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

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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. Mass-balance records from four modern glaciers (Ålfotbreen, Nigardsbreen, Storbreen and Gråsubreen) situated in maritime to continental climate regimes in southern Norway are used to test these equations. 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 Ålfotbreen, Nigardsbreen and Gråsubreen, while it deviates from the observed ELA on Storbreen; the latter is probably due to leeward 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, b_n, 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, b_w) and ablation during the summer season (summer balance, b_s). 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., Liestøl, 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

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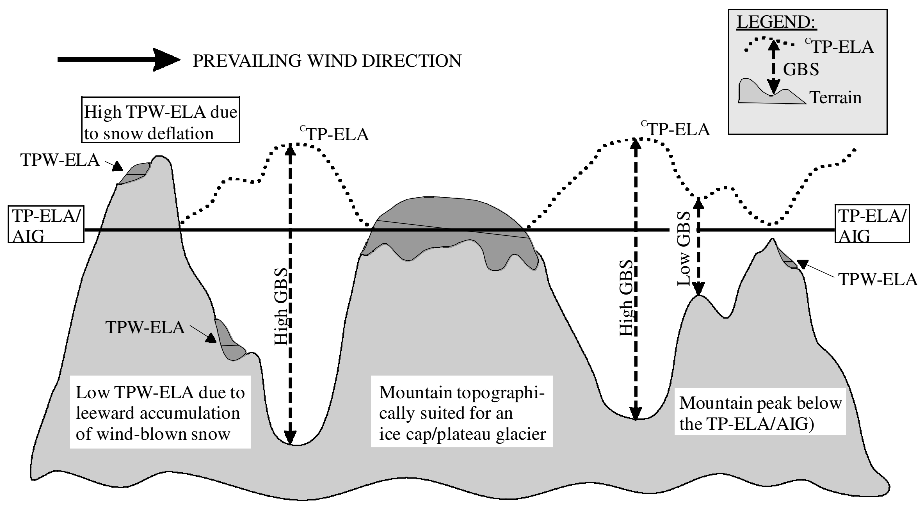

Figure 1 Schematic examples showing the difference between the temperature-precipitation equilibrium-line altitude (TP-ELA) on plateau glaciers and temperature-precipitation-wind equilibrium-line altitude (TPW-ELA) on cirque glaciers. The altitude of instantaneous glacierization (AIG) is the climatically calculated value of the observed TP-ELA. The glacial buildup sensitivity (GBS) defines the distance between the terrain and the altitude where conditions are favourable for glacier formation when taking the 'principle of terrain adaptation' into account. The climatic temperature-precipitation equilibrium-line altitude (CTP-ELA) defines the altitude where conditions are favourable for glacier formation. Note how the GBS and CTP-ELA are conditioned by the terrain due to the 'principle of terrain adaptation'. (Modified from: Dahl and Nesje, 1992; Dahl *et al*., 1997.)

present equations which enable theoretical calculations of 'altitude of instantaneous glacierization' (AIG), 'glacial buildup sensitivity' (GBS) and 'climatic temperature-precipitation equilibrium-line altitudes' (^{C}TP-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 (Ålfotbreen, 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.

Table 1 List of symbols and abbreviations

Symbol	Description	SI/units/ gradients
p_o	Mean winter precipitation (as water equivalent) at a known altitude (i.e., a climate station).	m
t_o	Mean ablation-season temperature at a known altitude (i.e., a climate station).	°C
A	Mean winter precipitation (as water equivalent) at the equilibrium-line altitude.	m
T	Mean ablation-season temperature at the equilibrium-line altitude.	°C
Δt	Adiabatic lapse rate.	0.65°C 100 m^{-1}
Δp	Precipitation-elevation gradient.	8% 100 m^{-1}
h	Altitude above climate station.	100 m
H	Altitude of the topography.	m
$H_{station}$	Altitude of climate station.	m
^{C}TP-ELA	Climatic temperature-precipitation equilibrium-line altitude.	m
AIG	Altitude of instantaneous glacierization.	m
GBS	Glacial buildup sensitivity (above topography).	m
T_i	Theoretical mean ablation-season temperature needed for instantaneous glacierization at H.	°C
T_s	Calculated mean ablation-season temperature at H.	°C
P_w	Calculated mean winter precipitation (water equivalent) based on equation (8).	m

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:

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

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:

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

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).

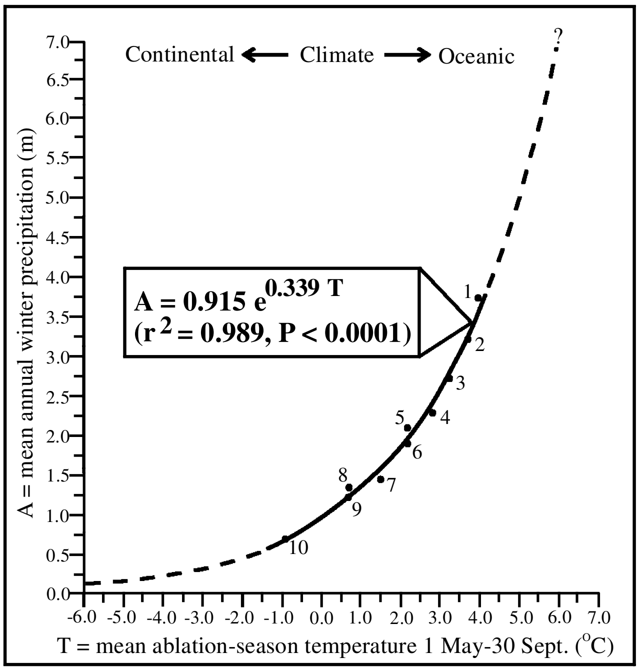

Figure 2 The non-linear (exponential) relationship between mean ablation-season temperature (1 May to 30 September) and winter precipitation (1 October to 30 April) at ELAs of 10 Norwegian glaciers: (1) Ålfotbreen; (2) Engabreen; (3) Folgefonna; (4) Nigardsbreen; (5) Tunsbergdalsbreen; (6) Hardangerjøkulen; (7) Storbreen; (8) Austre Memurubreen; (9) Hellstugubreen; (10) Gråsubreen. After Dahl *et al.* (1997).

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°C 100 m^{-1} have been found (Green and Harding, 1980). Thus, where t_0 is the temperature at a known altitude (such as a meteorological station), Δt is the adiabatic lapse rate and h is the height above the climate station in hundred metres:

$$T = t_0 - (\Delta t \times h) \quad (3)$$

Haakensen (1989) found a precipitation-elevation gradient (altitudinal increase) of 8% 100 m^{-1} at Ålfotbreen in outer Nordfjord, whereas Dahl and Nesje (1992) estimated this gradient to be 9% 100 m^{-1} for inner Nordfjord. On glaciers in different climate regimes, Laumann and Reeh (1993) estimated precipitation-elevation gradients of 7–8% 100 m^{-1}. In mountain areas of central southern Norway, Sælthun (1973) inferred that summer precipitation increased 5% 100 m^{-1}, while the value on Hardangervidda is estimated to 8% 100 m^{-1} (Skartveit, 1976). Based on data from climate stations, Førland (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:

$$A = p_0 \times (1 + \Delta p)^h \quad (4)$$

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

Table 2 Meteorological stations and glaciers with mass-balance observations used in the calculation of correlation coefficients and mean AIG/^{C}TP-ELA

Glacier/meteorological station	Station number	Years of observation used in this study	Type (P/T)[1]
Ålfotbreen		1963–1995	
Nigardsbreen		1962–1995	
Storbreen		1960–1996	
Gråsubreen		1962–1996	
Ålfoten II	5794	1957–1996	P
Grøndalen	5778	1978–1997	P
Sandane	5807	1962–1997	T
Briksdal	5848	1960–1996	P
Fjærland-Skarestad	5584	1957–1996	TP
Oppstryn	5870	1963–1997	T
Hafslo	5555	1957–1996	P
Bjørkehaug i Jostedal	5543	1965–1998	TP
Fortun	5516	1957–1996	TP
Bråtå[3]	1572	1965–1997	TP[2]
Sognefjell	5529	1979–1989	P
Fanaråken	5523	1961–1978	P
Elveseter	1536	1960–1968	TP
Bøverdal	1543	1960–1997	P
Øvre Tessa[3]	1469	1970–1997	TP
Vågåmo, Vågå[3]	1460	1962–1976	TP
Preststulen, Vågå[3]	1455	1961–1976	P
Beito	2356	1957–1997	P
Løken i Vollbu[3]	2350	1962–1986	T

[1]P = precipitation; T = temperature.
[2]P = 1957–1997.
[3]Vågå 'combined' temperature and precipitation based on linear regression between the meteorological stations used in combination with mass balance measurements on Gråsubreen 1962–1996.
References for meterological data: Aune (1993); Førland (1993); partly unpublished data from Meteorologisk Institutt, Oslo.
References for glacier mass-balance data: Kjøllmoen (1998).

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 (y1-axis) and precipitation controlled by the precipitation-elevation gradient (Δp) (y2-axis) (Figure 3). In this example, the temperature at an altitude of 1100 m is 6.0°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°C 100 m^{-1}, and equation (2) is plotted as line 2 based on the temperature at 1100 m and the calculated rise in precipitation with altitude ($\Delta p = 8\%$ 100 m^{-1}). 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 on the y1-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 (y2-axis) are also shown. This demonstrates that equation (1) can be solved for

Table 3 Periods, glaciers and climate stations used for calculation of correlation coefficients and mean AIGs/^{C}TP-ELAs

Period with climate data	Glacier	Temperature station[3]	Precipitation station[3]	Correlation coefficient[1]	Mean observed ELA	Mean calculated AIG/^{C}TP-ELA
1963–1996	Ålfotbreen	Fjærland-Skarestad	Ålfoten II	–0.81	*c.* 1200	1200
1963–1996[2]	**Ålfotbreen**[2]	**Sandane**[2]	**Ålfoten II**[2]	**–0.80**[2]		**1260**[2]
1978–1997	Ålfotbreen	Sandane	Grøndalen	–0.80		1130
1963–1996	Ålfotbreen	Fjærland-Skarestad	Fjærland-Skarestad	–0.80		1310
1962–1996	Nigardsbreen	Fortun	Fjærland-Skarestad	–0.79	*c.* 1560	1260
1963–1996	Nigardsbreen	Oppstryn	Briksdal	–0.76		1510
1965–1998[2]	**Nigardsbreen**[2]	**Bjørkehaug**[2]	**Bjørkehaug**[2]	**–0.82**[2]		**1570**[2]
1962–1996	Nigardsbreen	Fortun	Hafslo	–0.77		1540
1960–1996	Storbreen	Bråtå/Elveseter	Bøverdal	–0.78	*c.* 1710	2190
1961–1968	Storbreen	Bråtå/Elveseter	Elveseter	–0.91		2150
1961–1978	Storbreen	Bråtå/Elveseter	Fanaråken	–0.82		2230
1960–1996[2]	**Storbreen**[2]	**Bråtå/elveseter**[2]	**Bråtå**[2]	**–0.84**[2]		**2080**[2]
1979–1989	Storbreen	Bråtå/Elveseter	Sognefjell	–0.95		2170
1962–1996[2]	**Gråsubreen**[2]	**Løken i Vollbu**[2]	**Preststt./Øvre Tessa**[2]	**–0.81**[2]	*c.* 2130	**2190**[2]
1962–1995	Gråsubreen	Løken i Vollbu	Beito	–0.75		2080
1970–1995	Gråsubreen	Løken i Vollbu	Øvre Tessa	–0.80		2190
1966–1995	Gråsubreen	Bråtå/Øvre Tessa	Preststt./Øvre Tessa	–0.80		2200

[1]Pearson's correlation coefficient.
[2]Bold text refers to the calculations used in Figure 5.
[3]See Table 2 for more information.

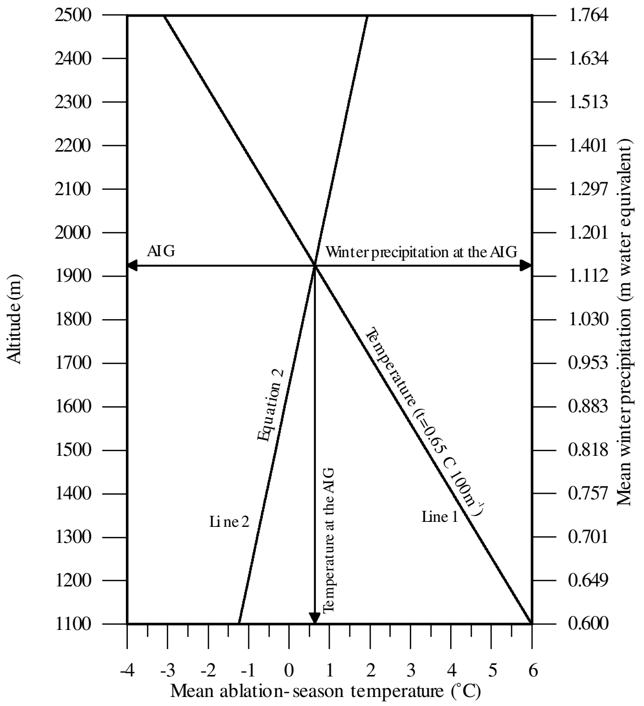

Figure 3 Three-variable xy-diagram, where the y1-axis shows altitude (m), and the y2-axis shows the calculated winter precipitation at the altitudes shown on the y1 axis based on a winter precipitation of 600 mm (0.6 m) water equivalent at an altitude of 1100 m and a Δp of 8% 100 m^{-1} (note that this axis is not linear). 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°C at an altitude of 1100 m and Δt = 0.65°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.

altitude (h), where Δt is the adiabatic lapse rate, Δp is the precipitation-elevation gradient, p_0 is the known winter precipitation, t_0 is the known ablation-season temperature at a known altitude ($H_{station}$) and h is the height of the AIG above the climate station in hundred metres. By substitution of equations (3) and (4) into equation (1), the expression becomes:

$$p_0 (1 + \Delta p)^h = 0.915e^{0.339(t_0 - \Delta t \times h)} \quad (5)$$

Solving equation (5) with respect to the only unknown parameter h gives:

$$h = \frac{\ln(0.915) + 0.339t_0 - \ln(p_0)}{\ln(1 + \Delta p) + 0.339 \times \Delta t} \quad (6)$$

Equation (6) shows the height (h) in hundred metres above the climate station where conditions are favourable for glacierization given measured t_0, p_0, Δt and Δp. The altitude of the meteorological station ($H_{station}$) added to h × 100 expresses the minimum terrain altitude needed to induce glacierization (AIG) in metres. As this calculation does not take into account the effect of additional wind deflation and/or accumulation, the AIG should give the same altitude as any measured TP-ELA (Figure 1), as defined by Dahl and Nesje (1992). The AIG is:

$$AIG = H_{station} + (h \times 100) \quad (7)$$

If the terrain is lower than the calculated AIG, the theoretical ELA lowering to induce glacierization 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 *et al.*, 1997). In Figure 3, including equations (6) 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 glacierization. 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 Δp and Δt, this can be calculated. A climate station is chosen with an altitude $H_{station}$ and the winter precipitation (P_w) at any altitude of the topography (H) can be calculated from:

$$P_w = p_0 \times (1 + \Delta p)^{\frac{H - H_{station}}{100}} \qquad (8)$$

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:

$$T_i = \frac{\ln\left(\frac{P_w}{0.915}\right)}{0.339} \qquad (9)$$

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

$$T_s = t_0 - \Delta t\left(\frac{H - H_{station}}{100}\right) \qquad (10)$$

To determine the temperature lowering needed to induce glacierization at H, T_i is subtracted from T_s. This result can be expressed as altitude, as temperature is dependent on Δt and altitude. By dividing 100 by Δt 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 fulfils the requirements for glacierization based on equation (1). The complete equation describing the GBS is given in equation (11).

$$GBS = \left[t_0 - \Delta t \times \left(\frac{H - H_{station}}{100}\right)\right] - \left[\frac{\ln\left[\frac{p_0 \times (1 + \Delta p)^{\frac{H - H_{station}}{100}}}{0.915}\right]}{0.339}\right] \times \frac{100}{\Delta t}\ (GBS \neq <0) \qquad (11)$$

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 ^{C}TP-ELA, the GBS must be added to the topographical altitude (H), as shown in equation (12):

$$^{C}\text{TP-ELA} = H + GBS\ (GBS \neq <0) \qquad (12)$$

The ^{C}TP-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 ^{C}TP-ELA and GBS, as they are all derived from the same basic relationship (equation 1). In the AIG calculations, an adiabatic lapse rate (Δt) of 0.65°C 100 m^{-1} and a vertical precipitation gradient (Δp) of 8% 100 m^{-1} have been used.

Ålfotbreen

Ålfotbreen (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 Ålfotbreen 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/continental 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 Bøverdalen, 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 Bråtå meteorological station (Table 3) indicates a winter precipitation of only *c*. 0.8 m water equivalent at the altitude of the TPW-ELA (Δp = 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.

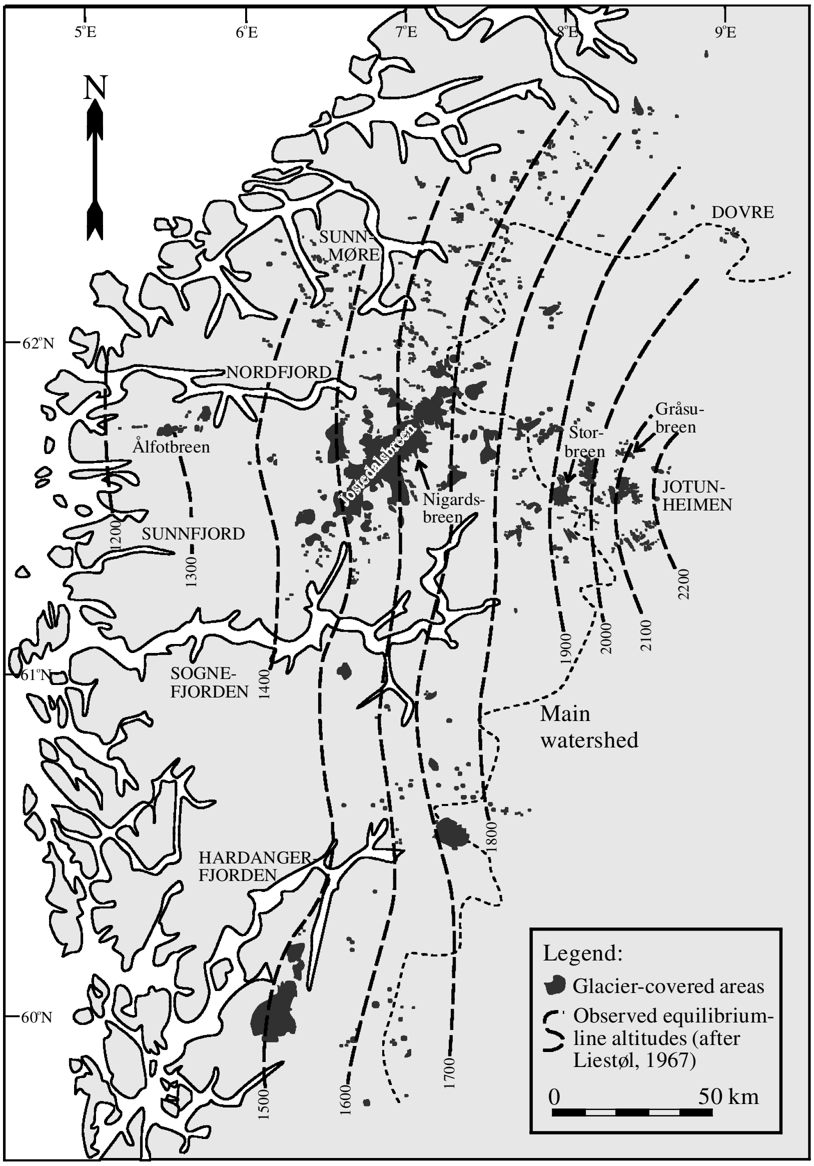

Figure 4 Location map showing the position of the glaciers with observed mass-balance records used in this study. The mean observed regional ELAs are shown (after Liestøl, 1967). Modified from Østrem *et al.* (1988).

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 on Gråsubreen m (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 Ålfotbreen 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 ($\Delta t = 0.65°C\ 100\,m^{-1}$), 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_0 or p_0, and may be used to calculate the

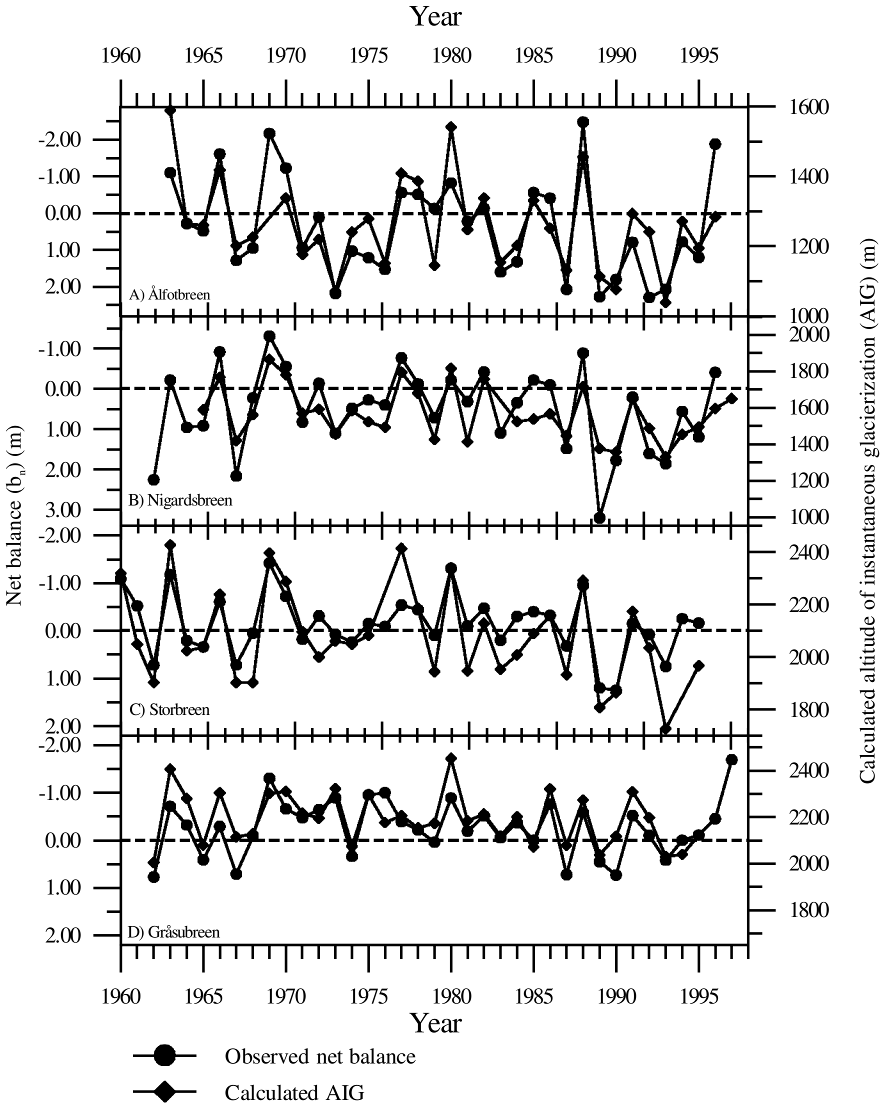

Figure 5 Plotted net-balance measurements (filled circles) (Kjøllmoen, 1998) versus calculated AIGs (filled squares) of: (A) Ålfotbreen (based on ablation-season temperature (t_O) from Sandane (station number 5807) and winter precipitation (p_O) from Ålfoten II (station number 5794); (B) Nigardsbreen based on t_O and p_O from Bjørkehaug (station number 5543); (C) Storbreen based on t_O from Bråtå (climate station 1572) and Elveseter (station number 1535) and p_O of Bråtå (station number 1572); (D) Gråsubreen based on t_O from Løken i Vollbu (station number 2350) and p_O from Vågå, Preststulen and Øvre Tessa ('Vågå combined', station numbers 1455, 1460 and 1469). Correlation coefficients show values of –0.80 at Ålfotbreen, –0.82 at Nigardsbreen, –0.84 at Storbreen and –0.81 at Gråsubreen. See Table 3 for correlation coefficients for other meteorological stations.

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), glacial buildup sensitivity (GBS) and 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 the 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 (Ålfotbreen, 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 Ålfotbreen in the west and the continental Gråsubreen 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 ^{C}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).

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Appendix: Example of calculation of AIG, GBS and ^{C}TP-ELA

Using the same parameters as in Figure 3, calculations of AIG, GBS and ^{C}TP-ELA can be done as follows: at an altitude of 1100 m t_0 is 6.0°C with a corresponding p_0 of 600 mm (0.6 m water equivalent), while Δt is 0.65°C 100 m^{-1} and Δp 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:

$$\mathrm{AIG} = 1100 + (8.26 \times 100) = 1926 \text{ m}$$

The calculation of GBS for 1600 m is shown below:

$$\mathrm{GBS}_{1600} = \left[6.0 - \left(0.65 \times \left\{\frac{1600 - 1100}{100}\right\}\right)\right] - \left[\frac{\ln\left(\dfrac{0.600 \times (1 + 0.08)^{\frac{1600-1100}{100}}}{0.915}\right)}{0.339}\right] \times \frac{100}{0.65} = 440$$

GBS_{1600} = 440 m – the ELA is 440 m above the terrain at 1600 m a.s.l. giving a ^{C}TP-ELA$_{1600}$ of:

$$1600 \text{ m} + 440 \text{ m} = 2040 \text{ m}.$$

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