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Abstract:

Improvements in our ability to model runoff from glaciers remain an important scientific goal. This paper describes a new temperature-radiation-index glacier melt model specifically enhanced for use in High-Arctic environments, utilising high temporal and spatial resolution datasets while retaining relatively modest data requirements. The model employs several physically constrained parameters and was tuned using a lidar-derived surface elevation model of Midtre Lovénbreen, meteorological data from sites spanning ~70% of the glacier's area-altitude distribution, and periodic ablation surveys during the 2005 melt season. The model explained 80% of the variance in observed ablation across the glacier, an improvement of ~40% on a simplified energy balance model (EBM), yet equivalent to the performance of a full EBM employed at the same location. Model performance was assessed further by comparing potential and measured runoff from the catchment, and through application to an earlier (2004) melt season. The additive model form and consideration of *a priori* parameters for the Arctic locality were shown to be beneficial, with a planimetry correction eliminating systematic errors in potential runoff. Further parameterisations defining modelled incident radiation failed to yield significant improvements to model output. Our results suggest that such enhanced melt models may perform well for singular melt seasons, yet are highly sensitive to the choice of lapse rates and their transferability to different locations and seasons may be limited. While modelling ablation requires detailed consideration of the transition between snow- and ice-melt, our study suggests that description of the ratio between radiative and turbulent heat fluxes may provide a useful step towards dynamic parameterisation of melt factors in temperature-index models.

1 Introduction

Small glaciers and ice caps account for ~ 14% of the terrestrial ice in the Arctic, and research has highlighted broadly persistent, negative mass balances of these ice masses in recent decades (Arendt *et al.*, 2002; Dowdeswell *et al.*, 1997; Koerner, 2005; Nuth *et al.*, 2010). This characteristic in high latitude glacier mass balance may be tentatively linked to apparent warming trends leading to lengthened ablation seasons (Christensen *et al.*, 2007; Kattsov *et al.*, 2005). Within this context, in glaciological terms, regional estimates of glacier mass balance sensitivity have been presented with values of between –0.5m a^{-1} K^{-1} and –0.63 m a^{-1} K^{-1} (Braithwaite and Raper, 2007; de Woul and Hock, 2005) for ablation season ice losses, suggestive of increasing seasonal meltwater fluxes from high-latitude, glacierised catchments in response to projected climate forcing. Nonetheless, significant issues remain in terms the spatial and temporal resolution and transferability of melt models for indicating both contemporary and future runoff volumes. There is still a need to trial models to discern which elements, if any, within more sophisticated schemes are beneficial to model output precision (Hock, 2005).

The amount of surface melting of snow or glacier ice during the ablation season is controlled by the energy fluxes, which are specific to local climatic and surface conditions. Thorough reviews of the surface energy balance can be found in Greuell and Genthon (2004) and Hock (2005) and numerous energy balance models (EBMs) have been thoroughly investigated for valley glaciers in temperate (e.g. Anslow *et al.*, 2008; Arnold *et al.*, 1996; Escher-Vetter, 2000; Klok and Oerlemans, 2002) and Arctic or sub-Arctic settings (e.g. Arnold *et al.*, 2006; Hock and Holmgren, 2005; Hock and Noetzli, 1997; MacDougall and Flowers, 2011; Rye *et al.*, 2010). Critically, EBMs all explicitly stress how variations in glacier surface conditions influence ablation and subsequent runoff patterns. There are considerable uncertainties involved in using EBMs due to their dependence on spatially and temporally variable factors such as cloud cover, albedo, wind speed and surface characteristics, which change with location, time and climate. The performance and accuracy of EBMs predictably decreases as the variability of the surface energy-balance increases through an ablation season (Kane *et al.*, 1997). The result of this uncertainty is that physical verification of the parameterisations used is not readily achieved, making it impossible to transfer EBMs from glacier to glacier without re-parameterisation which requires extensive, high resolution meteorological data and detailed information characterising of conditions at the melting surface, as highlighted in MacDougall & Flowers (2011).

As a result of the potentially problematic use of EBMs, alternative empirically based ‘index’ methods have been employed because snow and/or ice ablation is moderately well correlated to air temperature, a relationship long recognised (e.g. Finsterwalder and Schunk, 1887; Martinec, 1960). According to Ohmura (2001), the physical justification behind the temperature-index approach is that up to 75% of the energy available for ice melt may be derived from incoming longwave radiation and sensible heat. Consequently, temperature-index melt models (TIMs), with varied degrees of enhancement (e.g.Hock, 1999; Pellicciotti *et al.*, 2005), have become a widely used approach in glacial research programs (e.g. Braithwaite, 1995; de Woul *et al.*, 2006; Ebnet *et al.*, 2005; Hanna *et al.*, 2008; Klok *et al.*, 2001; Marshall and Sharp, 2009; Schneeberger *et al.*, 2003). Although Hock (2003) argued that TIMs yield lower accuracy over higher temporal resolution, these simplified models may hold advantages both in terms of parameterisation and potential transferability (e.g. Carenzo *et al.*, 2009). Therefore, using high resolution data sets, we extend, test and explore the use of a novel, yet distributed, temperature-index melt model at Midtre Lovénbreen, a valley glacier in Svalbard, to simulate seasonal glacier ablation and runoff. Specifically, model enhancements suitable for the study site are applied and their benefits for modelling ice ablation are assessed with comparison between models of varied complexity, over differing spatial resolution and at-a-point energy balance calculations.

2 Field site and data collection

2.1 Field site

Midtre Lovénbreen (hereafter, ML), located in the north-west of Spitsbergen (78°50'N 12°E; Figure 1), is one of the most studied glaciers in the High-Arctic. Local mean annual temperatures and precipitation at sea-level reach –6.2°C and ~370 mm respectively (Hanssen-Bauer *et al.*, 1990). The glacier occupies 49.5% of a 10.8 km^2 north-facing catchment and extends from approximately 50 to 650 m above sea level (masl) with a maximum thickness of 180 m (Björnsson *et al.*, 1996). The glacier's accumulation area ratio is ~ 30% with a long-term average equilibrium line altitude (ELA) of ~ 400 masl (Björnsson *et al.*, 1996). At lower elevations thin (< 0.2 m) winter superimposed ice forms on the glacier, but this is rapidly ablated during summer months, although annual accumulation of superimposed ice is found at elevations > 405 masl (König *et al.*, 2002). Measurements (Hagen *et al.*, 2003) and modelling (Rye *et al.*, 2010) have shown the glacier has exhibited predominantly negative mass balance over the last five decades and recent geodetic analyses have indicated the negative mass balance trend may be accelerating, with contemporary thinning rates of > 0.5 m water equivalent (w.e.) a^{-1} (Barrand *et al.*, 2010; Kohler *et al.*, 2007). Characteristically for the area, seasonal snow-cover below the ELA is removed within ~14 days from the onset of melt conditions (Bruland *et al.*, 2001). Further, during summer months (JJA) cloud cover is pronounced, with as few as one clear sky day per month and ~50 days with cloud below 300 masl (Hanssen-Bauer *et al.*, 1990). Summer meltwater is discharged from the glacier's catchment via two principal stream routes (MLW and MLE; Figure 1) over a hydrological season typically ~70-80 days in length.

2.2 Data collection

Ablation season field campaigns were conducted in 2004 and 2005, between July 9th and September 4th (DOY190-247) and June 24th and August 24th (DOY175-237), respectively. Hourly meteorological data were collected from automatic weather stations (AWS) positioned along ML's centre line (Figure 1, Table 1).

The Norsk Polarintitutt (NP) centre-line stakes (Figure 1) were used to monitor changes in the ice surface elevation (with precision of ±0.005 m) relative to a reference point (see Müller and Keeler, 1969). Concurrently, glacier surface albedo at representative locations close to each mass balance stake was recorded using a Middleton hemispheric pyranometer approximately 1 m to 1.2 m above the ice surface, and data errors were assumed to be negligible (see van der Hage, 1992).

Runoff data during the 2004 and 2005 observation periods were collected using standard hydrological methods, and uncertainties in discharge (Q) data were dominantly related to the forecasting procedure and were < 19% and < 16% for MLE and MLW, respectively (for full details, refer to Irvine-Fynn *et al.*, 2011a). Occasional missing data due to instrumental failure, typically less than 3 hours, were estimated statistically using other flow records or Q data from the adjacent Bayelva catchment, which is typically highly correlated ($r > 0.85$) with discharge from ML (Hodson *et al.*, 2005).

2.3 Glacier surface model

High-resolution digital elevation models of the ML glacier surface in 2003 and 2005 were derived from airborne laser scanning data (for details see Barrand *et al.* 2010; 2009). Data were collected with an Optech ALTM3033 scanning system and post-processed with inertial navigation system and onboard and ground-based differential GPS positioning data to yield raw point clouds with mean along- and across-track point spacing of 1.38 and 1.33 m, respectively, and average point density of 1.15 per m^2 (Barrand, 2008). DEMs of the glacier surface were constructed using an adapted Delauney triangulation gridding algorithm. Vertical elevation accuracy was +/-0.14 m based on comparison with ground-based differential GPS check data over the glacier surface (Barrand *et al*., 2009). To provide further model assessment opportunities, the DEM was resampled to horizontal resolutions of 2, 5, 10 and 20 m using bilinear interpolation.

3 Development of an Arctic temperature-index melt model (ArcTIM)

This section sequentially details the basis and parameterisations utilised in development of the enhanced temperature-index melt model described and assessed here.

3.1 The model's 'ABC' basis

The application of TIMs is advocated on the grounds of their computational simplicity, data availability and generally satisfactory performance (Hock, 2003). In the simplest form, for a given time-step a TIM defines melt (*M*) as a function of temperature (T):

$$M = a\mathrm{T} + b \quad \left(\text{for } \mathrm{T} > \mathrm{T}_{\mathrm{crit}}\right) \qquad [1]$$

in which parameter *a* is the degree of proportionality, or the melt factor which differs for snow or ice surfaces, given as m °C^{-1} per time interval while the threshold temperature for melt (T_{crit}) is, in most practical situations, taken to be the melting point of snow and ice (0°C) below which melt is zero. The variable T may be given as near-surface (~ 1.5 to 2.0 m) air temperature (T_a) or as the difference between T_a and T_{crit} (Martinec, 1960). As the value of *a* implicitly represents all the variables of the energy-balance, it is necessarily dynamic, and as Braithwaite (1995) demonstrated, as T increases to > 10°C, values for *a* converge, indicating a non-linearity between air temperatures and melt. Different TIM variants have utilised alternative approaches to parameterising *b*, with zero (e.g. Martinec, 1960) or non-zero (Braithwaite, 1995) values accounting for melt occurring when T < T_{crit}. Such formulations ensure melt rates are somewhat spatially uniform, neglecting the influence of topographic variations other than elevation.

In addressing these shortcomings, temperature-index models have, incorporated spatially and/or temporally variable components: parameter *b* has been given as a function of shortwave radiation (e.g. Kane *et al.*, 1997; Martinec, 1989; Pellicciotti *et al.*, 2005) or n*et* *al*l-wave radiation (e.g. Brubaker *et al.*, 1996; Kustas *et al.*, 1994) or, alternatively, *a* has been given as a function of incident radiation (e.g. Cazorzi and Dalla Fontana, 1996) with further model enhancement by using *b* parameterised as a function of T_a itself (e.g. Hock, 1999). Conversely, Shea *et al*. (2004) defined T using regression between radiation and temperature to identify a time-series of distributed residual temperatures (the difference between observed and modelled temperatures) and used incident radiation to define parameter *b*. The improvement in TIM performance using these varied approaches has differed markedly between locations (cf. references above).

The model presented by Hock (1999), given its applicability shown on Storglaciaren, has been adopted widely (e.g. de Woul *et al.*, 2006; Ebnet *et al.*, 2005; Flowers and Clarke, 2000; Huss *et al.*, 2008; Schneeberger *et al.*, 2003; Schuler *et al.*, 2007). However, limitations are apparent, particularly in the multiplication of T and the radiation component, which is counterintuitive in consideration of the independence of these two variables within the energy-balance (Greuell and Genthon, 2004; Pellicciotti *et al.*, 2005) and can result in overestimates of melt particularly during diurnal peak temperatures (Konya *et al.*, 2004). Instead, Williams and Tarboton (1999) elegantly demonstrated through simplification of terms in the energy-balance that a TIM can be better represented by:

$$M = az + b\mathrm{I} + c \qquad [2]$$

in which *z* is elevation and I is potential incident radiation. In the first instance, *z* provides a representation of the spatial distribution of T_a in response to an unknown lapse rate which is incorporated into parameters *a* and *c*; secondly, parameters *a* and *c* also include turbulent fluxes, again assumed to be linear with respect to T_a; thirdly, *b* is proportional to (1 – α); and finally, the parameter *c*, in addition to the constants arising from a linear lapse rate, relates to the turbulent fluxes and longwave emissivity. Therefore, here ArcTIM followed the form suggested by Konya *et al*., (2004):

$$\dot{M} = \begin{Bmatrix} a\mathrm{T} + b(1-\alpha)\mathrm{I} + c & : & \mathrm{T} > \mathrm{T}_{\mathrm{crit}} \\ b(1-\alpha)\mathrm{I} + c & : & \mathrm{T} \leq \mathrm{T}_{\mathrm{crit}} \end{Bmatrix} \qquad [3]$$

Unlike all previously published TIMs, here, melt ($\dot{M}$) is assumed to be normal to the ice surface, since potential incident radiation is defined as being perpendicular to a given surface slope. The use of constant *c* in the model allows a degree of correction for hitherto undefined boundary layer conditions (e.g. turbulent or subsurface energy exchanges etc.). Individual model parameters were defined using the dataset from 2005, as detailed in the following sections.

3.2 Temperature (T)

To apply a distributed TIM, it is necessary to extrapolate values for air temperature throughout the model domain. Numerous researchers have, for simplicity, assumed constant, linear lapse rates ranging from –0.004 °C m^{-1} to –0.0076 °C m^{-1} (cf. Bøggild *et al.*, 1994; Hock, 1999; Jóhannesson *et al.*, 1995; Konya *et al.*, 2004; Shea *et al.*, 2004). However, glaciers influence their local climate: boundary-layer processes cause high spatial and temporal temperature variability and phenomena including temperature inversions (e.g. Arendt and Sharp, 1999). Consequently, constant lapse rates are inappropriate, typically overestimating temperature gradients (Marshall *et al.*, 2007). Rather than optimise lapse rates within the model (e.g. Jóhannesson *et al.*, 1995), ArcTIM used a non-linear lapse rate derived from field observations, such that air temperature at elevation *z* (T_z) was given with respect to the AWS2 record:

$$T_z = T_{AWS2} - 1.8\ln(z) + 9.6 \qquad [4]$$

as derived from the four AWS sites deployed in 2005 (Figure 2a). The approach used by Shea *et al.*, (2005) was discounted because, although it was possible to remove the co-linearity between T_a and incident radiation for a single site, it was found that the relationship between these variables was not strong ($r < 0.33$) and varied across the glacier as well as in time.

3.3 Threshold melt temperature (T_{crit})

The typically used assumption that T_{crit} = 0°C (e.g. Hock, 1999) is not necessarily physically tenable in light of the actual energy balance, where energy is required to raise snow or ice temperature to melting point or when radiative fluxes lead to a temperature or energy maximum in the subsurface (e.g. Koh and Jordan, 1995; Liston *et al.*, 1999; Pellicciotti *et al.*, 2009). To define a value for T_{crit} suited for application at ML, the local threshold temperature of +1.62°C defined as equal probability of snow or rain was explored as a starting point (Førland and Hanssen-Bauer, 2003). To ascertain whether this choice of threshold temperature was valid, time-series of air temperatures at each stake in 2005 were developed using Eq. 4, and the respective cumulative above-threshold air temperatures for each ablation survey period were calculated for threshold temperatures incremented from 0°C to 3°C. These series were regressed against the corresponding ablation data (Figure 2b) illustrating a plateau in the coefficient of determination, and suggesting use of T_{crit} = +1.62°C was appropriate.

3.4 Potential incident radiation (I)

The incident radiation (I) at any point within the catchment was modelled using algorithms fully detailed by Iqbal (1983) but following Kreith and Kreider (1978):

$$I = I_0 E_0 \tau_b \cos\theta \qquad [5]$$

for which I_0 is the solar constant (~ 1368 W m^{-2}), E_0 is the orbital eccentricity correction factor calculated from the local day angle, θ is the angle of incidence on a tilted surface and transmissivity (τ_b) is given as:

$$\tau_b = 0.56\left(e^{-0.65 m_a} + e^{-0.0095 m_a}\right) \qquad [6]$$

for which, using the time varying solar altitude (ϖ) derived from the day angle and local latitude, the air mass ratio (m_a) is given as:

$$m_a = \sqrt{\left(1229 + (614 \sin \varpi)^2\right)} - 614 \sin \varpi \qquad [7]$$

The use of this variant of better-known algorithms is because clear-sky atmospheric transmissivity varies over both space and time and a secant exponent estimating air mass ratio using air pressure in order to adjust for local altitude is strictly only valid when solar zenith angles (θ_z) are less than 70° (Kreith and Kreider, 1978). When the zenith angle exceeds 70°, as is common at high-latitudes and is the case for ~90% of ML's ablation season, this atmospheric approximation underestimates solar energy by failing to account for atmospheric path length, refraction and curvature, yet this appears to be ignored in many high-latitude studies (e.g. Ebnet *et al.*, 2005; Schuler *et al.*, 2007). In testing the suitability of the derived I, modelled values for the level AWS sensors were compared with logged radiation during clear-sky conditions: results showed a systematic bias of < +11%, over thirteen full-day records r > 0.76 (AWS2: n = 312 and AWS4: n = 168). Using field notes to refine AWS2 data to periods with < 4 oktas of cloud coverage, the bias was reduced, with r = 0.84 (n = 124); and for periods with < 2 oktas, r = 0.92 (n = 41).

Several TIMs adjust I to account for topographic shading (e.g. Hock, 1999). However, the omission of shading has been shown to increase predicted radiation receipt by only ~6% (Arnold *et al.*, 2006) and the diurnally-averaged shadowed area of ML remains < 25% for ~65% of the ablation season, with greatest shadowing between 20:00 and 02:00 when radiative energy is at its lowest. Moreover, as noted in Section 2.1, the climatology of ML results in the prevalence of cloud covered conditions (Hanssen-Bauer *et al.*, 1990) and dominance of diffuse shortwave radiation, evidenced by a mean shortwave radiative energy flux of 141 Wm^{-2} at AWS2 during the summer of 2005 compared to a mean potentialreceipt of 273 Wm^{-2}. Relationships utilised to adjust incident radiation for cloud cover are typically unsatisfactory (e.g. Arnold *et al.*, 1996; Hock, 1999; Pellicciotti *et al.*, 2005). Furthermore, the basis of a TIM is that T provides a proxy for the dominant melt energy, and consequently, topographic shading and cloud cover was omitted from this model.

3.5 Albedo (α)

The TIM variant presented by Pellicciotti *et al*. (2005) indicated that inclusion of an albedo term can improve melt calculations. Therefore, keeping to a more physical basis an empirical albedo parameterisation was employed, the albedo at AWS2 (α_0) for the start day of the model period (t_0) was backcast using Ordinary Least Squares (OSL) regression against time in decimal days. An elevation dependency was then applied, such that albedo (α_{tz}) at decimal time *t* and for elevation *z* is given by:

$$\alpha_{tz} = \alpha_0 + (0.0006z - 0.1133) + \mathrm{d}\alpha(t - t_0) \qquad [8]$$

where

$$\alpha_0 = 1.3014 - 0.0041(t_0) \qquad [9]$$

and the term dα is the mean linear decay of α over time, as calculated from all the survey sites (Figure 3a). Linear regressions describing albedo variations at all the survey sites, illustrated time rather than cumulative temperature best explained the temporal trend observed (data not presented here).The linear relationship between *z* and α (Figure 3b) appeared stronger than that identified at Haut Glacier d'Arolla by Brock *et al.*, (2000) and was likely a function of supraglacial dust (cryoconite) distribution (cf. Hodson *et al.*, 2007) and assumed to be constant over the season. Overall, the albedo parameterisation yielded a significant correlation (r = 0.48, n = 64) with a mean overestimation of ~0.06.

3.6 Scale sensitivity

A common flaw with melt models is the tendency to assume that pixels or grid cells are planimetric (see Hopkinson (2010) for a review). This is significant for two reasons: first, calculations of energy fluxes are given normal to the surface slope. The modelled magnitude of radiation loading is, therefore, greatly influenced by changes in surface slope at differing DEM resolutions (Arnold and Rees, 2009; Chasmer and Hopkinson, 2001). Second, a systematic bias is introduced because melt occurs normal to the surface slope and the planimetric assumption underestimates the slope-variant surface area subject to melt and thus melt volumes (Hopkinson *et al.*, 2008). These two biases are more significant for models based on DEMs of higher resolution (Chasmer and Hopkinson, 2001; Hopkinson *et al.*, 2008). Thus, a scale correction factor was introduced, following Hopkinson *et al*., (2008), such that for any given grid cell:

$$M_z = \frac{0.9\dot{M}_z\left(L^2\right)}{\cos\beta}$$

[12]

where M_z is total melt in m^3 water equivalent (w.e.), L is the length dimension of each square grid cell, and $\dot{M}_z$ represents the modelled formulation of melt normal to the surface slope (β) based on the distributed values of temperature and radiation. An ice density of 0.9 g cm^{-3} for the near-surface of ML was assumed.

4 Comparative models

4.1 Melt models

Here, it is perhaps useful to summarise and describe the comparative model runs used to assess the formulation of ArcTIM described above. Melt simulations using the model form described in Equation 1 (hereafter, Model Tc), and in Equation 3 but omitting albedo (hereafter, Model TI) were run; for further comparison, melt was estimated with a modified version of Hock's (1999) model where:

$$\dot{M}_z = aT_z + bT_z(1-\alpha)I_m + c \quad (\text{for } T > T_{crit}) \qquad [13]$$

and in following Martinec (1960) and Kane *et al*. (1997):

$$\dot{M}_z = a(T_z - T_{melt}) + b(1-\alpha)I_m + c \qquad [14]$$

In this latter variant, T_{melt} was evaluated by manually adjusting its value, and regressing observed ablation against time-series of temperature ($T_z - T_{melt}$) which when iteratively adjusting T_{melt} gave an optimised value of +0.85°C. In all these model variants, we adhered to the same formulations and values for T_a, T_{crit} and I as described above, and melt below threshold was defined, as before, only for where $b(1-\alpha)I + c > 0$. The scale correction factor (Section 3.6) was used in all instances to determine melt volume and specific melt.

Many published temperature-index melt models distinguish between snow and ice covered surfaces which accounts for critical spatial differences in meltwater genesis (e.g. Hock, 1999; Jóhannesson *et al.*, 1995). However, unlike these examples, here, models did not account for a difference between snow and ice surface on the glacier. The reason for this was threefold: first, the TIM presented by Schneeberger *et al*. (2003) evidenced only subtle difference between melt factors (*a*) for ice and snow; second, snowline retreat on the shallow slopes characterising the majority of ML's ice area is typically rapid; and third, because the melt model which includes an elevation-defined albedo was to be tuned to ablation measurements, it was assumed optimisation would account for any differences potentially related to the snowline, especially at higher elevations. We recognise that this absence of differentiation between snow and ice may potentially represent an important source of model uncertainty. Ice topography was not evolved in time: with an estimated ceiling ablation of ~1.5 m (Hodson *et al.*, 2005), the maximum influence of an evolving surface on temperature and precipitation would be of the order of +0.03 °C and –0.3%, respectively, and therefore, can be considered negligible given the magnitude of uncertainties associated with the source data.

As an independent comparison, an adjusted EBM based on Brock and Arnold (2000) was run to model theoretical ablation along the glacier centre line. This model included the algorithms and parameterisations used within ArcTIM: the non-linear T_a lapse rate (Eq. 3), high-latitude atmospheric transmissivity (Eq. 6), and temporally evolving glacier albedo (Eq. 8). Received incident radiation and its variation with elevation was interpolated linearly from records at AWS2 and AWS4. In the absence of appropriate data, the calculations for turbulent energy fluxes were necessarily simplified, using a constant aerodynamic roughness length of 0.00253 (after Hodson *et al.*, 2005) and estimating absolute vapour pressure for each time-step, assuming that relative humidity recorded at AWS2 was spatially uniform, and that T_a varied non-linearly with elevation (Eq. 3) and held an empirical relationship with saturation vapour pressure over ice (after Tetens, 1930). Wind speed was distributed over elevation using linear gradients derived from the hourly data recorded at AWS2 and AWS4.

4.2 Melt model parameter calibration

Many glacier melt models are tuned to the time series of meltwater discharge, particularly when the period of interest is annual runoff cycles. In calibrating models to discharge using arbitrary routing algorithms, model output will consequently mask subtle water release or storage processes which, particularly for glaciers in Svalbard, may be significant (e.g. Hodgkins, 2001; Hodson *et al.*, 2005). Here, with the purpose of modelling potential runoff volumes without artificially forcing unverifiable flowpath delays, melt model parameters *a*, *b*, and *c* were calibrated using the time-normalised, observed centre-line stake ablation records for 2005. Models were based on the 5 m resolution DEM, which is much higher resolution than appears in most similar modelling studies. Parameter optimisation demands that the 'degree of fit' between measured data and modelled data is maximised (or minimised) for which there are a number of measures (e.g. Nash and Sutcliffe, 1970; Willmott, 1981). Experimental investigations into optimising the varied TIM parameter sets (such that *a* and *b* > 0.0) indicated there was indeed a strong interdependence between T and I, and the use of OLS multivariate regression was found to be the most effective optimisation process. Cross-validation analyses, repeating the OLS multivariate regression but withholding various datasets showed no bias in the determination of the parameters: the absolute variation in multivariate $R^2 < 2.4\%$ was not significant at $p = 0.05$.

4.3 Potential runoff and precipitation

Once optimised and executed over the glacier ice area (the model domain), summation of M_z provided an estimate of the potential runoff volume available at each time-step. Although precipitation was not considered as a significant process of ablation, liquid summer precipitation adds to the total water equivalent available as runoff. In the absence of *in situ* rain-gauge records, data collected by NP in Ny Ålesund (8 m asl) were used: hourly records (P_{NA}) were reconstructed using the twice-daily precipitation record and field notes of rainfall events. The local orography of Brøggerhalvøya creates precipitation lapse rates of 20% per 100 m for elevations < 300 m asl (Førland *et al.*, 1997) above which 10% per 100 m is more appropriate (Killingtveit *et al.*, 1994). Accordingly precipitation at elevation *z* (P_z) was described as:

$$P_z = \begin{cases} P_{NA}(1.00 + 0.002(z-8)) & : \quad z \leq 300 \\ P_{NA}(1.586 + 0.001(z-300)) & : \quad z > 300 \end{cases} \qquad [11]$$

Solid precipitation, implicit at T_a < 1.62°C, was assumed to refreeze and was accounted for by adding the equivalent ice depth to the measured ablation survey data. Liquid precipitation was assumed to fall on the planimetric grid cells, and therefore, when present, simply multiplied by the grid cell area and added to the scale corrected meltwater volume to yield a total available for runoff. The precipitation occurring on the surrounding mountain slopes within the glacier basin was excluded from the model because of the characteristic talus, lateral moraines and scree near the glacier margin for which the associated effects on water retention, percolation or routing were unknown.

5 Application of models on Midtre Lovénbreen

Here, we detail and discuss the melt model output(s) with reference to the primary data sets of observed ablation and meltwater runoff.

5.1 Ice ablation

The performance of the model parameterisations for 2005, comparing observed, slope-corrected and modelled ablation are detailed in Table 2. A slight improvement to modelling ablation was made by enhancing a simple TIM to ArcTIM, with the R^2 of ~80% matching similar model performance reported elsewhere for temperature-radiation index models (e.g. Brubaker *et al.*, 1996; de Woul *et al.*, 2006; Kane *et al.*, 1997). Furthermore, for 2005, ArcTIM equalled the performance of a more complex full EBM on ML presented by Arnold *et al.*, (2006). Interestingly, the inclusion of albedo within ArcTIM subtlety reduced its efficacy, while both the Hock and T_{melt} model variants appeared to show slightly improved agreement between modelled and observed ablation, with standard errors of ~0.17 mm hr^{-1}. The TIM models suggested a melt rate of 0.3 mm hr^{-1} °C^{-1} at ML which exceeds the 0.05 mm hr^{-1} °C^{-1} reported by Pellicciotti *et al.*, (2005) for a similar TIM formulation applied to an Alpine glacier.

To examine the sensitivity of individual parameter values, three numerical efficiency criteria were explored further: Figure 4 illustrates model sensitivity to changing parameter values, using the Nash-Sutcliffe (η^2: (Nash and Sutcliffe, 1970) and Willmott (d^2: (Willmott, 1981) indices of agreement and the root mean squared error (RMSE). Examination of these plots demonstrates (i) ArcTIM is highly sensitive to the values of both *b* and *c*, but less so for the value of *a* and (ii) the η^2 criteria is a more sensitive measure of model fit. To eliminate the potential for equifinality within the model, we varied all three parameters at random (Figure 4d) finding a single optimum combination with a minimum RMSE of ~0.2 mm hr^{-1}.

Crucially, comparison between the TIM variants and the adjusted EBM showed the latter to perform relatively poorly: the correlation between observed and modelled ablation was 0.6, with an RMSE of 0.4 mm hr^{-1}. The EBM typically under-predicted melt by 0.16 mm hr^{-1}, contrasting to the persistent over-prediction of melt using the EBM reported in Arnold *et al. (*2006). We ascribed this to the disparity in parameterisation of the EBM when compared to those presented by Arnold *et al*. (2006) and Rye *et al*. (2010), although this was not investigated further here.

5.2 *Comparison of potential and actual runoff*

With the parameters for ArcTIM successfully and robustly calibrated for 2005, and, ignoring the spatial uncertainty inherent in the input data, the model was applied to the ML catchment. The spatial distribution of ablation shown in Figure 5a is given as the modelled seasonal total, derived using ArcTIM. The companion map (Figure 5b) demonstrates the number of melt hours each grid cell was subject to during the 2005 modelled ablation period, highlighting the role the additive formulation of ArcTIM has on spatial melt distribution. This was considered important, particularly in the contributing cirques that are steeper sections of the glacier and are swept clear of snow by wind scour and sloughing (e.g. local slush flows or small-scale avalanches); ablation in these upper reaches was evident in observations of water filled crevasses and meltwater stream sources.

The time-series of potential runoff (W) generated by the various TIMs compared to total proglacial discharge (*Q*) for ML in 2005 is presented in Figure 6a. Despite the broadly comparable model skill in reproducing glacier ablation (Table 2), the visually noticeable difference between the Hock model and the TIM versions is twofold. Firstly, the Hock model results in much enhanced peak melt volumes for the entirety of the model run; and secondly, the Hock model during periods of low air temperature ($T < T_{crit}$) caused a drop to constant levels, which without the inclusion of *c* would have resulted in periods of zero melt. The TIM variant, accounting only for temperature with a constant, suffered similar under-prediction at low temperatures. Table 3 presents quantification of the degree of fit between the runoff volume time-series, clearly highlighting the disparity between the modified Hock model outputs and observed runoff (see η^2) when compared to ArcTIM and its variant, T_{melt}. Importantly, Table 3 demonstrates the need to consider model efficiency criteria with care: as Legates and McCabe(1999) emphasised, good correlation does not equate to time-series equivalence. Nonetheless, all model outputs emphasise that meltwater flow routing and the related time lags are not considered in transferring surface melt production to the proglacial discharge hydrograph simulation.

The uncertainty of the ArcTIM output was estimated as the standard error of the optimisation regression given as a percentage of the mean observed ablation (e.g. Hodgkins, 2001). This yielded an uncertainty in melt volumes of ±21% compared to the ±22% uncertainty in *Q*. A paired t-test showed no significant difference between the means of W and *Q* ($t = 0.39$, $p = 0.70$). However, visually at the diurnal scale (Figure 6a and b), there is no consistent over- or under-prediction: W consistently peaks above the actual *Q*, but consistently exhibits diurnal minima below those of *Q*, except following DOY228. The value of $\eta^2 = 0.18$ revealed only a relatively poor match between the modelled W and observed *Q*, despite the equivalence in total runoff volume during the time period (Figure 6c). Cross-correlation revealed a seasonal average lag-time of 3 hrs between W and *Q*, which increased η^2 to 0.29.

5.3 Evaluation of model components

Although the sensitivity of the respective TIMs was reported in Section 5.1, it was important to assess the suitability of the scale (Section 3.1) and precipitation (Section 4.3) algorithms by taking the time-series of W to be indicative of departures between model results.

ArcTIM was run for the three additional DEM grid resolutions (20, 10, and 2 m). Differences in the spatial distribution of melt were likely to be small, resulting primarily from the variation in location of the boundary of T_{crit} and values determined for melt from irradiance where the increasing resolution adds topographic texture (variations in z, β and γ; see also Arnold *et al*. 2009). The time-series of W for all four grid resolutions were near-perfectly correlated ($r > 0.99$), and an ANOVA test ($p < 0.001$) revealed no significant difference between the hourly outputs from the four differing resolutions, and although differences in total melt volumes were observed, there was no clearly systematic pattern between resolution and output (for DEM resolutions of 2, 5, 10 and 20 m, respective total melt values were 7.57, 7.57, 7.56, and 7.60 $\times 10^6$ m^3). It is thought these cumulative differences are likely to be due to small differences in glacier area due to changing resolution and DEM texture.

To explore whether uncertainty in glacier area was significant, the glacier margin was buffered at 10 m increments to ±20 m, and the model re-run with the differing glacier area, to determine the effect on total potential runoff. Averaged over the model domain, a ±1% error in glacier area resulted in ±1.6% error in W. Using the 5-m resolution control run, analysis of the W time-series indicated the larger change/error in glacier area resulted in more significant deviation in Q at low temperatures, especially below T_{crit}. Such errors at low air temperatures imply that reduction or increase in the glacier area, either below the threshold temperature elevation or subject to incident radiation driven melt, lead to substantial change in W. Such errors are, therefore, likely to explain the differences in total W noted between grids of differing resolutions as a result of the gridded area of a vector ice margin. Consequently, through the inclusion of the cosecant planimetric correction coefficient, ArcTIM can be treated as scale independent, although to verify this, further work would be required to test the model formulation on a more topographically varied glacier catchment.

To determine whether the parameterisation of precipitation caused systematic errors within the ArcTIM output, hourly errors in W (expressed as a percentage of the observed, unlagged Q) were plotted with precipitation (Figure 7) from which varied response could be identified: precipitation events were associated with both large (DOY195 and 207) and small errors (DOY220, 224 and 232). Times were also apparent when precipitation events appeared not to impact on the general trend of uncertainty (e.g. DOY203 and 209). Such varied response suggested the parameterisation was, on average, valid and did not lead to systematic and/or pervasive errors in estimates of W, and highlighted the event specificity of precipitation gradients. Note, however, summer precipitation at ML is typically very low.

5.4 Model validation

As a model validation exercise, ArcTIM optimised for 2005 was run using T_a data from 2004 and the 2003 DEM. This DEM change was justified because the lidar survey date was towards the close of the 2003 melt season thereby providing an estimate of the glacier topography for the commencement of the 2004 ablation season. The glacier margin as defined for 2005 was used to limit the output from ArcTIM. This was not considered a source of error since maximum glacier terminus recession was < 30 m between lidar surveys

(Barrand, 2008) which would contribute an increase of 0.5% in glacier area corresponding to an estimated mean error of < 1% in W. The bivariate values for goodness-of-fit between surveyed and modelled ablation during 2004 revealed $r^2 = 0.46$ ($\eta^2 = 0.45$), which is significantly less than for the equivalent comparison in 2005. The standard error in ablation was 0.5 mm hr^{-1}: a threefold increase from that observed in 2005. The decrease in model performance was also reflected in the time-series of W (from ArcTIM) and *Q* for 2004 (Table 3): $\eta^2 = -1.05$, indicating the poor nature of agreement between the series. Figure 8 clearly indicated over-prediction of potential runoff early in the melt season (prior to DOY210) during which time temperatures were persistently > 5 °C. Surprisingly, although visually for the remainder of the season (DOY211 onwards) W appeared a better fit to the observed *Q*, quantitatively this remained poor ($\eta^2 = 0.02$).

Examination of the parameters optimised for 2004 (Table 2) showed that values for *a*, *b*, and *c* for the respective models did not overlap, and negative values appeared to compensate for melt overestimation from T_a alone. Moreover, the use of a 'global optimisation' using data sets from both 2004 and 2005 yielded a much degraded model skill for all TIM variants (Table 3), emphasising the contrast between the two melt seasons.

6 Discussion

The results from the model runs, and comparisons, enabled further inferences to be made on the modelling strengths and weaknesses, more specifically for the two years considered.

6.1 Models for 2005

In considering the application of the ArcTIM for 2005, the model accounts for about 80% of the variance in ablation. The total ablation normalised by glacier area was 1.62 m (σ = 0.16), which agrees with but is slightly higher than the ranges of specific melt suggested for the glacier in preceding years (Hodson *et al.*, 2005).

Clearly, from the only minor improvements made to the Tc model performance with additional variables, air temperature was the forcing meteorological variable in ablation at ML, as reported for the adjacent Austre Brøggerbreen (Hodson *et al.*, 1998) but contrasting with previous energy balance considerations at ML where net shortwave radiation dominated ablation (Arnold *et al.*, 2006; Hodson *et al.*, 2005). Such a finding illustrates the interrelation between T_a and radiative fluxes, but also alludes to the potential interannual variability in energy balance considerations and validity of parameterisations within ArcTIM. Nonetheless, internal optimisation of a threshold melt value (the T_{melt} model variant) may provide a fruitful manner by which to improve TIM performance.

The small improvement by the inclusion of a radiation component compares well to similar models, but is considerably lower than those implied in results presented elsewhere (e.g. Kustas *et al.*, 1994; Pellicciotti *et al.*, 2005). Moreover, contrary to Pellicciotti *et al.*'s (2005) assertion, the inclusion of a simplistic albedo parameterisation was not beneficial in this instance. Data presented in Tables 2 and 3 suggest that a more complex and rigorous parameterisation of I (and thereby albedo) will likely have limited effect on improving model performance and partly justifies the exclusion of topographic shadowing here.

The time-series of potential and observed runoff during 2005 illustrated results analogous to those documented by Konya *et al.*, (2004) in comparing similar melt models: the additive approach to TIM model formulation appeared more suited to modelling glacier melt. The difference between the TIM formulations was small, with the modified version showing reductions in both peak and trough values; however, consistently diurnal amplitude of modelled melt exceeded that of *Q*. This was verified by an F-test showing significant difference between the variance of W and *Q* ($F = 1.7$, $p < 0.001$) despite the similarity in mean value. The daily under-prediction of melt was typically between 20:00 and 02:00 when shadowing across the glacier was greatest, which further implies that the influence of shadow was small and that there were factors involved in delaying runoff to the proglacial streams. The hydrological interpretation is that meltwater flowpaths regulate runoff, dampening the amplitude of the melt signal. This inference is emphasised by the difference in cumulative discharge series (Figure 6c) which, assuming the validity of the 2005 model run, suggests a period of storage and release within the glacier catchment, a process which has been documented previously (Hodson *et al.*, 2005); however, this is not explored further here. Nonetheless, the apparent lag time between W and Q of ~3hrs agrees with dye tracing experiments at ML which reveal transit times over and through the glacier of the order of 1 – 3hrs (Irvine-Fynn *et al.*, 2005).

The larger errors indicated in Figure 6a appeared more commonly linked to rain-free periods (cf. Figure 8), and were indicative of the overestimation and underestimation at the apexes of the diurnal cycle. Assuming rapid supraglacial runoff, this is suggestive of either changes in the threshold temperature triggering melt or variability in melt factors (a and b). In particular, the largest errors (DOY200) are seen following the cool period (DOY193-199) suggesting a potential link to thermal conditions where energy is required to raise ice temperatures prior to initiation of melting. Noticeably, the errors between DOY200 and DOY220 also exhibit a much more marked diurnal signal than at other times. Temporal variation in melt factors has been reported elsewhere (e.g. Singh and Kumar, 1996) but has seldom been explored. To examine the potential for such trends at ML, we consider melt factor a derived from Equation 1 assuming $c \neq 0$ for $T_{crit} = 1.62°C$ given the similarity in its value across the TIM variants (refer to Table 2).

For 2005, the elevation-averaged mean value of a was 0.28 mm hr^{-1} °C^{-1}, which compares well to the range of values reported from numerous locations (e.g. Hock, 2003; Zhang *et al.*, 2006). However, using AWS2 as an example, a showed variation across the observation periods (Figure 9): the increase during the middle of the ablation season then decrease thereafter is analogous to the results reported by Zhang *et al*., (2006). For glacier ice temporal changes in *a* may be attributable to changes in the distribution of supraglacial dust and cryoconite (Singh *et al.*, 2000); in the instance of ML and other Arctic glaciers, redistribution of cryoconite impacting upon surface albedo is known to occur (Hodson *et al.*, 2007; Irvine-Fynn *et al.*, 2011b).

Surface ice density provides an alternative mechanism enabling variations in *a*: rapid refreezing that occurs during the spring and very early melt season results in bubble-rich, low density ice, which may form atop the dense, bubble-free superimposed ice generated at the close of the ablation season and in early winter (Wadham *et al.*, 2006). Ice ablation is therefore likely to be reduced early in the melt season, with refreezing occurring initially, followed by a period demanding greater energy to melt the denser surface layer of winter-formed superimposed ice. Following the ablation of the dense superimposed ice layer, melt rates may increase for ice which represents the previous summer surface. The lowered porosity resulting from the previous year's melt processes, subsurface melting in response to direct irradiance, and the formation of a weathering crust layer resulting from impurities including cryoconite (e.g. Müller and Keeler, 1969) may accelerate ablation.

The presence of near-surface meltwater may also further increase melt rates and influence surface ice density. Not only does meltwater decrease albedo (e.g. Zuo and Oerlemans, 1996) but water in the liquid phase also requires less energy to raise its temperature such that a greater surface water volume may enhance ablation and enlarge void space between ice crystals. A variable water volume at the ice surface, particularly within the weathering crust (e.g. Larson, 1978), may also potentially contribute to changes in *a* throughout the season.

Critically, all the ice surface processes discussed above are likely to be linked to meteorological conditions, posing the question: do changes in *a* reflect variations in the energy balance? To assess this simply, despite the underestimation of ablation, we used the output from the adjusted EBM run (Section 4.1) to estimate the ratio between radiative and turbulent energy fluxes for each centre-line stake for all ablation survey periods. Despite the scatter, and given the uncertainties associated with both data series, comparison between the ratio of energies and *a* showed a significant positive relationship to *a* ($r^2 = 0.31$, $p < 0.05$; Figure 10). This result suggests that temporal (and spatial) variations in *a* may be described by changes in meteorology, which in turn controls ice surface characteristics.

6.2 Models for 2004

The relative failure of ArcTIM when applied for the 2004 data sets highlighted the weaknesses explored above. Table 2 illustrates the difference in optimised model parameters, and a marked contrast in spatial and temporal trends in *a* were evident between the two years (data not shown here). As detailed fully in Irvine-Fynn (2008), the meteorology of the two summer observation periods contrasted: statistically, at 99% confidence, significant annual differences existed in the mean and variance of T_a and wind speeds at both AWS2 and AWS4; during 2004, 63% of the monitoring period was significantly overcast compared to 50% of the 2005 summer; and multivariate analysis suggested low-pressure synoptic weather patterns were perhaps more important during 2004. Moreover, although directly comparable data are unavailable to validate the non-linear lapse rate observed in 2005 (Section 3.2), mean lapse rates between AWS2 and AWS4 were –0.005 °C m^{-1} and –0.004°C m^{-1} in 2004 and 2005, respectively, potentially reflecting contrasting meteorology or the changing prevalence of inversions. This lends credence to the suggestion that TIMs are sensitive to lapse rate values and demands longer-term analyses of lapse rates with respect to air temperatures (Gardner and Sharp, 2009; Gardner et al., 2009; Hodgkins et al., 2012). Such interannual contrasts in synoptic influences will certainly define the relationship between melt and T_a because clouds and inversions both have marked influence on longwave radiation fluxes

(Zhang *et al.*, 1997; Zhang *et al.*, 1996). Moreover, varying proportions of radiative energies can result in variability in ice surface characteristics (e.g. ice temperature, albedo and roughness) which furthers inability to confidently replicate ice ablation using parameters defined from a single year's observations, irrespective of TIM model formulation.

The importance of glacier surface condition is perhaps best emphasised over the first half of the 2004 season where ArcTIM over-predicted potential runoff prior to DOY210 during which time temperatures were persistently > 5 °C. An explanation of this is offered by field observations in 2004 which indicated that the early season was characterised by considerable volumes of slush, as is common on glaciers in Svalbard (e.g. Hodgkins, 2001): statistically, the mean pre-season (May) snow depths were greater in 2004 than 2005 despite a similarity in cross-glacier variance ($t = 4.06$, $p < 0.001$; $F = 1.19$, $p = 0.01$) and sea-level air temperatures consistently > 0°C commenced 10 days later than in 2005. The 'melt rate' of saturated slush is likely to be considerably different from that of glacier ice or snow, rendering predefined parameters *a* and *b* erroneous, as too is the use of ice density to convert melt to a water equivalent. Consequently, the model runs presented here emphasise how, for temporal transferability of melt models, incorporation of distinct firn, slush and snowpack elements within TIMs are beneficial (e.g. de Woul *et al.*, 2006; Hock, 1999). Indeed, existing snowpack retention (e.g. Bøggild, 2000; Janssens and Huybrechts, 2000) and refreezing (e.g. Gardner and Sharp, 2009; Hinzman and Kane, 1991) schemes to reduce or delay water release early in the melt season from TIM or EBM based models (e.g. Hanna *et al.*, 2005; Rye *et al.*, 2010) are advantageous for prediction of runoff, but may reciprocally impact on parameters used within a TIM context.

7 Conclusions

Here, a physically-based, high-resolution, distributed TIM was applied to an Arctic glacier to examine whether empirical enhancements can prove to be beneficial to model performance. The results suggested that a highly parameterised TIM, of an additive form, is successful in predicting potential melt volumes, which may be of use for predicting runoff in ungauged glacial catchments where limited ablation data is available. However, with strong correlation between ablation and T_a, inclusion of albedo to adjust potential incident radiation was ineffective in significantly enhancing accuracy of modelled ablation. The use of *a priori* knowledge of precipitation lapse rates was shown to be useful, but demonstrated the spatial distinctiveness of individual precipitation events. Moreover, the use of a scaling factor to correct between planimetric and inclined slope ablation appeared to eliminate systematic error in potential runoff volumes. While internal optimisation of the threshold temperature used within TIMs showed promise, the sensitivity of such models to the choice of temperature lapse rate is clear. The empirical parameterisation of the model did not improve model performance and certainly reduced model transferability, demonstrating the need to explore longer-term data sets linking, for example, T_a and lapse rates. These findings indicate model transferability may be limited, a conclusion contrasting to the assertions made for similar models by Carenzo *et al*. (2009). Detailed exploration of periodic measurements of ice ablation suggested that to improve TIM perfomance, a time-variant melt factor (*a*) based on the ratio of radiative to turbulent energy could be useful to explore. Critically, validation of the highly simplified melt model presented here demonstrated that, if physically-based strategies are to be considered, there is need for the inclusion of descriptors of surface and near-surface processes and flowpaths to better forecast melt and runoff. Researchers need to be vigilant in not simply assuming model or parameter transferability based on published studies examining temporally and spatially limited data sets, and in choosing appropriate models for the application in question.

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Figure Captions

Figure 1:
Map (UTM Projection, Zone 33) detailing locations of all monitoring sites utilised during field campaigns of 2004 and 2005. The mass balance stakes along the glacier centre line are numbered and referred to accordingly (e.g. MLB3). Note the change in configuration of proglacial stream channel MLW; initially active routes during 2004 (MLW04) were abandoned during the ablation season in preference for the routes reoccupied again in 2005 (MLW05).

Figure 2:
(a) Plot of mean daily air temperatures over ML from DOY200-235 (grey lines). The seasonal averages including error bars, the non-linear relationship describing the seasonal mean temperate lapse rate is shown by the bold black line and points. (b) Plot of coefficient of determination for regression between observed ablation and sum of air temperatures greater than $T_{crit.}$ T_{crit} = +1.62 indicated by black diamond.

Figure 3:
Graph displaying (a) periodic albedo variations at centre line stakes (MLB3/AWS2 shown in black) during 2005 and (b) the corresponding seasonal mean albedo at differing elevations. Linear trends for these data are shown by dashed lines.

Figure 4:
Respective ArcTIM parameter sensitivity plots to independent variations in the parameter values *a*, *b* and *c*. Graphs illustrate the responses for Nash-Sutcliff (η^2; black), Wilmott (d^2; dashed) and RMSE (grey) criteria with the 2005 OLS regression optimised parameter value indicated with '×' and the 95% parameter confidence limits shown with error bars at the top of each chart. Plot *d* shows the η^2 (black) and RMSE (grey) for 200 realisations of the ArcTIM modelled ablation with random variations of all parameters (*a*-*c*).

Figure 5:
Spatial plots determined using ArcTIM during 2005 for a) total seasonal ablation and b) number of days experiencing ablation during the observation period.

Figure 6:
Time-series of (a) total observed discharge (Q) and modelled potential runoff (W) in 2005 using four TIM variants ; (b) a plot of corresponding daily total water budget (W-Q) using ArcTIM; and (c) the cumulative daily total Q and predicted runoff W derived from ArcTIM. Dashed lines in (b) and (c) are uncertainty limits for the respective series (see body text for details).

Figure 7:
Time-series of the error (W – Q) expressed as a percentage of Q at hourly intervals and the NP precipitation record from Ny Ålesund in 2005.

Figure 8:
Time-series for 2004 of hourly total observed discharge (Q) and ArcTIM modelled potential runoff (W) using the 2005 parameterisations.

Figure 9:

Plot showing the values of *a* for AWS2 plotted against time. Dashed lines indicate the time-window over which *a* is calculated from periodic ablation stake measurements.

Figure 10:

Scatter plot showing relationship between *a* and the ratio of radiative to turbulent energy fluxes derived using an adjusted EBM.