Snow research in Svalbard—an overview

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This paper summarizes the most significant snow-related research that has been conducted in Svalbard. Most of the research has been performed during the 1990s and includes investigations of snow distribution, snowmelt, snow pack characteristics, remote sensing of snow and biological studies where snow conditions play an important role. For example, studies have shown regional trends with about 50% higher amounts of snow accumulation at the east coast of Spitsbergen compared to the west coast. Further, the accumulation rates are about twice as high in the south compared to the north. On average, the increase in accumulation with elevation is 97 mm water equivalents per 100 m increase in elevation. Several researchers reported melt rates, which are primarily driven by incoming short-wave radiation, in the range of 10-20 mm/day during spring. Maximum melt rates close to 70 mm/day have been measured. In addition to presenting an overview of research activities, we discuss new, unpublished results in areas where considerable progress is being made. These are i) modelling of snow distribution, ii) modelling of snowmelt runoff and iii) monitoring of snow coverage by satellite imagery. We also identify some weaknesses in current research activities. They are lacks of i) integration between various studies, ii) comparative studies with other Arctic regions, iii) applying local field studies in models that can be used to study larger areas of Svalbard and, finally, iv) using satellite remote sensing data for operational monitoring purposes.

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The Intergovernmental Panel on Climate Change concludes that the North Atlantic region is one of the most sensitive regions on Earth with respect to rapid climate change (IPCC 2001). Undoubtedly, a change in Svalbard's climate will affect snow cover characteristics, altering water and energy balances. Further, a change in snow distribution will have a great impact on the Arctic fauna and flora in this area. This paper presents an overview of snow research in Svalbard, which has increased significantly during the last decade. Available publications dedicated to snow research are relatively limited before the 1990s. Data on snow characteristics have often been by-products of glacier mass balance studies (e.g. Koriyakin & Troitskiy 1969; Palosuo 1987; Hagen & Liestøl

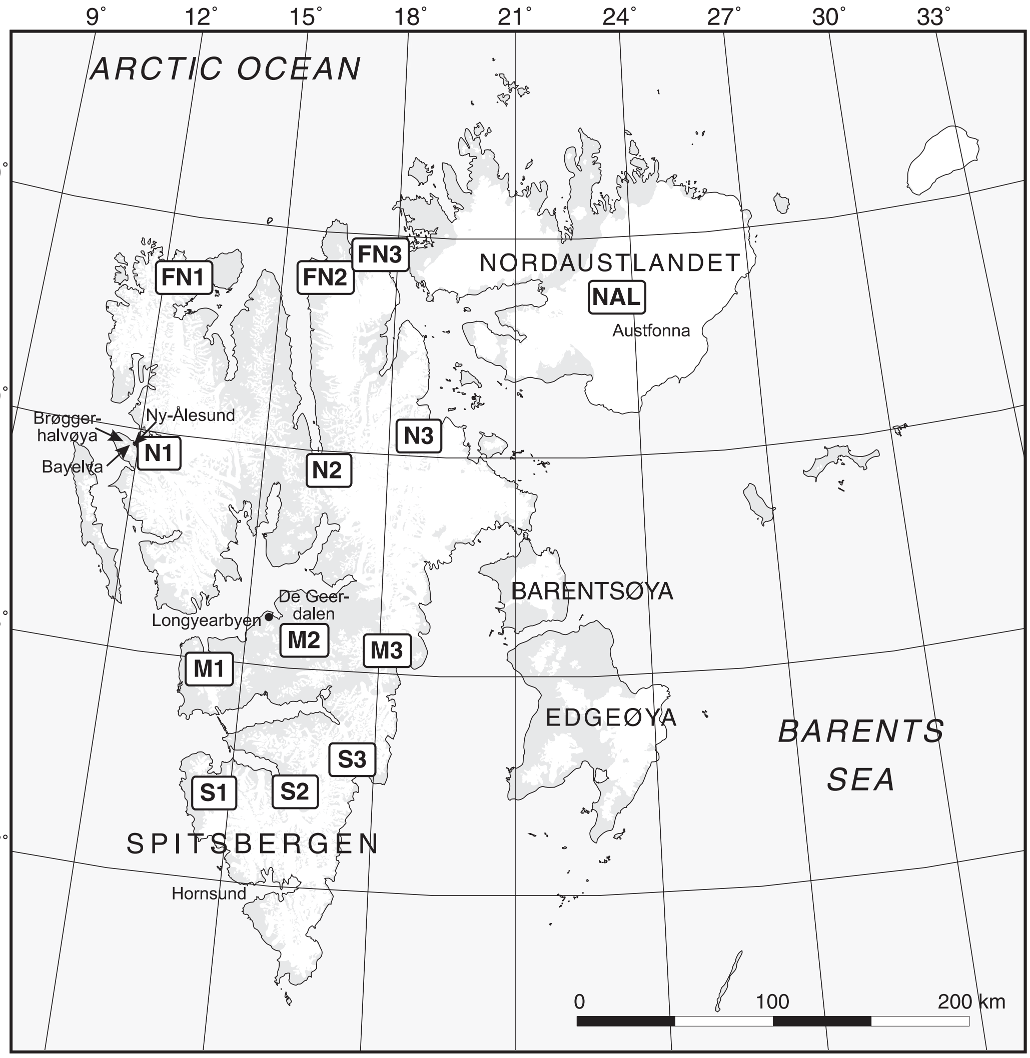


Fig. 1. The Svalbard archipelago (Bjørnøya, to the south, not shown). Study areas where snow accumulation measurements were collected during 1997–99 are indicated (after Sand et al. 2003).

1990) and some pre-1990s information relating to snow is not readily accessible because it was reported as grey literature.

Over the years, large amounts of data for studies of local distribution of winter snow accumulation have been collected (Troitskiy et al. 1980; Guskov & Troitskiy 1984). More recently, extensive data collection using ground penetrating radar (GPR) has enabled regional studies of snow accumulation (Winther et al. 1998; Bruland, Sand et al. 2001; Sand et al. 2003). These data shed new light on the amount and distribution of winter precipitation in Svalbard.

Observational and modelling studies of snowmelt have been conducted in the Ny-Ålesund area of western Spitsbergen (Fig. 1). Energy balance models like CROCUS and SNTHERM (Jordan 1991) have been used to model runoff during snowmelt. Model simulations have shown that it is crucial to correctly prescribe the surface albedo.

Extensive studies of the albedo of snow have been carried out during the spring melt (Gerland et al. 1999; Winther 1993a; Winther et al. 1999; Winther et al. 2002). These reflectance measurements enable comparison of ground measurements with (narrow band) satellite observations (Winther 1993b). The effects of snow metamorphism and surface contamination on snow albedo have also been carefully investigated. Additionally, routine albedo measurements have been taken at the Norwegian and German research stations in Ny-Ålesund since 1974 and 1992, respectively. There is therefore a comprehensive data set that can be used to ground-truth various other studies.

In all of the above-mentioned studies, data were also collected with respect to snow characteristics. These include snow density, snow temperature, liquid water content, stratigraphy, grain size and shape identification, snow depth and, in some cases, surface hardness and roughness.

Recent advances in Svalbard snow studies are also discussed. New, advanced satellite sensors such as Terra MODIS and ENVISAT MERIS enhance operational monitoring of snow albedo and snow coverage. Satellite remote sensing studies in Svalbard have generally been limited to glacier studies, often investigations of surface characteristics, glacier facies, front positions, crevasse patterns, mass balance and velocities (Dowdeswell 1986; Parrot et al. 1993; Winther 1993b; Rolstad et al. 1997; Engeset & Ødegård 1999; Bindschadler et al. 2001; König et al. 2002; Hagen et al. 2003). Recently, an algorithm for monitoring partial snow coverage using the MODIS albedo product in combination with the existing MODIS snow cover product has been suggested by Winther et al. (unpubl. ms.).

Finally, we identify research challenges and make recommendations for future work.

Snow distribution

Snow distribution, which displays high variation on local scales, tends to reveal significant variation in distribution patterns on regional scales as well.

Local snow distribution

Tveit & Killingtveit (1994) present the results of a snow survey programme carried out from 1991 to 1994 in the Endalselva catchment in the vicinity of Longyearbyen as well as in De Geerdalen, a valley about 11 km east–north-east of Longyearbyen. The measurement programme in De Geerdalen has been continued to date. Bruland, Sand et al. (2001) present the results of an extensive snow survey using a GPR (GSSI SIR-2) in May 1998 in the lower Bayelva catchment, near Ny-Ålesund. The measurements covered an area of about 3 km^2 and were carried out every metre along both north–south and east–west transects with about 100 m spacing. Similar measurements were conducted in 2000. Vonk (2001) carried out an extensive snow survey in De Geerdalen in 2001.

Bruland et al. (unpubl. ms.) modelled snow distribution using the snow transport model SnowTran-3D from Liston & Sturm (1998) in the Bayelva catchment area and in De Geerdalen. The results were validated for the years 1998 and 2000 in the Bayelva catchment and for 2000 in De Geerdalen by extensive snow surveys using GPR (GSSI SIR-2). The sites range from 3 km^2 in the Bayelva catchment to 250 km^2 in De Geerdalen and cover both rolling and mountainous terrain. In the Bayelva catchment, the modelling is based on high resolution topographical data (10 m resolution) and the simulated results have been compared to very detailed measurements of snow distribution. Two versions of the SnowTran-3D model were tested. In the WCurve model (Fig. 2) a more advanced curvature calculation is implemented compared to the original model (OrigCurve). The topographical data available for De Geerdalen has a lower spatial resolution of 100×100 m. Here, the snow depth measurements were carried out along selected snow courses since complete coverage is impossible over such large areas.

At both sites the snow is heavily redistributed. The snow depths ranged from 0 to about 3 m. In De Geerdalen, the average snow depth was about 0.50 m (2000) and in the Bayelva catchment it was 0.72 m in 1998 and and 0.60 m in 2000. The model was to a large degree capable of reproducing the observed snow distribution (Fig. 2).

Both the model and the observations show that the end of winter snow distribution depends strongly on the dominant wind direction at each site. Figure 3 shows the ratio between observed snow depths at different aspects and average snow depth in the Bayelva catchment for the two years with measurements—1998 and 2000—and for De Geerdalen for 2000. Typically, the accumulation areas in the Bayelva catchment are found on the leeward (western) side of mountains due to the prevailing easterly winds. Deeper snow is also found on the east–south-east facing slopes

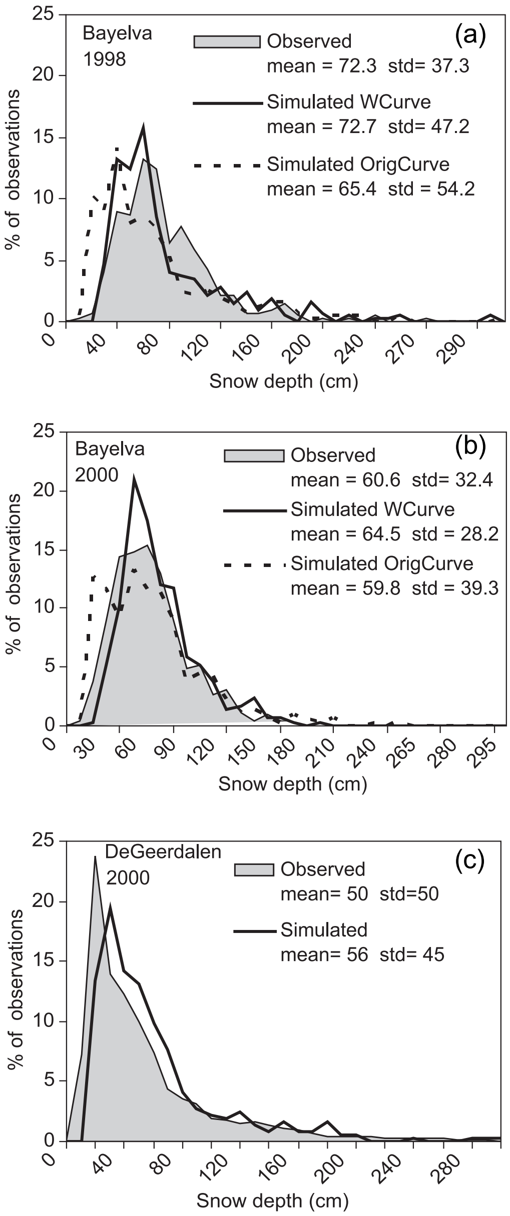


Fig. 2. Observed and modelled snow frequency distribution for Ny-Ålesund in (a) 1998 and (b) 2000 and for (c) De Geerdalen in 2000 (after Bruland 2002).

since the dominant wind direction during precipitation events is west–north-west. In De Geerdalen, snow distribution for different aspects are derived from the SnowTran-3D simulations. Here, the accumulation areas are found to be the north-facing slopes.

The snow distribution in the Bayelva catchment was compared to the vegetation distribution and a very clear correlation between snow depths and vegetation type was found (Bruland 2002). This illustrates how important snow depth is to the Arctic flora and how vulnerable these systems might be to changes induced by climatic change.

In 1998 an experiment investigating the redistribution of snow caused by man-made structures was carried out in Adventdalen using three model buildings made of 9 mm plywood (Thiis & Gjessing 1999). The experiment lasted for 14 days, which was sufficient to form significant snowdrifts on the buildings' lee sides. Meteorological measurements were performed during the experiments and at the end the snowdrifts were measured at 200 points with a surveying total station. In this area, the dominant wind direction is from the east (about 90% of the time during winter). All three buildings had the same floor space (2500 × 2500 mm) but with different rooftops. A flat roof produced the largest snowdrifts. A single pitch roof with an angle of 23° tilted towards the main wind direction gave larger snowdrifts than a similar single pitch roof tilted away from the main wind direction. Transport of snow was mainly by saltation.

Grześ & Sobota (2000) studied re-deposition of snow during events with high winds. They observed drifting snow amounting about 1000 kg over a 1 m^2 cross-section per day at wind velocities ranging between 4.8 and 5.0 m/s. When the wind velocity was reduced to 3.8 - 4.2 m/s the transport rates were about 300 - 350 kg per day over the same cross-section.

Regional snow distribution

Polish scientists started snow cover studies in the Hornsund fjord area in southern Spitsbergen in 1957 through investigations of mass balance of the Hansbreen and Werenskioldbreen glaciers (Kosiba 1960). The investigations showed that the snow accumulation on these glaciers was higher than in other places on Spitsbergen. Further, the snow in this area is significantly influenced by winds, mostly from an easterly direction. The mean snow accumulation on the Werenskioldbreen has decreased from 1570 mm water equivalent (w.e.) in 1956 to 1280 mm w.e. in the 1970s (Baranowski 1977). During the 1990s, snow accumulation rates did not exceed 1000 mm w.e. (Głowacki, pers. comm. 2002). The average increase of snow accumulation with altitude for

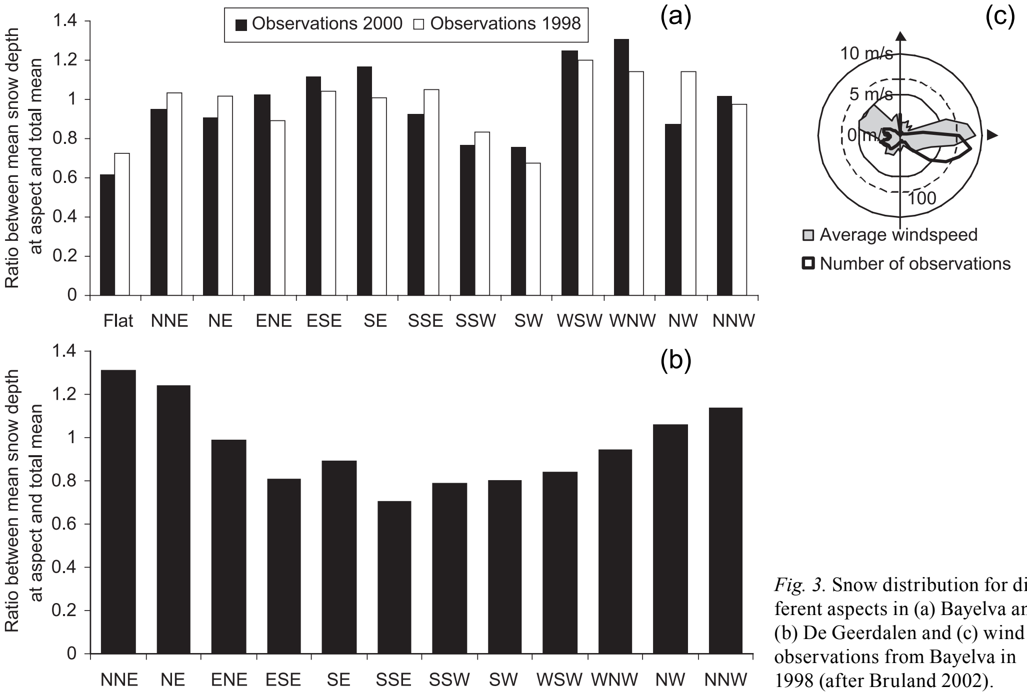


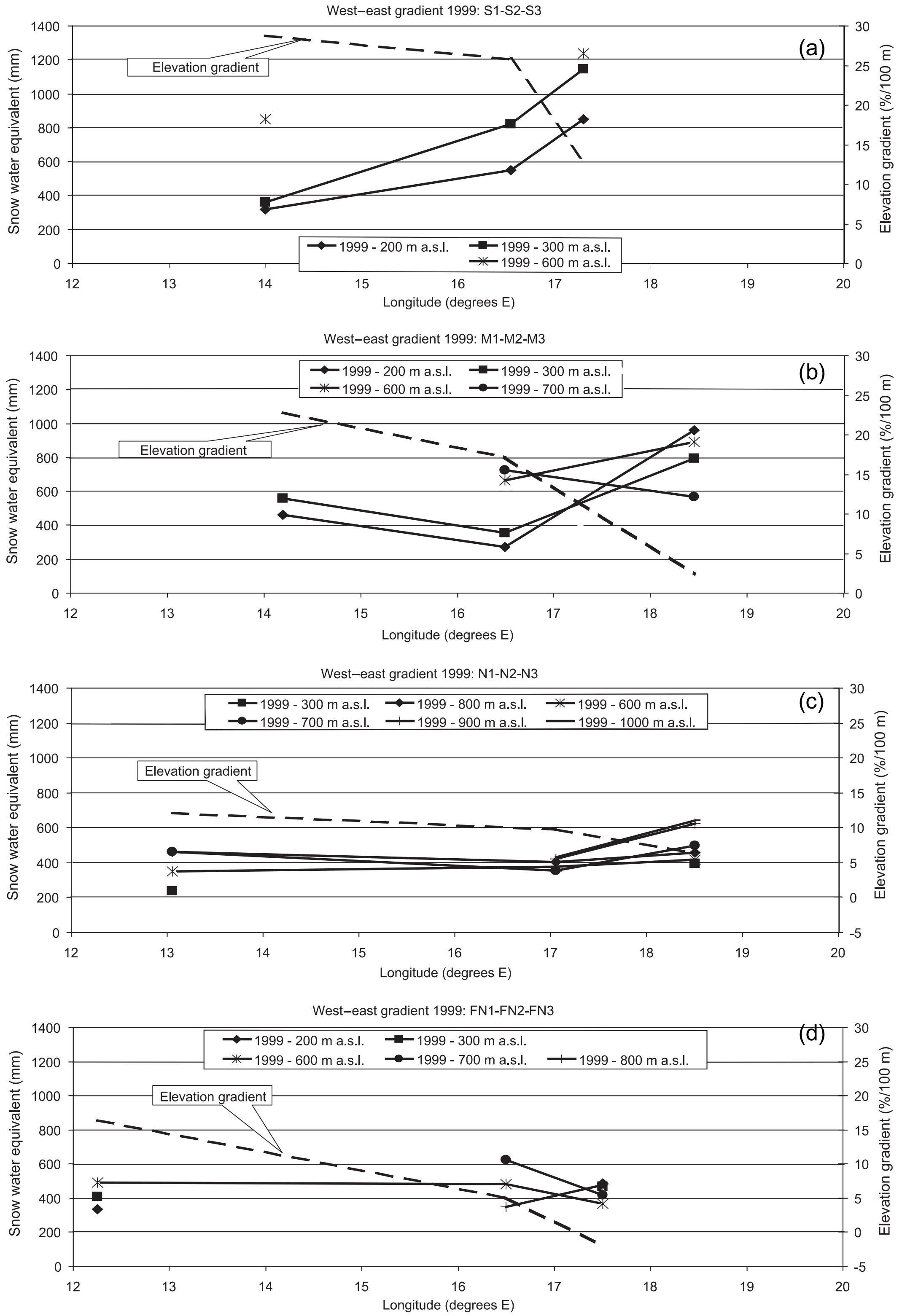
Fig. 3. Snow distribution for different aspects in (a) Bayelva and (b) De Geerdalen and (c) wind observations from Bayelva in 1998 (after Bruland 2002).

the glaciers in this region is found to lie between 110 and 190 mm w.e. per 100 m, except for altitudes below 100 m a.s.l., where snow erosion often takes place. Outside the glaciers, the mean thickness of snow is reported to be about 0.30 in valley outlets, 0.80 in upper parts of valleys and and 2.00 m in areas close to slopes (Migała et al. 1988).

More than 20 years ago, Russian scientists reported on the large variations in the regional distribution of snow accumulation in Svalbard (Troitskiy et al. 1980; Guskov & Troitskiy 1984). They found that the coastal areas of Spitsbergen received 2.5 to 3 times more winter accumulation than the central parts. Further, variation of snow accumulation from year to year was high at coastal sites but insignificant in the central areas.

More recently, Winther et al. (1998) collected large amounts of data on the winter snow accumulation in 1997 using GPR. They also found accumulation rates to be considerably higher (about two times) at the coasts compared to the central parts of Spitsbergen. Moreover, Winther et al. reported that snow accumulation was 39 to 49 % higher (depending on latitude) at the eastern coast than at the western coast of Spitsbergen. A south-to-north gradient was identified: 55 % less snow accumulation was measured at the northern locations compared to the southern locations.

The work by Winther et al. (1998) was expanded during the field seasons in 1998 and 1999 to include more areas in northern Spitsbergen and on Nordaustlandet (Fig. 1; Sand et al. 2003). This comprehensive data set on winter snow accumulation covering three years (1997–99) confirmed that the eastern coast receives approximately 50 % more snow w.e. than the western coast (Fig. 4). A continental effect, with lower accumulation rates, can be seen in central parts of Spitsbergen at middle and northern latitudes, but not in the south (Fig. 4). In the southern part of Spitsbergen accumulation rates are about twice as high as in the north (Fig. 5). Elevation gradients vary, but are on average 97 mm w.e. accumulation increase per 100 m increase in elevation. Data from the summit of Austfonna, the ice cap in Nordaustlandet, revealed large variations in accumulation (200 mm - 800 mm w.e., i.e. by a factor of 4) over a few tens of kilometres. Most likely, the subtle surface topography causes drifting snow to settle on the lee side of the summit of Austfonna.



West–east gradient 1999: S1-S2-S3
(a)
Elevation gradient
1999 - 200 m a.s.l.
1999 - 300 m a.s.l.
1999 - 600 m a.s.l.
Snow water equivalent (mm)
Elevation gradient (%/100 m)
Longitude (degrees E)
West–east gradient 1999: M1-M2-M3
(b)
1999 - 200 m a.s.l.
1999 - 300 m a.s.l.
1999 - 600 m a.s.l.
1999 - 700 m a.s.l.
Elevation gradient
Snow water equivalent (mm)
Elevation gradient (%/100 m)
Longitude (degrees E)
West–east gradient 1999: N1-N2-N3
(c)
1999 - 300 m a.s.l.
1999 - 800 m a.s.l.
1999 - 600 m a.s.l.
1999 - 700 m a.s.l.
1999 - 900 m a.s.l.
1999 - 1000 m a.s.l.
Elevation gradient
Snow water equivalent (mm)
Elevation gradient (%/100 m)
Longitude (degrees E)
West–east gradient 1999: FN1-FN2-FN3
(d)
1999 - 200 m a.s.l.
1999 - 300 m a.s.l.
1999 - 600 m a.s.l.
1999 - 700 m a.s.l.
1999 - 800 m a.s.l.
Elevation gradient
Snow water equivalent (mm)
Elevation gradient (%/100 m)
Longitude (degrees E)

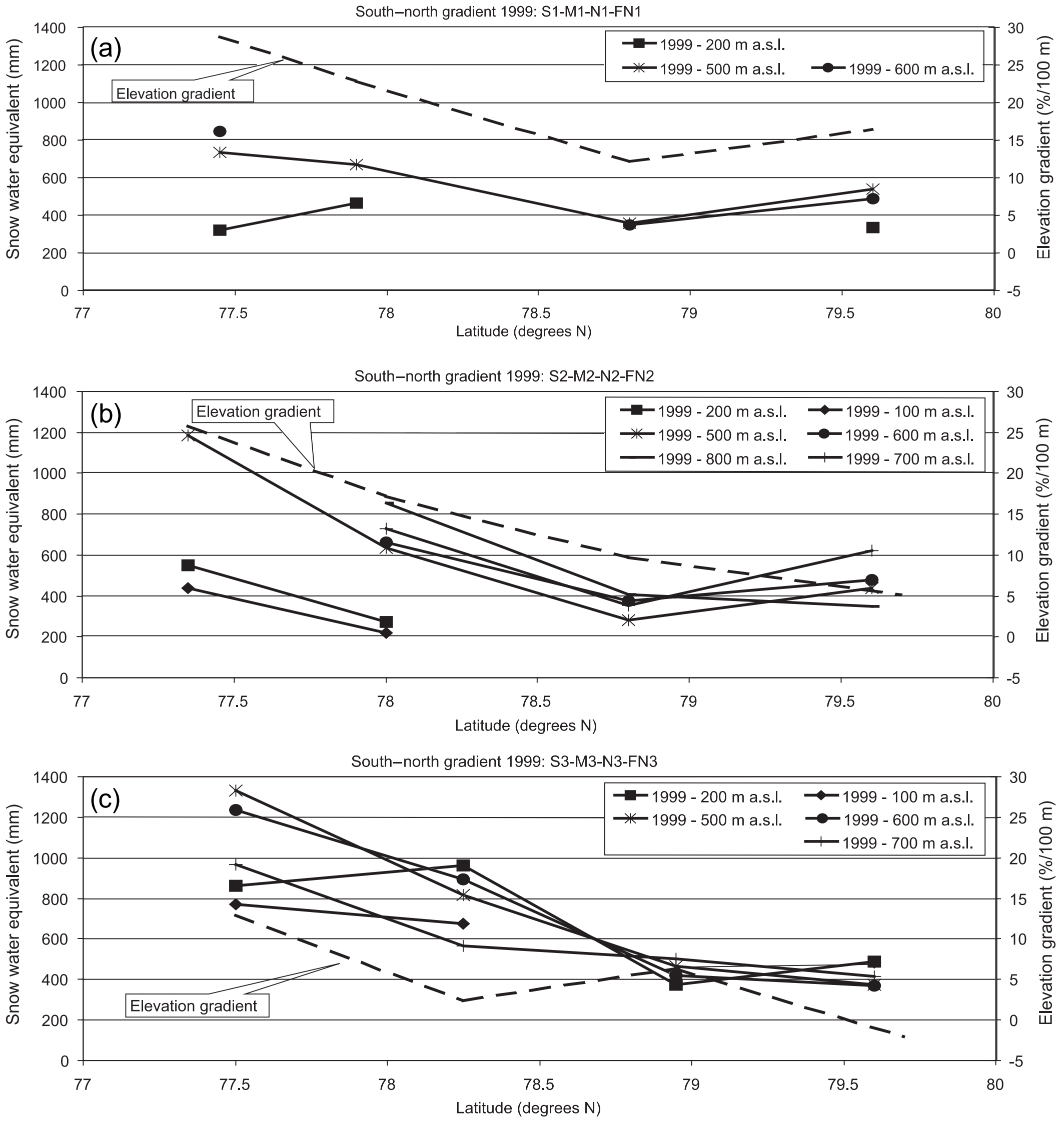


Fig. 5. Snow accumulation and accumulation/elevation gradients along south to north transects along (a) the west side of Spitsbergen, (b) the central part and (c) the eastern side as measured in 1999 (after Sand et al. 2003).

Surface energy balance

Most processes occurring in the snow cover are strongly related to energy fluxes. Hence, the energy balance of the snow cover (Eq. 1) is of primary importance:

Fig. 4 (opposite page). Snow accumulation and accumulation/elevation gradients along west to east transects on (a) the southern part of Spitsbergen, (b) the middle part, (c) the northern part and (d) the far northern part, as measured in 1999 (after Sand et al. 2003).

$$Q_m = Q_{sn} + Q_{ln} + Q_h + Q_e + Q_g + Q_p - \frac{dU}{dt}, \quad (1)$$

where Q_m is energy available for heating or melting of the snow pack, Q_{sn} is the net short-wave radiation flux, Q_{ln} is the net long-wave radiation flux, Q_h is the sensible heat flux, Q_e is the latent heat flux, Q_g is the ground heat flux, Q_p is heat transferred from rain with temperature higher than the snow's temperature and dU/dt is the rate of change of internal energy per unit area of the snow cover.

During the long Arctic winter, precipitation is stockpiled as snow. Winter radiation balance is dominated by long-wave radiation losses to space. There is hardly any short-wave radiation at all during winter, as the sun is below the horizon for about four months (depending on latitude). In the spring, net radiation turns positive and energy becomes available for snowmelt. Measurements by Harding & Lloyd (1998) show that the snomelt period is dominated by radiation fluxes, while sensible and latent heat fluxes keep very low. Normally, the snow cover disappears within a period of three to four weeks (Winther et al. 2002). After snowmelt much of the net radiation flux goes to melting of the frozen ground. Hence, the rate of evaporation from the tundra surfaces is generally low.

Snowmelt—observations

In the Hornsund area, Głowicki (1975) estimated melt rates of 11.8 mm snow/°C (equals 7.75 mm w.e./°C). The maximum decrease during one day was measured to 102 mm (67.2 mm w.e.). No ablation took place at elevations above 900 m a.s.l. The onset of melting normally occurs in the second half of May. Correlation between precipitation, measured at the Hornsund meteorological station, and the amount of water accumulated in the tundra snow cover is moderate (K = 0.42), while correlation with snow on glaciers in the region is very low (K = 0.07). This indicates that snowmelt is not dominated by rain-on-snow events; solar radiation is more likely the dominant factor.

Sand (1990) measured snowmelt intensities in the Endalen valley near Longyearbyen in June 1988. The monitoring period was mostly cool, with air temperatures around 0 °C, interrupted by one short event of strong winds, air temperatures up to 6 °C and high relative humidity. The melt rates observed in this study ranged between 0 and nearly 6 mm/h. Such events are rather characteristic for the climate in Spitsbergen; inflow of mild and moist air from the ocean may occur at any time of the year.

Bruland, Maréchal et al. (2001) observed snowmelt intensities from runoff plots on the tundra in the lower part of the Bayelva river basin west of Ny-Ålesund during spring snowmelt periods from 1992 to 1998. They found an average snowmelt rate of 14 mm/day, and a maximum of 68 mm/day. The highest snowmelt rates were observed a few days after the start of the ablation period. This is because it takes time before the snow pack over the entire runoff plot area becomes saturated and yields runoff. Also, drainage channels need some time to develop before the drainage becomes fully efficient.

Snowmelt—modelling

Sand (1990) modelled the snowmelt regime by applying a temperature index model as well as a full energy balance model. He found that the energy balance model was the only model capable of describing the snowmelt rates during the observation period. It should be noted, however, that the observation period was largely dominated by a single event of mild and moist air and high winds. Normally, temperature index snowmelt models will not give reliable results under such conditions.

Bruland, Maréchal et al. (2001) tested three snow cover models for the spring melt periods in 1992 and 1996 at the site in the Bayelva river basin. The three models were a temperature index model, a simple energy balance model and a complex energy balance model (CROCUS). They found that the dominant source of energy for melting during spring in Bayelva is the net shortwave radiation. This result was also verified by the modelling work of Maréchal (pers. comm. 2001) for 1998 at the same site and by Boike et al. (unpubl. ms.) for 1999 and 2000 at a proximal site. Results from the temperature index model strongly indicated that temperature alone is not a robust and representative index of the melting processes in Svalbard.

In fact, melt may be a response to varying meteorological conditions. Bruland, Maréchal et al. (2001) and Boike et al. (unpubl. ms.) found that a simple energy balance model can be suitable for spring melt modelling as long as the melting process is continuous and an isothermal snow pack at 0 °C can be assumed. In the case of periods with re-freezing, it is necessary to take into account the thermal energy of the snow pack. Of the three models tested by Bruland, Maréchal et al. (2001), the complex energy balance model (CROCUS) predicted spring melt best. However, Maréchal (pers. comm. 2001) tested two physically based models based on the same principles (CROCUS and SNTHERM) and found that both models poorly described the variations of surface albedo throughout the melting season. Through the insufficient simulation of albedo, melt rates are modelled inaccurately. Boike et al. (unpubl. ms.) also tested their energy balance model for melting due to warm weather events occurring during the winter period. They concluded that the melting processes during such events were dominated by sensible heat fluxes and heat fluxes from rain.

Surface fluxes of heat and water

Harding & Lloyd (1998) and Lloyd et al. (2001) conclude that the snowmelt in Ny-Ålesund is dominated by radiation input. Harding & Lloyd reported melt rates early in the melt season of 17 and 20 mm/day in 1995 and 1996, respectively. Later in the 1995 season, the melt rates decreased to less than 10 mm/day. It might be expected that snow at a patchy site would melt faster with the local advection of sensible heat from surrounding snow-free areas. However, Harding & Lloyd found no evidence that the development of patches affected the rate of snowmelt. This may happen because of the very stable air temperature that exists in this area, probably controlled for the most part by the surrounding sea, snow and ice fields.

The sensible heat flux is very low and negative in the main part of the melt period, only becoming positive at the end of the melt (Harding & Lloyd 1998). Harding & Lloyd found that evaporation uses between 20 and 30 % of the net radiation, but this is small in terms of quantity of water evaporated. Similar values for the snowmelt energy balance in Ny-Ålesund were found by Takeuchi et al. (1995) and Nakabayashi et al. (1996). Lloyd et al. (2001) report that an approximately equal balance between sensible and latent heat occurs early in the season, while larger latent heat fluxes occur during the second half of the melt period.

Snowmelt and slush flows

During snowmelt water can accumulate in low gradient slopes or valley sections, often resulting in the release of snow and water masses due to hydraulic effects. These mass displacements, normally following a narrow path, have been termed slush flows or slush avalanches. They are primarily released due to the hydraulic gradient evolving from an increasingly inclined meltwater table within the snow pack (Gude & Scherer 1998; Scherer et al. 1998). Scherer et al. (1998) studied slush flows in Liefdefjorden, north-western Spitsbergen. They observed that slush flows could be released due to energy input in the snow cover by the net radiation and sensible heat within the ordinary range of a high latitude snowmelt period, i.e. without rainfall events. The sensible heat flux is particularly important for slush flow initiation. Also crucial for slush flow initiation is the timing of energy input and meltwater flow through the snow pack. Further, infiltration losses can be disregarded, even when permafrost is not present.

Snow cover characteristics

Long-term albedo measurements

Surface albedo has been measured routinely at the Norwegian Polar Institute's (NPI) research station in Ny-Ålesund since 1974 (Hisdal et al. 1992; Hisdal & Finnekåsa 1996; Ørbæk et al. 1998; Winther et al. 1999; Winther et al. 2002). During 1974 to 1981 only monthly values were recorded. Since 1981, hourly values are available. The staff of the German Koldewey Station (Alfred Wegener Institute for Polar and Marine Research) in Ny-Ålesund has measured surface albedo since 1992.

Winther et al. (2002) statistically analysed the surface albedo data from NPI for the period 1981–1997. Regression analysis showed no significant trends in the 17-year period with respect to changes in the timing of the onset of snowmelt, melt rates, the timing of snow arrival in the autumn, and the period without snow cover. The analysis revealed that 5 June is the date with the highest likelihood for the onset of snowmelt; the start of snow formation is most likely to fall on 17 September. The highest probability for the length of the snow-free season in Ny-Ålesund is 94 days. Winther et al. (2002) found no correlation between the North Atlantic Oscillation (NAO) index and albedo nor temperature or precipitation. However, the existence of a correlation on decadal time scales could not be ruled out because of the limited length of the albedo data.

Extinction of solar radiation in snow pack

The extinction of photosynthetically active radiation (PAR) over time in a snow pack in the Bayelva basin has been investigated (Gerland et al.

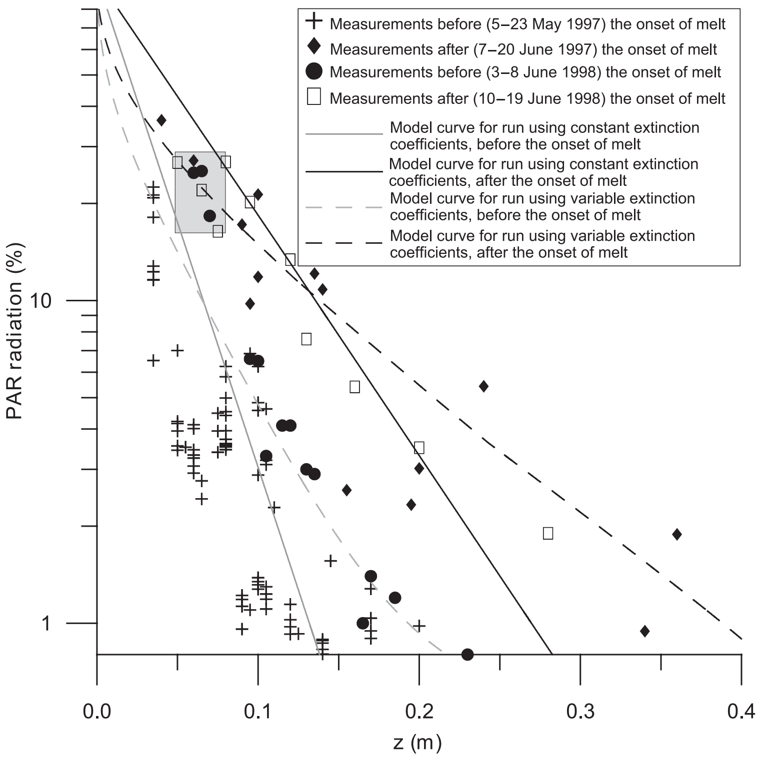


Fig. 6. PAR radiation data and model curves. Data from various snow pits and days at Ny-Ålesund and Bayelva are shown for measurements before and after the onset of melting. The relative instrument reading I(z)/I(surface) is plotted at logarithmic scale versus snow thickness z above sensor. Model curves are displayed for runs using constant and variable extinction coefficients. In 1998, the data indicated "before melt" (circles) were actually obtained while melting had already begun near the surface. The lower snow layers were not affected. This explains different characteristics of data from near surface and deeper snow. Squares (after onset of melt) indicate values measured when the entire snow pack was isothermal (0 °C) where PAR measurements were obtained (after Gerland et al. 2000; printed with permission of *Annals of Glaciology*).

1999; Gerland et al. 2000). Analyses of the bulk PAR showed that temperate snow which underwent metamorphosis appeared more transparent than colder, less changed snow, which is in agreement with other studies (e.g. Grenfell & Warren 1999). A comparison with a one-dimensional two-stream model showed that the snow grain's microstructure might play an important role in the attenuation of solar radiation in snow (Fig. 6). However, by enhancing the snow transparency, a dark ground surface becomes visible earlier from above the snow surface, which again reduces surface albedo and this further enhances melt. This feedback may play a crucial role in melting processes in polar land areas. Consequently, moderate increases in temperature and precipitation could result in substantial changes of the duration and intensity of snowmelt.

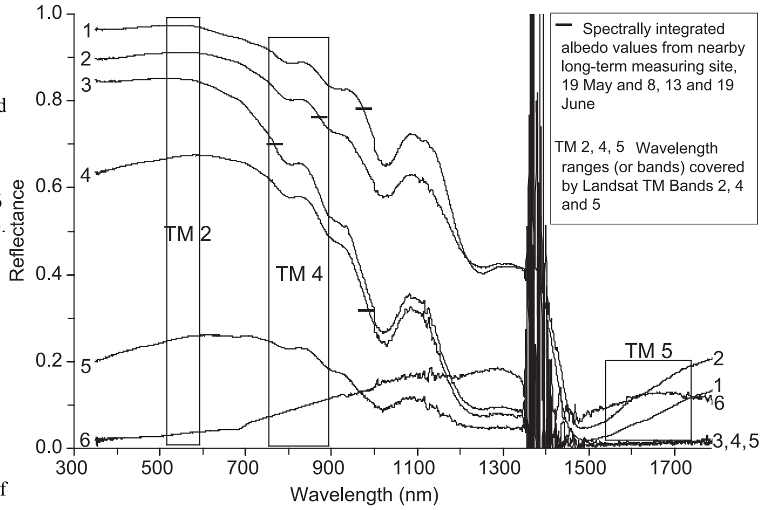
When looking at the spectral composition of solar radiation penetrating the snow pack (Gerland et al. 2000), one can see that during the melting period, the transparency increases most in the early phase of melting (measurements on 25 May and 8 June 1998). Later, no significant change in snow transparency was registered (12 June 1998). However, the transparency in the lower part of the PAR range (400 - 450 nm) was less enhanced than at higher wavelengths (450 - 700 nm).

Spectral reflectance of melting snow

Field experiments where snow reflectance has been measured have been carried out during several spring seasons in the Ny-Ålesund area (Winther 1993b; Gerland et al. 1999; Ørbæk et al. 1999; Winther et al. 1999; Cassacchia et al. 2001; Svendsen et al. 2002). Typically, reflectance in the visible part of the wavelength region is high for dry, freshly fallen snow (>0.80) but drops rapidly when melting starts (<0.50). While the reflectance of snow is fairly stable within the visible region, it drops quickly in the near-infrared region (Fig. 7). Snow becomes almost non-reflective at wavelengths in the mid-infrared region.

Factors that influence the reflectance of snow are surface roughness, surface contamination, degree of metamorphosis (crystal grain size as well as shape), liquid water content and snow depth (especially when the snow pack becomes

Fig. 7. Examples of snow reflectance measured at Ny-Ålesund. Curve 1 (19 May, clear skies), represents dry snow that had undergone minor metamorphosis and had an apparently unpolluted surface. Curves 2 and 3 (8 and 13 June, respectively, cloudy) were taken at melting snow surfaces. Curves 4 - 6 were measured on 19 June (cloudy). Curve 4 represents snow that had undergone about one month of snow metamorphosis and had some surface blackening. Curve 5 was acquired where a 0.5 - 3 cm snow layer covered a 5 - 10 cm ice layer on the ground. The surface was strongly blackened by soil particles. Curve 6 was measured over bare tundra (after Winther et al. 1999; printed with permission of John Wiley & Sons Limited).



shallow). Most of these parameters change rapidly during spring melt and it can be complicated to isolate the contribution from each one of them (Cassacchia et al. 2001). However, snow metamorphosis (crystal size growth, rounding and increasing liquid water content) is most effective in lowering the albedo of snow at the beginning of the melt season (Winther et al. 2001). This reduction in snow albedo is most prominent in the near-infrared wavelength regions (Fig. 7). Later in the melt season, contamination of the snow surface (mostly by transportation of organic material from surrounding snow-free areas) tends to most effectively reduce snow albedo, especially in the visible wavelength region (Fig. 7). This shift from early season reduction in the infrared region to late season reduction in the visible region has implications for satellite-derived albedo measurements because normally a specific satellite (narrow) band is used to calculate the broadband surface albedo (Winther et al. 1999).

Winther (1993b) investigated the effect of cloud cover on snow albedo. The integrated albedo (300 - 900 nm) increased from 0.81 to 0.87 (i.e. by about 7 %) when the weather condition changed from clear sky to 100 % overcast within two hours on 9 June 1992. This occurs because clouds absorb a higher proportion of infrared than visible radiation. Thus, a relatively high proportion of the visible radiation reaches the ground under cloudy conditions. Since the visible snow albedo is very high (> 0.90) compared to the near-infrared albedo (about 0.50), surface albedo increases (Winther 1993b).

Snow density

Sand et al. (2003) report snow density measurements from 12 locations on Spitsbergen taken during the end-of-the-winter snow surveys in 1997, 1998 and 1999 (Fig. 1). The locations were distributed from approximately 77 - 80° N and 12 - 19° E and covered an altitude range from 100 to 600 m a.s.l. The snow density varied from 271 kg m^{-3} to 572 kg m^{-3} with an average value of 374 kg m^{-3}. They found no significant correlation between density and elevation. Grześ & Sobota (2000) found an average snow density of 420 kg m^{-3} at the Kaffløyra glaciers on Spitsbergen (about halfway between Ny-Ålesund and Longyearbyen). Johnsen (2000) did vertical snow density profiles in six different snow pits in the Bayelva basin during the snowmelt periods in 1998 and 1999—in the same area as the runoff plots described by Bruland, Maréchal et al. (2001). Snow density measurements were taken every 0.1 m along a vertical profile from top to bottom of the snow cover. Very different snow texture in different layers of the snow column showed very high variability of density along the vertical profile.

Snow temperature

Vertical temperature profiles in snow pits give a history of the local temperature regime. Temperature variations in the snow are affected by time-dependent changes of solar radiation, air temperature, wind, ground temperature and thermal snow properties. Snow temperature profiles have been collected during spring by different research projects in the vicinity of Ny-Ålesund (Gerland et al. 1999; Gerland et al. 2000; Gerland et al. 2001; Boike et al. unpubl. ms.). Gerland et al. (1999) report that the temperature profile changed from a linear decrease with depth to a constant snow temperature at 0 °C throughout the snow pack during 9 days in June 1997, a typical picture during the peak melting period when air temperature and intensities of solar radiation increase.

Continuous snow temperature measurements are also possible using fixed installations with temperature sensors and data loggers (e.g. Gerland et al. 2001). This method is not as destructive as digging snow pits but the permanent presence of the sensors and other equipment affect the formation and decay of snow. For example, a well-known problem is solar radiation heating of thermistor chains, which both artificially increases the observed temperature and also replaces snow with air in the near vicinity of the chains because of melting. The latter can produce enhanced ventilation, creating a different temperature profile than under undisturbed conditions.

Liquid water content

Bruland, Maréchal et al. (2001) and Maréchal (unpubl. data) measured liquid water content (LWC) in the Bayelva river basin during 1992–99. They used an electronic device (LEAS TEL05.1) to measure the real part of the dielectric constant every 0.10 m along a vertical profile in snow pits. Knowing the dielectric constant and the bulk density of the snow sample, LWC could be calculated. During the snowmelt season 1999, LWC varied from 0 to 10.9 % (by volume). Boike et al. (unpubl. ms.) measured LWC throughout the winters 1998/99 and 1999/2000 from a proximal site in the Bayelva river basin. They measured LWC using time domain reflectometry probes installed vertically in the snow pack. Their measurements showed that LWC was rather constant during the winter, approximately 1 % (by volume) during the winter 1998/99 and approximately 4 % during the winter 1999/2000. Their measurements also show that LWC is highly variable during the beginning of the snow-covered season as well as during the snowmelt period. During the snowmelt period in the spring of 2000, LWC reached values of approximately 20 %. Casacchia et al. (2001) measured LWC values at 0 for all measurements in April 1998 using a snow fork.

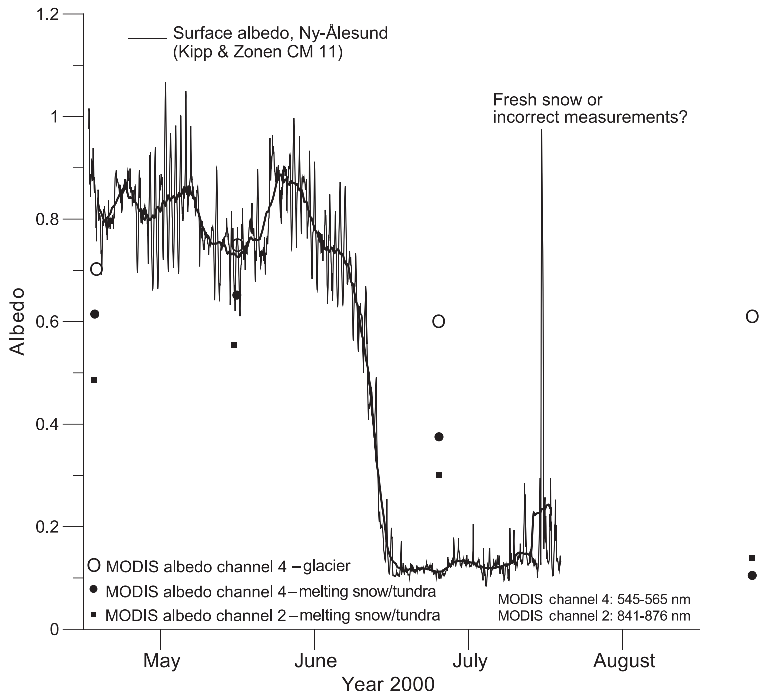
Basal ice

Gerland et al. (1999) and Bruland, Maréchal et al. (2001) report on basal ice forming either during rain-on-snow events in winter or when melt water percolates through the snow pack in spring, freezing when meeting the almost impermeable frozen ground. This ice layer will continue to grow until the temperature of the uppermost ground has warmed to 0 °C. Bruland, Maréchal et al. (2001) present data from 1996, when the basal ice grew to a thickness of approximately 0.10 m (average value over the 100 m^2 runoff plot area) during the snowmelt period. Maréchal (unpubl. data) found that the basal ice on the same runoff plot reached up to 0.19 m during the snowmelt period in 1999. Gerland et al. (1999) found basal ice thickness varying between 0 and 0.20 m from measurements during spring in 1997 and 1998.

Deposition of marine salt in snow

In the Hornsund area, investigations of the deposition of marine salts in the snow pack have been carried out over the last 15 years. These studies show that large amounts of salt can be deposited in the snow cover during winter. Salt that is released from the snow pack during the melting period can reach amounts of 40 t/km^2. They can significantly influence biological and geo-morphological processes. Studies have been also made to determine velocity of the washing out of salts from the snow cover during melting. The effectiveness of such washing out depends on the amount and velocity of water percolating to deeper layers of the snow pack and to the ground. It also depends on snow pack stratigraphy. For example, ice and crust layers can delay the flushing. Washing out of salts is much faster than reduction of the snow cover thickness. It often exceeds 90 % of the whole accumulated load during the first phase of snowmelt (Głowacki 1997). Not all ions are washed out with the same velocity.

Fig. 8. Comparison of broadband albedo (285 - 2500 nm) measured in Ny-Ålesund and MODIS Bands 2 and 4 reflectance data from one melting snow site at the tundra and one on a nearby glacier (Kongsvegen) in 2000. The tundra site corresponds to where the broadband albedo measurements are taken. The glacier site is located about 20 km east of Ny-Ålesund and at an elevation of about 500 m a.s.l. Thus, no blue ice is exposed during the time of observations (Winther et al. unpubl. ms.).



Remote sensing of snow albedo

The reflectance of snow can vary significantly in time and space during snowmelt (König et al. 2001). Winther (1993b) reported that satellite-derived Landsat TM Band 4 surface albedo varied between 0.65 (snow) and 0.19 (blue ice) at Midre Lovénbreen on 31 August 1988. In situ measurements showed a drop in daily mean albedo from 0.88 to 0.13 during four days at a fixed location on the glacier Austre Brøggerbreen (on Brøggerhalvøya) in August 1991 (Winther 1993b). At this location, the snow was relatively dry at the beginning of the measurement period while most of the snow had melted away at the end, leaving an exposed blue ice surface.

The bi-directional reflectance of snow was also measured by taking spectral scans for viewing angles starting at nadir and ending at 60°, with steps of 15° for viewing directions facing the sun and at azimuths 90° and 180° away from the sun. The increase in albedo relative to the nadir albedo was found to be 8, 15, 19 and 26 % for viewing angles 15°, 30°, 45° and 60°, respectively. The largest anisotropy was found for metamorphosed snow in measurements facing the sun. It is therefore important to correct satellite-derived surface reflectance for the specular properties of snow if it is going to be used as absolute albedo values.

A new generation of satellite imaging spectroradiometers such as the Moderate Imaging Spectroradiometer (MODIS) onboard NASA's Terra (EOS AM-1), launched in December 1999 (a second MODIS sensor was launched in May 2002), represent an improved tool for global monitoring of snow cover and surface albedo. Imaging spectrometers are able to record a continuous spectral range, in contrast to multispectral scanners, for example on Landsat, which only record distinct, selected bands within a spectral region.

Winther et al. (unpubl. ms.) used the MOD09A1 albedo product that is an 8-day composite of daily surface reflectance data and compared these with ground-truth measurements from Ny-Ålesund. The main goal was to check whether (absolute) albedo values recorded by MODIS corresponded well with ground measurements. In Fig. 8, narrow-band albedo derived from MODIS Bands 2 (841 - 876 nm) and 3 (459 - 479 nm) and the corresponding broadband albedo recorded in Ny-Ålesund are plotted. Winther et al. (unpubl. ms.) selected the pixel in the MODIS images that corresponded best to the location where ground measurements are recorded. Additionally, they

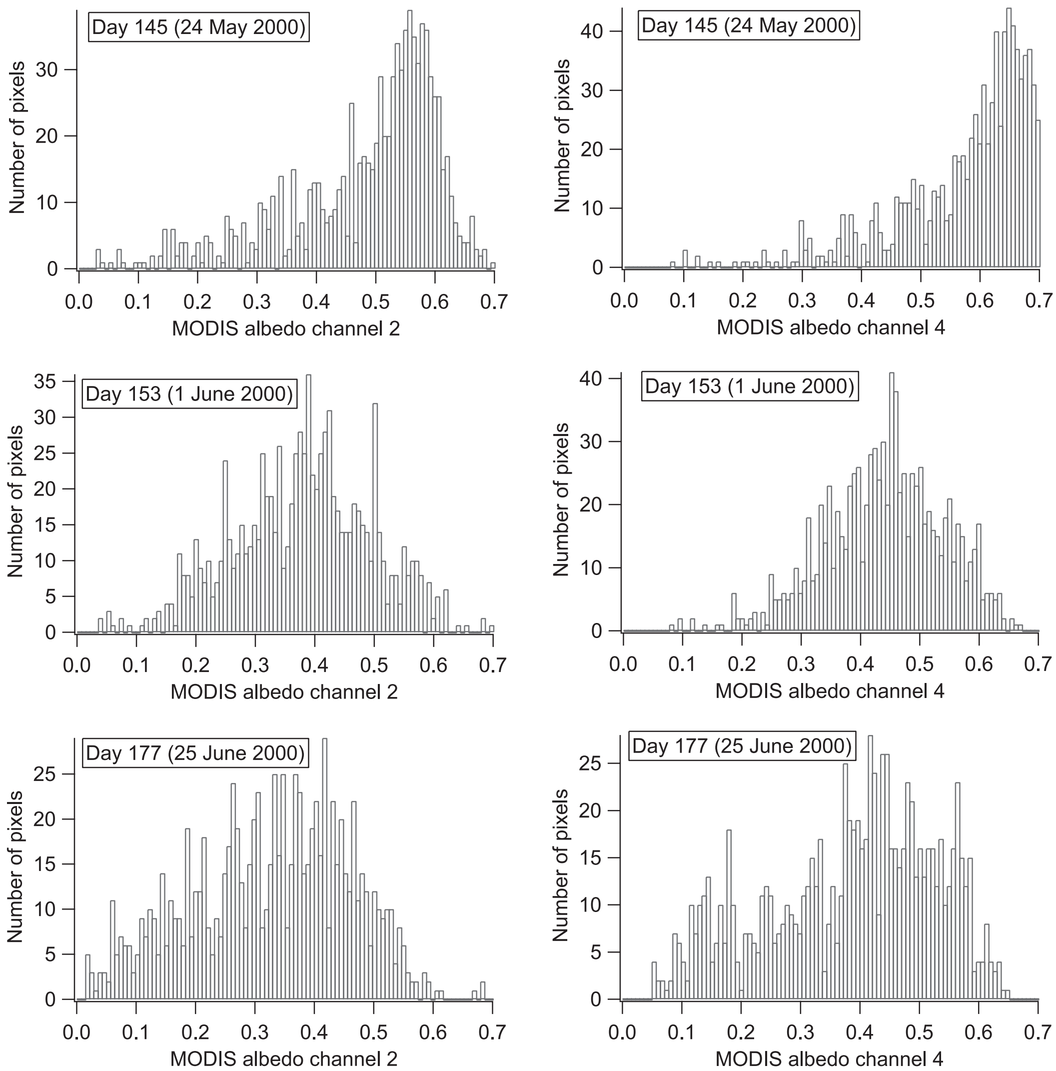


Fig. 9. Histograms showing the frequency of occurrence of albedo values at the Brøggerhalvøya peninsula on 24 May, 1 June, and 25 June in 2000 for MODIS Bands 2 and 4.

plotted MODIS albedo from a nearby glacier, Kongsvegen, at an elevation well above the equilibrium line altitude that was certainly not snow-free during this period. This site was used as an independent location where the albedo was expected to decrease due to metamorphosis of snow and not drop down to albedo levels expected for bare ground (as for the tundra site) or in this case, more realistically, albedo values of blue ice. In this way, they could check the performance of the MODIS albedo product for two sites that exhibited different snow melt-off regimes.

Winther et al. (unpubl. ms.) also plotted histograms of the MODIS Bands 2 and 4 albedo products for three different dates at different stages in the snowmelt season covering the Brøggerhalvøya peninsula, i.e. over an area much larger than the test area used in Fig. 8 (Fig. 9). On the first date (day 145, 24 May) the snow is still dry. Next, on 1 June (day 153) snowmelt has started and the surface albedo decreases (i.e. histogram is shifted towards lower albedo values). Late in the melt season, when the snow cover is unevenly distributed, surface albedo continues to decrease

(Fig. 9). Further, we can see a bimodal distribution in the histogram on day 177 for MODIS Band 4, reflecting the fact that there are two main groups of surfaces in the catchment: snow-covered ground and snow-free ground. In addition, a third group is identified as mixed pixels.

The MODIS snow albedo data considerably underestimates the absolute value of snow albedo that is measured in situ (Fig. 8). The most likely explanation for this deviation is that the MODIS pixel surrounding the ground measurement site includes not only snow but also a smaller portion of snow-free ground. It is quite common that several surface types occur within 500 × 500 m during spring melt in this area. In spite of the above discrepancy, the albedo decrease during the transition from dry snow, via wet snow, to bare ground is captured very well by the MODIS albedo data (Fig. 8). Also, the values of bare ground MODIS data correspond very well with in situ measurements. At last, the albedo reduction on a snow-covered glacier seems to be recorded realistically by MODIS.

The influence of snow cover on biota

Svalbard reindeer population

Unlike most populations at lower latitudes, the Svalbard reindeer population seems not to be dependent on the summer climate (Aanes et al. 2000; Solberg et al. 2001). In contrast, winter conditions appear to be critical. Aanes et al. (2000) and Solberg et al. (2001) found that the amount of winter precipitation was negatively correlated with the annual variation in population growth: low growth rates when winter precipitation was high. Most likely, this occurs because snow covers the food or increases the costs of movement, leading to starvation and the death of individuals in poor condition (Aanes et al. 2000).

Solberg et al. (2001) found winter precipitation to be negatively correlated with population growth rate, calves per female and mortality index. Further, an icing index based on the sum of precipitation during days with temperatures above 0 °C was negatively correlated with calves per female and mortality index. The icing index may indicate a more solid snow pack (due to re-freezing), ice layers within the snow pack or ice formation at the ground. Solberg et al. found a positive correlation between the NAO index and calves per female but no correlation between NAO and the population growth rate, which is in agreement with Aanes et al. (2000).

Interestingly, Solberg et al. (2001) point out the importance of snow distribution for survival of the reindeers in winter. Reindeer feed where wind uncovers the ground or where the snow pack is shallow. This leads the reindeer to aggregate in smaller areas, typically on mountain slopes and at higher altitudes. This seems like a area where snow hydrologists and biologists can potentially collaborate.

Barnacle geese

Prop & Vries (1993) studied the impact of snow conditions on the reproductive performance of barnacle geese (*Branta leucopsis*). They observed that when snowmelt was late it affected (in descending order of importance): nest success, brood size at hatching and the proportion of pairs that initiated breeding. Also, the parts of the tundra with the highest production of food were the last to become available for the geese. This shows again that snow distribution, and not only snow depth, is vital when studying the ecosystem.

Vegetation

Rieley et al. (1995) reported on the effect of snow cover thickness on the leaf wax characteristics of two Arctic species: *Dryas octopetala* and *Saxifraga oppositifolia*. To test whether biochemical and anatomical attributes might change in response to short-term alterations in winter climate, snow fences were erected on ridge sites. The wax attributes of ridge plants exposed to a single year of increased winter snow thickness showed that the *n*-alkane composition of leaf waxes were more like those of plants growing in adjacent swale areas than for those of ridge plants growing in unmanipulated areas. This shift in leaf wax composition implies that climatic changes during leaf development can have an influence on final leaf wax composition (Rieley et al. 1995).

As reindeer preferentially forage in places with little or no snow cover during winter, changes in snow distribution could impact vegetation. Laboratory experiments by Wegener & Odasz (1997) suggest that reindeer forage grasses respond differently to increased grazing. With increased clipping, *Poa arctica* and *Festuca rubra* increased their below-ground biomass and decreased their above-ground biomass. *Deschampsia alpina*, in contrast, increased its above-ground biomass and decreased its below-ground biomass, rendering this species more vulnerable to grazing and thus to changes in snow distribution caused by climate change.

Snow algae

The identification and distribution of snow algae in north-west Spitsbergen has been studied during four seasons—1995–98 (Müller et al. 1998; Müller et al. 2001). The snow algal fields are mostly found near bird colonies on steep slopes. Persistent fields are mainly the result of suitable local conditions where the following is essential: availability of nutrients, presence of wetness in snow, and the quantity and quality of solar radiation. In early summer, meltwater streams concentrate cells in the upper part of snow slopes, below the mountain edge. In addition to water availability, horizontal ice occurring less than 0.40 m from the surface is favourable for snow algae blooming. Long-distance dispersion is dependent on dry snow conditions and strong winds between autumn and spring or cells carried by birds. Resting cells, deposited and frozen near or in the ground ice are probably not the carriers of the next year's algal bloom (Müller et al. 2001).

Research challenges

Snow distribution

Extensive snow surveys have been carried out over the latest years in Svalbard using georadar (Winther et al. 1998; Bruland, Sand et al. 2001; Jaedicke 2001; Bruland et al. unpubl. ms.; Sand et al. 2003). These snow surveys have both proven the efficiency of the georadar as a tool to measure snow depths and provided large data sets with detailed information on snow distribution. These data sets have been analysed and correlations between snow depths, location and topography have to some degree been studied.

Bruland, Sand et al. (2001) describe the local snow distribution in lower parts of the Bayelva catchment area. To transfer the result found in this study to other areas of Svalbard more general analyses of the correlation between topography and snow distribution would be valuable. Comparable studies in other parts of Svalbard would also be beneficial in validating the results from Bayelva. The studies in Bayelva have been of a local nature, while the studies by Winther et al. (1998) and Sand et al. (2003) were of a regional nature.

There are additional measurements on local scales but these are not extensive or dense enough to study the effect of local topography on snow distribution, for example, how far a large topographical feature, such as a mountain, exerts influence on snow transport, or what fetch distance is necessary to obtain snow saturated air masses (balance between erosion and sedimentation of snow from the airflow). This is important information for modelling snow transport. These models can be valuable tools in studies of snow redistribution and consequences of potential climate change.

The snow threshold against erosion is another matter of great importance for snow transport modelling. The description of this snow property is based on early studies but is still not fully understood. Bruland, Sand et al. (2001) state that the climate and snow properties in Svalbard reduce the snow transport rate compared to Norwegian mountainous areas. Implementation of the SnowTran-3D model shows how important this property is for the redistribution (Bruland et al. unpubl. ms.). Detailed studies and better description of this property would improve the model parameterization and thereby improve the model performance.

Snow distribution has a great influence on snowmelt, directly through location and exposition of the snowdrifts and indirectly through its influence on sensible heat transfer by heating of air masses over bare spots and the lower reflectance of solar radiation from bare slopes. Studies of the effect of these latter impacts of snow distribution could improve snowmelt modelling and snow cover depletion.

Snow distribution has also a large influence on the fauna and flora. The snow distribution data sets in Bayelva open a range of possibilities for investigating the impact of snow distribution on vegetation distribution and the foraging behaviour of Svalbard reindeer, as suggested earlier.

Snowmelt modelling

The snowmelt studies of Sand (1990) and Bruland, Maréchal et al. (2001) confirm that simplified snowmelt models such as temperature index models are not capable of describing snowmelt processes satisfactory. Some kind of physically based energy balance model is necessary to parameterize the different energy components contributing to the snowmelt process. The energy balance models SNTHERM and CROCUS used by Maréchal (unpubl. data) even suffer from insufficient parameterization of albedo. Since snowmelt in this area is mainly driven by radiation, albedo is a very important parameter and it is crucial to improve its description in the snowmelt models.

All the known snowmelt modelling studies in Svalbard to date have been limited to one-dimensional or plot scale models—a correct approach toward understanding how to describe the snowmelt process. Such models have been the key to development of global climate models. However, to provide snowmelt models for watershed scale hydrology modelling or quantifying sub-grid scale variability we need to develop these models towards spatially distributed models. Several challenges will be faced. One is snowmelt modelling of unevenly distributed snow covers where the surface is a mosaic of snow-covered ground and snow-free patches. The energy exchange over such an area with mixed surface types creates significant lateral fluxes of energy.

Remote sensing of snow coverage

Since the MODIS reflectance product has a continuous representation of snow albedo, while the MODIS snow cover product only gives two values (snow or no snow), the former can be used to produce a continuous representation of snow coverage. In it simplest version, if the albedo of snow and bare ground are 0.80 and 0.10, respectively, then an albedo value of 0.45 indicates snow coverage of 50%, assuming a linear relationship between albedo and snow cover:

$$\text{Snow coverage in \%} = [(\text{snow albedo} - \text{ground albedo})/2 + \text{ground albedo}] \times 100 \qquad (2)$$

In this way, we would be able to attribute a percentage of snow cover to every pixel using its albedo value. Thus, a potential methodology of how the MODIS reflectance product can be used to improve the MODIS snow cover product is suggested in the following. First, reflectance values of all pixels in an area of interest could be plotted to create a histogram showing the frequency of occurrence of different values of surface albedo. Then, results from experimental studies such as Winther et al. (1999) could be used to restrict the range over which the albedo of snow can be expected to vary. Winther et al. (1999) defined the maximum snow albedo as a_{max} (in this example a_{max} for MODIS Band 2 is equal to 0.87; Table 1) and the minimum snow albedo as a_{min} (here, a_{min} equals 0.55; Table 1). Bare ground albedo was defined as a_0 (here, a_0 equals 0.10; Table 1).

Then, for a given date, MODIS-derived albedo values can vary over the full range between 0 and 1. Also, the maximum albedo value occurring in a particular image might exceed a_{max}, for example due to bi-directional surface reflectance (Winther & Hall 1999). Hence, Winther et al. (unpubl. ms.) suggest that if the albedo in a pixel exceeds a_{max}, i.e. $a > a_{max}$, then the albedo is assigned to a_{max}, giving a snow coverage of 100%. Next, pixels with values between a_{max} and a_0 are assigned to a snow coverage (in %) distributed linearly between a_{max} (100%) and a_0 (0%). Pixels with values equal to or less than a_0 are defined as bare ground (0% snow coverage).

This basic concept of calculating partially snow-covered areas using MODIS reflectance products can be useful to improve the existing MODIS snow cover product. In future, we will refine our algorithms in Svalbard, aiming at establishing a snow coverage monitoring product covering Svalbard.

Conclusions and recommendations

Snow-related research in Svalbard has intensified considerably over the last 10-15 years. New methods that have been applied, such as modelling of snow distribution as well as radar and satellite technology, have improved the means of data col-

Table 1. Empirical data on snow albedo based on Winther et al. (1999). a_{max} and a_{min} represent the maximum and minimum snow albedo measured in wavelengths corresponding to MODIS Bands 2 and 4, respectively. a_0 represents the albedo of bare ground (Winther et al. unpubl. ms.).

	a_{max}	a_{min}	a_0
MODIS Band 4	0.97	0.67	0.04
MODIS Band 2	0.87	0.55	0.10

lection, analysis and interpretation. For example, observations of snowmelt rates from various researchers have shown melt rates between 10 and 20 mm/day, short-wave radiation being the most important variable affecting melt rates. Extreme values of close to 70 mm/day have been measured. Further, snowmelt modelling studies indicate clearly how important—and difficult—the correct estimation of snow albedo is during spring melt, when large variations in albedo occur. To improve this, as well as to ground-truth satellite-derived data, detailed studies of spectral reflectance of melting snow have been undertaken. These show a rapid decrease of snow albedo after onset of melt, typically dropping from 0.80 to less than 0.50. Reduction occurs first most prominently in the infrared region due to snow metamorphosis processes. Later, the visible albedo reduces more strongly as a consequence of surface contamination. Recent studies of regional snow distribution have shown 50 % higher snow accumulation rates on the east coast of Spitsbergen compared to the west coast. Accumulation rates are about twice as high in the south compared to the north. Finally, there is an increasing number of biological investigations that include snow pack characteristics to explain factors such as reindeer population growth, reproduction of barnacle geese, vegetation growth and distribution of snow algae.

In spite of a number of successful project studies related to snow research in Svalbard, we identify some shortcomings and make suggestions for future directions. First, there are few examples of true integration among different projects. We recommend much stronger integration and collaboration between various snow-related projects in the future. Further, it is an obvious challenge for snow scientists to collaborate with biologists that work with issues where snow pack properties have an important effect on the biological system (Sturm et al. 2001).

A second striking weakness of today's snow research in Svalbard is the lack of comparative studies with other Arctic regions. We recommend the establishment of a major circumpolar snow hydrology project, for example, within the framework of the UNESCO International Hydrological Programme's scientific network *Northern Research Basins.* A potential task could be to make comparative studies of snow distribution and snowmelt to address the important issue of freshwater runoff into the Arctic Ocean.

Third, we wish to address the importance of applying field observations to validate model results and thereafter to refine and implement these models. Today, many local observational studies of high quality have been performed. However, there is a need for implementing models, such as a snow distribution model, and applying them to larger areas of Svalbard. Thus, we recommend establishment of a regional atmospheric model of Svalbard (including parameters relevant for snow research). Such a model will give scientists, and the environmental authorities, an advanced tool for regional studies of (complex) changes in the physical environment, for example, as a consequence of climate change.

Finally, we recognize that satellite remote sensing is a highly relevant method for monitoring remote areas such as Svalbard. Some early results using the MODIS sensor have shown that this satellite is capable of monitoring surface albedo (important for energy balance studies) and snow coverage on a regular basis for the whole of the archipelago. We recommend that a satellite-based operational snow monitoring system for Svalbard be established.

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