

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

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1 **Examination of a physically-based, high-resolution,**
2 **distributed Arctic temperature-index melt model, on Midtre**
3 **Lovénbreen, Svalbard**

4
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28 **Abstract:**

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

50

51

52 1 Introduction

53 Small glaciers and ice caps account for ~ 14% of the terrestrial ice in the Arctic, and
54 research has highlighted broadly persistent, negative mass balances of these ice masses in
55 recent decades (Arendt *et al.*, 2002; Dowdeswell *et al.*, 1997; Koerner, 2005; Nuth *et al.*,
56 2010). This characteristic in high latitude glacier mass balance may be tentatively linked to
57 apparent warming trends leading to lengthened ablation seasons (Christensen *et al.*, 2007;
58 Kattsov *et al.*, 2005). Within this context, in glaciological terms, regional estimates of glacier
59 mass balance sensitivity have been presented with values of between $-0.5\text{ m a}^{-1}\text{ K}^{-1}$ and -0.63
60 $\text{ m a}^{-1}\text{ K}^{-1}$ (Braithwaite and Raper, 2007; de Woul and Hock, 2005) for ablation season ice
61 losses, suggestive of increasing seasonal meltwater fluxes from high-latitude, glacierised
62 catchments in response to projected climate forcing. Nonetheless, significant issues remain in
63 terms the spatial and temporal resolution and transferability of melt models for indicating
64 both contemporary and **future runoff volumes. There is still a need** to trial models to discern
65 which elements, if any, within more sophisticated schemes are beneficial to model output
66 precision (Hock, 2005).

67 The amount of surface melting of snow or glacier ice during the ablation season is
68 controlled by the energy fluxes, which are specific to local climatic and surface conditions.
69 Thorough reviews of the surface energy balance can be found in Greuell and Genthon (2004)
70 and Hock (2005) and numerous energy balance models (EBMs) have been thoroughly
71 investigated for valley glaciers in temperate (e.g. Anslow *et al.*, 2008; Arnold *et al.*, 1996;
72 Escher-Vetter, 2000; Klok and Oerlemans, 2002) and Arctic or sub-Arctic settings (e.g.
73 Arnold *et al.*, 2006; Hock and Holmgren, 2005; Hock and Noetzli, 1997; MacDougall and
74 Flowers, 2011; Rye *et al.*, 2010). Critically, EBMs all explicitly stress how variations in
75 glacier surface conditions influence ablation and subsequent runoff patterns. There are
76 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.

103 2 Field site and data collection

104 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 $\sim 30\%$ 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.

125 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):

$$162 \quad M = aT + b \quad (\text{for } T > T_{\text{crit}}) \quad [1]$$

163 in which parameter a is the degree of proportionality, or the melt factor which differs for
164 snow or ice surfaces, given as $\text{m } ^\circ\text{C}^{-1}$ 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 $> 10^\circ\text{C}$, 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 $T < T_{\text{crit}}$. 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 *net*
178 *all-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:

$$195 \quad M = az + bI + c \quad [2]$$

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

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):

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

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 $-0.004 \text{ }^\circ\text{C m}^{-1}$ to $-0.0076 \text{ }^\circ\text{C m}^{-1}$ (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:

222 $T_z = T_{\text{AWS2}} - 1.8 \ln(z) + 9.6$ [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 $T_{\text{crit}} = 0^\circ\text{C}$ (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 $+1.62^\circ\text{C}$ 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 $T_{\text{crit}} = +1.62^\circ\text{C}$
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 ($\sim 1368 \text{ W m}^{-2}$), 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:

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

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:

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

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 $\sim 90\%$ 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 $\sim 6\%$
267 (Arnold *et al.*, 2006) and the diurnally-averaged shadowed area of ML remains $< 25\%$ for
268 $\sim 65\%$ 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 potential receipt
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_z) at
283 decimal time t and for elevation z is given by:

$$284 \alpha_z = \alpha_0 + (0.0006z - 0.1133) + d\alpha(t - t_0) \quad [8]$$

285 where

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

287 and the term $d\alpha$ 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:

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

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.

313 4 Comparative models

314 4.1 Melt models

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

$$320 \dot{M}_z = aT_z + bT_z(1-\alpha)I_m + c \quad (\text{for } T > T_{\text{crit}}) \quad [13]$$

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

$$322 \dot{M}_z = a(T_z - T_{\text{melt}}) + b(1-\alpha)I_m + c \quad [14]$$

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

329 Many published temperature-index melt models distinguish between snow and ice
330 covered surfaces **which accounts for critical spatial differences in meltwater genesis** (e.g.
331 Hock, 1999; Jóhannesson *et al.*, 1995). However, unlike these examples, here, models did not
332 account for a difference between snow and ice surface on the glacier. The reason for this was
333 threefold: first, the TIM presented by Schneeberger *et al.* (2003) evidenced only subtle
334 difference between melt factors (a) for ice and snow; second, snowline retreat on the shallow
335 slopes characterising the majority of ML's ice area is typically rapid; and third, because the
336 melt model which includes an elevation-defined albedo was to be tuned to ablation
337 measurements, it was assumed optimisation would account for any differences potentially

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

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:

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

387 Solid precipitation, implicit at $T_a < 1.62^\circ\text{C}$, 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 subtly 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 $\sim 0.17 \text{ mm hr}^{-1}$.
408 The TIM models suggested a melt rate of $0.3 \text{ mm hr}^{-1} \text{ }^\circ\text{C}^{-1}$ at ML which exceeds the 0.05 mm
409 $\text{hr}^{-1} \text{ }^\circ\text{C}^{-1}$ 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 $\sim 0.2 \text{ mm hr}^{-1}$.

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 ($T < T_{\text{crit}}$) 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 $\pm 21\%$ compared to the $\pm 22\%$
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 $7.60 \times 10^6 \text{ m}^3$). 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 $\pm 1\%$ error
484 in glacier area resulted in $\pm 1.6\%$ 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 $> 5 \text{ }^\circ\text{C}$. 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 ~ 3 hrs 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 $T_{\text{crit}} = 1.62^\circ\text{C}$ 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 $0.28 \text{ mm hr}^{-1} \text{ }^\circ\text{C}^{-1}$, 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 $-0.005\text{ }^\circ\text{C m}^{-1}$ and $-0.004\text{ }^\circ\text{C m}^{-1}$ 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

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 $> 5\text{ }^{\circ}\text{C}$. 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 $> 0^{\circ}\text{C}$ 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

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

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966 Figure Captions

967

968 **Figure 1:**

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

975

976 **Figure 2:**

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

982

983 **Figure 3:**

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

987

988 **Figure 4:**

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

995

996 **Figure 5:**

997 Spatial plots determined using ArcTIM during 2005 for a) total seasonal ablation and b)
998 number of **days** experiencing ablation during the observation period.

999

1000 **Figure 6:**

1001 Time-series of (a) total observed discharge (Q) and modelled potential runoff (W) in 2005
1002 using four TIM variants ; (b) a plot of corresponding daily total water budget (W-Q) using
1003 ArcTIM; and (c) the cumulative daily total Q and predicted runoff W derived from ArcTIM.
1004 Dashed lines in (b) and (c) are uncertainty limits for the respective series (see body text for
1005 details).

1006

1007 **Figure 7:**

1008 Time-series of the error (W – Q) expressed as a percentage of Q at hourly intervals and the
1009 NP precipitation record from Ny Ålesund in 2005.

1010

1011 **Figure 8:**

1012 Time-series for 2004 of hourly total observed discharge (Q) and ArcTIM modelled potential
1013 runoff (W) using the 2005 parameterisations.

1014

1015 **Figure 9:**

1016 Plot showing the values of a for AWS2 plotted against time. Dashed lines indicate the time-
1017 window over which a is calculated from periodic ablation stake measurements.

1018

1019 **Figure 10:**

1020 Scatter plot showing relationship between a and the ratio of radiative to turbulent energy
1021 fluxes derived using an adjusted EBM.

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