Mapping glacier velocities on Svalbard using ERS tandem DInSAR data

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Introduction

Detailed knowledge of glacier velocity fields is important to increase knowledge of glacier dynamics and to verify models dealing with this topic. Glacier surface velocities are also an important part of mass balance modelling of glaciers and therefore constitute an important parameter for monitoring how glaciers respond to a changing climate. Many of Svalbard’s glaciers are of the surge type, which makes glacier velocity monitoring even more important. Estimates of the percentage of surge-type glaciers vary from 15 to 90 (Liestøl 1969, Lefauconnier & Hagen 1991, Hagen et al. 1993, Hamilton & Dowdeswell 1996, Jiskoot et al. 2000). Satellite remote sensing of glacier velocities is less expensive and covers larger areas than traditional surveying in remote areas such as the Arctic. Other remote sensing techniques, such as feature tracking in optical satellite imagery or aerial photos, require areas of high contrast in order to work (Lefauconnier et al. 1994, Kääb & Funk 1999). Hence, feature tracking only works in crevassed or debris-covered areas on glaciers, and serves as a complimentary technique to InSAR, which works best in more homogenous and stable areas.

Glacier surface velocities were first measured with data from satelliteborne InSAR in 1993 (Goldstein et al. 1993). The InSAR technique has been extensively used for mapping glacier surface velocity; see Massonnet & Feigl (1998) and König et al. (2001) for an overview. The launch of ERS-2 in 1995 made available tandem scenes from ERS-1 and ERS-2 taken only one day apart. The short time interval between acquisitions of the SAR scenes increases the possibility for detection and mapping of glacier movements as there is less influence from temporal factors such as melting. The scenes used in this study are from the period of the tandem mode in 1995 and 1996.

The main goals of this study were to develop a method for unwrapping existing DInSAR data of three Svalbard glaciers, to estimate glacier velocities from the unwrapped DInSAR data, and to decompose these velocities into the direction of glacier surface flow using a photogrammetrically derived DEM. All these tasks were performed using the ESRI GIS software Arc and no interferometry software was used.

The glaciers investigated in this study are Isachsenfonna, Nordbreen and Akademikberreen (Fig. 1). Isachsenfonna is the only one of these for which previous velocity data exists and the InSAR scenes of this glacier were therefore chosen to develop the method.

Interferometric SAR

InSAR uses the phase information in radar acquisitions from two receiving antennas to produce an interferogram (InSAR-scene). The two antennas can be separated in either time (repeat track acquisition) or space (along- or across-track). An interferogram is a 2-dimensional representation of the measured radar phase difference between the two antenna positions and corresponding positions on the Earth’s surface. Several factors contribute to the phase difference, with ground surface topography and ground surface deformation being the two most important.

Geometry

The geometry of a satellite InSAR acquisition is shown in Fig. 2. The first SAR scene is acquired in orbit O1 and the second in orbit O2. These two orbits are separated by the spatial baseline B, which can be decomposed into Bx and By. The difference in distance from a point on the ground to the two antennas (r2 and r1) can be expressed as a certain number of wavelengths and a fraction of a wavelength: a phase difference. This is the phase difference depicted in the
InSAR scene and can be written as the following (Gens & van Genderen 1996):

\[ D_f = \frac{4 \pi p l}{r_2 / C_0 - r_1} = \frac{4 \pi p l}{B_x \sin \theta - B_y \cos \theta} \]  

(Eq 1)

Here, \( \theta \) is the look angle and \( \lambda \) the wavelength (Fig. 2).

**Unwrapping**

The phase difference is measured modulo \( 2\pi \); in intervals from 0 to \( 2\pi \). It is the shifts from \( 2\pi \) to 0, or 255 to 0 in terms of grey values, that create the characteristic fringe pattern. Converted to wavelengths, one fringe equals half a wavelength due to the double pathway. The fringes represent ambiguities and should therefore be unwrapped. The principal method of all-phase unwrapping is that the phase differences are integrated by adding multiples of \( 2\pi \) when a fringe transition is crossed (i.e. when the phase difference wraps from \( 2\pi \) to 0). A more thorough description of the different unwrapping methods and algorithms is given in Gens & van Genderen (1996). In principle, after unwrapping, it is possible to create a DEM by using one known elevation point together with a given spatial baseline to deduce the elevations for every pixel in the whole scene.

**Differential interferometry**

Satellite InSAR scenes are usually based on two scenes acquired some time apart (temporal baseline). If surface deformation has taken place during this time interval it will also cause phase differences. The phase difference as noted in Eq. 1 could therefore be expanded with a second part for the deformation (Kwok & Fahnestock 1996):

\[ D_f = \frac{4 \pi p l}{B_x \sin \theta - B_y \cos \theta} + \frac{4 \pi p l}{\Delta \rho} \]  

(Eq 2)

where \( \Delta \rho \) is the deformation component in range during the temporal baseline. As can be seen from Eq. 2, the contribution from the deformation is only dependant on the temporal baseline and not the spatial baseline. In an interferogram where deformation has taken place one has to isolate the contribution from one of the two components topography or deformation to utilize the other. This is called differential interferometry or DInSAR and can be done by using two interferograms (three SAR-scenes) (Kwok & Fahnestock 1996) or one interferogram and a DEM (Eldhuset et al. 2003).

**Limitations**

A suitable interferogram can only be obtained if the two signals received from the same object are coherent (i.e. not decorrelated). Decorrelation of the spatial baseline is a factor that is governed by acquisition parameters, and can be avoided by choosing image pairs with a spatial baseline shorter than the critical spatial baseline, and for ERS-1 and ERS-2 this is \( \approx 1.1 \) km.

Another effect of the spatial baseline length is that shorter baselines will result in less contribution from topography. There are examples of applying interferograms with spatial
baselines of only a few meters or for nearly flat ice-sheet surfaces, and thereby completely disregarding the topographic effect (Goldstein et al. 1993, Michel & Rignot 1999).

Temporal effects are changes that occur at the Earth’s surface during the time between the acquisitions and these will also have an impact on the coherence of the interferogram. These could be changes in the reflective properties of the material at the surface. Changes in the water content will drastically alter the dielectric properties of the ground and, hence, also the backscatter of the electromagnetic waves. Motion on the Earth’s surface could also decorrelate the image if it is too chaotic, having a discrete rather than a continuous fashion. Glaciers moving with high velocities (several metres per day) and that are heavily crevassed will often decorrelate (Weydahl 2001).

Mapping glacier velocities by satellite SAR interferometry

Goldstein et al. (1993) demonstrated the possibilities of satellite SAR interferometry for mapping the glacier velocity of an Antarctic ice stream. The technique has been widely used for mapping glacier movement on Greenland and Svalbard, and in the Alps (Joughin et al. 1995, Eldhuset et al. 1996, Rott & Siegel 1996, Unwin & Wingham 1997). All of these studies have only measured the component of the glacial motion in the direction towards the satellite, also called the line of sight component. Traditional ground measurements of glacier velocity are done in the direction of movement. Rignot et al. (1995) used an existing altimetric DEM to compute slope gradient and slope direction in Western Greenland. They also made the normal assumption of glacier movement being surface parallel and calculated velocities that were within 6% of ground measurements. Kwok & Fahnestock (1996) used two interferograms to separate motion and topography. The resulting DEM was used to decompose the glacier movement into the direction of slope.

When measuring movement with InSAR data, only the component of motion in the slant range (i.e. radar beam or line of sight) direction is measured. If all motion is taking place in the azimuth direction, there will be no component in slant range, and hence no motion-induced contribution to the interferogram. Mohr et al. (1998) combined interferograms from ascending and descending orbits to map an ice stream on Greenland in three dimensions, thereby omitting the problem of measuring motion in the azimuth direction. They also assumed surface parallel flow.

Study sites

All the glaciers used in this study are situated at Spitsbergen, the largest island of the arctic archipelago Svalbard (Fig. 1). Svalbard is situated between 74°N–84°N and 10°E–35°E in the North Atlantic, and has permafrost conditions down to sea level (Liestøl 1977). Glaciers of different types cover 60% of the archipelago.

All the following data about the glaciers are taken from Glacier Atlas of Svalbard and Jan Mayen (Hagen et al. 1993). The method was developed using a DInSAR scene of Isachsenfonna (78°85’N, 13°11’E), shown as site 1 in Fig. 1. This glacier is situated close to Kongsfjorden and the research settlement in Ny-Ålesund. Isachsenfonna is classified as an outlet from an ice cap with a composite firn area and a fast-retreating calving front. The equilibrium line altitude (ELA) is c.670 m a.s.l. The ELA is in the lower part of the area where the velocity mapping was performed. Isachsenfonna merges with Holtdahlsfonna and is called Kronebreen (the latter two are not named on Fig. 1) further down.

The second glacier, Nordbreen (‘2’ in Fig. 1), is considerably smaller than the other two. It is situated in the north-eastern part of Spitsbergen (79°38’N, 16°10’E). Nordbreen is an outlet from the Asgårdsfonna ice cap, with one firm area and a calving front in Wijdefjorden. The front is probably retreating slowly. The ELA is c.450 m a.s.l.

For site 3 in Fig. 1, the name Akademikerbreen is used as a joint name for Transparentbreen, Opalbreen (the latter two are not named on Fig. 1), the lower part of Akademikerbreen and parts of Negribreen and Lomonosovfonna. Since Akademikerbreen occupies the greater part of the InSAR scene, this name is used for reference purposes. The centre of the scene is situated in the eastern part of Spitsbergen (78°40’N, 18°40’E) and the glacier is classified as an outlet from an ice cap with more than one firm area and a fast-retreating calving front. The area of the whole Negribreen glacier system is 1180 km², the length is 41 km and the volume 250 km³. The ELA is c.360 m a.s.l.

Due to their position close to the research settlement in Ny-Ålesund, the glaciers in the Kongsfjorden area are among those that have been most investigated on Svalbard. The glaciers of site 1 are situated within this area. The lower part of Kronebreen, which drains Isachsenfonna, is heavily crevassed and moves at a very high velocity. Due to this, the interferograms decorrelate in this area. Using SPOT-imagery, Lefauconnier et al. (1994) found velocities between 0 and 2.15 m d⁻¹ on the tongue of Kronebreen. Maximum summer velocities of 4.5 m d⁻¹ have been measured at the very front (Voigt 1965, Lefauconnier et al. 1994) and this is much higher than for other similar glaciers on Svalbard. However, the high velocity is not sufficiently high compared with the calving rate and the front has therefore been retreating over the last 200 years. Due to the high velocity, the glacier is not believed to be frozen to its bed (Liestøl 1988). The velocity measurements referred above were done below the areas where velocities are mapped in this study.

Lefauconnier et al. (2001) have reported on a GPS-measured displacement for three stakes at Holtdahlsfonna from May 1996 to May 1997 that corresponds to 0.10 to 0.15 m d⁻¹. Unfortunately, the stakes are situated just outside the area mapped in this project. Lefauconnier et al. (2001) also report on nearly constant velocities ranging from 0.20 to 0.22 m d⁻¹ for Isachsenfonna, measured with satellite SAR interferometry from ERS-1 scenes acquired six and nine days apart in September and October 1991. Velocities ranging from 0.08 5 to 0.52 m d⁻¹ were obtained from the same ERS-1 scenes for Holtdahlsfonna, having the highest velocities in the lower part were the Holtdahlsfonna becomes Kronebreen.
Table 1. Data on the InSAR-scenes used in the study. $B_n$ is the orthogonal satellite baseline in metres.

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Date (ERS1/ERS2)</th>
<th>Orbit (ERS1/ERS2)</th>
<th>Frame</th>
<th>Track</th>
<th>$B_n$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Isachsenfonna</td>
<td>10/11.10.96</td>
<td>23472/3799</td>
<td>1989</td>
<td>438</td>
<td>−31</td>
</tr>
<tr>
<td>Isachsenfonna</td>
<td>05/06.04.96</td>
<td>24703/5030</td>
<td>1989</td>
<td>166</td>
<td>−10</td>
</tr>
<tr>
<td>Akademikerbreen</td>
<td>20/21.10.95</td>
<td>22989/2625</td>
<td>1971</td>
<td>266</td>
<td>−22</td>
</tr>
<tr>
<td>Nordbreen</td>
<td>27/28.09.95</td>
<td>21969/2296</td>
<td>1971</td>
<td>438</td>
<td>−130</td>
</tr>
<tr>
<td>Nordbreen</td>
<td>05/06.04.96</td>
<td>24703/5030</td>
<td>1971</td>
<td>166</td>
<td>−12</td>
</tr>
</tbody>
</table>

Data set

From the launch of ERS-2 in 1995 until ERS-1 ended its eight-year service in 1999 it was possible to obtain InSAR-scenes combined from the two satellites. Both satellites have a repeat cycle of 35 days. During the nine-month tandem mode mission from August 1995 to May 1996, the satellites had the same orbit and repeat cycle, but followed each other one day apart. It is data from this tandem period that is used in this study.

The InSAR scenes were processed at the Norwegian Defence Research Establishment (FFI) with data from an ESA AO (European Space Agency Announcement of Opportunity) Tandem project. Using a DEM delivered by the Norwegian Polar Institute, the topographic influence was removed during the processing at FFI to produce height differentiated InSAR scenes (DInSAR). The method is more thoroughly described by Eldhuset et al. (2003). This DEM was also used in the decomposition of glacier velocities. The DEM is interpolated from photogrammetric data with an equidistance of 50 m. The relative error of the DEM is believed to be $c.10 \text{ m}$ for glacier surfaces (see Discussion). Implications of this error on the resulting velocities will be discussed later in this article. The DInSAR scenes used in this study were chosen from a larger number of scenes available. Scenes from several glaciers were chosen in order to be able to test the method on more than one glacier. The InSAR-scenes used in the study are listed in Table 1. The spatial baseline $B_n$ corresponds to the orthogonal baseline defined by the European Space Agency (ESA) and shown in Fig. 2. ESA is also the source of the baseline data shown in Table 1 (http://odisseo.esrin.esa.it; accessed September 2005). Spatial baselines with decimetre accuracy were used during the InSAR processing at FFI. These precise orbit (PRC) data were obtained from ESA.

The ground resolution of the three first scenes in Table 1 is $16 \text{ m} \times 20 \text{ m}$. The three last scenes were processed in a different way and have a ground resolution of $64 \text{ m} \times 80 \text{ m}$. Since both the DInSAR-scenes and the DEM are in SAR-geometry, all the calculations were done in this geometry. Georeferencing and transformation to a map projection was not done since it was not necessary in order to demonstrate the method.

Methods

The method for calculating glacier velocities was developed as a semi-automatic algorithm using DInSAR-scenes of Isachsenfonna. The method was then applied to the rest of the InSAR-scenes. The flow chart of the method is shown in Fig. 3. All of the steps have been done using the GIS package ARC (Version 8.2).

In order to unwrap the DInSAR-scene, the discrete fringe transitions had to be identified. First, a $3 \times 3$ median filter is applied twice to reduce the characteristic speckle noise that is present in SAR and InSAR images. The result is a smoother InSAR image with the fringe transitions still intact. The raw and median-filtered DInSAR are shown in Fig. 4a and 4b respectively. The scenes from Nordbreen have been processed with 16-looks during the SAR processing at FFI, rather than 4-looks (as for the other scenes). This reduces the speckle noise considerably (at the cost of spatial resolution), and the noise level is consequently also reduced in the interferograms. It was therefore unnecessary to apply the median filter to the Nordbreen interferograms.

An edge-detecting filter is applied after the median filtering. It is a derivative filter that uses the Prewitt-operator.
to calculate the gradient of the pixel values in the directions of azimuth and range. The gradient filtered image is shown in Fig. 4c. Lin et al. (1994) also used the Prewitt-operator in addition to two similar masks to detect lines at an angle of 45°. Since the Prewitt-operator implemented in Arc gave satisfying results for the 45° lines, further detection of these was not pursued.

There is still some noise in the image after the Prewitt-filtering and the image is therefore thresholded. The user can define the threshold after looking at the histogram of the image. All pixels with values less than the threshold are set to 0, and the others are kept as they are. A threshold of 0.150 has been shown to be suitable. Determining the threshold is a choice between the amount of noise reduction and preservation of fringe lines. The thresholded image is shown in Fig. 4d. This process could also have been done automatically by using the histogram. This requires a bimodal histogram, which means that the grey-level
distribution has two peaks, representing the noise and lines respectively. The threshold could then have been chosen at the frequency minima between the peaks; see Gonzales & Woods (1993). However, in the histogram of this image there is a smooth transition between the noise pixels with the greatest values and line pixels with the lowest values, and hence the histogram is not bimodal.

The thresholded image is then vectorized with an existing algorithm in ARC. Points that are closer than a certain distance are connected with vectors. The resulting data have to be edited to some extent. This is similar to the edge-segment linking performed by Lin et al. (1994). During the automatic editing, short individual line segments are removed and small gaps in longer lines are snapped. Some of the holes in the image segments are longer than the smallest distance between two different fringe lines, and cannot be snapped. It is therefore not possible to close all the gaps automatically and some manual editing has to be done. The vectorized fringe lines are shown in Fig. 4e.

It is important for the calculations to determine the right boundary for the glacier. The boundary is digitized manually and used in all the steps following the median filtering to limit the amount of calculations that have to be done. It also serves as a reference line of zero velocity in the unwrapping. It is best to use one of the amplitude images to determine the glacier boundary; the DEM and the DInSAR scene are more difficult to apply.

When the lines are edited, the topology for the polygons that mark the transitions between the fringes can be created. The polygons are numbered according to their relative position. The outermost polygon is given the value 0, the next 1, etc. This information is used in the unwrapping of the InSAR image. Where half a wavelength (2.83 cm) is added to the median-filtered InSAR-scene values inside fringe one, one wavelength is added inside fringe 2, etc. The median-filtered InSAR-scene is used because it contains less noise and is believed to be closer to the real phase differences than the original DInSAR scene. In other studies it has been common to use the unfiltered interferogram (Lin et al. 1994). The generation of topology could ideally have been done automatically, but since some interpretation is needed, it is done manually. The unwrapped scene, showing displacements in centimetres towards the satellite, is shown in Fig. 4f.

The fringe lines found by edge-detecting techniques will never hit all the fringe transitions exactly. Some pixels that should have been on the outside are therefore caught on the inside, and vice versa. Pixels that end up on the wrong side will differ greatly in pixel value from their neighbours after the unwrapping. They will have pixel values that differ by about half a wavelength from their neighbours. It is therefore possible to remove these wrongly classified pixels by statistical comparison of a pixel and its neighbours. For reasons of simplicity, a median filter is used for this task as well. The median filtering will alter all the pixels in the image. However, since the variation of the data showed a rather continuous trend, the altering will be relatively small for the correctly classified pixels.

Since the scenes are already height differentiated, only one part of the expression in Eq. 2 is left, namely the contribution of glacier motion to the phase difference. It could then be written as (Kwok & Fahnestock 1996):

$$\Delta \phi_{\text{mot}} = \frac{4\pi}{\lambda} \Delta \rho = \frac{4\pi}{\lambda} \bar{v} \Delta T$$

where \(\bar{v}\) is the surface ground velocity in the direction of the radar beam, \(\Delta T\) the time between the acquisitions (temporal baseline) and \(\vec{r}\) the unit vector in direction of range. The unwrapped values, explained in the proceeding section, represent the displacement towards the satellite. A displacement calculated towards a satellite is of limited use from a glaciological point of view. It is therefore common to assume surface parallel flow and decompose the motion into the direction of surface slope.

Since the Norwegian Polar Institute had DEMs available, the velocities were decomposed in the flow direction of the glacier. Terrain slope is calculated together with aspect (i.e. direction of steepest surface slope). This is done automatically with a gradient filter. Before the terrain parameters could be calculated, substantial smoothing of the DEM had to be done, since the glacier flow direction is governed by large-scale changes in slope and aspect, and not small undulations in surface topography. Kamb & Echelmeyer (1986) found that, for modelling glacier velocities using stress-gradient coupling, the filters had to be 4 to 10 times the ice thickness. Using the 273 m depth found by Bamber (1987) in the area between Kronebreen and Isachsenfonna, the work by Kamb & Echelmeyer (1986) implies the application of smoothing filters covering up to 2.7 km. The radar look angle was also calculated. It varies between 20° and 26° from near-range to far-range.

When these three parameters, the slope of the glacier, the aspect angle between glacier slope and radar range direction, and the radar look angle, have been calculated, together with the unwrapped interferogram, it is possible to calculate the velocity, \(V_{\text{glac}}\), in the flow direction of the glacier (K. Eldhuset, personal communication 1998):

$$V_{\text{glac}} = \frac{U_w}{\cos z \cos \phi \sin \theta + \cos \theta \sin z}$$

\(U_w\) is the displacement values in the unwrapped scene, \(z\) is the slope of the glacier, \(\phi\) is the aspect angle between glacier slope and radar range direction, and \(\theta\) is the radar look angle. Generally, there should also have been a \(\Delta T\) (temporal baseline) term in the denominator of Eq. 4, but since it is tandem data and velocities are calculated in metres per day (m d\(^{-1}\)), this term is omitted. The geometry for decomposition of glacier movement is shown in Fig. 5. Since the glacier velocity in this study is defined as positive in the downward direction of the steepest slope, the denominator of Eq. 4 has a plus sign differing from the minus sign found in corresponding equations used in other studies, defining the velocity to be positive upwards (Kwok & Fahnestock 1996). This difference is only due to the definition of velocity direction and has no implications for the result. All the parameters are calculated for every pixel at the glacier surface, resulting in a complete velocity field for the whole glacier.
Results

The described method has been used to calculate the surface velocities along the surface-flow direction for three glaciers at Svalbard. Data are presented in Table 2. The velocity field based on the April 1996 scene of Isachsenfonna is shown in Fig. 6a. The maximum velocity is 0.42 m d$^{-1}$ in the images acquired in January 1996, having an average velocity for the whole glacier surface of 0.23 m d$^{-1}$. For the images acquired in April 1996, the maximum velocity is 0.42 m d$^{-1}$ and the average velocity is 0.18 m d$^{-1}$.

For Nordbreen, the maximum velocities vary between 0.33 and 0.36 m d$^{-1}$ and the average velocities vary between 0.10 and 0.17 m d$^{-1}$. The maximum velocities are found in the areas of the greatest slope. The velocity field derived from the InSAR-scene of Nordbreen acquired in April 1996 is shown in Fig. 6b.

The velocity field derived from the InSAR-scene of Akademikerbreen is shown in Fig. 6c. The maximum velocity is 0.41 m d$^{-1}$ and the average velocity is 0.07 m d$^{-1}$. There is a significant increase in the velocity in the narrow passage between the two nunataks in the centre of the scene. In this passage, the surface slope is also greater than for the rest of the glacier.

Discussion

Since the goal of this study was to develop a method, a detailed glaciological discussion of the results is beyond the scope of this article. However, the results can be used to assess the accuracy and the limitations of the method in glaciological applications.

Accuracy

The accuracy of the phase measurements with satellite SAR interferometry is believed to be 1.5 mm for vertical movements and 4 mm for horizontal movements (Goldstein et al. 1993). These accuracy values are in the direction of the radar beam (range) and will be larger for decomposed velocities. The errors due to unwrapping are considered negligible since local errors do not propagate across the fringe lines, and the filtering has removed most of the local noise. Errors in the DEM caused by interpolation and co-registration will affect the resulting interferogram. Eldhuset et al. (2003) found RMS differences between the DEM and the InSAR-derived DEMs, using the same tandem scenes as in this study, of 7 m to 20 m. This RMS difference is an expression of errors in the DEM, errors in the InSAR-derived DEM, and also surface elevation changes between the acquisitions of the two different DEMs. The average relative error in the DEM used for height differentiation is believed to be between 5 and 10 m, based on air photo scale and general knowledge on the map compilation and DEM interpolation procedure (T. Eiken, personal communication 2004). A relative error of 10 m in the DEM will give different errors for the end result for each glacier depending on baseline, terrain slope and glacier-satellite geometry. The errors for the different glaciers are in the range 1–10% of the calculated average velocities (Table 2).

Variation in water vapour in the atmosphere between the two acquisitions can also have an impact on the phase measurements, and hence also on the deformation measurements. Zebker et al. (1997) found that temporal and spatial variation in air humidity of 20% could cause errors in the deformation calculation of up to 10 cm. If such a variation of humidity is the case on Svalbard the interferogram will probably be unusable due to decorrelation. Several of the DinSAR-scenes available were unusable since a small change in humidity also can cause big differences in the backscatter properties of snow. Due to Svalbard’s geographical position, large and fast fluctuations in temperature

Table 2. Data from the velocity calculations showing maximum ($V_{\text{max}}$) and average ($V_{\text{aver}}$) velocities, the calculated average terrain slope and average aspect (i.e. angle between range and the direction of steepest slope), and error of calculated glacier velocities caused by relative errors in the DEM ($E_{\text{topo}}$).

<table>
<thead>
<tr>
<th>Glacier</th>
<th>Location</th>
<th>Date (ERS1/ERS2)</th>
<th>$V_{\text{max}}$ (m/d)</th>
<th>$V_{\text{aver}}$ (m/d)</th>
<th>Slope (deg.)</th>
<th>Aspect (deg.)</th>
<th>$E_{\text{topo}}$ (%)</th>
</tr>
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<tr>
<td>Isachsenfonna</td>
<td>79°N 13°E</td>
<td>10/11.10.96</td>
<td>0.42</td>
<td>0.23</td>
<td>0.46</td>
<td>128.0</td>
<td>2.3</td>
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<tr>
<td>Isachsenfonna</td>
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<td>0.18</td>
<td>0.45</td>
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<tr>
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<td>79°N 19°E</td>
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<td>1.19</td>
<td>176.0</td>
<td>3.1</td>
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<td>1.02</td>
<td>43.7</td>
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<td>Nordbreen</td>
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<td>0.36</td>
<td>0.13</td>
<td>1.14</td>
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<td>1.2</td>
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<td>Nordbreen</td>
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<td>0.33</td>
<td>0.10</td>
<td>1.07</td>
<td>17.0</td>
<td>6.4</td>
</tr>
</tbody>
</table>
and weather conditions occur. This can be the reason for the high percentage of decorrelated DInSAR-scenes. The penetration depth of radar signals varies a lot. The ERS-1 and ERS-2 signal with a frequency of 5.3 GHz could penetrate 10 m of ice and some tens of metres of dry snow (Ulaby et al. 1982). Hoen & Zebker (2000) reported penetration depths of 12 m to 35 m from ERS data from the Greenland Ice Sheet. It could therefore be stated that, in fact, it is not surface velocity that is measured. However, in accordance with glaciological theory, most of the vertical velocity change in a glacier takes place close to the bed. Bamber (1987) found an ice thickness of 273 m in the area between Kronebreen and Isachsenfonna. This means that there is practically no difference in the surface velocity that is measured. However, in accordance with glaciological theory, most of the vertical velocity change in a glacier takes place close to the bed. Bamber (1987) found an ice thickness of 273 m in the area between Kronebreen and Isachsenfonna. This means that there is practically no difference in the surface velocity and the velocity found some tens of metres down in the glacier (Paterson 1994, 251). If the penetration causes the signal to be reflected from a level lower than the level of the external DEM, this will create errors in the DInSAR. However, most Svalbard glaciers will experience epochs of melting even during the winter. Ice layers and ice bodies within the snow will therefore cause most of the reflection to happen near the surface.

The estimation of the terrain parameters also introduces inaccuracies, but because of the substantial smoothing of the DEM it is not influenced much by the relative errors in the DEM described previously. However, surface slope estimation accuracy has an impact on the end result but not as drastically as the direction of slope. The average surface slopes in the three InSAR-scenes vary between 0.45° and 1.19° (Table 2) and for these small angles an error will not cause much harm according to Eq. 4. Yet when the direction of slopes approaches the azimuth direction the accuracy drastically decreases. This is due to the fact that a DInSAR-scene only measures the displacement component in range. Fig. 7 shows how the calculated velocity will vary with an increased angle between the direction of slope and range. An error of 10° if the direction of slope is 0° will cause small errors (1.54%) for the calculated velocities. On the other hand, an error of 10° for a glacier moving at an angle of 50° with range will be much higher (up to 46.2%). These
estimates are also based on Eq. 4. Due to this, Akademikerbreen will have the most accurate results, since the average angle between range and the direction of slope is rather small, as shown in Table 2. For the same reason, the accuracy of the estimated velocity for Nordbreen and Isachsenfonna will be lower, and calculation for the lower parts of Isachsenfonna is omitted completely. The greatest mean aspect angle is 52°. Since it is difficult to establish the accuracy of the slope direction calculation, it is also difficult to say anything more precisely about the effect on the calculated velocities. If the slope direction angle of 52° has a 5% error, the error after decomposition is 6.3%, while a 10% error will cause a 13.7% error in the end result. Rignot et al. (1995) determined the error in the direction of slope in their study to be ± 4°. If it is assumed that the accuracy of slope calculation in this study is 0.5°, the look angle 0.2°, and direction of slope 5%, then the total error from decomposition will be just under 7%. The error for glaciers moving in the range direction will be smaller, close to 3%, since the error from the direction of slope calculation will be much smaller. In addition, there will still be errors from the aforementioned phase measurements, of 1.5 mm for vertical and 4 mm for horizontal movement, and also errors from differentiating the InSAR scenes. Rignot et al. (1995) found InSAR-calculated velocities to be within 6% of ground measured ones.

Glacier velocity

Comparing the results from Lefauconnier et al. (2001), it is evident that the velocities differ somewhat from the velocities obtained for Isachsenfonna in this study. They found a nearly constant velocity for the whole glacier, ranging from 0.20 to 0.22 m d⁻¹, while the results presented here show a change along the centre line from c.0.20 m d⁻¹ in the higher parts (lower left in Fig. 6a), via maximum velocities of over 0.40 m d⁻¹ around the middle, and down to 0.15–0.20 m d⁻¹ in the lower parts (upper right in Fig. 6a). Calculations further down the glacier are omitted in this study since the flow direction is turning more southwards and more or less parallel to the azimuth direction, and hence velocity measurements will be heavily influenced by errors or even impossible. On the lower left side of the glacier there is an expected decrease in velocity. This is an area in between the ice flows of Holtedahlsfonna and Isachsenfonna where one would expect stagnant ice masses. Lefauconnier et al. (2001) also report on GPS- and InSAR-measured velocities close to this area of 0.10 and 0.8 m d⁻¹ respectively.

The difference in measured velocities at the mid to upper part of Isachsenfonna could be caused by the fact that the glacier is at a relatively small angle with the azimuth direction in the InSAR scene used in Lefauconnier et al. (2001). They also state that they had some problems with fringe counting and that only the velocities measured in the central basin (i.e. where Isachsenfonna and Holtedahlsfonna merge) should be considered valid. Looking at the geometry of the glacier, a velocity increase in the narrow part of the glacier and a decrease further downstream could be expected where the glacier is widening, in common with those presented here. More ground measured velocity data have to be gathered to confirm this.

Generally, all the velocity maps in this study show velocity patterns that conform to glaciological theory; velocity increases with surface slope and decreases with increasing width. Thus the glaciers show extending and compressive flow due to changes in surface slope and width as expected. These factors overrun the climatic ones and therefore there is no good correlation between the position of the equilibrium line and the maximum velocities. The greatest change in velocity is found along the edges of the glacier, since the edges exert a drag on the glacier.

For Nordbreen, two late winter scenes and one autumn scene were chosen. The highest velocities are found in the September 1995 scene, which also has a higher mean velocity than the other scenes. This could indicate higher velocities during summer due to more meltwater. Since the scene was acquired in late September in the Arctic, this is probably not the case. Hence, there has to be an alternative explanation for the increase. The autumn scene was also acquired the year before the spring scenes, so the explanation could equally be one other than seasonal variation. The uncertainty in determining the glacial border for Nordbreen can also have implications for the result. To be able to determine a seasonal variation, a scene from July or August is needed together with a better determination of the glacier border.
Conclusions

The results from this study show that glacier velocities on Svalbard can be calculated with DEMs and height-differentiated tandem InSAR data by the use of modules in ESRI’s Arc GIS software. A semi-automatic algorithm is developed in order to calculate glacier velocities in the flow direction, resulting in maps of the velocity fields for the glaciers. Velocity fields for the three glaciers Isachsenfonna, Nordbreen and Akademikerbreen are presented, with maximum values of 0.42 m d\(^{-1}\), 0.36 m d\(^{-1}\) and 0.41 m d\(^{-1}\) respectively. The accuracy is greatest when the angle between direction of slope and range is small. The results seem to be reliable and in agreement with other ground and satellite-borne observations.

The results also show that the ERS-1/ERS-2 SAR tandem archive from 1995–1996 can reliably be used to establish velocity fields for most of the larger glaciers on Svalbard. Such a database showing the state of the glaciers in the mid-1990s will work as a unique reference when making new field measurements or remote sensing acquisitions in the future.

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