Comparison of different retrieval techniques for melt ponds on Arctic sea ice from Landsat and MODIS satellite data

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ABSTRACT. Melt ponds are regularly observed on the surface of Arctic sea ice in late spring and summer. They strongly reduce the surface albedo and accelerate the decay of Arctic sea ice. Until now, only a few studies have looked at the spatial extent of melt ponds on Arctic sea ice. Knowledge of the melt-pond distribution on the entire Arctic sea ice would provide a solid basis for the parameterization of melt ponds in existing sea-ice models. Due to the different spectral properties of snow, ice and water, a multispectral sensor such as Landsat 7 ETM+ is generally applicable for the analysis of distribution. An additional advantage of the ETM+ sensor is the very high spatial resolution (30 m). An algorithm based on a principal component analysis (PCA) of two spectral channels has been developed in order to determine the melt-pond fraction. PCA allows differentiation of melt ponds and other surface types such as snow, ice or water. Spectral bands 1 and 4 with central wavelengths at 480 and 770 nm, respectively, are used as they represent the differences in the spectral albedo of melt ponds. A Landsat 7 ETM+ scene from 19 July 2001 was analysed using PCA. The melt-pond fraction determined by the PCA method yields a different spatial distribution of the ponded areas from that developed by others. A MODIS subset from the same date and area is also analysed. The classification of MODIS data results in a higher melt-pond fraction than both Landsat classifications.

INTRODUCTION

During the Arctic summer, the occurrence of melt ponds on Arctic sea ice is a periodic event. Melt ponds can cover up to 50% of the sea-ice area (Fetterer and Untersteiner, 1998). The albedo of a melt-pond-covered surface is substantially lower than that of a snow-covered or bare sea-ice surface. Typical wavelength-integrated albedo values for mature melt ponds are 0.2–0.3. The values for younger melt ponds are higher and the value for fresh snow is 0.8 (Perovich, 1996). Melt processes on the surface decrease the overall averaged wavelength-integrated albedo of sea-ice area from ~0.8 to 0.5 (Perovich and Tucker, 1997). Melt ponds therefore play an important role in the ice-albedo feedback. Increasing temperatures and radiation during summer cause a reduction in the snow and ice extent, resulting in a decrease in the surface albedo and increasing the amount of sunlight absorbed by the surface. This in turn results in additional melting (Curry and others, 1995). Since the ice-albedo feedback has a significant influence on Earth’s radiation balance, it is important to represent the fraction of melt ponds and albedo of the sea-ice-covered area in sea-ice and climate models.

In the Arctic region, melting of sea ice generally starts with the increasing solar radiation at the beginning of June. By mid-June, a significant fraction of pack ice is covered by melt ponds. Due to the increasing heat transmission in ponded areas, the ponds deepen and can even melt through the ice layer. Freezing starts in late August or early September and melt-pond fraction decreases (Fetterer and Untersteiner, 1998).

The Arctic sea-ice cover is a small-fractioned inhomogeneous construct with surfaces varying from snow-covered ice, bare ice and melt ponds to open water; ice thickness ranges from zero (open water) to several metres (perennial level ice) to tens of metres (ridges). All these categories have different physical and optical properties. To determine the albedo changes of a larger area during the melt season, knowledge of the temporal and spatial variability of albedo in each of these ice categories is essential (Perovich and others, 2002b).

In the past, several field experiments have been conducted at different locations on the Arctic sea ice to study the spectral characteristics of melt ponds (Morassutti and LeDrew, 1996; Perovich and others, 2002b) as well as their distribution and size (S. El Naggar and others, http://epic.awi.de/epic/Main?puid=18004; Perovich and others, 2002a, 2009). Knowledge of the spectral behaviour of ponded sea ice in comparison to bare or snow-covered sea ice was used by Tschudi and others (1997, 2001, 2008) and Markus and others (2002, 2003) to determine the fraction of melt ponds from video imagery or satellite data.

In this study we develop a new technique based on principal component analysis (PCA) to identify the melt-pond fraction from Landsat satellite data. The results are compared with a classification method based on tie points utilized by Markus and others (2003) and with analysis of a MODIS (Moderate Resolution Imaging Spectroradiometer) scene developed by Tschudi and others (2008). We discuss the feasibility of archiving an accurate and complete melt-pond dataset throughout the seasonal cycle from Arctic sea-ice coverage.

DATA

Satellite data from Landsat 7, Terra and Aqua are used for this analysis. We selected an area in the northern Beaufort Sea. This area is covered by a Landsat scene acquired on 18 July 2001 (Fig. 1). In the corresponding true-colour subset (Fig. 2a) open water and individual ice floes can be distinguished by eye. The reflectivities of the ice fraction...
The ETM+ sensor on board the Landsat 7 satellite is an opto-
mechanical multispectral scanner. The incoming reflected
radiation passes through a telescope and is separated into
seven spectral bands (reflective bands 1–5 and 7 with 30 m
spatial resolution and thermal band 6 with 60 m spatial
resolution) and a panchromatic band (band 8) with 15 m
spatial resolution. The received electromagnetic energy is
converted into 256 discrete levels referred to as digital
numbers (DNs). The revisit period of the satellite is 16 days.
The inclination is 98.2°; this implies that acquisitions up to a
latitude of 82° N are possible.

An extensive archive search of the US Geological Survey
(USGS) Landsat archive was conducted to determine
whether an overall seasonal coverage of the Arctic region
would be possible with Landsat data. Unfortunately, the
archive mainly covers coastal areas and not the open sea-ice
area. The long revisit time of 16 days and the high number of
cloudy days further limit the number of usable scenes.

The 8-bit Landsat sensor has severe problems with the
saturation of the sensor over snow- and ice-covered surfaces.
The saturation of the sensor is also related to the solar
elevation (Bindschadler and others, 2008); saturation prob-
lems increase with sun elevation. The Landsat 7 Enhanced
Thematic Mapper Plus (ETM+) scene used in this study is
located in the upper Beaufort Sea at 80–82° N from 18 July
2001 with a sun elevation angle of 28.7°. This dataset
contains no saturated pixels. A 25 km × 25 km subset of the
full scene was analysed. Within this subset, training areas
covering all necessary features such as open water (A),
snow-covered ice (B) and two types of melt ponds (C and D)
are visually identified from the true-colour composite
(Fig. 2a). We assume that they reasonably represent the
specific classes.

The ETM+ sensor is calibrated on a regular basis to
maintain an accurate conversion of the DN values to the at-
sensor radiance. The calibration coefficients are included in
the metadata file of each Landsat scene (NASA, http://
landsathandbook.gsfc.nasa.gov/handbook.html).

To calibrate the DN to absolute radiances, we use

\[
L(\lambda) = \frac{L_{\text{max}}(\lambda) - L_{\text{min}}(\lambda)}{255} \text{DN} + L_{\text{min}}(\lambda),
\]

where \( L(\lambda) \) (W (m² µm)⁻¹) is the spectral at-sensor radiance
and \( L_{\text{max}}(\lambda) \) and \( L_{\text{min}}(\lambda) \) are the spectral radiances for each
band corresponding to DN = 0 or 255, respectively. These
have to be extracted for each band and are also included in
the metadata file of each scene. To account for some
between-scene variations, the radiance data were converted
into planetary or top-of-the-atmosphere reflectance, \( \alpha_{\text{TOA}} \)
(dimensionless), using

\[
\alpha_{\text{TOA}} = \pi L(\lambda) d^2 / E_s(\lambda) \cos \theta_s,
\]

where \( L(\lambda) \) is the spectral radiance at the aperture of the
sensor, \( d \) is the Earth–Sun distance in astronomical units,
\( E_s(\lambda) \) is the mean solar exoatmospheric irradiances (W (m²
µm)⁻¹) and \( \theta_s \) is the solar zenith angle in degrees (90° – solar
elevation angle). This information, as well as the Earth–Sun
distance, the spectral band radiances and the solar spectral
irradiances, is found in the Landsat 7 Science Data Users
handbook.html).

Landsat data are delivered as GeoTiff files in Universal
Transverse Mercator (UTM) map projection. For comparison
with different sea-ice products, the data are reprojected to a
polar stereographic grid according to the specifications of
the US National Snow and Ice Data Center (NSIDC). For
comparison with MODIS data, a grid size of 50 m × 50 m
spatial resolution was chosen and the data were resampled
with a nearest-neighbour algorithm.

MODIS

The MODIS sensor is an optical instrument on board the
Terra and Aqua platforms. The satellites were launched on
18 December 1999 and on 4 May 2002, respectively, in a
sun-synchronous near-polar orbit at an altitude of 705 km.
Terra operates on a descending node with a mean equator-
crossing time of 1030 h; Aqua operates on an ascending
node with an equatorial overpass at 1330 h. MODIS has 36
spectral bands ranging from 459 to 14 385 nm with a spatial
resolution of 250 m × 250 m to 1 km × 1 km.

The MODIS Level 1B dataset used for this study was
acquired on 18 July 2001 at 2315 h UTC. It contains calibrated
and geolocated at-aperture radiances (W (m² µm)⁻¹) and is

Fig. 1. Overview of the study area with the location of the selected
Landsat 7 scene from 18 July 2001 and the subset area (black
square) used for this study.

vary significantly. Snow-covered areas and/or bare white ice
appear white. With the onset of melting and wetting of the
snow and the ice surface, the colour spectrum varies across
shades of grey and blue. Shallow melt ponds appear in light
blue and develop from a green-blue tone to a darker blue
colour when they have reached their full extent (Markus
and others, 2002).

Landsat 7 ETM+

The ETM+ sensor on board the Landsat 7 satellite is an opto-
mechanical multispectral scanner. The incoming reflected
radiation passes through a telescope and is separated into
seven spectral bands (reflective bands 1–5 and 7 with 30 m
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generated from MODIS level 1A data (http://modis.gsfc.nasa.gov/data/dataprod/). After removing the bow-tie effect, the MODIS data were also reprojected to the NSIDC polar stereographic grid and resampled with a nearest-neighbour algorithm to 250 m\(^2\) cell size. For this study, a 25 \(\times\) 25 km subset, covering the same coordinates as the Landsat subset, was used.

**METHODS**

The optical properties of ice and snow are a strong function of the wavelength of the incident solar radiation, \(\lambda\) (Fig. 3). Furthermore, the sea-ice albedo is affected by the incident solar zenith and azimuth angle and the viewing geometry of the satellite sensor (Warren, 1982). The influence of incidence and detector angles on the reflectance, expressed as bidirectional reflectance distribution function (BRDF), is the subject of future investigations and is therefore not considered further in this analysis.

The spectral- as well as the wavelength-integrated albedos of bare ice are fairly constant during the melting period. On the other hand, the albedo of ponded ice depends on the pond depth and varies throughout the melting period (Perovich and others, 2002b).

It is evident from Figure 3 that the curves for snow-covered ice, bare ice and melting bare ice demonstrate a smaller reduction in spectral albedo at longer wavelengths than the curves for two types of melt pond. At short wavelengths the reflectivity remains high compared to the longer wavelengths. These spectral differences can be used to separate melt ponds from bare or snow-covered ice.

To calculate the albedo value on a larger scale, the surface-based albedos are weighted with the fraction of their corresponding surface component (Perovich and others, 2009). The so-called areally averaged albedo, \(\overline{\alpha}\), can be expressed:

\[
\overline{\alpha} = \alpha_l A_l + \alpha_p A_p + \alpha_i a_i
\]

where \(A\) is the area fraction, \(\alpha\) is the wavelength-integrated albedo and \(l, p\) and \(i\) represent values for leads, ponds and (bare) ice, respectively. Wavelength-integrated or total albedo values of different surface types have been extensively reported in the literature (e.g. Perovich, 1996; Perovich and others, 2002b; Tschudi and others, 2008).

**Fig. 2.** (a) 25 km \(\times\) 25 km true-colour subset (band combination 3–2–1) from the Landsat scene displayed in Figure 1 with the selected training areas. This subset is used for all analysis. The red box shows the zoom for comparison (c). The coordinates are the gridcell numbers. (b) Melt-pond fraction of the Landsat subset determined with PCA. Black areas are identified as open water. (c) Melt-pond fraction of the Landsat subset determined using the method described by Markus and others (2003). Black areas are open water. (d) Melt-pond fraction determined from a MODIS subset of the same date and area. (e) From left to right: true-colour zoom; results of the PCA; results using the method of Markus and others (2003); differences between the two methods; and results from MODIS. The colour bar represents the values of the differences.
Perovich and others (2009) report albedo values of 0.07 for leads/open water, of 0.65 for bare ice and 0.25 for melt ponds.

Algorithm based on principal component analysis

The spectral differences of melt ponds in comparison to bare or snow-covered sea ice and open water are used to detect melt ponds. From Figure 3 it is clear that these differences increase with longer wavelengths. We therefore used the spectral bands that could best resolve these differences. Figure 4 shows a two-dimensional (2-D) histogram plot of band 1 vs band 4 of the above-described Landsat scene. Included are pixels of the determined training areas. In this plot, a maximum of pixels can be detected at high reflectances of both bands. A second maximum is located at the low reflectances of both bands.

Area A represents open water, B represents snow-covered ice and C and D represent two types of melt ponds. It can be observed that the main quantity of the pixels, including classes A and B, lies on a main axis. The values are highly correlated between the two bands on this main axis; they represent the first principal component. As well as the principal component, some pixels form another cluster: the values in band 4 are lower than their corresponding values in band 1. They describe pixels with an included melt-pond fraction and represent the second principal component. Classes C and D can therefore be found in this cluster.

To separate the pixels with a melt-pond fraction from the pixels that only have a snow and water signature, PCA is applied. First, the mean vector of the first principal component \( g_1 \) and a rotation matrix must be defined. The second principal component \( g_2 \) is perpendicular to \( g_1 \). Generally, \( g_1 \) is evaluated from all pixels of the dataset; for this analysis, however, the mean vector is defined only for the main axis (open-water/ice pixels) using the two maxima in the 2-D histogram plot (Fig. 4).

The mean vector can be described with the linear equation

\[
f(x) = mx + b, \tag{4}\]

where \( m \) is the slope and \( b \) the offset. The rotation matrix is defined:

\[
R = \begin{bmatrix}
\cos \gamma & -\sin \gamma \\
\sin \gamma & \cos \gamma
\end{bmatrix}, \tag{5}
\]

where

\[
\gamma = -\arctan m. \tag{6}
\]

The individual components \( g_1 \) and \( g_2 \) (also called eigenvectors) describe the principal component axis of the original coordinate system. After the transformation matrix is applied, the pixels are positioned in a new coordinate system where the eigenvectors \( g_1 \) and \( g_2 \) become the new axes \( G_1 \) and \( G_2 \) (Fig. 5). The data now exhibit no discernible correlation between the main axis and the cluster of melt ponds. Furthermore, the transformed dataset allows discrimination of the water-ice axis and the cluster.

A threshold to separate the water/ice pixels from the melt-pond pixels is defined. This runs parallel to the open-water/ice axis and is defined as:

\[
G_1 < 0.2 \text{ and } G_2 = \max(G_2). \tag{7}\]

For the analysed Landsat subset, the threshold value is set as \( G_2 = 0.3 \). All pixels greater than the given threshold will be assigned as pixels containing melt ponds (melt-pond pixels).

In the present case, melt-pond areas are very well developed and it can be assumed that at least one pixel contains 100% melt pond. All other pixels are only
fractionally ponded. The melt-pond pixels are scaled between 0% and 100% (Fig. 5).

To determine the melt-pond fraction (the fraction of ponds on the sea ice), the open-water area needs to be defined and subtracted from the overall area. To determine the open-water area in Landsat data, a threshold technique was applied as described by Steffen and Schweiger (1991) and Cavalieri and others (2006).

For the error estimation, we selected all mixed pixels at the border of open water and the ice edge and assigned them 50% misclassification. This led to an error of $\pm 0.35\%$ for the open-water area for this subset.

RESULTS

The results from the classification method based on PCA are depicted in Figure 2b. Comparison of the melt-pond fraction with the true-colour composite shows good agreement.

Melt ponds cover 106.3 $\pm$ 23.9 km$^2$ of a total sea-ice area of 623.2 km$^2$ $\pm$ 2.2 km$^2$; this is $17.1 \pm 3.8\%$ of the sea-ice area.

Markus and others (2002, 2003) showed that the differences between spectral bands 2 and 3 of the Landsat 7 ETM+ sensor can also be used to detect melt ponds. They discovered that differences between bands 2 and 3 ($\Delta_{23} = \alpha_{b2} - \alpha_{b3}$) above a certain threshold can be classified as ponded area. Values below this threshold can be determined as open water and unponded ice areas.

To separate between open water and ice, the information from band 1 ($\alpha_{b1}$) is used. The high values are classified as ice and the low values as water. For our Landsat 7 subset, $\Delta_{23}$ is plotted against $\alpha_{b1}$ and the scatter plot is divided into three regions using the following threshold and tie points: $\Delta_{23} = 0.06$ separates melt ponds from open water/ice; additionally, $\alpha_{b1} = 0.16$ is defined as 100% open water and $\alpha_{b3} = 0.82$ as 100% ice. If $\Delta_{23} \geq 0.06$ and $\alpha_{b1} = 0.45$, the pixel is 100% melt pond (Fig. 6).

Figure 2c depicts the results of the classification using the method of Markus and others (2003). The melt-pond fraction conforms with the results of PCA and the true-colour composite. The coefficient of determination is $r^2 = 0.94$. Melt ponds cover $88.8 \pm 10.6$ km$^2$ of a total sea-ice area of $23.2 \pm 2.2$ km$^2$; this is $14.3 \pm 3.3\%$ of the sea-ice area.

Note from Figure 2e that the results show differences in the spatial distribution of the melt ponds. In areas with an apparently high melt-pond fraction, the results of the method developed by Markus and others (2003) give lower values than for PCA (red areas). By contrast, in areas with a lower melt-pond fraction, the method of Markus and others yields higher values (blue areas). The blue pixels in the difference plot (Fig. 2e, right) represent the greyish pixels in the true-colour subset (Fig. 2e, left).

The threshold in this analysis which divides pixels into ponded and unponded sea ice is the input parameter. The uncertainty of this method is mainly defined by the threshold value $G = 0.3$. The threshold in this procedure is dependent upon the solar zenith angle, the geographical latitude and subsequent spectral reflectance of the surface, and must be defined for each satellite scene. It is therefore important to investigate how sensitive the results are to a variation of $G_{2}$ and $\Delta_{23}$ by $\pm 1\%$ and $\pm 5\%$ for the method of Markus and others (2003).

For $G_{2} = \pm 1\%$ and $\pm 5\%$, the total melt-pond area varies by 4.75% and 22.45%, respectively. The linear increase of the error in both procedures shows that the distribution of the melt-pond fraction is in this case uniform. The calculated errors for the PCA method are about 0.5% lower than for the method of Markus and others (2003). The lower values for the PCA method are due to a better-defined intersection for the separation of the melt-pond fraction and the snow/water fraction. It can be seen from Figure 6 that the ice/open-water axis is not exactly parallel to the threshold. This effect causes the slightly more diffuse intersection between the two fractions.

Tests in the full Landsat scene implied that a variation of $\pm 5\%$ for the threshold is realistic. This value was therefore used for the error calculation.

Tschudi and others (2008) describe a technique which uses a set of linear equations to analyse MODIS data in order to estimate different fractions for each pixel of a MODIS scene. Given the relatively coarse resolution of the MODIS sensor, we can assume that each MODIS pixel could contain
open water, ice and melt ponds.

$$\left[ \sum a_i r_i = R \right]_k$$ where \( \sum a_i = 1 \),

(8)

where \( R \) is the reflectance, \( a_i \) are the fractional coverages for each MODIS pixel for band \( k (k = 1, 3, 4) \) and \( r_i \) are the spectral reflectances of each surface type.

Since linear equations do not always give a definite result, we introduced a condition to restrict the interval of the solution between 0 and 1. This also avoids negative values for the different classes.

The surface types determined in the Landsat 7 ETM+ scene are used for the classification. The spectral reflectance values for each surface type, \( r_i \), also evaluated from this Landsat scene, are listed in Table 1.

For their analysis, Tschudi and others (2008) use an atmospheric-corrected MODIS surface-reflectance product (MOD09). For better comparison of the results from the MODIS subset with the Landsat scene, however, bands 1, 3 and 4 from the MODIS Level 1B product are used here.

Figure 2d depicts the result of the classification using the method introduced by Tschudi and others (2008). The maximum error for this method is \( \approx 10\% \). Melt ponds cover \( 183.9 \pm 18.4 \text{ km}^2 \) of a total sea-ice area of \( 587.8 \pm 58.8 \text{ km}^2 \); this is \( 29.4 \pm 2.9\% \) of the sea-ice area.

**DISCUSSION**

Table 2 lists the results of the calculated melt-pond fraction from three different methods and two different datasets. The Landsat methods are consistent. The main difference appears in the comparison of the Landsat methods with the MODIS classification. The melt-pond fraction is clearly higher for the MODIS classification than for both Landsat classifications. This behaviour has been described by Tschudi and others (2008).

From examination of the MODIS classification (Fig. 2d and e), we note that the results over sea-ice areas are consistent but that a misclassification occurs at the border with open-water areas. Instead of a water and a snow/ice fraction, the algorithm detects a very high melt-pond fraction and a very low snow/ice fraction. This needs to be investigated in detail in the future.

The differences between the two Landsat methods could result from the fact that three spectral bands (bands 1–3) are used by Markus and others (2003). They only cover the short wavelength range of the spectral reflectances of the surface features. For the PCA method, only bands 1 and 4 are used.

**CONCLUSION**

Very high-resolution satellite data can be used to determine melting features on Arctic sea ice. The limitations are many, however. An extensive search of the USGS archive for suitable Landsat scenes results in only a few usable images. On closer inspection of the data, it became clear that nearly all scenes that contain melt ponds have problems with saturated pixels. Saturation is related to sun elevation and the surface type. Unfortunately, melt-pond development increases when the sun elevation is high. There are procedures to interpolate between the saturated and unsaturated bands, but they could not be easily applied since the different spectral behaviour of the surface types has to be incorporated. Other disadvantages are the revisit time of 16 days and the fact that the Arctic is an area with an above-average proportion of cloudy days.

The above-listed points demonstrate that very high-resolution satellite data can be used to detect melt ponds or provide a basis for comparison with lower-resolution satellite data, but not as the only source for a seasonal melt-pond dataset over the entire Arctic. Lower-resolution sensors such as MODIS obtain a full coverage of the area on a daily basis. To acquire a melt-pond dataset of the entire Arctic, a combination of high- and medium-resolution sensors is recommended. **ACKNOWLEDGEMENTS**

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