A theoretical approach to glacier equilibrium-line altitudes using meteorological data and glacier mass-balance records from southern Norway

Øyvind Lie,1,3,* Svein Olaf Dahl2,3 and Atle Nesje1,3

(1Department of Earth Science, University of Bergen, Allégt. 41, N-5007 Bergen, Norway; 2Department of Geography, University of Bergen, Breiviksveien 40, N-5045 Bergen, Norway; 3Bjerknes Centre for Climate Research, Allégt. 55, N-5007 Bergen, Norway)

Abstract: Based on a close exponential relationship between mean ablation-season temperature and winter precipitation at the equilibrium-line altitude (ELA) of 10 Norwegian glaciers, three equations are derived. The first equation enables calculation of the minimum altitude of areas climatically suited for glacier formation, and is termed the altitude of instantaneous glacierization (AIG). Equation (2) is derived based on the ‘principle of terrain adaptation’, enabling quantification of the glacial buildup sensitivity (GBS) in presently non-glaciated areas. The theoretical climatic temperature-precipitation ELA (C-TP-ELA) in presently non-glaciated areas is calculated in equation three by combining GBS with terrain altitude. Correlation between AIG and net balance measurements (bn) yielded correlation coefficients of r = –0.80 to r = –0.84. Calculated AIGs correspond well with observed ELAs on Ålafotbreen, Nigardsbreen and Gråsubreen, while it deviates from the observed ELA on Storbreen; the latter is probably due to lee accumulation of wind-blown snow on this cirque glacier. Based on this approach, regional representative climatic ELAs can be calculated for non-glaciated areas with instrumental records of ablation-season temperature and winter precipitation.

Key words: Glaciers, equilibrium-line altitude, ELA, mass-balance measurements, climate, Norway.

Introduction

The equilibrium-line altitude (ELA) on a glacier is a theoretical line which defines the altitude where annual accumulation equals the ablation (i.e., net balance, bn, is zero). Thus, the ELA is regarded as the most useful parameter to quantify the influence of climatic variability on glaciers, and it is widely used to infer present and past climatic conditions (e.g., Andrews, 1975; Porter, 1975; 1977). The ELA is generally dependent on the accumulation of snow during the winter season (winter balance, bw) and ablation during the summer season (summer balance, bs). Processes related to ablation on glaciers include evaporation, melting of snow and ice, radiation and heat exchange with the air. Accumulation is generally influenced by the regional distribution of precipitation as snow and local redistribution of snow by wind (e.g., Sissons and Sutherland, 1976; Sutherland, 1984; Robertson, 1989; Dahl and Nesje, 1992; Dahl et al., 1997). In addition, surface topography, glacier hypsometry and aspect may have a local influence on the ELA (e.g., Liestol, 1967; Porter, 1975; 1977; Leonard, 1984; Kuhn et al., 1985; Dahl and Nesje, 1992; Nesje, 1992). However, the main parameters controlling the ELA are the regional ablation-season temperature and winter precipitation as snow.

Due to the pronounced effect of wind-blown snow on ELAs, Dahl and Nesje (1992) introduced the terms temperature-precipitation equilibrium-line altitude (TP-ELA) and temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) to distinguish between glacier ELAs reflecting the general winter precipitation and ablation-season temperature in a region (such as ice caps) and glacier ELAs that are influenced by either snow deflation or accumulation (such as on cirque glaciers) (Figure 1). The TP-ELA can thus be regarded as synonymous with the lowest altitude of ‘instantaneous glacierization’ on a plateau as defined by Ives et al. (1975), Dahl and Nesje (1992) and Dahl et al. (1997).

To investigate the potential lowering of the regional ELA necessary to induce glacierization in presently non-glaciated areas in southern Norway, Dahl et al. (1997) formulated a theoretical approach to calculate present ELAs based on observed winter precipitation and ablation-season temperature. In this paper, we
present equations which enable theoretical calculations of ‘altitude of instantaneous glacieriation’ (AIG), ‘glacial buildup sensitivity’ (GBS) and ‘climatic temperature-precipitation equilibrium-line altitudes’ (TPW-ELA) based on records of mean ablation-season temperature and winter precipitation from meteorological stations combined with established adiabatic lapse rates, precipitation-elevation gradients and topography. The validity of the equations are tested against mass-balance records from four modern glaciers (Ålfbreen, Nigardsbreen, Storbreen and Gråsubreen) existing in maritime to continental climate regimes in southern Norway. A list of symbols and abbreviations used in the calculations is given in Table 1.

### The relation between winter precipitation and ablation-season temperature at the ELA

Several investigations have calculated and used relationships between the ablation-season temperature and winter precipitation, or similar terms, on glaciers in steady-state (e.g., Liestøl, 1967; Porter, 1977; Liestøl in Sissons, 1979; Sutherland, 1984; Leonard, 1984; 1989; Ballantyne, 1989; Ohmura et al., 1992), and most show non-linear relationships. Based on 10 modern Norwegian glaciers located in maritime to continental climate regimes in southern Norway, a close exponential relationship between mean ablation-season temperature (1 May to 30 September) and winter precipitation (1 October to 30 April) at the ELA has been demonstrated (Liestøl in Sissons, 1979; Sutherland, 1984). This relationship is expressed by the regression equation:

$$ A = 0.915 e^{0.339T} \quad (r^2 = 0.989, P < 0.0001) \quad (1) $$

where $A$ is winter precipitation (metres water equivalent) and $T$ is ablation-season temperature ($°C$) (Ballantyne, 1989; Figure 2). Solving for $T$ the equation is:

$$ T = \ln \left( \frac{A}{0.915} \right) \div 0.339 \quad (2) $$

### Table 1 List of symbols and abbreviations

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>SI/units/ ( P )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( p_0 )</td>
<td>Mean winter precipitation (as water equivalent) at a known altitude (i.e., a climate station).</td>
<td>m</td>
</tr>
<tr>
<td>( t_0 )</td>
<td>Mean ablation-season temperature at a known altitude (i.e., a climate station).</td>
<td>°C</td>
</tr>
<tr>
<td>( A )</td>
<td>Mean winter precipitation (as water equivalent) at the equilibrium-line altitude.</td>
<td>m</td>
</tr>
<tr>
<td>( T )</td>
<td>Mean ablation-season temperature at the equilibrium-line altitude.</td>
<td>°C</td>
</tr>
<tr>
<td>( \Delta t )</td>
<td>Adiabatic lapse rate.</td>
<td>0.65°C ( 100 \text{m}^{-1} )</td>
</tr>
<tr>
<td>( \Delta P )</td>
<td>Precipitation-elevation gradient.</td>
<td>8% ( 100 \text{m}^{-1} )</td>
</tr>
<tr>
<td>( h )</td>
<td>Altitude above climate station.</td>
<td>100 m</td>
</tr>
<tr>
<td>( H )</td>
<td>Altitude of the topography.</td>
<td>m</td>
</tr>
<tr>
<td>( H_{\text{mean}} )</td>
<td>Altitude of climate station.</td>
<td>m</td>
</tr>
<tr>
<td>TPW-ELA</td>
<td>Climatic temperature-precipitation equilibrium-line altitude.</td>
<td>m</td>
</tr>
<tr>
<td>AIG</td>
<td>Altitude of instantaneous glacieriation.</td>
<td>m</td>
</tr>
<tr>
<td>GBS</td>
<td>Glacial buildup sensitivity (above topography).</td>
<td>m</td>
</tr>
<tr>
<td>( T_i )</td>
<td>Theoretical mean ablation-season temperature needed for instantaneous glacieriation at ( H ).</td>
<td>°C</td>
</tr>
<tr>
<td>( T_e )</td>
<td>Calculated mean ablation-season temperature at ( H ).</td>
<td>°C</td>
</tr>
<tr>
<td>( P_w )</td>
<td>Calculated mean winter precipitation (water equivalent) based on equation (2).</td>
<td>m</td>
</tr>
</tbody>
</table>

Equations (1) and (2) imply that, if either the winter precipitation or the ablation-season temperature at the ELA is known, the other variable can be calculated. This relationship can thus be used to calculate variations in winter precipitation based on knowledge of ELA variations and independent records of ablation-season temperature as demonstrated by Dahl and Nesje (1996).
Altitudinal relationships between ablation-season temperature and winter precipitation

In southern Norway, regional relationships between altitude and both ablation-season temperature and winter precipitation have been demonstrated. For the mean ablation-season temperature, linear adiabatic lapse rates of $0.6–0.7^\circ C$ per 100m have been found (Green and Harding, 1980). Thus, where $t_0$ is the temperature at a known altitude (such as a meteorological station), $\Delta t$ is the adiabatic lapse rate and $h$ is the height above the climate station in hundred metres:

$$T = t_0 - (\Delta t \times h)$$

(3)

Haakensen (1989) found a precipitation-elevation gradient (altitudinal increase) of $8\%$ per 100m at Ålfotbreen in outer Nordfjord, whereas Dahl and Nesje (1992) estimated this gradient to be $9\%$ per 100m for inner Nordfjord. On glaciers in different climate regimes, Laumann and Reeh (1993) estimated precipitation-elevation gradients of $7–8\%$ per 100m. In mountain areas of central southern Norway, Sæ lthun (1973) inferred that summer precipitation increased $5\%$ per 100m, while the value on Hardangervidda is estimated to be $8\%$ per 100m (Skartveit, 1976). Based on data from climate stations, Forland (1979) found that the regional precipitation pattern in Norway depends mainly on distance from the coast and on topography. Owing primarily to local orographic conditions, it was not possible to establish a general precipitation-elevation gradient for Norway. However, the precipitation-elevation gradients presented above are given as exponential relationships, and can thus be described on the generalized form:

$$A = p_0 \times (1 + \Delta p)^h$$

(4)

where $p_0$ is the measured precipitation at any altitude (such as at a meteorological station), $\Delta p$ is the precipitation-elevation gradient and $h$ is the height above the meteorological station in hundred metres.

As both $T$ and $A$ in equations (1) and (2) are functions of altitude, where the temperature drops as a result of adiabatic cooling, and the precipitation increases with altitude, both can be plotted in a three-variable xy-diagram of temperature ($x$-axis), altitude ($y_1$-axis) and precipitation controlled by the precipitation-elevation gradient ($y_2$-axis) (Figure 3). In this example, the temperature at an altitude of 1100 m is $6.0^\circ C$ with a corresponding winter precipitation of 600 mm ($0.6$ m water equivalent). The temperature at higher elevation is plotted as line 1 based on an adiabatic lapse rate of $0.65^\circ C$ per 100m, and equation (2) is plotted as line 2 based on the temperature at 1100 m and the calculated rise in precipitation with altitude ($\Delta p = 8\%$ per 100m). As a consequence, the right-hand ordinate axis (winter precipitation) is not linear. Where lines 1 and 2 intersect in Figure 3, the exponential relationship between ablation-season temperature and winter precipitation at the glacier ELA (equation (1)) is met based on the observed temperature and precipitation and applied vertical gradients. The AIG is determined by the $y_1$-axis, defining the theoretical minimum altitude at which a glacier can form. If the terrain is higher than the AIG, equation (1) also indicates how the ELA fluctuates based on observed instrumental variations in ablation-season temperature and winter precipitation. The related ablation-season temperature ($x$-axis) and winter precipitation ($y_2$-axis) are also shown. This demonstrates that equation (1) can be solved for
The x-axis shows the corresponding mean ablation-season temperature. Line 1 shows the adiabatic cooling of the mean ablation-season temperature (\( t_0 = -0.80 \)). Line 2 shows the adiabatic cooling of the mean ablation-season temperature (\( t_0 = -0.80 \)).

Table 3: Periods, glaciers and climate stations used for calculation of correlation coefficients and mean AIG/TP-ELAs

<table>
<thead>
<tr>
<th>Period with climate data</th>
<th>Glacier</th>
<th>Temperature station (^a)</th>
<th>Precipitation station (^b)</th>
<th>Correlation coefficient (^c)</th>
<th>Mean observed ELA</th>
<th>Mean calculated AIG/TP-ELA</th>
</tr>
</thead>
<tbody>
<tr>
<td>1963–1996</td>
<td>Åfotbreen</td>
<td>Fjærbjord Skarestad</td>
<td>Åfoten II</td>
<td>-0.81</td>
<td>c. 1200</td>
<td>1200</td>
</tr>
<tr>
<td>1969–1996</td>
<td>Åfotbreen</td>
<td>Sandane</td>
<td>Åfoten II(^d)</td>
<td>-0.80</td>
<td>c. 1200</td>
<td>1200</td>
</tr>
<tr>
<td>1978–1997</td>
<td>Åfotbreen</td>
<td>Sandane</td>
<td>Grøndalen</td>
<td>-0.80</td>
<td>c. 1200</td>
<td>1200</td>
</tr>
<tr>
<td>1963–1996</td>
<td>Åfotbreen</td>
<td>Fjærbjord Skarestad</td>
<td>Fjærbjord Skarestad</td>
<td>-0.80</td>
<td>c. 1200</td>
<td>1200</td>
</tr>
<tr>
<td>1962–1996</td>
<td>Nigardsbreen</td>
<td>Fortun</td>
<td>Fjærbjord Skarestad</td>
<td>-0.79</td>
<td>c. 1560</td>
<td>1200</td>
</tr>
<tr>
<td>1966–1996</td>
<td>Nigardsbreen</td>
<td>Oppstyn</td>
<td>Briskdal</td>
<td>-0.76</td>
<td>c. 1510</td>
<td>1510</td>
</tr>
<tr>
<td>1965–1998(^e)</td>
<td>Nigardsbreen</td>
<td>Bjørkehaug</td>
<td>Bjørkehaug(^d)</td>
<td>-0.82(^f)</td>
<td>c. 1210</td>
<td>1570(^g)</td>
</tr>
<tr>
<td>1962–1996</td>
<td>Nigardsbreen</td>
<td>Fortun</td>
<td>Hafslo</td>
<td>-0.77</td>
<td>c. 1540</td>
<td>1540</td>
</tr>
</tbody>
</table>

\(^a\) Pearson’s correlation coefficient.
\(^b\) Bold text refers to the calculations used in Figure 5.
\(^c\) See Table 2 for more information.

### Figure 3
Three-variable xy-diagram, where the y1-axis shows altitude (m), and the y2-axis shows the calculated winter precipitation at the altitudes shown on the y1 axis based on a winter precipitation of 600 mm (0.6 m) water equivalent at an altitude of 1100 m and an \( \Delta p \) of 8% 100 m\(^{-1}\). The x-axis shows the corresponding mean ablation-season temperature. Line 1 shows the adiabatic cooling of the mean ablation-season temperature (\( t_0 = 6.0^\circ C \)) at an altitude of 1100 m and \( \Delta t = 0.65^\circ C 100 m^{-1} \)). Line 2 is plotted using equation (2) based on the predicted precipitation at 100 m intervals and the corresponding calculated mean ablation-season temperature. The figure shows that, for the used values of winter precipitation and mean ablation-season temperature at an altitude of 1100 m, the AIG is c. 1920 m.

**Equation 1**
\[
\text{T}_{\text{TP-ELA}} = h + \frac{\ln(0.915)}{0.339t} + \frac{\ln(p_0) - \ln(p)}{0.339 \Delta t} + \frac{0.339 \Delta p}{0.339 \Delta t} \tag{1}
\]

**Equation 2**
\[
\text{a} = \frac{\ln(0.915) + 0.339 \Delta p}{0.339} \tag{2}
\]

**Equation 3**
\[
\text{b} = \frac{\ln(p_0) - \ln(p)}{0.339 \Delta t} \tag{3}
\]

**Equation 4**
\[
\text{AIG} = H_{\text{station}} + (h \times 100) \tag{4}
\]

If the terrain is lower than the calculated AIG, the theoretical ELA lowering to induce glaciation at this altitude can be investigated. A slightly different approach must be taken, however, as the required ELA lowering depends on the ‘principle of terrain adaptation’ (Dahl, 1997). In Figure 3, including equations (4) and (7), the precipitation is allowed to increase unlimited with altitude, regardless to the elevation of the topography. Obviously, the precipitation-elevation gradient is irrelevant above the terrain, as it is the mean winter precipitation falling on the terrain surface that will define the accumulation and thus the ELA lowering necessary to induce glaciation. The ‘principle of terrain adaptation’ therefore states that the precipitation falling onto the terrain
surface is determining the potential for glacierization with the corresponding ablation-season temperature. Hence, in presently non-glaciated regions the equations must be solved with respect to the climatic conditions at the terrain surface, and topography must consequently be included in the equations. Based on equation (1) and for any $\Delta t$ and $\Delta t$, this can be calculated. A climate station is chosen with an altitude $H_{\text{station}}$, and the winter precipitation ($P_w$) at any altitude of the topography ($H$) can be calculated from:

$$P_w = p_0 \times (1 + \frac{H - H_{\text{station}}}{100})$$  \tag{8}

As $P_w$ is identical to $A$ at an altitude ($H$), equation (1) can be solved for the theoretical temperature needed for instantaneous glacierization ($T_i$) at $H$:

$$T_i = \frac{\ln (\frac{P_w}{0.915})}{0.339}$$  \tag{9}

The present ablation-season temperature ($T_o$) at $H$ based on the nearest climate station is:

$$T_o = t_0 - \frac{(H - H_{\text{station}})}{100}$$  \tag{10}

To determine the temperature lowering needed to induce glacierization at $H$, $T_i$ is subtracted from $T_o$. This result can be expressed as altitude, as temperature is dependent on altitude and latitude. By dividing 100 by $\Delta t$ and multiplying this with the above result, the theoretical difference between the theoretical ELA and $H$ is obtained. This is defined as the ‘glacial buildup sensitivity’ (GBS). Hence, the GBS gives the height above $H$ which fulfills the requirements for glacierization based on equation (1). The complete equation describing the GBS is given in equation (11).

$$\text{GBS} = \left[t_0 - \Delta t \times \frac{(H - H_{\text{station}})}{100}\right] - \ln \left(\frac{P_w}{0.915}\right)$$  \frac{H - H_{\text{station}}}{100}  \tag{11}

If a low value is calculated, the area is likely to have a glacier buildup, while a high value indicates that a large ELA-lowering (climatic deterioration) is necessary to induce glacierization. Consequently, the GBS at the calculated AIG will be 0 m. Values of GBS<0 are invalid from the principle of terrain adaptation as negative values describe conditions below the terrain surface. To calculate the GBS, the GBS must be added to the topographical altitude ($H$), as shown in equation (12):

$$\text{GTP-ELA} = H + \text{GBS} \ (\text{GBS} \neq 0)$$  \tag{12}

The GTP-ELA is also conditioned by the principle of terrain adaptation and cannot be calculated for areas where the GBS is <0, and in these circumstances the AIG, not taking terrain-elevation into account, must be used to describe the climatic ELA.

**Test of the equations against glacier mass-balance observations**

To test the validity of the equations, four glaciers with continuous mass-balance measurements in southern Norway have been investigated (Kjøllmoen, 1998), and the AIG has been calculated using data from adjacent meteorological stations (Tables 2 and 3). The AIG represents fluctuations of the ELA on a topographical surface for each balance year, and this is correlated with mass-balance data. A close relationship between measured annual ELAs and mass-balance fluctuations exists on glaciers (e.g., Liestøl, 1967; Andrews, 1975; Porter, 1975), and allows the AIG to be tested against observed annual net balance ($b_n$) variations on glaciers. The test of the AIG equation, however, is also representative for the GTP-ELA and GBS, as they are all derived from the same basic relationship (equation 1). In the AIG calculations, an adiabatic lapse rate ($\Delta t$) of 0.65°C 100 m $^{-1}$ and a vertical precipitation gradient ($\Delta p$) of 8% 100 m $^{-1}$ have been used.

**Åfotbreen**

Åfotbreen (Figure 4) is a small ice cap (c. 17 km$^2$) that is representative for glaciers in a maritime climate regime in southern Norway. Based on climate data from adjacent meteorological stations (Table 3), the annual AIG was tested against the annual net balance of Åfotbreen between 1963 and 1995 (Kjøllmoen, 1998). The AIG calculations reproduced the net mass balance with a correlation coefficient of $r = 0.80$ (Figure 5A). The calculated AIG of 1260 m is also close to the observed mean ELA of c. 1200 m (Østrem et al., 1988). As plateau glaciers are less influenced by wind-drift of snow, they represent the TP-ELA in the scheme of Dahl and Nesje (1992), hence being representative for the regional winter precipitation and measured precipitation (Figure 1).

**Nigardsbreen**

Nigardsbreen (Figure 4) is a southeasterly outlet glacier (48.2 km$^2$) from the Jostedalsbreen ice cap (487 km$^2$) located in a semi-continental climate regime. The observed mean ELA is c. 1560 m (Østrem et al., 1988). Based on adjacent meteorological stations (Table 3), the correlation coefficient between the calculated AIG and the measured net mass balance from 1962 to 1995 (Kjøllmoen, 1998) was calculated to $r = 0.82$ (Figure 5B). The calculated mean AIG of 1570 m is close to the observed mean ELA of 1560 m as expected on a plateau glacier like Jostedalsbreen.

**Storbreen**

To test the equations for a semi-continental/glacial glacier, the AIG of Storbreen, a cirque glacier of c. 5.16 km$^2$ in central Jotunheimen (Figure 4), was calculated using adjacent meteorological stations (Table 3). The AIG calculation reproduced the measured net mass balance ($b_n$) variations of the glacier for the period 1960–1996 (Kjøllmoen, 1998), with a correlation coefficient of $r = 0.84$ (Figure 5C). However, the observed mean ELA of c. 1700 m of Storbreen (Østrem et al., 1988) contrasts with the calculated mean AIG of 2080 m for the same period. Storbreen is a cirque glacier situated on the leeward side from the prevailing wind direction in southern Norway, and may receive substantial amounts of wind-blown snow (Figure 1). In addition, the local meteorological station (Bøverdalen) is located in one of the driest valleys of southern Norway, thus probably underestimating the absolute amount of winter precipitation at Storbreen. The estimated deviation of c. 380 m between the observed mean TPW-ELA and the calculated mean AIG probably reflects a combination of these two factors. Based on equation (1) and the mean ablation-season temperature from the local climate station Bøverdal, Storbreen requires c. 1.9 m water-equivalent winter precipitation at the observed mean TPW-ELA of 1710 m to be in steady-state. However, the Brætå meteorological station (Table 3) indicates a water-equivalent of only c. 0.8 m water equivalent at the altitude of the TPW-ELA ($\Delta p = 8\%$ 100 m $^{-1}$). Hence, Storbreen receives an additional amount of winter precipitation of c. 1.1 m water equivalent. Consequently, the additional accumulation at this glacier has a factor of c. 2.4 of the mean regional precipitation, probably due to wind-blown snow, but regional precipitation gradients may also contribute to this figure.
Gråsubreen
Gråsubreen (Figure 4) is a continental, high-altitude polythermal cirque glacier located in eastern Jotunheimen. Using local meteorological stations (Table 3), the calculated mean annual AIG yields a correlation coefficient of $r = -0.81$ when compared with measured annual mass-balance fluctuations for the period 1962–1996 (Kjøllmoen, 1998; Figure 5D). The mean AIG of 2190 m for the same period was somewhat higher than the observed mean ELA of c. 2130 m on Gråsubreen (Østrem et al., 1988). This may, however, be explained by a small additional amount of wind-blown snow on Gråsubreen; but horizontal climate gradients between the glacier and the climate stations used in the calculation may as well explain this deviation.

Discussion
All correlations between calculated AIGs and measured net balance ($b_n$) are significant, demonstrating that the equations are valid for both maritime and continental climate regimes (Table 3). In addition, the correlations are rather high, considering that the factors controlling the ELA are reduced to only two as neither factors like wind-blown snow nor aspect are taken into account. Variations in vertical and regional precipitation-elevation and temperature gradients are probably the most important uncertainties. However, in remote areas of Norway with few meteorological stations this problem can only be partly solved by using statistical methods.

The calculations demonstrate how the ELA rises from west to east with increasing continentality. Hence, the increase in the calculated mean AIG of 950 m between Alflotbreen and Gråsubreen (Table 3) can be explained climatically. Reduced to sea level, the relative difference between mean ablation-season temperatures at the climate stations of Fortun and Løken i Vollbu is 1.5°C, while the relative difference between Fjærland and Løken i Vollbu is 1.4°C (Aune, 1993). This indicates that the temperature alone can explain only 215–230 m of the eastward rise in the AIG ($\Delta t = 0.65°C \ 100m^{-2}$), while the additional 70% rise (c. 720–735 m) is a result of lower winter precipitation. This demonstrates the importance of taking both summer temperature and winter precipitation into account when inferring climate from reconstructed glacier fluctuations (Dahl and Nesje, 1996).

The derivation of the expressions introduced here has been solved with respect to the ELA-related definitions. However, due to their ‘open-ended’ nature, the equations can easily be solved for any of the factors $t_o$ or $p_o$, and may be used to calculate the

Figure 4 Location map showing the position of the glaciers with observed mass-balance records used in this study. The mean observed regional ELAs are shown (after Liestøl, 1967). Modified from Østrem et al. (1988).
distribution of, for example, winter precipitation when data is available for regional ELAs and temperature. To further investigate the regional pattern of glacierization potential and ELAs, the expressions will be implemented in a geographical information system to calculate the introduced terms as ‘surfaces’ covering both presently glaciated and non-glaciated regions of southern Norway (Lie et al. (second paper), this issue).

Conclusions

(1) Based on a close exponential relationship between mean ablation-season temperature and winter precipitation at the ELA of Norwegian glaciers, three equations linked to climate-glacier interaction are derived: the altitude of instantaneous glacierization (AIG), climatic temperature-precipitation equilibrium-line altitude ($C_{TP}$-ELA). The $C_{TP}$-ELA and GBS are directly determined by mean winter precipitation, ablation-season temperature and surrounding topography, while AIG simply represents the minimum regional altitude of terrain suitable for glacierization. None of the presented equations include correction for wind-blown snow, and thus reflect the TP-ELA as defined by Dahl and Nesje (1992) (Figure 1). The presented equations enable climate regimes to be expressed as ELA-related values, even in presently non-glaciated areas. This makes it possible to investigate the sensitivity of glacierization in any area and the response of ELAs to climatic change (winter precipitation, ablation-season temperature) and topography.

(2) The close exponential relationship between glaciers and climate used in this paper (equation (1)) is based on Norwegian glaciers. However, any regional relationship between temperature and precipitation at the ELA can be derived using this approach, and any regional precipitation-elevation and temperature gradients can be applied.

(3) Four modern glaciers localized in maritime to continental climate regimes in southern Norway (Ålfofthem, Nigardsbreen, Storbreen and Gråsubreen) with long mass-balance records are used to test the equations. All correlation coefficients are between –0.80 and –0.84 without adjusting the climate data for any regional climatic gradients, aspect and leeward accumulation of snow by prevailing wind directions, or modifying the equations and vertical climatic gradients to local conditions.
(4) At present, the difference in calculated AIG between the maritime Ålftobreen in the west and the continental Gråsurbreen in the east is c. 950 m, which is close to the range of observed ELAs across the region (Liestøl, 1967). Less than 30% (215–230 m) can be explained by differences in observed mean ablation-season temperature, whereas lower winter precipitation accounts for the additional 70% (720–735 m).

(5) To investigate where glaciers first will form, the presented approach can be used to calculate different TP-ELA scenarios in areas with a known topography and representative meteorological data. Based on southern Norway this will be demonstrated by the use of GIS (geographical information systems) in the following paper (Lie et al., this issue).

Acknowledgements

Partly unpublished climate data were made available by Meteorologisk Institutt. This paper is a contribution from the Bjerknes Centre for Climate Research. We express our gratitude to Mel Reasoner and an anonymous referee, whose comments helped to improve and clarify the manuscript.

Appendix: Example of calculation of AIG, GBS and \(^{13}TP\)-ELA

Using the same parameters as in Figure 3, calculations of AIG, GBS and \(^{13}TP\)-ELA can be done as follows: at an altitude of 1100 m \(t_o\) is 6.0°C with a corresponding \(p_o\) of 600 mm (0.6 m water equivalent), while \(\Delta t\) is 0.65°C 100 m\(^{-1}\) and \(S_p\) is 8% 100 m\(^{-1}\). By substitution in equation (6), the AIG is:

\[
h = \frac{\ln(0.915) + (0.339 \times 6.0) - \ln(0.600)}{\ln(1 + 0.08) + (0.339 \times 0.65)} = 8.26
\]

This gives an AIG of:

\[\text{AIG} = 1100 + (8.26 \times 100) = 1926\text{ m}\]

The calculation of GBS for 1600 m is shown below:

\[
\text{GBS}_{1600} = \left[ 6.0 \times \left( \frac{1600 - 1100}{100} \right) \right] - \left( \frac{0.600 \times (1 + 0.08)}{0.915} \times 0.339 \right) \left( \frac{1600 - 1100}{100} \right)
\]

\[\text{GBS}_{1600} = 440\text{ m} - \text{the ELA is 440 m above the terrain at 1600 m a.s.l. giving a }^{13}TP\text{-ELA}_{1600}\]

\[1600\text{ m + 440 m} = 2040\text{ m}\]

References


Sutherland, D.G. 1984: Modern glacier characteristics as a basis for inferring former climates with particular reference to the Loch Lomond stadial. Quaternary Science Reviews 3, 291–309.