5 Satellite-retrieval of mass balance: comparing SAR images with albedo images and in situ mass balance observations*

Abstract

We present an analysis of many ERS SAR images of Vatnajökull (Iceland) by comparing them with AVHRR images, mass balance observations and modeled firn stratigraphy. Summer SAR and AVHRR images both detect the surface firn line as a distinct boundary. Winter SAR images of Vatnajökull display sub-surface firn-ice transitions, but these do not correspond to the late summer surface firn line. We found no differences in backscatter between melting snow and firn and for most years no differences in reflectance either. Hence, the equilibrium line is mostly not visible when it lies above the firn line and we only identified it on SAR images for one out of nine years and on AVHRR images for only three years. For Vatnajökull, equilibrium line altitude is therefore not a particular good estimator of mean specific mass balance \( B_m \). However, for some drainage basins, mean firn line altitude during the melting season can be used to infer \( B_m \). Unlike albedo images, SAR images do not display inter-annual variations of the signal within the accumulation area that are clearly related to \( B_m \). Hence, for glaciers and ice caps like Vatnajökull that display no dry snow during the summer, the only advantage of SAR images over albedo images lies in the fact that they display the firn line irrespective of weather and illumination conditions.

5.1 Introduction

Glaciers and ice caps often lie in remote areas, are difficult to access and can have vast surface areas. Direct observation of the mean specific mass balance \( B_m \) and related quantities, such as the equilibrium line altitude (ELA) and accumulation area ratio, is therefore a time-consuming and often costly procedure. As a consequence, these quantities have been measured on relatively few glaciers worldwide. Larger parts of the cryosphere can be studied on a regular basis with satellite sensors. Østrem (1975) proposed that the often-found linear relation between \( B_m \) and ELA can be used to infer \( B_m \) by using satellite-derived ELA. In recent years much work has been devoted to inferring the snow line at the end of the melting season. Note that we define snow as being younger than one year and firn as being at least one year old, which implies that the equilibrium line equals the snow line at the end of the melting season (if there is no superimposed ice).

Some authors successfully applied Østrem's method by using satellite albedo images (e.g. Rott and Markl, 1989) or Synthetic Aperture Radar (SAR) images (e.g. Demuth and Pietroniro, 1999), but the method often fails. This can occur for years when the equilibrium line lies above its position of the previous year(s). In such cases the snow line is sometimes visible (e.g. Rott and Markl, 1989), but often the difference in albedo ($\alpha$) between firn and snow of several months old is too small to be detectable (e.g. Rott and Markl, 1989; Hall et al., 1995; De Ruyter de Wildt et al., 2002b). De Ruyter de Wildt et al. (2002b) developed a method that does not suffer from this problem. The net potential shortwave radiation, defined as the net shortwave radiation at the top of the atmosphere, can be calculated from NOAA AVHRR albedo images. This quantity, integrated over the glacier surface and over the melting season, both depends on and influences summer melt. It also depends on winter mass balance and was found to be linearly related to $B_m$ for Vatnajökull (Iceland). Unfortunately, in some years high cloudiness limited the availability of albedo images and hence introduced uncertainty in this linear relationship.

Clouds do not pose a problem if SAR backscatter images are used. Backscatter ($\sigma_0$) images can display several boundaries, which correspond to surface or subsurface facies transitions. Microwaves penetrate through dry snow so the main signal over dry snow surfaces originates from subsurface material (e.g. Rott et al., 1985). The underlying material in the ablation area is ice, which produces less backscatter than the underlying firn in the accumulation area. Firn contains scattering elements such as ice structures and internal surfaces. SAR images acquired in winter or early spring can therefore be used to detect the approximate location of the firn line (e.g. Fahnstock et al., 1993; Hall et al., 1995; Partington, 1998; König et al., 2001b). When the surface is melting, the $\sigma_0$ signal is much lower and stems from the top few cm (e.g. Stiles and Ulaby, 1982) and only surface features can be observed. The boundary between dry and wet snow is detectable (e.g. Steffen et al., 1993), as well as the boundary between bare ice and wet snow or firn (e.g. Rott and Mätzler, 1987; Adam et al., 1997). Smith et al. (1997) also observed a fourth zone, marked by high backscatter over wet snow (phase 2 melt or P2), which is most likely caused by roughness due to suncups (Ramage et al., 2000). These four zones (dry snow, wet snow, metamorphosed wet snow and ice) were seen to move up-glacier and replace each other during the melting season. In spite of these observations, SAR images have been compared to in situ mass balance observations only a few times (e.g. Demuth and Pietroniro, 1999; Hall et al., 2000; König et al., 2001b), while a detailed comparison of radar images with albedo images and mass balance measurements from a number of years yet has to be made. From the latter two studies it appears that, just as is often the case for albedo images, the $\sigma_0$-signal cannot distinguish between snow and firn. Consequently, the snow line appears only to be visible on radar images when no firn is present at the surface.

For Vatnajökull (figure 5.1), one of the largest temperate ice caps in the world, a wealth of information is available with which SAR images can be analyzed and interpreted. Its mass balance has been regularly measured (e.g. Björnsson et al., 1998a) and recently a mass balance model was developed and calibrated to the conditions on the ice cap (De Ruyter de Wildt et al., 2002a). The mass balance measurements have been compared to NOAA AVHRR images (De Ruyter de Wildt et al., 2002b). In situ mass balance measurements, NOAA AVHRR images, and a long-term modeled mass balance series form a
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unique data set with which SAR images can be compared. In this paper, we compare these data sets with SAR images that were acquired during the same period as these data sets, i.e. the melting seasons of seven years in the 1990s. We also use a few images that were acquired in the winter during dry-snow situations. First we directly compare SAR images with AVHRR images to identify the glacier facies that are visible on SAR images. Then we reconstruct the stratigraphy of a part of Vatnajökull from measured and modeled mass balance values, in order to explain several boundaries and structures that are visible on the SAR images and on AVHRR images. On the basis of these investigations we discuss some possibilities of retrieving the mass balance from SAR images.
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5.2 Data

5.2.1 ERS SAR images

The European Space Agency (ESA) launched the ERS-1 satellite in 1991 and the ERS-2 satellite in 1995. Both satellites carry a C-band (5.3 Ghz) SAR with vertical transmit and receive (VV) polarization. We use Precision Image ERS SAR data which are three-look (speckle-reduced) and which we calibrate and correct for the SAR antenna pattern and range-spreading loss (Laur et al., 1998). The $\sigma_0$ signal is dimensionless and given in dB (i.e. on a logarithmic scale). The horizontal resolution is 33 m in range and 30 m in azimuth while the pixel size is 12.5 m. The ERS scenes are very large (130 MB in binary format) and therefore unmanageable, so we scale up the pixel size with a factor of 10 (up to a pixel size of 125 m). In view of the fairly large size (8200 km$^2$) and low surface slopes (2.8° on average) of Vatnajökull, this has no consequences for the results. ESA provides the images' horizontal locations with an accuracy of 100 m in range and 200 m in azimuth. We purchased images that were acquired during the melting seasons of 1993, 1995, 1996, 1997 and 1998, which are years with positive, negative and near-neutral mean specific mass balances. Typically, the time interval between subsequent images is two to three weeks. Most images display only a part of Vatnajökull so the number of images varies per drainage basin. For each of these five years, we have six to ten images for the drainage basins Brúarjökull, Dyngjujökull, Köldukvíslarjökull and Tungnaárjökull (figure 5.1). Furthermore, for 1991 and 1994 we have one image acquired at the end of the melting season, and we have one additional image for each of the winters of 1992/1993, 1993/1994 and 1998/1999.

5.2.2 NOAA AVHRR images

We purchased NOAA AVHRR images from the Dundee Satellite Receiving Station in the U.K. The images were acquired during the melting seasons (April - September) of the years 1991-1999 inclusive. The horizontal resolution at nadir is 1.1 km. In the first four years, the NOAA-11 satellite provided the data. From 1995 onwards the images were acquired by the NOAA-14 satellite. Because Iceland lies in the north Atlantic Ocean, where storm activity is high, the skies are often overcast and most of the images display some clouds. It was therefore not always possible to find cloud-free images for the end of the melting season (e.g. in 1994 and 1999). We found nine to fifteen images for each melting season with a typical time interval of two weeks between images. Nearly all images were acquired near solar noon when irradiance was high. The process of retrieving the surface albedo from the images has been discussed in detail elsewhere (De Ruyter de Wildt et al., 2002b), which is why we restrict ourselves here to a short description of the successive processing steps:

- Clouds are discriminated from snow and ice by making use of the reflective and thermal differences in AVHRR channel 3 (3.5-3.9 µm) and the thermal differences in
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channel 4 (10.5-11.5 µm) that often, but not always, exist. Not all clouds are detected in this way, so we had to check the images for errors manually. A convenient way of doing this is to compare subsequent images and look at textural characteristics.

- We apply a geolocation to each image by comparing the images to a Digital Elevation Model (DEM), depicted in figure 5.1. The horizontal resolution of the DEM is smaller than the AVHRR pixel size, namely 500 m. For most images we were able to locate the ice margin and features like mountain peaks and steep ridges with an accuracy of one pixel. The accuracy of the geolocation for a few images with a large satellite viewing angle is estimated to be two pixels.

- Measured radiation intensities are calibrated with the calibration formulas of Rao and Chen (1995) and Rao and Chen (1999) for the AVHRR instruments aboard the NOAA-11 and NOAA-14 satellites, respectively. These coefficients take instrument drift into account. The resulting radiances are converted into planetary albedos.

- We use a radiative transfer model (Koelemeijer et al., 1993) to convert the planetary albedos into surface albedos. This model takes Rayleigh scattering and attenuation by ozone and water vapor into account but neglects the effect of aerosols. Because the results are not very sensitive to the atmospheric profiles (Reijmer, 1997), we use the standard Sub Arctic Summer Profile of McClatchey et al. (1972) as input.

- We employ an empirical expression to convert the surface narrowband albedos in AVHRR channels 1 and 2 into broadband albedos. This expression (De Ruyter de Wildt et al., 2002b; W. Greuell, personal communication) is based on many (8,000) point measurements made simultaneously in AVHRR channels 1 and 2 and over the entire solar spectrum (Greuell et al., 2002). The measurements were made over a broad range of surface types, ranging from dirty glacier ice to melting snow and having broadband albedos between 0.05 and 0.80.

- To correct for the anisotropic nature of reflection at snow surfaces, we apply an empirical parameterization (Koks, 2001), which is based on measurements over melting snow of two to three weeks old, and is valid for a broad range of solar zenith angles (15.9° - 65.5°). During the summer, virtually all surface snow of Vatnajökull melts and becomes metamorphosed, so the parameterization is likely to be valid for the average summer conditions on Vatnajökull. For glacier ice no usable parameterization is available.

- In the above, the fluxes are calculated with respect to a horizontal plane. If the surface is inclined, the fluxes through a plane parallel to the surface differ from those through the horizontal plane and a correction needs to be applied to obtain the surface albedo. We calculate the surface albedo with the expression of Knap et al. (1999).

5.2.3 Mass balance observations

The mass balance of Vatnajökull has been measured with good spatial resolution since 1992 (Björnsson et al., 1997, Sigurdsson, 1997; Björnsson et al., 1998a, b, c, 1999; O. Sigurdsson, personal communication). The data have mainly been obtained over the drainage basins of Eyjabakkajökull, Brúarjökull, Dyngjujökull, Köldukvisljarjökull and
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Tungnaárjökull (figure 5.1). For Eyjabakkajökull, data from 1991 are also available. Most measurements were taken at the end of September or the beginning of October. On each outlet, the mass balance has been measured along one or two profiles that capture the altitudinal variation. Because most of Vatnajökull is quite flat, the profiles can be used together with a few additional measurement sites to describe the lateral variation. We obtain the mean specific balances of the drainage basins and the ELAs by interpolating between the measurement sites. For this interpolation, we developed an algorithm that takes vertical gradients in the mass balance into account, even when these are not resolved by the measurement sites. A DEM is needed for this feature to work. For each grid point of the DEM, the algorithm determines the n closest measurement sites within 500 m in altitude from the grid point. Then, because of the limited height differences, a linear relation between mass balance and altitude is found for the n measurement sites, with which the mass balance at the grid point can be calculated. To smooth discontinuities in the resulting mass balance field, the contribution of each measurement site is weighted with the inverse of its distance to the grid point. The resulting values of $B_m$ are not very sensitive to the value of n. We therefore use a value of 6, which is the lowest value that gives smooth mass balance fields. Table 5.1 displays the resulting values of $B_m$. The data clearly include years with a highly positive $B_m$ (1992, 1993) and years with a highly negative $B_m$ (1997, 1998). We estimate the uncertainty in $B_m$ to be 0.25 m w.e.

5.3 Interpretation of radar boundaries

A first and direct way to interpret the different $\sigma_0$ signatures observed on the SAR images of Vatnajökull, is to make a scatter plot of $\sigma_0$ against $\alpha$ (figure 5.2). In this plot we clearly see three clusters of data points, representing three kinds of surface facies. Dry snow has high $\alpha$ and high $\sigma_0$ signatures. The difference between dry snow lying directly on top of glacier ice and dry snow lying on firn is small (3 to 4 dB) and the two groups merge in figure 5.2

<table>
<thead>
<tr>
<th>year</th>
<th>western Brúarjökull</th>
<th>eastern Brúarjökull</th>
<th>Eyjabakkajökull</th>
<th>Dyngjujökull</th>
<th>Köldukvislarjökull</th>
<th>Tungnaárjökull</th>
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<tr>
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<td></td>
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<td>0.35</td>
</tr>
<tr>
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<td>0.89</td>
<td>0.69</td>
<td>1.58</td>
<td>0.03</td>
<td>-0.11</td>
<td>0.31</td>
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<td>0.42</td>
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<td>0.03</td>
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<td>-0.11</td>
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<td>0.04</td>
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<td></td>
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<td>-2.18</td>
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<td>-0.58</td>
<td>-0.99</td>
<td>-0.99</td>
<td>-0.26</td>
</tr>
</tbody>
</table>

Table 5.1. Mean specific mass balance of various drainage basins of Vatnajökull (in m w.e.). The values were obtained by interpolation. The weighted mean for the whole northwestern part of Vatnajökull (all) is shown when the mass balance was measured over the largest part of this area.
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Figure 5.2. Radar backscatter against satellite-derived surface albedo. Each point represents an AVHRR pixel; for this plot the backscatter images have been resampled up to the same resolution as the AVHRR images (1.1 km). The maximum time gap between radar and albedo images used to make this plot is three days. Sixteen of such pairs were found, distributed randomly over spring and summer of five years.

Figure 5.2 shows no cluster of P2 melt facies, consisting of roughened melting snow with high $\sigma_0 (>12$ dB) as Smith et al. (1997) and Ramage et al. (2000) observed. Only five out of 43 SAR images of Vatnajökull with melting surfaces display wet snow with high $\sigma_0$, but for these images no simultaneous AVHRR images are available. Figure 5.3 shows two of these five images. All zones in these images follow the height contours of the ice cap (figure 5.1). The lowermost zone is bare glacier ice, as confirmed by AVHRR images acquired at most one week later. The uppermost zone (only in figure 5.3a) lies in the highest part of Vatnajökull, approximately above the 1600 m contour. From the daily mean temperature in Kirkjubæjarklaustur and assuming a lapse rate of 6°C/km, we find the daily mean 0°C isotherm to be at 1700 m. The correspondence of these two heights and the high $\sigma_0$ values suggest that the uppermost zone represents dry snow. Zones I, II and III must consist of melting snow and/or melting firn. Zone I has low backscatter and corresponds to
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The initial-melt zone ("M") of Smith et al. (1997). Zone II has high $\sigma_0$ (-6 to -10 dB), but in both images it is bounded on both sides by zones of low $\sigma_0$. If the high backscatter in zone II is caused by roughening elements like suncups (Ramage et al., 2000), then these must be absent in zone III. In this respect zone II differs from the P2 zone found by Smith et al. (1997), which lies directly above the bare-ice zone. In any case, most images do not display zone II and Hall et al. (2000), who studied many SAR images of Hofsjökull (Iceland), do not mention such a zone either. The reason for this absence may be the regular occurrence of snowfall in the accumulation area during the summer, which prevents the formation of suncups.

For further interpretation of the different boundaries present in the SAR images of Vatnajökull, we compare $\sigma_0$ profiles with $\alpha$ profiles and, following König et al. (2001b), with modeled stratigraphy. We do so for the western part of Brúarjökull, one of the large northern outlets, because this outlet mostly displays a rather sharp firn line. On other outlets the situation is more complicated with a gradual and patchy firn-ice transition, which makes it difficult to compare SAR images with model results.

Figure 5.3. ESA ERS $\sigma_0$ images of part of Vatnajökull, acquired on September 20th, 1993 (a), and acquired on August 2nd, 1998 (b). Latin numerals indicate different zones of surface facies.
5.3.1 Stratigraphic modeling

To model the stratigraphy we used mass balance observations (since 1992) and mass balance values (from 1960 until 1992) obtained with the mass balance model of De Ruyter de Wildt et al. (2002a), which was especially constructed for Vatnajökull. It describes the energy fluxes between atmosphere and glacier, which are tuned with *in situ* measured data (Oerlemans et al., 1999), in a detailed way. The most important model specifications are:

- Temperature in the katabatic surface layer is related, but not equal, to the temperature in the free atmosphere just above the surface layer (in the surface layer temperature is mostly lower and temperature variations are smaller than in the free atmosphere).
- Incoming longwave radiation is a function of temperature in the free atmosphere, which is justified by the relatively thin katabatic layer over Vatnajökull.
- Sensible heat flux is a function of temperature in the katabatic layer.
- Snow albedo depends upon the number of days since the last snowfall.
- Ice albedo depends upon location and varies from very low in the northwest (0.10 due to volcanic ash layers) to 0.30 at some locations in the south and southeast.
- Subsurface processes, such as refreezing of melt water, are neglected and whenever the surface energy flux is positive, the surface is assumed to be at the melting point (“zero-degree assumption”). For a temperate ice cap like Vatnajökull, this is a reasonable assumption.
- Free-air temperature and vapor pressure are assumed to be always horizontally (but not vertically) homogeneous over the ice cap. Meteorological data show that this is mostly the case. Cloudiness and relative variations in precipitation are also assumed to be horizontally homogeneous. This allows us to force the energy balance over the entire ice cap with data from one meteorological station.
- The mean spatial distribution of precipitation over Vatnajökull is not well known, which is why we use this variable to calibrate the model to observations of the mass balance (Björnsson et al., 1997, 1998a, b, c, 1999). These observations were mainly made over the central and northeastern parts of the ice cap (i.e. the drainage basins of Tungnaárjökull, Kolduvisljarjökull, Dynjújökull and Brúarjökull). Precipitation data from coastal weather stations in the south and southeast are available to prescribe a spatial distribution.

Daily mean temperature, humidity and cloudiness measured at Kirkjubæjarklaustur (figure 5.1) are used to drive the model. It was found that the observed mass balance is best simulated when the model is driven with the precipitation from Fagurhólsmyri (figure 5.1).

We modeled the stratigraphy at many points along the transect over western Brúarjökull (figure 5.1) which enables us to plot the stratigraphy as a function of surface elevation. We do not take ice flow into account, because Brúarjökull is in the quiescent phase of its surging cycle; the last surge took place in 1963 (about 4 km; Sigurdsson, 1998). Furthermore, the surface slope of Brúarjökull in the vicinity of the equilibrium line is very low (0.5° to 1°), meaning that mass is transported to lower surface elevations only very slowly.
Figure 5.4. Modeled stratigraphy, backscatter profiles ($\sigma_0$, solid lines) and albedo profiles ($\alpha$, dashed lines) along the transect shown in figure 5.1. For each of the years 1991-1998, annual layering at the end of the melting season (September 21$^{st}$) is shown. The $\sigma_0$ and $\alpha$ profiles are named by year and day of the year. For each year the profiles that display the highest firn line are shown. Some additional $\sigma_0$ profiles acquired during freezing conditions are shown as bold lines. 10 m in surface altitude approximately corresponds to 800 m in horizontal distance.
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5.3.2 Comparison of SAR data with AVHRR data and stratigraphy

Figure 5.4 compares backscatter profiles, albedo profiles, and modeled stratigraphy along a flow line of western Brúarjökull. Note that in each stratigraphy plot, the equilibrium line is given by the lower boundary of the most recent annual layer (the uppermost layer). First of all we notice that near the end of the melting season, $\sigma_0$ and $\alpha$ images display the same Firn Line Altitude (FLA). This is the case in all years (note that we have no $\sigma_0$ images for 1992), although in 1998 there is a small difference that may be due to a geolocation error. In all years except 1992, the equilibrium line lies above the firn line and cannot be seen on the $\sigma_0$ profiles. The $\alpha$ profiles only display it for the years 1992, 1997 and 1998. In 1993 and 1994, both years with a positive mean specific mass balance, the ELA is obscured by firn from 1992, although in 1993 the firn line and the snow line lie closely together. From 1992 till 1996 the late-summer firn line was formed by relatively young firn from 1992. In September 1991, 1997 and 1998 the firn line corresponded to annual layers of 15 to 20 years old (between 1150 and 1180 m). Note that annual layer thickness is given in m w.e. so the compaction of firn into ice is not taken into account. On a temperate ice cap like Vatnajökull the compaction process is complicated by melting and strong metamorphism of firn layers, which is very difficult to model. On other temperate glaciers, the firm-ice transition occurs in layers of 4 to 13 years old, at depths of 13 to 32 m (Paterson, 1994). In 1991, the firn line on Vatnajökull lay at a slightly lower elevation than in 1997 and 1998, which is caused by melting of the lowermost surface firn layers in these two years.

Only at the end of the melting season of 1992, all firn was covered by snow and for this year it should be possible to detect the snow line on satellite images. Unfortunately, we have no radar images for this year and we found no cloud-free AVHRR images for the end of the melting season either. Images from the winter and spring of 1993 might also display the equilibrium line from 1992, but this appears not to be the case (figure 5.4, panel for 1992). Near the end of August 1992, there was a considerable amount of snowfall, which contributed to the low ELA of 1992 (Björnsson et al., 1998). This snow may have undergone some melting in September and October, but the $\sigma_0$ signal can still penetrate several meters through refrozen stratified snow and firn (e.g. Mätzler, 1987). Therefore, just above the equilibrium line from 1992 the main signal seems to have come from the underlying ice and it seems unlikely that the late-summer snow can be distinguished from winter snow on the images from the winter and spring of 1993. The same appears to be the case for the two following years. Winter images from late November 1993 (day 333, see panel for 1993) and April 1995 (day 118, see panel for 1994) do not display the late summer firn line. In these winters, as in the winter of 1992/1993, the firn pack remained shallow below a surface altitude of about 1150-1180 m. Below this altitude the $\sigma_0$ signal was only partially influenced by the firn and thus reached the underlying ice. Therefore, near the late-summer firn line, the effect of the firn seems to have been very small. Images acquired during freezing conditions often display two boundaries on western Brúarjökull, for example, the SAR images from January 18th, 1993 (figure 5.5a) and from January 23rd, 1999 (figure 5.6a). Figure 5.4 shows in red all $\sigma_0$ profiles from images that were acquired during freezing conditions. These images display high backscatter values and correspond to low daily mean temperatures in Kirkjubæjarklaustur. The upper boundary lies between 1230 and
Figure 5.5. ESA ERS $\sigma_0$ images of the northern part of Vatnajökull, acquired on January 18th, 1993 (a), and acquired on September 13th, 1995 (b). The latter image is compared with $\sigma_0$ boundaries from plot a (labeled B1 and B2).
Figure 5.6. ESA ERS $\sigma_0$ images of the northern part of Vatnajökull, acquired on January 23rd, 1999 (a), and acquired on September 3rd, 1998 (b). The latter image is compared with $\sigma_0$ boundaries from plot a (labeled B1 and B2).
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1240 m (hereafter called B1) and the lower boundary between 1160 and 1190 m (hereafter called B2). In some images, B1 is the most obvious boundary, but B2 is clearer in others. Both boundaries most likely correspond to subsurface transitions because they are not visible on late summer images (figures 5.5b and 5.6b), which only display the surface firn line. B1 did not noticeably change position between 1993 and 1999 (unlike the late summer firn line) and lay at a higher elevation than the late summer firn line. B2 also lay above the late summer firn line of 1995 but below the late summer firn line of 1998. Hence, both wintertime images (figures 5.5a and 5.6a) do not display the late-summer surface firn line. All stratigraphy plots in figure 5.4 show an increase in subsurface firn age close to the altitude of B1 (1225 m). This increase most likely corresponds to an increase in density and 1220-1240 m seems to be the lowest altitude where, during freezing conditions, the SAR sensor only ‘sees’ firn and no ice. Just below this altitude, the firn-ice transition lies closer to the surface and the $\sigma_0$ signal probably stems both from ice and from the overlying firn. The winter $\sigma_0$ profiles in figure 5.4, panels 1993 and 1995, do not show B1. On these images, it may be obscured by the thick pack of relatively recent snow and firn. Note that the stratigraphy plots do not display fresh winter snow, which overlies the displayed stratigraphy on wintertime images. Furthermore, the stratigraphy in figure 5.4 is displayed in m w.e. so the actual annual layers are thicker, especially the young layers.

Figures 5.5 and 5.6 show that the situation on Dyngjujökull is more complicated than on western Brúarjökull. On Dyngjujökull, there is a transition zone, rather than a line, between firn and ice. Only one wintertime $\sigma_0$ boundary is clearly present (corresponding to B1 on western Brúarjökull). When we compare this boundary to the late summer firn lines, the results resemble those of western Brúarjökull: in both years it lay above the late summer firn line.

![Graph](image)

Figure 5.7. Firn Line Altitude (FLA) at the end of the melting season and Equilibrium Line Altitude (ELA) as a function of the mean specific mass balance for western Brúarjökull and Dyngjujökull. Both radar and albedo images were used to derive FLAs. ELAs are derived from the mass balance measurements.
Figure 5.8. FLA along the transect shown in figure 5.1 (western Brúarjökull) as a function of the day of the year. Both backscatter images (solid symbols) and albedo images (open symbols) were used to obtain FLAs. Each FLA was manually determined as the altitude at which the change in backscatter or albedo was largest. Low FLAs in the latter part of the summer, indicative of summer snowfalls, are not shown.
5.4 Mass balance retrieval

SAR images can be used to trace changes in the long-term surface firn line, but are of limited use for annual detection of the equilibrium line. Figure 5.7 shows that, whereas the ELA is linearly related to $B_m$, the FLA is not related to $B_m$ for negative values of $B_m$. However, $B_m$ may be related to other quantities that can be detected on SAR images. Recent work (De Ruyter de Wildt et al., 2002b) has shown that the average surface albedo correlates well with $B_m$, whereas melting snow and firn tend to have higher $\sigma_0$ values when the albedo is lower (figure 5.2). In addition, figure 5.8 shows that in years with a positive mass balance (e.g. 1993), the firn line rises more slowly and later in the melting season than in years with a negative mass balance (e.g. 1997). The snow line position contains information about the accumulation and melt history, as long as the snow line is located below the equilibrium line in previous years. This is always the case during the first part of the melting season and for some years also at the end of the melting season. These observations suggest that during the melting season, the average $\sigma_0$ ($\langle \sigma_0 \rangle$) over the glacier surface may be lower (more negative) in years with a positive mass balance than in years with a negative mass balance. This can of course only be the case when dry snow and firn, and the P2 facies are excluded from the analysis, as these facies disturb the linear relation between $\sigma_0$ and $\alpha$. Figure 5.9 shows that for western Brúarjökull, $\langle \sigma_0 \rangle$ generally increases during the summer, corresponding to increasingly higher firn lines and/or snow and firn surfaces that are increasingly wet and rough. Only in 1993 there was no obvious increase and $\langle \sigma_0 \rangle$ remains relatively low during the summer. In this year, the summer was cold with regular snowfalls. In general, the signal strength changes with surface wetness and

![Figure 5.9](image.png)

*Figure 5.9. Average $\sigma_0$ over western Brúarjökull as a function of day of the year, for each of the ERS SAR images. Only images that do not display dry snow and firn were used.*
Comparing radar images with albedo images and in situ mass balance observations

roughness, which both are highly dependent on temporal circumstances. This is shown, for example, by the data from 1998. Also, in 1995 $\langle \sigma_0 \rangle$ was quite high in the summer, while $B_m$ had an intermediate value (Table 5.1). For this year, no suitable images are available for a large part of the summer and the high $\langle \sigma_0 \rangle$ values may not be representative. By interpolating $\langle \sigma_0 \rangle$ between the successive images, we can compute the mean value of $\langle \sigma_0 \rangle$ over the summer. We do so for the period from day 195 till day 245, which is the longest period for which $\langle \sigma_0 \rangle$ values are available in our data set. We then find a correlation coefficient of 0.72 for $\langle \sigma_0 \rangle$, averaged over the summer, and $B_m$. There is no significant correlation between $\langle \sigma_0 \rangle$ and $B_m$ for Dyngjujökull and Köldukvíslarjökull.

Figure 5.8 shows that the course of the FLA during the summer varies per year. A simple way to quantify this is to compute the average FLA during the melting season. We do this for the second half of the melting season (days 200 – 264), because in the year with the lowest $B_m$ (1993), no ice was visible and hence no firm line existed before day 200. Dips in the FLA in the later part of the summer, indicative of summer snowfalls, are not used. These are often highly temporal events that are not representative of the previous and following weeks. When no images for the end of the melting season are available (e.g. 1992 and 1994), we linearly extrapolated the FLA from FLAs earlier in the melting season. In such cases the highest FLA from other years forms an upper limit to the extrapolated FLA. Figure 5.10 shows that the average FLA during the second half of the melting season correlates well with $B_m$ on several drainage basins. We have no ERS images for 1994 and, due to clouds, only a few AVHRR images from which the FLA can be derived. For this year, there is an outlier for western Brúarjökull ($B_m=0.55$) and we expect that the correlation

Figure 5.10. Mean FLA during the second half of the melting season (day 200-264) as a function of the mean specific mass balance for western Brúarjökull, Dyngjujökull and Eyjabakkajökull. Both radar and albedo images were used to derive the FLA.
would be better if we had more satellite-derived FLAs for this year. In view of the correlation between the mean albedo and $B_m$ (De Ruyter de Wildt et al., 2002b), the correlation between $B_m$ and the average FLA was to be expected. The mean albedo over the ice cap depends both on the albedo of the snow- and firn-covered area, and on the position of the firn line (because the firn line corresponds to a large change in albedo).

### 5.5 Discussion and conclusions

We have analyzed many ERS SAR images by comparing them with AVHRR images, mass balance observations, and modeled firn stratigraphy. Both SAR and AVHRR images detect the surface firn line as a distinctive boundary because the reflective and backscattering properties of melting snow and firn are different from those of bare glacier ice. However, we found no differences in backscatter and mostly no differences in reflectance between snow and firn. This implies that the equilibrium line can mostly not be detected when surface firn is present below the equilibrium line. As a consequence, the retrieval of the mass balance through detection of the equilibrium line is often not possible. During nine subsequent years, the equilibrium line of western Brúarjökull could be detected on SAR images for only one year and on AVHRR images for only three years.

Apart from the boundary between ice and melting snow and firn, SAR images can detect several boundaries that are not visible on albedo images. The intensity of the $\sigma_0$ signal is highly sensitive to the liquid water content of snow, which makes it possible to detect the extent of surface melt (e.g. Steffen et al., 1993). For Vatnajökull this is of little use because all snow quickly starts to melt in late spring. A boundary between newly melting snow and roughened melting snow has been found elsewhere (Smith et al., 1997; Ramage et al., 2000). Although it may be present on some SAR images of Vatnajökull, most images do not display such a boundary. All of the boundaries mentioned above are surface boundaries. Sub-surface boundaries can be detected when the surface is not melting, because then microwaves can penetrate several meters into the snow and scatter on subsurface structures. This is for example the case on images that display a dry winter snow pack. These images often display a distinctive boundary, and sometimes a second and less distinctive boundary, which lies in the vicinity of the late-summer surface firn line of warm years. However, it always lies at a somewhat higher elevation and seems to represent a vertical subsurface structure, possibly a subsurface firn-ice transition. The late-summer surface firn line cannot be detected on these winter images, probably because the firm-ice transition often slopes only little with respect to the surface. This allows the penetrating microwaves to reach the underlying ice in the area directly above the late-summer FLA. Hence, the late-summer situation is not visible in winter images and these images cannot be related to annual mass balance, but only to long-term changes in glacier zonation. On glaciers with steeper slopes than Brúarjökull, the surface firn line and the subsurface firm-ice transition are probably not separable from each other. Such glaciers will tend to have a firm-ice transition that dips more with respect to the surface.

The equilibrium line is often not visible on SAR images, but nevertheless SAR
images may be used in other ways to infer the mass balance. The intensity of the $\sigma_0$-signal is related to surface wetness (and roughness) and therefore to the amount of surface melt. For western Brúarjökull, we found a weak correlation between $B_m$ and the mean $\sigma_0$ over the surface area and during part of the melting season, but for two other drainage basins no such correlation was found. It seems that, due to variable surface roughness and/or wetness, $\sigma_0$ is too variable in time to be a good proxy for $B_m$, but to draw a firm conclusion data from more years are needed. Now we have three to five years with both mass balance data and SAR data for each drainage basin. The mean FLA during part of the melting season may be a better estimator of $B_m$. This is the case for western Brúarjökull, Eyjabakkajökull and Dyngjujökull. On Tungnaárjökull, the transition from ice to snow and/or firn is often extremely patchy and gradual, making it impossible to derive a FLA. For Köldukvíslarjökull, we found too few usable images to obtain a usable correlation with $B_m$.

This is due to its hypsometry: the ice margin lies relatively high and bare ice appears relatively late in the summer. The period during which the transient snow line lies between the ice margin and the long-term firn line is therefore short; once the transient snow line lies above the firn line, the FLA is unrelated to the annual mass balance. On the other hand, in 1992 the transient snow line remained at a low altitude all summer and hardly reached the ice margin.

It can be concluded therefore that, for a temperate ice cap like Vatnajökull with no or little dry surface snow during the summer, SAR images contain less information about $B_m$ than albedo images. Variations in the albedo of snow and firn, which are visible on albedo images, can be related to $B_m$ (De Ruyter de Wildt et al., 2002b), but SAR images of melting surfaces do not seem to display usable variations of the signal within the accumulation area. The equilibrium line is often not visible, and the firn line is the only feature that can be unambiguously detected on summer SAR images of Vatnajökull. The average FLA during the melting season is for some drainage basins of Vatnajökull related to $B_m$, but for others this quantity cannot be derived. For glaciers and ice caps like Vatnajökull, the only advantage of SAR images over albedo images lies in the fact that they display the firn line irrespective of weather and illumination conditions, which for some glaciers can be used to infer $B_m$.  

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