR and SSM/I EASE-Grid brightness temperature data for this research project.

References