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Examination of a physically-based, high-resolution, distributed Arctic temperature-index melt model, on Midtre Lovénbreen, Svalbard.

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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 $-0.5 \text{m a}^{-1} \text{K}^{-1}$ and $-0.63 \text{ m a}^{-1} \text{K}^{-1}$ (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

77 temporally variable factors such as cloud cover, albedo, wind speed and surface

78 characteristics, which change with location, time and climate. The performance and accuracy

79 of EBMs predictably decreases as the variability of the surface energy-balance increases

80 through an ablation season (Kane *et al.*, 1997). The result of this uncertainty is that physical

81 verification of the parameterisations used is not readily achieved, making it impossible to

82 transfer EBMs from glacier to glacier without re-parameterisation which requires extensive,

83 high resolution meteorological data and detailed information characterising of conditions at

84 the melting surface, as highlighted in MacDougall & Flowers (2011).

85   As a result of the potentially problematic use of EBMs, alternative empirically based

86 'index' methods have been employed because snow and/or ice ablation is moderately well

87 correlated to air temperature, a relationship long recognised (e.g. Finsterwalder and Schunk,

88 1887; Martinec, 1960). According to Ohmura (2001), the physical justification behind the

89 temperature-index approach is that up to 75% of the energy available for ice melt may be

90 derived from incoming longwave radiation and sensible heat. Consequently, temperature-

91 index melt models (TIMs), with varied degrees of enhancement (e.g.Hock, 1999; Pellicciotti

92 *et al.*, 2005), have become a widely used approach in glacial research programs (e.g.

93 Braithwaite, 1995; de Woul *et al.*, 2006; Ebnet *et al.*, 2005; Hanna *et al.*, 2008; Klok *et al.*,

94 2001; Marshall and Sharp, 2009; Schneeberger *et al.*, 2003). Although Hock (2003) argued

95 that TIMs yield lower accuracy over higher temporal resolution, these simplified models may

96 hold advantages both in terms of parameterisation and potential transferability (e.g. Carenzo

97 *et al.*, 2009). Therefore, using high resolution data sets, we extend, test and explore the use of

98 a novel, yet distributed, temperature-index melt model at Midtre Lovénbreen, a valley glacier

99 in Svalbard, to simulate seasonal glacier ablation and runoff. Specifically, model

100 enhancements suitable for the study site are applied and their benefits for modelling ice

101   ablation are assessed with comparison between models of varied complexity, over differing

102   spatial resolution and at-a-point energy balance calculations.

2   Field site and data collection

2.1   Field site

105   Midtre Lovénbreen (hereafter, ML), located in the north-west of Spitsbergen

106   (78°50'N 12°E; Figure 1), is one of the most studied glaciers in the High-Arctic. Local mean

107   annual temperatures and precipitation at sea-level reach –6.2°C and ~370 mm respectively

108   (Hanssen-Bauer *et al.*, 1990). The glacier occupies 49.5% of a 10.8 km^2 north-facing

109   catchment and extends from approximately 50 to 650 m above sea level (masl) with a

110   maximum thickness of 180 m (Björnsson *et al.*, 1996). The glacier's accumulation area ratio

111   is ~ 30% with a long-term average equilibrium line altitude (ELA) of ~ 400 masl (Björnsson

112   *et al.*, 1996). At lower elevations thin (< 0.2 m) winter superimposed ice forms on the glacier,

113   but this is rapidly ablated during summer months, although annual accumulation of

114   superimposed ice is found at elevations > 405 masl (König *et al.*, 2002). Measurements

115   (Hagen *et al.*, 2003) and modelling (Rye *et al.*, 2010) have shown the glacier has exhibited

116   predominantly negative mass balance over the last five decades and recent geodetic analyses

117   have indicated the negative mass balance trend may be accelerating, with contemporary

118   thinning rates of > 0.5 m water equivalent (w.e.) a^{-1} (Barrand *et al.*, 2010; Kohler *et al.*,

119   2007). Characteristically for the area, seasonal snow-cover below the ELA is removed within

120   ~14 days from the onset of melt conditions (Bruland *et al.*, 2001). Further, during summer

121   months (JJA) cloud cover is pronounced, with as few as one clear sky day per month and ~50

122   days with cloud below 300 masl (Hanssen-Bauer *et al.*, 1990). Summer meltwater is

123   discharged from the glacier's catchment via two principal stream routes (MLW and MLE;

124   Figure 1) over a hydrological season typically ~70-80 days in length.

2.2 Data collection

126   Ablation season field campaigns were conducted in 2004 and 2005, between July 9th

127   and September 4th (DOY190-247) and June 24th and August 24th (DOY175-237),

128   respectively. Hourly meteorological data were collected from automatic weather stations

129   (AWS) positioned along ML's centre line (Figure 1, Table 1).

130   The Norsk Polarintitutt (NP) centre-line stakes (Figure 1) were used to monitor changes

131   in the ice surface elevation (with precision of ± 0.005 m) relative to a reference point (see

132   Müller and Keeler, 1969). Concurrently, glacier surface albedo at representative locations

133   close to each mass balance stake was recorded using a Middleton hemispheric pyranometer

134   approximately 1 m to 1.2 m above the ice surface, and data errors were assumed to be

135   negligible (see van der Hage, 1992).

136   Runoff data during the 2004 and 2005 observation periods were collected using

137   standard hydrological methods, and uncertainties in discharge (Q) data were dominantly

138   related to the forecasting procedure and were < 19% and < 16% for MLE and MLW,

139   respectively (for full details, refer to Irvine-Fynn *et al.*, 2011a). Occasional missing data due

140   to instrumental failure, typically less than 3 hours, were estimated statistically using other

141   flow records or Q data from the adjacent Bayelva catchment, which is typically highly

142   correlated (r > 0.85) with discharge from ML (Hodson *et al.*, 2005).

143   *2.3 Glacier surface model*

144   High-resolution digital elevation models of the ML glacier surface in 2003 and 2005 were

145   derived from airborne laser scanning data (for details see Barrand *et al.* 2010; 2009). Data

146   were collected with an Optech ALTM3033 scanning system and post-processed with inertial

147   navigation system and onboard and ground-based differential GPS positioning data to yield

148   raw point clouds with mean along- and across-track point spacing of 1.38 and 1.33 m,

149   respectively, and average point density of 1.15 per m^2 (Barrand, 2008). DEMs of the glacier

150   surface were constructed using an adapted Delauney triangulation gridding algorithm.

151   Vertical elevation accuracy was +/-0.14 m based on comparison with ground-based

152   differential GPS check data over the glacier surface (Barrand *et al*., 2009). To provide further

153   model assessment opportunities, the DEM was resampled to horizontal resolutions of 2, 5, 10

154   and 20 m using bilinear interpolation.

155   **3   Development of an Arctic temperature-index melt model (ArcTIM)**

156   This section sequentially details the basis and parameterisations utilised in development

157   of the enhanced temperature-index melt model described and assessed here.

158   *3.1   The model's 'ABC' basis*

159   The application of TIMs is advocated on the grounds of their computational simplicity,

160   data availability and generally satisfactory performance (Hock, 2003). In the simplest form,

161   for a given time-step a TIM defines melt (*M*) as a function of temperature (T):

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

163   in which parameter *a* is the degree of proportionality, or the melt factor which differs for

164   snow or ice surfaces, given as m $°\mathrm{C}^{-1}$ per time interval while the threshold temperature for

165   melt (T_{crit}) is, in most practical situations, taken to be the melting point of snow and ice (0°C)

166   below which melt is zero. The variable T may be given as near-surface (~ 1.5 to 2.0 m) air

167   temperature (T_a) or as the difference between T_a and T_{crit} (Martinec, 1960). As the value of *a*

168   implicitly represents all the variables of the energy-balance, it is necessarily dynamic, and as

169   Braithwaite (1995) demonstrated, as T increases to > 10°C, values for *a* converge, indicating

170   a non-linearity between air temperatures and melt. Different TIM variants have utilised

171   alternative approaches to parameterising *b*, with zero (e.g. Martinec, 1960) or non-zero

172   (Braithwaite, 1995) values accounting for melt occurring when T < T_{crit}. Such formulations

173 ensure melt rates are somewhat spatially uniform, neglecting the influence of topographic

174 variations other than elevation.

175   In addressing these shortcomings, temperature-index models have, incorporated

176 spatially and/or temporally variable components: parameter b has been given as a function of

177 shortwave radiation (e.g. Kane *et al.*, 1997; Martinec, 1989; Pellicciotti *et al.*, 2005) or n*et*

178 *al*l-wave radiation (e.g. Brubaker *et al.*, 1996; Kustas *et al.*, 1994) or, alternatively, *a* has

179 been given as a function of incident radiation (e.g. Cazorzi and Dalla Fontana, 1996) with

180 further model enhancement by using b parameterised as a function of T_a itself (e.g. Hock,

181 1999). Conversely, Shea *et al*. (2004) defined T using regression between radiation and

182 temperature to identify a time-series of distributed residual temperatures (the difference

183 between observed and modelled temperatures) and used incident radiation to define

184 parameter b. The improvement in TIM performance using these varied approaches has

185 differed markedly between locations (cf. references above).

186   The model presented by Hock (1999), given its applicability shown on Storglaciaren,

187 has been adopted widely (e.g. de Woul *et al.*, 2006; Ebnet *et al.*, 2005; Flowers and Clarke,

188 2000; Huss *et al.*, 2008; Schneeberger *et al.*, 2003; Schuler *et al.*, 2007). However, limitations

189 are apparent, particularly in the multiplication of T and the radiation component, which is

190 counterintuitive in consideration of the independence of these two variables within the

191 energy-balance (Greuell and Genthon, 2004; Pellicciotti *et al.*, 2005) and can result in

192 overestimates of melt particularly during diurnal peak temperatures (Konya *et al.*, 2004).

193 Instead, Williams and Tarboton (1999) elegantly demonstrated through simplification of

194 terms in the energy-balance that a TIM can be better represented by:

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

196 in which z is elevation and I is potential incident radiation. In the first instance, z provides a

197 representation of the spatial distribution of T_a in response to an unknown lapse rate which is

8

198   incorporated into parameters *a* and *c*; secondly, parameters *a* and *c* also include turbulent

199   fluxes, again assumed to be linear with respect to T_a; thirdly, *b* is proportional to $(1 - \alpha)$; and

200   finally, the parameter *c*, in addition to the constants arising from a linear lapse rate, relates to

201   the turbulent fluxes and longwave emissivity. Therefore, here ArcTIM followed the form

202   suggested by Konya *et al*., (2004):

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

204   Unlike all previously published TIMs, here, melt ($\dot{M}$) is assumed to be normal to the ice

205   surface, since potential incident radiation is defined as being perpendicular to a given surface

206   slope. The use of constant *c* in the model allows a degree of correction for hitherto undefined

207   boundary layer conditions (e.g. turbulent or subsurface energy exchanges etc.). Individual

208   model parameters were defined using the dataset from 2005, as detailed in the following

209   sections.

210   *3.2   Temperature (T)*

211   To apply a distributed TIM, it is necessary to extrapolate values for air temperature

212   throughout the model domain. Numerous researchers have, for simplicity, assumed constant,

213   linear lapse rates ranging from –0.004 °C m^{-1} to –0.0076 °C m^{-1} (cf. Bøggild *et al.*, 1994;

214   Hock, 1999; Jóhannesson *et al.*, 1995; Konya *et al.*, 2004; Shea *et al.*, 2004). However,

215   glaciers influence their local climate: boundary-layer processes cause high spatial and

216   temporal temperature variability and phenomena including temperature inversions (e.g.

217   Arendt and Sharp, 1999). Consequently, constant lapse rates are inappropriate, typically

218   overestimating temperature gradients (Marshall *et al.*, 2007). Rather than optimise lapse rates

219   within the model (e.g. Jóhannesson *et al.*, 1995), ArcTIM used a non-linear lapse rate derived

220   from field observations, such that air temperature at elevation *z* (T_z) was given with respect to

221   the AWS2 record:

9

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

223   as derived from the four AWS sites deployed in 2005 (Figure 2a). The approach used by Shea

224   *et al.*, (2005) was discounted because, although it was possible to remove the co-linearity

225   between T_a and incident radiation for a single site, it was found that the relationship between

226   these variables was not strong ($r < 0.33$) and varied across the glacier as well as in time.

227   *3.3   Threshold melt temperature (T_{crit})*

228   The typically used assumption that $T_{crit} = 0°C$ (e.g. Hock, 1999) is not necessarily

229   physically tenable in light of the actual energy balance, where energy is required to raise

230   snow or ice temperature to melting point or when radiative fluxes lead to a temperature or

231   energy maximum in the subsurface (e.g. Koh and Jordan, 1995; Liston *et al.*, 1999;

232   Pellicciotti *et al.*, 2009). To define a value for T_{crit} suited for application at ML, the local

233   threshold temperature of $+1.62°C$ defined as equal probability of snow or rain was explored

234   as a starting point (Førland and Hanssen-Bauer, 2003). To ascertain whether this choice of

235   threshold temperature was valid, time-series of air temperatures at each stake in 2005 were

236   developed using Eq. 4, and the respective cumulative above-threshold air temperatures for

237   each ablation survey period were calculated for threshold temperatures incremented from $0°C$

238   to $3°C$. These series were regressed against the corresponding ablation data (Figure 2b)

239   illustrating a plateau in the coefficient of determination, and suggesting use of $T_{crit} = +1.62°C$

240   was appropriate.

241   *3.4   Potential incident radiation (I)*

242   The incident radiation (I) at any point within the catchment was modelled using algorithms

243   fully detailed by Iqbal (1983) but following Kreith and Kreider (1978):

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

245 for which I_0 is the solar constant (~ 1368 W m^{-2}), E_0 is the orbital eccentricity correction

246 factor calculated from the local day angle, θ is the angle of incidence on a tilted surface and

247 transmissivity (τ_b) is given as:

248 $$\tau_b = 0.56\left(e^{-0.65m_a} + e^{-0.0095m_a}\right) \qquad [6]$$

249 for which, using the time varying solar altitude (ϖ) derived from the day angle and local

250 latitude, the air mass ratio (m_a) is given as:

251 $$m_a = \sqrt{\left(1229 + \left(614\sin\varpi\right)^2\right)} - 614\sin\varpi \qquad [7]$$

252   The use of this variant of better-known algorithms is because clear-sky atmospheric

253 transmissivity varies over both space and time and a secant exponent estimating air mass ratio

254 using air pressure in order to adjust for local altitude is strictly only valid when solar zenith

255 angles (θ_z) are less than 70° (Kreith and Kreider, 1978). When the zenith angle exceeds 70°,

256 as is common at high-latitudes and is the case for ~90% of ML's ablation season, this

257 atmospheric approximation underestimates solar energy by failing to account for atmospheric

258 path length, refraction and curvature, yet this appears to be ignored in many high-latitude

259 studies (e.g. Ebnet *et al.*, 2005; Schuler *et al.*, 2007). In testing the suitability of the derived I,

260 modelled values for the level AWS sensors were compared with logged radiation during

261 clear-sky conditions: results showed a systematic bias of $< +11\%$, over thirteen full-day

262 records r > 0.76 (AWS2: $n = 312$ and AWS4: $n = 168$). Using field notes to refine AWS2

263 data to periods with < 4 oktas of cloud coverage, the bias was reduced, with r $= 0.84$ ($n =$

264 124); and for periods with < 2 oktas, r $= 0.92$ ($n = 41$).

265   Several TIMs adjust I to account for topographic shading (e.g. Hock, 1999). However,

266 the omission of shading has been shown to increase predicted radiation receipt by only ~6%

267 (Arnold *et al.*, 2006) and the diurnally-averaged shadowed area of ML remains $< 25\%$ for

268 ~65% of the ablation season, with greatest shadowing between 20:00 and 02:00 when

269 radiative energy is at its lowest. Moreover, as noted in Section 2.1, the climatology of ML

270   results in the prevalence of cloud covered conditions (Hanssen-Bauer *et al.*, 1990) and

271   dominance of diffuse shortwave radiation, evidenced by a mean shortwave radiative energy

272   flux of 141 Wm^{-2} at AWS2 during the summer of 2005 compared to a mean potentialreceipt

273   of 273 Wm^{-2}. Relationships utilised to adjust incident radiation for cloud cover are typically

274   unsatisfactory (e.g. Arnold *et al.*, 1996; Hock, 1999; Pellicciotti *et al.*, 2005). Furthermore,

275   the basis of a TIM is that T provides a proxy for the dominant melt energy, and consequently,

276   topographic shading and cloud cover was omitted from this model.

277   *3.5   Albedo (α)*

278   The TIM variant presented by Pellicciotti *et al*. (2005) indicated that inclusion of an

279   albedo term can improve melt calculations. Therefore, keeping to a more physical basis an

280   empirical albedo parameterisation was employed, the albedo at AWS2 (α_0) for the start day

281   of the model period (t_0) was backcast using Ordinary Least Squares (OSL) regression against

282   time in decimal days. An elevation dependency was then applied, such that albedo (α_{tz}) at

283   decimal time *t* and for elevation *z* is given by:

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

285   where

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

287   and the term dα is the mean linear decay of α over time, as calculated from all the survey

288   sites (Figure 3a). Linear regressions describing albedo variations at all the survey sites,

289   illustrated time rather than cumulative temperature best explained the temporal trend

290   observed (data not presented here).The linear relationship between *z* and α (Figure 3b)

291   appeared stronger than that identified at Haut Glacier d'Arolla by Brock *et al*., (2000) and

292   was likely a function of supraglacial dust (cryoconite) distribution (cf. Hodson *et al.*, 2007)

293   and assumed to be constant over the season. Overall, the albedo parameterisation yielded a

294   significant correlation ($r = 0.48$, $n = 64$) with a mean overestimation of ~0.06.

295   *3.6   Scale sensitivity*

296   A common flaw with melt models is the tendency to assume that pixels or grid cells are

297   planimetric (see Hopkinson (2010) for a review). This is significant for two reasons: first,

298   calculations of energy fluxes are given normal to the surface slope. The modelled magnitude

299   of radiation loading is, therefore, greatly influenced by changes in surface slope at differing

300   DEM resolutions (Arnold and Rees, 2009; Chasmer and Hopkinson, 2001). Second, a

301   systematic bias is introduced because melt occurs normal to the surface slope and the

302   planimetric assumption underestimates the slope-variant surface area subject to melt and thus

303   melt volumes (Hopkinson *et al.*, 2008). These two biases are more significant for models

304   based on DEMs of higher resolution (Chasmer and Hopkinson, 2001; Hopkinson *et al.*,

305   2008). Thus, a scale correction factor was introduced, following Hopkinson *et al.*, (2008),

306   such that for any given grid cell:

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

308   [12]

309   where M_z is total melt in m^3 water equivalent (w.e.), L is the length dimension of each square

310   grid cell, and $\dot{M}_z$ represents the modelled formulation of melt normal to the surface slope (β)

311   based on the distributed values of temperature and radiation. An ice density of 0.9 g cm^{-3} for

312   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:

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

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

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

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

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338   related to the snowline, especially at higher elevations. We recognise that this absence of

339   differentiation between snow and ice may potentially represent an important source of model

340   uncertainty. Ice topography was not evolved in time: with an estimated ceiling ablation of

341   ~1.5 m (Hodson *et al.*, 2005), the maximum influence of an evolving surface on temperature

342   and precipitation would be of the order of +0.03 °C and –0.3%, respectively, and therefore,

343   can be considered negligible given the magnitude of uncertainties associated with the source

344   data.

345   As an independent comparison, an adjusted EBM based on Brock and Arnold (2000)

346   was run to model theoretical ablation along the glacier centre line. This model included the

347   algorithms and parameterisations used within ArcTIM: the non-linear T_a lapse rate (Eq. 3),

348   high-latitude atmospheric transmissivity (Eq. 6), and temporally evolving glacier albedo (Eq.

349   8). Received incident radiation and its variation with elevation was interpolated linearly from

350   records at AWS2 and AWS4. In the absence of appropriate data, the calculations for

351   turbulent energy fluxes were necessarily simplified, using a constant aerodynamic roughness

352   length of 0.00253 (after Hodson *et al.*, 2005) and estimating absolute vapour pressure for

353   each time-step, assuming that relative humidity recorded at AWS2 was spatially uniform, and

354   that T_a varied non-linearly with elevation (Eq. 3) and held an empirical relationship with

355   saturation vapour pressure over ice (after Tetens, 1930). Wind speed was distributed over

356   elevation using linear gradients derived from the hourly data recorded at AWS2 and AWS4.

357   *4.2   Melt model parameter calibration*

358   Many glacier melt models are tuned to the time series of meltwater discharge,

359   particularly when the period of interest is annual runoff cycles. In calibrating models to

360   discharge using arbitrary routing algorithms, model output will consequently mask subtle

361   water release or storage processes which, particularly for glaciers in Svalbard, may be

362   significant (e.g. Hodgkins, 2001; Hodson *et al.*, 2005). Here, with the purpose of modelling

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363　potential runoff volumes without artificially forcing unverifiable flowpath delays, melt model

364　parameters a, b, and c were calibrated using the time-normalised, observed centre-line stake

365　ablation records for 2005. Models were based on the 5 m resolution DEM, which is much

366　higher resolution than appears in most similar modelling studies. Parameter optimisation

367　demands that the 'degree of fit' between measured data and modelled data is maximised (or

368　minimised) for which there are a number of measures (e.g. Nash and Sutcliffe, 1970;

369　Willmott, 1981). Experimental investigations into optimising the varied TIM parameter sets

370　(such that a and $b > 0.0$) indicated there was indeed a strong interdependence between T and

371　I, and the use of OLS multivariate regression was found to be the most effective optimisation

372　process. Cross-validation analyses, repeating the OLS multivariate regression but

373　withholding various datasets showed no bias in the determination of the parameters: the

374　absolute variation in multivariate $R^2 < 2.4\%$ was not significant at $p = 0.05$.

375　*4.3　Potential runoff and precipitation*

376　　　　Once optimised and executed over the glacier ice area (the model domain), summation

377　of M_z provided an estimate of the potential runoff volume available at each time-step.

378　Although precipitation was not considered as a significant process of ablation, liquid summer

379　precipitation adds to the total water equivalent available as runoff. In the absence of *in situ*

380　rain-gauge records, data collected by NP in Ny Ålesund (8 m asl) were used: hourly records

381　(P_{NA}) were reconstructed using the twice-daily precipitation record and field notes of rainfall

382　events. The local orography of Brøggerhalvøya creates precipitation lapse rates of 20% per

383　100 m for elevations < 300 m asl (Førland *et al.*, 1997) above which 10% per 100 m is more

384　appropriate (Killingtveit *et al.*, 1994). Accordingly precipitation at elevation z (P_z) was

385　described as:

386
$$P_z = \begin{cases} P_{NA}(1.00 + 0.002(z - 8)) & : \quad z \leq 300 \\ P_{NA}(1.586 + 0.001(z - 300)) & : \quad z > 300 \end{cases}$$
[11]

16

387   Solid precipitation, implicit at $T_a < 1.62°C$, was assumed to refreeze and was accounted for

388   by adding the equivalent ice depth to the measured ablation survey data. Liquid precipitation

389   was assumed to fall on the planimetric grid cells, and therefore, when present, simply

390   multiplied by the grid cell area and added to the scale corrected meltwater volume to yield a

391   total available for runoff. The precipitation occurring on the surrounding mountain slopes

392   within the glacier basin was excluded from the model because of the characteristic talus,

393   lateral moraines and scree near the glacier margin for which the associated effects on water

394   retention, percolation or routing were unknown.

395   **5**   **Application of models on Midtre Lovénbreen**

396   Here, we detail and discuss the melt model output(s) with reference to the primary data

397   sets of observed ablation and meltwater runoff.

398   *5.1*   *Ice ablation*

399   The performance of the model parameterisations for 2005, comparing observed, slope-

400   corrected and modelled ablation are detailed in Table 2. A slight improvement to modelling

401   ablation was made by enhancing a simple TIM to ArcTIM, with the R^2 of ~80% matching

402   similar model performance reported elsewhere for temperature-radiation index models (e.g.

403   Brubaker *et al.*, 1996; de Woul *et al.*, 2006; Kane *et al.*, 1997). Furthermore, for 2005,

404   ArcTIM equalled the performance of a more complex full EBM on ML presented by Arnold

405   *et al.*, (2006). Interestingly, the inclusion of albedo within ArcTIM subtlety reduced its

406   efficacy, while both the Hock and T_{melt} model variants appeared to show slightly improved

407   agreement between modelled and observed ablation, with standard errors of ~0.17 mm hr^{-1}.

408   The TIM models suggested a melt rate of 0.3 mm hr^{-1} °C^{-1} at ML which exceeds the 0.05 mm

409   hr^{-1} °C^{-1} reported by Pellicciotti *et al.*, (2005) for a similar TIM formulation applied to an

410   Alpine glacier.

411   To examine the sensitivity of individual parameter values, three numerical efficiency

412   criteria were explored further: Figure 4 illustrates model sensitivity to changing parameter

413   values, using the Nash-Sutcliffe (η^2: (Nash and Sutcliffe, 1970) and Willmott (d^2: (Willmott,

414   1981) indices of agreement and the root mean squared error (RMSE). Examination of these

415   plots demonstrates (i) ArcTIM is highly sensitive to the values of both b and c, but less so for

416   the value of a and (ii) the η^2 criteria is a more sensitive measure of model fit. To eliminate

417   the potential for equifinality within the model, we varied all three parameters at random

418   (Figure 4d) finding a single optimum combination with a minimum RMSE of ~0.2 mm hr^{-1}.

419   Crucially, comparison between the TIM variants and the adjusted EBM showed the

420   latter to perform relatively poorly: the correlation between observed and modelled ablation

421   was 0.6, with an RMSE of 0.4 mm hr^{-1}. The EBM typically under-predicted melt by 0.16 mm

422   hr^{-1}, contrasting to the persistent over-prediction of melt using the EBM reported in Arnold *et*

423   *al. (*2006). We ascribed this to the disparity in parameterisation of the EBM when compared

424   to those presented by Arnold *et al*. (2006) and Rye *et al*. (2010), although this was not

425   investigated further here.

426   *5.2   Comparison of potential and actual runoff*

427   With the parameters for ArcTIM successfully and robustly calibrated for 2005, and,

428   ignoring the spatial uncertainty inherent in the input data, the model was applied to the ML

429   catchment. The spatial distribution of ablation shown in Figure 5a is given as the modelled

430   seasonal total, derived using ArcTIM. The companion map (Figure 5b) demonstrates the

431   number of melt hours each grid cell was subject to during the 2005 modelled ablation period,

432   highlighting the role the additive formulation of ArcTIM has on spatial melt distribution. This

433   was considered important, particularly in the contributing cirques that are steeper sections of

434   the glacier and are swept clear of snow by wind scour and sloughing (e.g. local slush flows or

435   small-scale avalanches); ablation in these upper reaches was evident in observations of water

436   filled crevasses and meltwater stream sources.

437   The time-series of potential runoff (W) generated by the various TIMs compared to

438   total proglacial discharge (Q) for ML in 2005 is presented in Figure 6a. Despite the broadly

439   comparable model skill in reproducing glacier ablation (Table 2), the visually noticeable

440   difference between the Hock model and the TIM versions is twofold. Firstly, the Hock model

441   results in much enhanced peak melt volumes for the entirety of the model run; and secondly,

442   the Hock model during periods of low air temperature ($T < T_{crit}$) caused a drop to constant

443   levels, which without the inclusion of c would have resulted in periods of zero melt. The TIM

444   variant, accounting only for temperature with a constant, suffered similar under-prediction at

445   low temperatures. Table 3 presents quantification of the degree of fit between the runoff

446   volume time-series, clearly highlighting the disparity between the modified Hock model

447   outputs and observed runoff (see η^2) when compared to ArcTIM and its variant, T_{melt}.

448   Importantly, Table 3 demonstrates the need to consider model efficiency criteria with care: as

449   Legates and McCabe(1999) emphasised, good correlation does not equate to time-series

450   equivalence. Nonetheless, all model outputs emphasise that meltwater flow routing and the

451   related time lags are not considered in transferring surface melt production to the proglacial

452   discharge hydrograph simulation.

453   The uncertainty of the ArcTIM output was estimated as the standard error of the

454   optimisation regression given as a percentage of the mean observed ablation (e.g. Hodgkins,

455   2001). This yielded an uncertainty in melt volumes of ±21% compared to the ±22%

456   uncertainty in Q. A paired t-test showed no significant difference between the means of W

457   and Q (t = 0.39, $p = 0.70$). However, visually at the diurnal scale (Figure 6a and b), there is

458   no consistent over- or under-prediction: W consistently peaks above the actual Q, but

459   consistently exhibits diurnal minima below those of Q, except following DOY228. The value

460 of $\eta^2 = 0.18$ revealed only a relatively poor match between the modelled W and observed Q,

461 despite the equivalence in total runoff volume during the time period (Figure 6c). Cross-

462 correlation revealed a seasonal average lag-time of 3 hrs between W and Q, which increased

463 η^2 to 0.29.

464 *5.3   Evaluation of model components*

465 Although the sensitivity of the respective TIMs was reported in Section 5.1, it was

466 important to assess the suitability of the scale (Section 3.1) and precipitation (Section 4.3)

467 algorithms by taking the time-series of W to be indicative of departures between model

468 results.

469 ArcTIM was run for the three additional DEM grid resolutions (20, 10, and 2 m).

470 Differences in the spatial distribution of melt were likely to be small, resulting primarily from

471 the variation in location of the boundary of T_{crit} and values determined for melt from

472 irradiance where the increasing resolution adds topographic texture (variations in z, β and γ;

473 see also Arnold *et al*. 2009). The time-series of W for all four grid resolutions were near-

474 perfectly correlated (r > 0.99), and an ANOVA test ($p < 0.001$) revealed no significant

475 difference between the hourly outputs from the four differing resolutions, and although

476 differences in total melt volumes were observed, there was no clearly systematic pattern

477 between resolution and output (for DEM resolutions of 2, 5, 10 and 20 m, respective total

478 melt values were 7.57, 7.57, 7.56, and 7.60 $\times 10^6$ m^3). It is thought these cumulative

479 differences are likely to be due to small differences in glacier area due to changing resolution

480 and DEM texture.

481 To explore whether uncertainty in glacier area was significant, the glacier margin was

482 buffered at 10 m increments to ±20 m, and the model re-run with the differing glacier area, to

483 determine the effect on total potential runoff. Averaged over the model domain, a ±1% error

484 in glacier area resulted in ±1.6% error in W. Using the 5-m resolution control run, analysis of

485   the W time-series indicated the larger change/error in glacier area resulted in more significant

486   deviation in Q at low temperatures, especially below T_{crit}. Such errors at low air temperatures

487   imply that reduction or increase in the glacier area, either below the threshold temperature

488   elevation or subject to incident radiation driven melt, lead to substantial change in W. Such

489   errors are, therefore, likely to explain the differences in total W noted between grids of

490   differing resolutions as a result of the gridded area of a vector ice margin. Consequently,

491   through the inclusion of the cosecant planimetric correction coefficient, ArcTIM can be

492   treated as scale independent, although to verify this, further work would be required to test

493   the model formulation on a more topographically varied glacier catchment.

494   To determine whether the parameterisation of precipitation caused systematic errors

495   within the ArcTIM output, hourly errors in W (expressed as a percentage of the observed,

496   unlagged Q) were plotted with precipitation (Figure 7) from which varied response could be

497   identified: precipitation events were associated with both large (DOY195 and 207) and small

498   errors (DOY220, 224 and 232). Times were also apparent when precipitation events appeared

499   not to impact on the general trend of uncertainty (e.g. DOY203 and 209). Such varied

500   response suggested the parameterisation was, on average, valid and did not lead to systematic

501   and/or pervasive errors in estimates of W, and highlighted the event specificity of

502   precipitation gradients. Note, however, summer precipitation at ML is typically very low.

503   *5.4   Model validation*

504   As a model validation exercise, ArcTIM optimised for 2005 was run using T_a data

505   from 2004 and the 2003 DEM. This DEM change was justified because the lidar survey date

506   was towards the close of the 2003 melt season thereby providing an estimate of the glacier

507   topography for the commencement of the 2004 ablation season. The glacier margin as

508   defined for 2005 was used to limit the output from ArcTIM. This was not considered a source

509   of error since maximum glacier terminus recession was < 30 m between lidar surveys

510 (Barrand, 2008) which would contribute an increase of 0.5% in glacier area corresponding to

511 an estimated mean error of < 1% in W. The bivariate values for goodness-of-fit between

512 surveyed and modelled ablation during 2004 revealed $r^2 = 0.46$ ($\eta^2 = 0.45$), which is

513 significantly less than for the equivalent comparison in 2005. The standard error in ablation

514 was 0.5 mm hr^{-1}: a threefold increase from that observed in 2005. The decrease in model

515 performance was also reflected in the time-series of W (from ArcTIM) and Q for 2004 (Table

516 3): $\eta^2 = -1.05$, indicating the poor nature of agreement between the series. Figure 8 clearly

517 indicated over-prediction of potential runoff early in the melt season (prior to DOY210)

518 during which time temperatures were persistently > 5 °C. Surprisingly, although visually for

519 the remainder of the season (DOY211 onwards) W appeared a better fit to the observed Q,

520 quantitatively this remained poor ($\eta^2 = 0.02$).

521   Examination of the parameters optimised for 2004 (Table 2) showed that values for a,

522 b, and c for the respective models did not overlap, and negative values appeared to

523 compensate for melt overestimation from T_a alone. Moreover, the use of a 'global

524 optimisation' using data sets from both 2004 and 2005 yielded a much degraded model skill

525 for all TIM variants (Table 3), emphasising the contrast between the two melt seasons.

526 **6   Discussion**

527 The results from the model runs, and comparisons, enabled further inferences to be made

528 on the modelling strengths and weaknesses, more specifically for the two years considered.

529 *6.1   Models for 2005*

530   In considering the application of the ArcTIM for 2005, the model accounts for about

531 80% of the variance in ablation. The total ablation normalised by glacier area was 1.62 m (σ

532 = 0.16), which agrees with but is slightly higher than the ranges of specific melt suggested for

533 the glacier in preceding years (Hodson *et al.*, 2005).

534   Clearly, from the only minor improvements made to the Tc model performance with

535   additional variables, air temperature was the forcing meteorological variable in ablation at

536   ML, as reported for the adjacent Austre Brøggerbreen (Hodson *et al.*, 1998) but contrasting

537   with previous energy balance considerations at ML where net shortwave radiation dominated

538   ablation (Arnold *et al.*, 2006; Hodson *et al.*, 2005). Such a finding illustrates the interrelation

539   between T_a and radiative fluxes, but also alludes to the potential interannual variability in

540   energy balance considerations and validity of parameterisations within ArcTIM. Nonetheless,

541   internal optimisation of a threshold melt value (the T_{melt} model variant) may provide a fruitful

542   manner by which to improve TIM performance.

543   The small improvement by the inclusion of a radiation component compares well to

544   similar models, but is considerably lower than those implied in results presented elsewhere

545   (e.g. Kustas *et al.*, 1994; Pellicciotti *et al.*, 2005). Moreover, contrary to Pellicciotti *et al.*'s

546   (2005) assertion, the inclusion of a simplistic albedo parameterisation was not beneficial in

547   this instance. Data presented in Tables 2 and 3 suggest that a more complex and rigorous

548   parameterisation of I (and thereby albedo) will likely have limited effect on improving model

549   performance and partly justifies the exclusion of topographic shadowing here.

550   The time-series of potential and observed runoff during 2005 illustrated results

551   analogous to those documented by Konya *et al.*, (2004) in comparing similar melt models:

552   the additive approach to TIM model formulation appeared more suited to modelling glacier

553   melt. The difference between the TIM formulations was small, with the modified version

554   showing reductions in both peak and trough values; however, consistently diurnal amplitude

555   of modelled melt exceeded that of *Q*. This was verified by an F-test showing significant

556   difference between the variance of W and *Q* (F = 1.7, p < 0.001) despite the similarity in

557   mean value. The daily under-prediction of melt was typically between 20:00 and 02:00 when

558   shadowing across the glacier was greatest, which further implies that the influence of shadow

559   was small and that there were factors involved in delaying runoff to the proglacial streams.

560   The hydrological interpretation is that meltwater flowpaths regulate runoff, dampening the

561   amplitude of the melt signal. This inference is emphasised by the difference in cumulative

562   discharge series (Figure 6c) which, assuming the validity of the 2005 model run, suggests a

563   period of storage and release within the glacier catchment, a process which has been

564   documented previously (Hodson *et al.*, 2005); however, this is not explored further here.

565   Nonetheless, the apparent lag time between W and Q of ~3hrs agrees with dye tracing

566   experiments at ML which reveal transit times over and through the glacier of the order of 1 –

567   3hrs (Irvine-Fynn *et al.*, 2005).

568   The larger errors indicated in Figure 6a appeared more commonly linked to rain-free

569   periods (cf. Figure 8), and were indicative of the overestimation and underestimation at the

570   apexes of the diurnal cycle. Assuming rapid supraglacial runoff, this is suggestive of either

571   changes in the threshold temperature triggering melt or variability in melt factors (*a* and *b*). In

572   particular, the largest errors (DOY200) are seen following the cool period (DOY193-199)

573   suggesting a potential link to thermal conditions where energy is required to raise ice

574   temperatures prior to initiation of melting. Noticeably, the errors between DOY200 and

575   DOY220 also exhibit a much more marked diurnal signal than at other times. Temporal

576   variation in melt factors has been reported elsewhere (e.g. Singh and Kumar, 1996) but has

577   seldom been explored. To examine the potential for such trends at ML, we consider melt

578   factor *a* derived from Equation 1 assuming $c \neq 0$ for $T_{crit} = 1.62°C$ given the similarity in its

579   value across the TIM variants (refer to Table 2).

580   For 2005, the elevation-averaged mean value of *a* was 0.28 mm hr^{-1} °C^{-1}, which

581   compares well to the range of values reported from numerous locations (e.g. Hock, 2003;

582   Zhang *et al.*, 2006). However, using AWS2 as an example, *a* showed variation across the

583   observation periods (Figure 9): the increase during the middle of the ablation season then

584 decrease thereafter is analogous to the results reported by Zhang *et al*., (2006). For glacier ice

585 temporal changes in *a* may be attributable to changes in the distribution of supraglacial dust

586 and cryoconite (Singh *et al.*, 2000); in the instance of ML and other Arctic glaciers,

587 redistribution of cryoconite impacting upon surface albedo is known to occur (Hodson *et al.*,

588 2007; Irvine-Fynn *et al.*, 2011b).

589   Surface ice density provides an alternative mechanism enabling variations in *a*: rapid

590 refreezing that occurs during the spring and very early melt season results in bubble-rich, low

591 density ice, which may form atop the dense, bubble-free superimposed ice generated at the

592 close of the ablation season and in early winter (Wadham *et al.*, 2006). Ice ablation is

593 therefore likely to be reduced early in the melt season, with refreezing occurring initially,

594 followed by a period demanding greater energy to melt the denser surface layer of winter-

595 formed superimposed ice. Following the ablation of the dense superimposed ice layer, melt

596 rates may increase for ice which represents the previous summer surface. The lowered

597 porosity resulting from the previous year's melt processes, subsurface melting in response to

598 direct irradiance, and the formation of a weathering crust layer resulting from impurities

599 including cryoconite (e.g. Müller and Keeler, 1969) may accelerate ablation.

600   The presence of near-surface meltwater may also further increase melt rates and

601 influence surface ice density. Not only does meltwater decrease albedo (e.g. Zuo and

602 Oerlemans, 1996) but water in the liquid phase also requires less energy to raise its

603 temperature such that a greater surface water volume may enhance ablation and enlarge void

604 space between ice crystals. A variable water volume at the ice surface, particularly within the

605 weathering crust (e.g. Larson, 1978), may also potentially contribute to changes in *a*

606 throughout the season.

607   Critically, all the ice surface processes discussed above are likely to be linked to

608 meteorological conditions, posing the question: do changes in *a* reflect variations in the

609   energy balance? To assess this simply, despite the underestimation of ablation, we used the

610   output from the adjusted EBM run (Section 4.1) to estimate the ratio between radiative and

611   turbulent energy fluxes for each centre-line stake for all ablation survey periods. Despite the

612   scatter, and given the uncertainties associated with both data series, comparison between the

613   ratio of energies and a showed a significant positive relationship to a ($r^2 = 0.31$, $p < 0.05$;

614   Figure 10). This result suggests that temporal (and spatial) variations in a may be described

615   by changes in meteorology, which in turn controls ice surface characteristics.

616   *6.2   Models for 2004*

617   The relative failure of ArcTIM when applied for the 2004 data sets highlighted the

618   weaknesses explored above. Table 2 illustrates the difference in optimised model parameters,

619   and a marked contrast in spatial and temporal trends in a were evident between the two years

620   (data not shown here). As detailed fully in Irvine-Fynn (2008), the meteorology of the two

621   summer observation periods contrasted: statistically, at 99% confidence, significant annual

622   differences existed in the mean and variance of T_a and wind speeds at both AWS2 and

623   AWS4; during 2004, 63% of the monitoring period was significantly overcast compared to

624   50% of the 2005 summer; and multivariate analysis suggested low-pressure synoptic weather

625   patterns were perhaps more important during 2004. Moreover, although directly comparable

626   data are unavailable to validate the non-linear lapse rate observed in 2005 (Section 3.2), mean

627   lapse rates between AWS2 and AWS4 were -0.005 °C m^{-1} and -0.004°C m^{-1} in 2004 and

628   2005, respectively, potentially reflecting contrasting meteorology or the changing prevalence

629   of inversions. This lends credence to the suggestion that TIMs are sensitive to lapse rate

630   values and demands longer-term analyses of lapse rates with respect to air temperatures

631   (Gardner and Sharp, 2009; Gardner et al., 2009; Hodgkins et al., 2012). Such interannual

632   contrasts in synoptic influences will certainly define the relationship between melt and T_a

633   because clouds and inversions both have marked influence on longwave radiation fluxes

26

634 (Zhang *et al.*, 1997; Zhang *et al.*, 1996). Moreover, varying proportions of radiative energies

635 can result in variability in ice surface characteristics (e.g. ice temperature, albedo and

636 roughness) which furthers inability to confidently replicate ice ablation using parameters

637 defined from a single year's observations, irrespective of TIM model formulation.

638   The importance of glacier surface condition is perhaps best emphasised over the first

639 half of the 2004 season where ArcTIM over-predicted potential runoff prior to DOY210

640 during which time temperatures were persistently > 5 °C. An explanation of this is offered by

641 field observations in 2004 which indicated that the early season was characterised by

642 considerable volumes of slush, as is common on glaciers in Svalbard (e.g. Hodgkins, 2001):

643 statistically, the mean pre-season (May) snow depths were greater in 2004 than 2005 despite

644 a similarity in cross-glacier variance (t = 4.06, $p < 0.001$; F = 1.19, $p = 0.01$) and sea-level air

645 temperatures consistently > 0°C commenced 10 days later than in 2005. The 'melt rate' of

646 saturated slush is likely to be considerably different from that of glacier ice or snow,

647 rendering predefined parameters *a* and *b* erroneous, as too is the use of ice density to convert

648 melt to a water equivalent. Consequently, the model runs presented here emphasise how, for

649 temporal transferability of melt models, incorporation of distinct firn, slush and snowpack

650 elements within TIMs are beneficial (e.g. de Woul *et al.*, 2006; Hock, 1999). Indeed, existing

651 snowpack retention (e.g. Bøggild, 2000; Janssens and Huybrechts, 2000) and refreezing (e.g.

652 Gardner and Sharp, 2009; Hinzman and Kane, 1991) schemes to reduce or delay water

653 release early in the melt season from TIM or EBM based models (e.g. Hanna *et al.*, 2005;

654 Rye *et al.*, 2010) are advantageous for prediction of runoff, but may reciprocally impact on

655 parameters used within a TIM context.

656 **7   Conclusions**

657   Here, a physically-based, high-resolution, distributed TIM was applied to an Arctic

658 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 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 '$\times$' 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.