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1 **Modelling the feedbacks between mass balance, ice flow and debris**
2 **transport to predict the response to climate change of debris-covered**
3 **glaciers in the Himalaya**

4

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15 **Abstract**

16 Many Himalayan glaciers are characterised in their lower reaches by a rock debris layer. This
17 debris insulates the glacier surface from atmospheric warming and complicates the response
18 to climate change compared to glaciers with clean-ice surfaces. Debris-covered glaciers can
19 persist well below the altitude that would be sustainable for clean-ice glaciers, resulting in
20 much longer timescales of mass loss and meltwater production. The properties and evolution
21 of supraglacial debris present a considerable challenge to understanding future glacier
22 change. Existing approaches to predicting variations in glacier volume and meltwater
23 production rely on numerical models that represent the processes governing glaciers with
24 clean-ice surfaces, and yield conflicting results. We developed a numerical model that
25 couples the flow of ice and debris and includes important feedbacks between debris
26 accumulation and glacier mass balance. To investigate the impact of debris transport on the
27 response of a glacier to recent and future climate change, we applied this model to a large
28 debris-covered Himalayan glacier—Khumbu Glacier in Nepal. Our results demonstrate that
29 supraglacial debris prolongs the response of the glacier to warming and causes lowering of
30 the glacier surface *in situ*, concealing the magnitude of mass loss when compared with
31 estimates based on glacierised area. Since the Little Ice Age, Khumbu Glacier has lost 34%
32 of its volume while its area has reduced by only 6%. We predict a decrease in glacier volume
33 of 8–10% by AD2100, accompanied by dynamic and physical detachment of the debris-
34 covered tongue from the active glacier within the next 150 years. This detachment will
35 accelerate rates of glacier decay, and similar changes are likely for other debris-covered
36 glaciers in the Himalaya.

37

38 **1. Introduction**

39 Glaciers in the Himalaya are rapidly losing mass (Bolch et al., 2012). However, data
40 describing past and present glacier volumes are scarce, resulting in varying predictions of
41 future glacier volumes (Cogley, 2011; Kääb et al., 2012). To improve predictions of how
42 Himalayan glaciers will decline through the 21st Century and the impact on Asian water
43 resources, we need to quantify the processes that drive glacier change (e.g. Immerzeel et al.,
44 2013; Pellicciotti et al., 2015; Ragettli et al., 2015; Shea et al., 2015). Changes in glacier
45 volume are driven by climate variations, particularly changes in atmospheric temperature and
46 precipitation amount, and modified by ice flow (Bolch et al., 2012; Kääb et al., 2012). The
47 lower portions of clean-ice glaciers lose mass rapidly during periods of warming. As glaciers
48 recede to higher elevations, a new equilibrium state between this smaller glacier and the

49 warmer climate may be established. Numerical modelling is required to understand the
50 processes that cause glaciers to change because we cannot rely simply on the extrapolation of
51 present-day trends. Previous studies of Himalayan glaciers using models designed for clean-
52 ice glaciers resulted in predictions of widespread rapid deglaciation (e.g. Shea et al., 2015).
53 However, debris-covered glaciers respond differently to warming because debris insulates the
54 ice surface (Jouvet et al., 2011; Kirkbride and Deline, 2013; Pellicciotti et al., 2015; Østrem,
55 1959). Debris-covered glaciers lose most mass by surface lowering rather than terminus
56 recession (Hambrey et al., 2008). Debris-covered glaciers can persist at lower elevations than
57 would be possible for an equivalent clean-ice glacier even when dramatically out of
58 equilibrium with climate (Anderson, 2000; Benn et al., 2012). As glaciers lose mass
59 preferentially from areas of clean ice and mass loss results in the melt-out of englacial debris,
60 debris coverage will increase as a glacier shrinks (Bolch et al., 2008; Kirkbride and Deline,
61 2013; Thakuri et al., 2014). Therefore, predicting the future of the Himalayan cryosphere and
62 water resources depends on understanding the impacts of climate change on debris-covered
63 glaciers.

64

65 Debris on glacier tongues is derived from surrounding hillslopes and is transported
66 englacially before resurfacing in the ablation zone (Fig. 1a). In times negative mass balance,
67 velocities decline and debris thickness at the ice surface increases (Kirkbride and Deline,
68 2013) (Fig. 1b). A thin layer of rock debris (0.01–0.1 m) enhances glacier surface ablation by
69 reducing albedo, whereas thicker rock debris reduces ablation by insulating the surface
70 (Mihalcea et al., 2008; Nicholson and Benn, 2006; Østrem, 1959). Thick supraglacial debris
71 causes a reversal of the mass balance gradient, with higher ablation rates upglacier than at the
72 terminus leading to reduced driving stresses and ice flow (Jouvet et al., 2011; Quincey et al.,
73 2009). Spatial heterogeneity in debris thickness results in differential surface ablation and the
74 formation and decay of ice cliffs and supraglacial ponds that locally enhance ablation (Reid
75 and Brock, 2014; Sakai et al., 2000). An obstacle to understanding the behaviour of debris-
76 covered glaciers lies in quantifying the highly variable distribution of debris across the
77 glacier surface and how this differs between glaciers. Supraglacial debris distribution and
78 thickness are difficult to determine remotely and laborious to measure directly (Mihalcea et
79 al., 2008; Nicholson and Benn, 2006; Reid et al., 2012), particularly over more than one
80 glacier (Pellicciotti et al., 2015). A further challenge to predicting the response of debris-
81 covered glaciers to climate change is understanding not only the distribution of debris on a
82 glacier surfaces, but also how this varies over time.

83

84 In the Himalaya, 14–18% of the total glacierised area is debris-covered (Kääb et al., 2012)
85 increasing to about 36% in the Everest region of Nepal which contains some of the longest
86 debris-covered glacier tongues in the world are found (Nuimura et al., 2012; Thakuri et al.,
87 2014). Where debris cover on an individual glacier exceeds 40% of the total area mass loss is
88 mainly by terminus stagnation rather than recession (which requires a loss of mass whilst
89 maintaining flow towards the migrating terminus) (Immerzeel et al., 2013; Quincey et al.,
90 2009). Some Himalayan glaciers are over 50% debris covered (Ragettli et al., 2015) and
91 debris is sufficiently thick to reduce rather than enhance ablation (Benn et al., 2012; Bolch et
92 al., 2008; Nicholson and Benn, 2006). In the Everest region of Nepal, 70% of the glacierised
93 area comprises just 40 of 278 glaciers, and these large glaciers are generally debris covered
94 (Thakuri et al., 2014) (Fig. 2a). Glaciers in the Everest region last advanced around 0.5 ka, a
95 period referred to as the Little Ice Age (LIA) but distinct from the European event of the
96 same name (Owen et al., 2009; Richards et al., 2000). Since the LIA, Everest-region glaciers
97 have consistently lost mass (Kääb et al., 2012; Nuimura et al., 2012). Between 1962 and
98 2011, the proportion of Everest region glaciers covered by rock debris has doubled due to
99 ongoing mass loss (Thakuri et al., 2014).

100

101 The future of debris-covered glaciers worldwide is uncertain due to the limitations of our
102 knowledge about the distribution of supraglacial debris and how this evolves over time.
103 Existing models designed for clean-ice glaciers or static assumptions that describe only the
104 present state of the glacier are difficult to extrapolate under a changing climate. Here, we use
105 a novel glacier model that includes the self-consistent development of englacial and
106 supraglacial debris and reproduces the feedbacks among mass-balance, ice-flow and debris
107 transport to investigate how debris modifies the behaviour of a Himalayan glacier in response
108 to climate change. As an example of how many debris-covered Himalayan glaciers respond
109 to climate change, we applied this model to the evolution of Khumbu Glacier in the Everest
110 region of Nepal from the Late Holocene advance (1 ka) to AD2200.

111

112 **2. Khumbu Glacier, Nepal**

113 Khumbu Glacier is a large debris-covered glacier in the Everest region (Fig. 2), with a length
114 of 15.7 km and area of 26.5 km². The Changri Nup and Changri Shar Glaciers were
115 tributaries of Khumbu Glacier during the LIA but have since detached. The equilibrium line
116 altitude (ELA) estimated from mass balance measurements made in 1974 and 1976 is 5600 m

117 (Benn and Lehmkuhl, 2000; Inoue, 1977; Inoue and Yoshida, 1980). More recent studies
118 have placed the ELA of Khumbu Glacier at 5700 m around AD2000 (Bolch et al., 2011)
119 within the icefall that links the accumulation area in the Western Cwm to the glacier tongue
120 (Fig. 3). The ELA may have increased due to atmospheric warming of about 0.9°C between
121 1994 and 2013 (Salerno et al., 2014).

122

123 The active part of Khumbu Glacier (the area exhibiting ice flow) receded towards the base of
124 the icefall since the end of the LIA while the total glacier length remained stable. Feature-
125 tracking observations of velocities define the length of the active glacier as 10.3 km (62% of
126 the LIA glacier length) (Fig. 4). Decaying ice at the terminus beneath debris several metres
127 thick indicates terminus recession of less than 1 km since the LIA (Bajracharya et al., 2014).
128 We divide Khumbu Glacier into two parts based on observations of glacier dynamics; (1) the
129 active glacier where velocities range from 10 m to 70 m a⁻¹ and mass is replenished from the
130 accumulation zone, and (2) the decaying tongue that no longer exhibits ice flow of more than
131 a few m a⁻¹. Similar behaviour is reproduced by our glacier model and observed for many
132 glaciers in the Everest region (Quincey et al., 2009).

133

134 **3. Methods**

135 **3.1 Bed topography**

136 Ice thickness for Khumbu Glacier (Fig. 3) has been measured along seven transects down-
137 glacier from the icefall using radio-echo sounding was 440 ± 20 m at 0.5 km below the icefall
138 close to Everest Base Camp, decreasing to less than 20 m at 4930 m at 2 km up-glacier of the
139 terminus (Gades et al., 2000). Gravity observations gave an ice thickness of 110 m adjacent
140 to Lobuche and 440 m adjacent to Gorak Shep (Moribayashi, 1978). No data exist above the
141 icefall. Ice thickness can be estimated by assuming that glacier ice behaves as a perfectly
142 plastic material such that thickness (h) is determined by surface slope (α) and basal shear
143 stress (τ_b) (Nye, 1952):

144

$$145 \quad h = \lambda * (\tau_b / f * \rho * g * \sin(\alpha))$$

146

147 where ρ is the density of glacier ice, and g is acceleration due to gravity. A shape factor (f)
148 describes the aspect ratio of the cross-section of a valley glacier (Nye, 1952), and a down-
149 glacier thinning factor (λ) describes the long profile:

150

151 $\lambda = 1 - a * x^b$

152

153 where a is a constant accounting for the length of the glacier, x is the flowline distance from
154 the headwall and b describes where thinning first occurs along the flowline. We estimated the
155 thickness of Khumbu Glacier at 35 regularly-spaced transects perpendicular to the central
156 flowline. Glacier topography was described using the ASTER GDEM 2011 Digital Elevation
157 Model (DEM) and the GLIMS outline (GLIMS et al., 2005). Values for τ_b, f and λ were
158 determined by tuning against observations resulting in a mean τ_b value of 150 kPa. Subglacial
159 bedrock topography was described by subtracting the interpolated ice thickness from the
160 DEM, smoothing and resampling to 100-m grid spacing.

161

162 **3.2 Glacier topography**

163 Topographic profiles were measured using a DEM with a 10-m grid spacing generated from
164 ALOS PRISM imagery acquired in 2006 (Fig. 2b). Glacier topography was calculated
165 perpendicular to the central flowline by taking the mean of a 200-m wide moving window.
166 The LIA glacier surface was reconstructed from the elevation of lateral and terminal moraine
167 crests which are preserved below the icefall (Fig. 2a). The LIA moraine crest was defined by
168 taking the maximum of a 300-m wide moving window centred on the moraine, and verified
169 in the field using a Garmin GPSmap 62s handheld unit (Fig 2c). There are no indicators of
170 past glacier topography above the icefall, so model simulations were fitted to the available
171 data from the ablation zone.

172

173 **3.3 Glacier dynamics**

174 Glacier velocities (i.e. surface displacements) were calculated using the panchromatic bands
175 of multi-temporal Landsat Operational Land Imager imagery and a Fourier-based cross-
176 correlation feature tracking method (Luckman et al., 2007). The images were first co-
177 registered with sub-pixel accuracy using large feature (128 x 128 pixels; 1920 m square) and
178 search (256 x 256 pixels; 3840 m square) windows focusing on non-glacierised areas. Glacier
179 displacements were then calculated using finer feature and search windows of 48 x 48 pixels
180 (720 m square) and 64 x 64 pixels (960 m square). Sufficiently robust correlations were
181 accepted on the strength of their signal-to-noise ratio and matches above an extreme
182 threshold of 100 m a^{-1} were removed as blunders. The remaining displacements were
183 converted to annual velocities assuming no seasonal variability in flow. Errors in the velocity
184 data comprise mismatches associated with changing surface features between images, and

185 any inaccuracy in the image co-registration. Given that the glacier is slow-flowing (and thus
186 features do not change rapidly), and that the images were co-registered to a fraction of a
187 pixel, we estimate a maximum theoretical error of one pixel per year (i.e. 15 m). Empirically
188 measured displacements in stationary areas adjacent to the glacier suggest the real error is
189 around half this (i.e. 7–8 m a⁻¹).

190

191 **3.4 Numerical modelling**

192 We used the ice model iSOSIA (Egholm et al., 2011) with a novel description of debris
193 transport that represents the self-consistent development of englacial and supraglacial debris
194 and reproduces the feedbacks amongst mass-balance, ice-flow and debris accumulation.
195 iSOSIA is a higher-order shallow-ice model, which in contrast to standard shallow-ice
196 approximation (SIA) models includes the effects of longitudinal and transverse stress
197 gradients. This makes iSOSIA more accurate than SIA models in settings where flow
198 velocities can vary over short distances, such as in steep and rugged terrains of alpine glaciers
199 (Egholm et al., 2011). Supraglacial debris across Himalayan glaciers is generally decimetres
200 to metres thick and acts to reduce rather than enhance ablation. Moreover, where debris cover
201 is thin in the upper part of the ablation zone of Khumbu Glacier, similar ablation rates are
202 observed for surfaces both with and without debris (Inoue and Yoshida, 1980). Therefore,
203 ablation beneath supraglacial debris was calculated using an exponential function that gave a
204 halving of ablation beneath 0.5 m of debris and assuming minimal ablation beneath a debris
205 layer with a thickness exceeding 1.0 m, in line with values calculated for neighbouring
206 Ngozumpa Glacier (Nicholson and Benn, 2006).

207

208 Transport of debris within and on top of the glacier was modelled as an advection problem
209 assuming that the ice passively transports the debris. Internal ice deformation and basal
210 sliding drive ice flow in iSOSIA and the depth-averaged flow velocity is therefore

211

$$212 \quad \bar{\mathbf{u}} = \bar{\mathbf{u}}_d + \mathbf{u}_b$$

213

214 The velocity due to ice deformation, $\bar{\mathbf{u}}_d$, is approximated as a tenth-order polynomial
215 function of ice thickness with coefficients that depend on ice surface slope and bed slope as
216 well as longitudinal stress and stress gradients (Egholm et al., 2011).

217

218 Basal sliding is assumed to scale with the basal shear stress according to the following
219 empirical sliding model (Budd and Keage, 1979):

220

$$u_b = \frac{B_s \tau_b^m}{N_e}$$

221

222 where N_e is the effective pressure at the bed, and $B_s = 4 \times 10^{-4} \text{ m a}^{-1} \text{ Pa}^{-1}$ and $m = 2$ are
223 constants. The basal shear stress is the bed-parallel stress vector at the base of the ice, which
224 is computed by projecting the full stress tensor at the base of the ice onto the glacier bed. The
225 shear stress is therefore sensitive to ice thickness, ice surface slope, local ice velocity
226 variation, as well as bed slope orientation (Egholm et al., 2011). The effective pressure was
227 assumed to be 20% of the ice overburden pressure. This standard approach (e.g.
228 Bindschadler, 1983; Braun et al., 1999; Egholm et al., 2012; Kessler et al., 2008) to
229 modelling basal sliding in alpine glaciers ignores the detailed distribution of water pressure as
230 well as ice-bed cavitation, which are both elements that we have no means of calibrating
231 empirically for Khumbu Glacier. We note that the distribution of sliding is thus considered
232 uncertain, also because the two sliding parameters B_s and m are difficult to constrain
233 empirically; according to Budd et al. (1979), m should vary between 1 and 3. On the other
234 hand, variations in sliding rate do not significantly influence our modelling results as long as
235 ice, and thus also englacial debris, is transported from the accumulation zone to the ablation
236 zone by either basal sliding or internal ice creep.

237

238 The debris concentration, c , at any point within the ice was updated through time, t , using the
239 following equation for debris advection:

240

$$\frac{\partial c}{\partial t} = -\nabla \cdot \{c\mathbf{u}\}$$

241

242 where \mathbf{u} is the three-dimensional ice velocity vector. The equation is based on the assumption
243 that debris is transported passively with the ice, and hence that any change in debris
244 concentration in a point is controlled by the flux of debris and ice to and from that point. For
245 example, at the surface in the ablation zone, debris concentration generally increases over
246 time because melting of ice causes the total influx of ice by flow to be positive. Debris may
247 also accumulate along the base of the ice, because basal melting, controlled by the excess

248 heat at the glacier bed (Egholm et al., 2012), drives ice towards the bed. However, most
249 debris follows a concave path from the ice surface in the accumulation zone, down to some
250 depth within the glacier, and then back to the glacier surface in the ablation zone. As a
251 boundary condition to the above equation, we assumed that debris is fed to the surface of the
252 glacier in the accumulation zone and that $c_{sa}=0.001$ (the concentration of debris at the ice
253 surface) is constant across the accumulation area. Debris in the high parts of Khumbu Glacier
254 is likely transported to the glacier by avalanches, and the high energy of the avalanches can
255 spread snow and debris across wide areas of the glacier surface. Without detailed knowledge
256 of the distribution and frequency of avalanches, we used a constant surface debris
257 concentration in the accumulation zone as the simplest possible boundary condition. We note,
258 however, that because localised quantities of debris in the accumulation zone have a tendency
259 of diffusing during transport in the glacier, the wide-spread distribution of debris near the
260 terminus of the glacier is largely insensitive to variations in the debris input distribution of
261 the accumulation zone. The order of c_{sa} was roughly estimated by considering the total area
262 of the surrounding ice-free hillslopes and assuming that the mean erosion rate is about 1 mm
263 a^{-1} . The total hillslope sediment production was then uniformly distributed across the area of
264 the ice accumulation zone. We note that sediment production from these hillslopes varies
265 through time in response to variations in rock uplift and climate change (Scherler et al.,
266 2011). However, our model experiments focus on the spatial patterns of debris distribution
267 and disregard any temporal evolution of debris production. The rate of debris input used here
268 should consequently only be regarded as a first-order estimate.

269

270 Debris transport was modelled using a three-dimensional grid. iSOSIA is a depth-integrated
271 2-D model, but for the purpose of tracking the three-dimensional debris transport, the
272 thickness of the ice was divided into 20 layers representing the vertical dimension of the 3-D
273 grid structure. iSOSIA only computes depth-averaged velocity components. However, to
274 capture velocity variations at depth within the ice we reconstructed in every time step the full
275 three-dimensional velocity field of the glacier. The vertical variation of velocity components
276 was derived from the assumption that the horizontal ice velocity caused by viscous ice
277 deformation decays as a fourth-order polynomial down through the ice, which is valid for
278 laminar flow of ice with a stress exponent of 3 (Van der Veen, 2013; p. 77). We calibrated
279 the fourth-order polynomial to yield the correct depth-averaged velocity:

280

$$\mathbf{u}(z) = \frac{5}{4} \left[1 - \left(\frac{z}{h} \right)^4 \right] \bar{\mathbf{u}} + \mathbf{u}_b$$

281

282 where $\bar{\mathbf{u}}$ is the depth-averaged horizontal velocity and \mathbf{u}_b is basal sliding velocity. z is burial
 283 depth below the ice surface and h is ice thickness. The internal vertical component of the ice
 284 velocity, u_v , was scaled linearly with accumulation/ablation at the surface (\dot{m}_s) and melting at
 285 the glacier bed (\dot{m}_b):

$$u_v(z) = \frac{h-z}{h} \dot{m}_s + \frac{z}{h} \dot{m}_b$$

286

287 Melting at the bed is computed from the heat available at the bed:

288

$$\dot{m}_b = \frac{q_b + \mathbf{u}_b \cdot \boldsymbol{\tau}_b - q_c}{\rho L}$$

289

290 where $q_b = 0.045 \text{ W m}^{-2}$ is the heat flux from the underlying crust; $\mathbf{u}_b \cdot \boldsymbol{\tau}_b$ is the heat
 291 produced at the bed by friction due to basal sliding; $\rho = 980 \text{ kg m}^{-3}$ is the density of glacier
 292 ice and $L = 334 \text{ kJ kg}^{-1}$ is the latent heat of ice. q_c is the heat transported away from the
 293 glacier bed by heat conduction in the overlying ice. It is estimated from the thermal gradient
 294 at the glacier bed:

$$q_c = -k \frac{\partial T}{\partial z}$$

295

296 and the thermal conductivity of ice, $k = 2.4 \text{ W m}^{-1} \text{ K}^{-1}$. The temperature field within the ice
 297 was computed using the three-dimensional semi-implicit algorithm described by Egholm et
 298 al. (2012). The rates of basal melting were typically limited to the order of 0.01 m a^{-1} , which
 299 is 1–2 orders of magnitude smaller than the rates of surface ablation.

300

301 The advection equation was integrated through time using explicit forward time stepping in
 302 combination with a three-dimensional upwind finite-difference scheme. The size of the time
 303 step was restricted by the Courant-Friedrichs-Lewy condition:

304

$$\Delta t \leq \frac{1}{2} \frac{\Delta_{\min}}{u_{\max}}$$

305

306 where Δ_{\min} is the smallest cell-dimension (along the x, y and z axes), and u_{\max} is the
307 maximum ice velocity component. Time steps were by this condition restricted to 1–5 model
308 days. The iSOSIA and debris transport algorithms were parallelised using OpenMP
309 (Chapman et al., 2007), and run on 12-core CPU servers. Each simulation typically lasted 8–
310 12 hours.

311

312 **3.5 Experimental design**

313 Simulations were made for the catchment upstream from the base of the LIA terminal
314 moraine. The DEM was constructed from data collected between AD2001 and AD2010 so
315 we place the present day at the start of this window as AD2000. Mass balance was calculated
316 assuming linear temperature-dependent rates of accumulation and ablation following those
317 measured in 1974 and 1976 (Benn and Lehmkuhl, 2000; Inoue, 1977; Inoue and Yoshida,
318 1980). An atmospheric lapse rate of $-0.004^{\circ}\text{C m}^{-1}$ was calculated by linear regression of
319 MODIS Terra land surface temperatures (24/02/00–31/12/06) (NASA, 2001) for the Central
320 Himalayan region (Fig. 4). Glacier advance and recession were simulated by varying ELA
321 over time. Extreme topography results in the majority of glacier mass gain by avalanching
322 rather than direct snowfall, and the avalanche contribution to mass balance was estimated as
323 75% (Benn and Lehmkuhl, 2000). We removed snow and ice from slopes exceeding 28° and
324 redistributed the total volume uniformly on the accumulation area of the glacier surface. The
325 critical slope of 28° was selected because this threshold is low enough to prevent ice
326 accumulation on slopes that are clearly ice-free today, but high enough to not limit ice
327 accumulation on the glacier surface.

328

329 *3.5.1 Initial Late Holocene simulation*

330 Prior to the LIA (0.5 ka), Khumbu Glacier had a slightly greater extent during the Late
331 Holocene (~1 ka) and is likely to have reached the LIA extent by the formation of high
332 moraines that enclosed the glacier and drove the ice mass to thicken (Owen et al., 2009) (Fig.
333 2a). As a starting point for our transient simulations, we reconstructed the Late Holocene
334 glacier from an ice-free domain using an ELA of 5325 m over a 5000-year period. This
335 simulation was optimised to result in a steady-state glacier that provided a good fit to the Late
336 Holocene moraines (Fig. 5). A minor recession, inferred from the position of the LIA
337 moraines inside the Late Holocene moraines, was imposed after the Late Holocene advance
338 equivalent to an increase in ELA of 50 m to 5375 m over 500 years, and supraglacial debris
339 thickened due to the reduction in debris export as glacier velocities decreased.

340

341 *3.5.2 Simulation from the LIA to the present day*

342 To simulate the LIA advance, maximum and recession, the ELA was increased from 5375 m
343 to 6000 m over 500 years. The distribution of englacial and supraglacial debris simulated for
344 the Late Holocene was used as a starting point for the LIA simulation. A range of present-day
345 ELA values (Fig. 3) was tested by comparing the simulated ice volume with observed glacier
346 topography; the best fit to the present-day ice thickness was an ELA of 6000 m. This places
347 the ELA of Khumbu Glacier at the top of the icefall rather than in the lower half as indicated
348 by recent measurements (Fig. 3). The simulated ice thicknesses were optimised to the LIA
349 moraines and the present-day glacier. This simulation ran to steady state to indicate how the
350 glacier would continue to evolve without any further change in climate.

351

352 *3.5.3 Simulation from the present day to AD2200*

353 Simulation of glacier change from the present day until AD2200 continued from the present-
354 day simulation where the glacier was out of balance with climate. We imposed a linear rise in
355 ELA over 100 years from AD2000 to AD2100 equivalent to predicted minimum and
356 maximum warming relative to 1986–2005 by 2080–2099 of 0.9°C (increase in ELA of 225 m
357 assuming an atmospheric lapse rate of $-0.004^{\circ}\text{C m}^{-1}$) and 1.6°C (increase in ELA of 400 m),
358 in line with IPCC model ensemble predictions (CMIP5 RCP 4.5 scenario) (Collins et al.,
359 2013). The simulation continued until AD2200 without any further change in climate.

360

361 **3.6 Mass balance sensitivity**

362 We tested the sensitivity of Khumbu Glacier to mass balance parameter values through the
363 LIA to the present day to assess the impact of these uncertainties on our projections for
364 AD2100. A range of present-day ELA values equivalent to a change in ELA of 150 m
365 (equivalent to $\pm 0.3^{\circ}\text{C}$) produced a difference in glacier volume of $0.3 \times 10^9 \text{ m}^3$ (14% of
366 present-day volume) with no change in glacier length beyond the cell size of the model
367 domain (100 m). Lapse rates between $-0.003^{\circ}\text{C m}^{-1}$ and $-0.006^{\circ}\text{C m}^{-1}$ and no change in ELA
368 produced a difference in glacier volume of $0.4 \times 10^9 \text{ m}^3$ (19%) with no change in length.
369 Maintaining the relationship with temperature between rates of accumulation and ablation
370 whilst varying maximum values by $\pm 10\%$ produced a difference in glacier volume of $4.0 \times$
371 10^6 m^3 (0.2%) with no change in length.

372

373 **3.7 Comparison with simulations that do not transport debris**

374 To verify the effect of supraglacial debris on glacier change, the LIA to the present day was
375 simulated: (1) without the modification of ablation beneath the debris layer, that is, assuming
376 a clean rather than debris-covered surface, and (2) with maximum ablation reduced by 50%
377 (as in Section 3.6) to compare the impact of a uniform reduction in ablation, as sometimes
378 used when clean-ice glacier models are applied to debris-covered glaciers (Fig. 6). Mass loss
379 from the clean-ice glacier greatly exceeded that from the debris-covered glacier, resulting in a
380 glacier with 16% of the present-day volume and a 6.7 km reduction in length compared to the
381 dynamic debris-covered glacier simulated for the same period. A reduction in ablation of
382 50% resulted in dramatic mass loss to 27% of present-day volume and a 4.4 km reduction in
383 length compared to the dynamic debris-covered glacier simulated for the same period. Our
384 results highlight that the change in terminus position of debris-covered glaciers in response to
385 climate change is slower than for clean-ice glaciers. Similar behaviour is observed using 1-D
386 modelling (Banerjee and Shankar, 2013) and remote-sensing observations (Kääb et al., 2012).
387 Therefore, models developed for clean-ice glaciers using a uniform reduction in ablation do
388 not reliably simulate the evolution of debris-covered glaciers.

389

390 **4. Results**

391 **4.1 Glacier morphology**

392 Reconstruction of Khumbu Glacier using moraine crests showed that, since the LIA, glacier
393 area has decreased from 28.1 km² to 26.5 km² (a reduction of 6%). If the glacier is considered
394 only in terms of active ice, then area has declined to 20.3 km² (a reduction of 28%) (Fig. 4).
395 These values exclude the change in area attributed to the dislocation of the Changri Nup and
396 Changri Shar tributaries (Fig. 4). The volume of the active glacier is 1.7 x 10⁹ m³ (50% of the
397 LIA volume). The lack of dynamic behaviour in the tongue can be observed from the relict
398 landslide material on the true left of the glacier that has not moved between 2003 and 2014
399 (Fig. 2a). Comparison of swath topographic profiles of the glacier surface and the LIA lateral
400 moraine crests (Fig. 2c) indicate mean surface lowering across the debris-covered tongue of
401 25.5 ± 10.6 m, or 0.05 ± 0.02 m a⁻¹ since the LIA. Glacier volume decreased from 3.4 x 10⁹
402 m³ to 2.3 x 10⁹ m³ (66% of the LIA volume), a loss of 1.2 x 10⁹ m³ and equivalent to 2.3 x 10⁶
403 m³ a⁻¹. Mean surface lowering observed between 1970 and 2007 across the ablation area was
404 13.9 ± 2.5 m (Bolch et al., 2011) suggesting that rates of mass loss have accelerated over the
405 last 50 years compared to the last 500 years, and consistent with the observed decrease in the
406 active glacier area (Quincey et al., 2009).

407

408 **4.2 Glacier modelling**

409 The initial simulation representing the Late Holocene maximum was computed from an ice-
410 free domain using an ELA of 5325 m (-2.7°C relative to the present day). Debris
411 accumulated at the ice margins rather than on the glacier surface to form lateral moraines
412 (Fig. 5).

413

414 *4.2.1 The Little Ice Age to the present day*

415 Khumbu Glacier initially advanced during the LIA for 150 years despite the rise in ELA as
416 decreasing velocity in the tongue (Table 1) resulted in thickening supraglacial debris (Fig. 7a
417 and 7c). The large LIA moraines suggest that debris export from the glacier to the ice
418 margins declined because the glacier was impounded following the construction of these
419 moraines. This simulation reproduced this observation, and resulted in the formation of a
420 thick debris layer (Fig. 7d). The simulated LIA glacier surface provided a good fit to the LIA
421 moraine crests (Fig. 7e). The simulated glacier then lost mass by surface lowering
422 accompanied by minor terminus recession, despite the reduction in ablation beneath
423 supraglacial debris (Fig 7b and Table 1). Simulated present-day ice thicknesses were in good
424 agreement with the observed glacier surface (Fig. 7f). The maximum simulated present-day
425 ice thickness was 345 m. The mean flowline ice thickness was 168 m for the whole glacier,
426 88 m in the accumulation area and 210 m for the debris-covered tongue. Simulated velocities
427 (Table 1 and Fig. 8) reproduced the pattern and absolute values measured from remote-
428 sensing observations (Fig. 4).

429

430 After the LIA maximum, simulated ice thickness declined most rapidly for the first 200 years
431 of warming followed by slightly less rapid mass loss for the following 300 years. Mean ice
432 thickness across the entire glacier decreased by 0.01 m a^{-1} , and surface lowering was greatest
433 between 1.8 km and 3.2 km upglacier from the terminal moraine. The active glacier shrunk to
434 the observed active ice extent but did not reach steady state. The response time to reach
435 equilibrium with the present-day ELA was 1150 years, 500 years longer than the time elapsed
436 between the LIA maximum and the present day, indicating that Khumbu Glacier is out of
437 balance with climate. According to our model, Khumbu Glacier will continue to respond to
438 post-LIA warming until about AD2500 and will lose a further $0.4 \times 10^9\text{ km}^3$ (18%) of ice
439 without any further change in climate.

440

441 *4.2.2 Present day to AD2200*

442 To predict glacier volume at AD2100 and AD2200, we imposed a linear rise in ELA from the
443 present day following IPCC minimum and maximum warming scenarios for AD2100
444 (Collins et al., 2013). These simulations were driven by an increase in ELA of 225 m to 6225
445 m (equivalent to warming of 0.9°C) and 400 m to 6400 m (equivalent to warming of 1.6°C)
446 over 100 years, and without a further change in climate until AD2200. Warming of 0.9°C by
447 AD2100 will result in mass loss of $0.17 \times 10^9 \text{ km}^3$ and warming of 1.6°C will result in mass
448 loss of $0.21 \times 10^9 \text{ km}^3$ (Fig. 9a and 9c), a decrease in glacier volume of between 8% and 10%
449 (Table 1). Simulated mass loss will be greatest close to the base of the icefall, where ablation
450 exceeds that occurring down-glacier beneath thicker supraglacial debris and up-glacier in the
451 Western Cwm. Supraglacial debris will expand and thicken across the glacier tongue,
452 particularly between the confluence with Changri Nup Glacier and the icefall (Fig. 9e
453 compared to Fig. 7d), reaching 1.5 m thickness at the base of the icefall. The debris-covered
454 tongue could physically detach from the base of the icefall within 150 years and persist *in*
455 *situ* while the active glacier recedes (Fig. 9b and 9d). After the physical detachment of the
456 debris-covered tongue, supraglacial debris will develop on the tongue of the active glacier
457 near the upper part of the icefall (Fig. 9f).

458

459 **5. Discussion**

460 **5.1 Validation of model simulations**

461 The present-day simulation was validated by comparison with observations of velocities,
462 mean surface elevation change and geodetic mass balance derived from satellite imagery. The
463 simulated present-day maximum flowline velocity was 59 m a^{-1} and the mean was 9 m a^{-1}
464 (Fig. 8a and 8b). The mean simulated velocity above the base of the icefall was 24 m a^{-1} , and
465 the mean velocity of the debris-covered tongue below the icefall was 2 m a^{-1} . These
466 simulated velocities are in good agreement with those measured using feature tracking (Fig.
467 4), which give a maximum flowline velocity of 67 m a^{-1} and a mean of 16 m a^{-1} . The mean
468 measured velocity above the base of the icefall was 25 m a^{-1} , and the mean velocity of the
469 debris-covered tongue was 9 m a^{-1} . The measured velocity of the tongue is within the
470 uncertainty of the feature tracking method due to the 15-m grid spacing of the imagery used,
471 and the actual displacement could be less than 9 m a^{-1} .

472

473 The decrease in the elevation of the simulated glacier surface over the 40 years prior to the
474 present day was close to zero at the terminus and increased to 8–10 m in the upper part of the
475 ablation area. The simulated surface lowering shows good agreement both in terms of the

476 absolute values and the distribution of surface lowering to that observed for a similar period
477 (1970–2007) which gave an elevation difference of -13.9 ± 2.5 m across the ablation area
478 (Bolch et al., 2011). Integrated mass balance for the simulated present-day glacier was -0.22
479 m w.e. a^{-1} , slightly less negative than but not dissimilar to geodetic mass balance values
480 estimated between 1970 and 2007 as of -0.27 ± 0.08 m w.e. a^{-1} (Bolch et al., 2011) and
481 between 1992 and 2008 as -0.45 ± 0.52 m w.e. a^{-1} (Nuimura et al., 2012).

482

483 ***5.2 Equilibrium Line Altitude***

484 The ELA of Khumbu Glacier could be placed in a range from 5200 m to 5600 m assuming
485 that the integrated mass balance is zero (Benn and Lehmkuhl, 2000). However, methods for
486 calculating ELA such as the accumulation-area ratio are difficult to apply to avalanche-fed,
487 debris-covered glaciers for which values appear to be lower (around 0.1–0.4) than those for
488 clean-ice glaciers (0.5–0.6) (Anderson, 2000). Snowline altitude is not a reliable indicator of
489 ELA in high mountain environments, because avalanching, debris cover and high relief affect
490 mass balance such that ELA may differ by several hundred metres from the mean snowline
491 (Benn and Lehmkuhl, 2000). Simulations using the lower estimated ELAs and assuming a net
492 mass balance of zero produced a glacier equivalent to the Late Holocene extent. Simulations
493 optimised to the present-day glacier indicate that ELA is probably about 5800–6000 m (Fig.
494 3b).

495

496 **5.3 Sources of uncertainty associated with modelling debris-covered glaciers**

497 We used a simple approach to represent the relationship between climate and glacier mass
498 balance to avoid introducing additional uncertainties by making assumptions about the
499 response of meteorological parameters such as monsoon intensity to climate change.
500 Therefore, our results indicate the sensitivity of a debris-covered Himalayan glacier to
501 climate change over the Late Holocene period (1 ka to present). Although iSOSIA captures
502 the dynamics of mountain glaciers, the interaction of high topography with atmospheric
503 circulation systems will affect mass balance (Salerno et al., 2014). Future studies could use
504 downscaled climate model outputs or energy balance modelling to better capture these
505 variables. However, mass balance and meteorological data to support these approaches are
506 scarce for the majority of Himalayan glaciers.

507

508 Differences in the estimated and simulated volume of the present-day glaciers were due to
509 differences in simulated glacier extents. Simulations were designed to give a best fit to

510 Khumbu Glacier and produced less extensive ice than observed for Changri Nup and Changri
511 Shar Glaciers (Fig. 7a and b). Sensitivity experiments showed that a range of mass balance
512 values and lapse rates had little impact on these tributaries, suggesting that the mass balance
513 of Khumbu Glacier does not precisely represent that of the tributaries. This mismatch could
514 be due to the differences in hypsometry between glaciers and model calibration to the
515 extreme altitudes in the Western Cwm.

516

517 There are no measurements with which to constrain ice thickness in the Western Cwm, so our
518 estimate of ice thickness is based solely on the slope of the glacier surface derived from the
519 DEM and tuning of values for basal shear stress (τ_b) and glacier shape to match geophysical
520 observations (Fig. 3). The τ_b values initially used to determine bed topography are within the
521 range simulated using iSOSIA (Fig. 8a and 8b) suggesting that our estimate of bed
522 topography is appropriate. However, calculation of bed topography beneath glaciers and ice
523 sheets remains an outstanding challenge in glaciology, and one that is difficult to resolve in
524 the absence of data describing the basal properties of the glacier.

525

526 The addition of debris to the glacier surface by rock avalanching from the surrounding
527 hillslopes is not represented in our glacier model, but previous studies have demonstrated that
528 large rock avalanches can perturb the terminus position of mountain glaciers (e.g. Menounos
529 et al., 2013; Vacco et al., 2010). Sub-debris ablation is modified by the physical properties of
530 the debris layer, particularly variations in water content and grain size (Collier et al., 2014).
531 Exposed ice cliffs can enhance ablation locally on debris-covered glaciers; at Miage Glacier
532 in the European Alps ice cliffs occupy 1% of the debris-covered area and account for 7% of
533 total ablation (Reid and Brock, 2014). Previous work has hypothesised that ice cliff ablation
534 may be responsible for the comparable rates of mass loss observed for debris-covered and
535 clean-ice glaciers in the Himalaya and Karakoram (Gardelle et al., 2012; Kääb et al., 2012).
536 Mapping the area of debris-covered surfaces occupied by ice cliffs and increasing ablation
537 accordingly could refine our predictions of future glacier change. This would require more
538 detailed topographic data than the 30-m DEM and parameterisation of the processes by which
539 ice cliffs form and decay. As we do not incorporate ice cliffs or supraglacial ponds into our
540 modelling, and as these features are likely to become more widespread as surface lowering
541 continues, our estimates of mass loss from the present day to AD2200 are likely to be
542 cautious.

543

544 **6. Conclusions**

545 Predictions of debris-covered glacier change based either on assumptions about clean-ice
546 glaciers or including static adjustments of ablation rates do not capture the feedbacks
547 amongst mass balance, ice dynamics and debris transport that govern the behaviour of these
548 glaciers, and are unlikely to give reliable results. We present the first dynamic model of the
549 evolution of a debris-covered glacier and demonstrate that including these important
550 feedbacks simulated glacier mass loss by surface lowering rather than terminus recession, and
551 represents the observed response to climate change of debris-covered glaciers. Models such
552 as this that represent the transient processes governing the behaviour of debris-covered
553 glaciers, supported by detailed direct and remotely-sensed observations, are needed to
554 accurately predict glacier change in mountain ranges such as the Himalaya.

555

556 The development of supraglacial debris on Khumbu Glacier in Nepal promoted a reversed
557 mass balance profile across the ablation area, resulting in greatest mass loss after the Little
558 Ice Age (0.5 ka) where debris was absent close to the icefall and least mass loss on the
559 debris-covered tongue. The reduction in ablation across the debris-covered section of the
560 glacier resulted in reduced ice flow and debris export. Khumbu Glacier extends to a lower
561 altitude (4870 m a.s.l. compared to 5160 m a.s.l.) and greater length (15.7 km compared to
562 10.3 km) than would be possible without supraglacial debris. We predict a loss of ice
563 equivalent to 8–10% of the present-day glacier volume by AD2100 with only minor change
564 in glacier area and length, and physical detachment of the debris-covered tongue from the
565 upper active part of the glacier before AD2200. Regional atmospheric warming is likely to
566 result in a similar response from other debris-covered glaciers in the Everest region over the
567 same period.

568

569

570

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578

579

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724 **Table caption**

725 Table 1. Simulated glacier volume, ice thickness and velocity during the Little Ice Age (LIA),

726 the present day and predicted for AD2100 under a maximum IPCC warming scenario of
727 1.6°C.

728

729 **Figure captions**

730 Figure 1. Conceptual model of the development of a debris-covered Himalayan glacier; (a) in
731 balance with climate, and (b) during net mass loss under a warming climate.

732

733 Figure 2. Debris-covered Khumbu Glacier. (a) Photograph of the ablation area of Khumbu
734 Glacier looking downglacier from Kala Pathar showing the elevation difference between the
735 Little Ice Age lateral moraine crests and the glacier surface and relict landslide material that
736 has remained *in situ* from at least 2003 to 2014. (b) Long profile of 100-m mean swath
737 topography of Khumbu Glacier, and (c) long profile of 100-m mean swath topography of
738 Khumbu Glacier below the confluence with the Changri Nup tributary showing the elevation
739 difference between the glacier surface and the Little Ice Age lateral moraine crest due to mass
740 loss by surface lowering. The lowest point of the terminal moraine is 4670 m a.s.l.

741

742 Figure 3. (a) Present-day ice thickness of Khumbu Glacier estimated from a DEM
743 constructed for the year 2011 as described in Section 3.1, tuned with available ice thickness
744 measurements from radio-echo sounding (Gades et al., 2000) and gravity observations
745 (Moribayashi, 1978), draped over a shaded-relief map of the estimated bedrock topography
746 used to describe the model domain (EBC = Everest Base Camp). (b) Long profile of Khumbu
747 Glacier showing the range of the ELA based on morphometric calculations and
748 measurements of mass balance from previous studies (Benn and Lehmkuhl, 2000; Bolch et
749 al., 2011; Inoue, 1977; Inoue and Yoshida, 1980), the ELA simulated for the present day
750 glacier, and the range of ELA used to represent IPCC scenarios for AD2100.

751

752 Figure 4. Present-day velocities of Khumbu Glacier calculated using feature tracking of the
753 panchromatic bands of multi-temporal Landsat Operational Land Imager (OLI) imagery. The
754 underlying image was acquired by the Landsat OLI on 4th May 2013 and the white areas
755 denote areas with no data. The glacier outline is defined according to the Randolph Glacier
756 Inventory (GLIMS et al., 2005). Note the termination of the measured active ice (i.e. above
757 the uncertainty in the method) is 5.4 km upglacier from the terminus. The location of this
758 figure and the extent of the Central Himalayan region (red dashed line) are shown in the inset
759 map.

760

761 Figure 5. Initial steady-state simulation of Khumbu Glacier during the Late Holocene
762 advance (1 ka) used as a starting point for the Little Ice Age simulations, showing (a) ice
763 thickness, (b) debris thickness, (c) mass balance (in metres of water equivalent per year), and
764 (d) velocities.

765

766 Figure 6. Simulations of Khumbu Glacier as a clean-ice rather than debris-covered glacier.
767 (a) present day ice thickness simulated without a supraglacial debris layer, and (b) present-
768 day ice thickness simulated without ablation beneath a debris layer with a 50% reduction in
769 ablation.

770

771 Figure 7. Simulations of Khumbu Glacier during the Little Ice Age (LIA; 0.5 ka) and present
772 day. Results from the iSOSIA model for; (a) ice thickness during the LIA, (b) ice thickness at
773 the present day, and simulated supraglacial debris (c) during the LIA, and (d) at the present
774 day. The fit between the simulated glaciers, the LIA lateral moraine crest, and the present day
775 glacier surface are shown for (e) the LIA and (f) the present day.

776

777 Figure 8. Simulated velocities for Khumbu Glacier (a) during the Little Ice Age, and (b) at
778 the present day [Note log scale for velocity], and simulated basal shear stress (τ_b) for Khumbu
779 Glacier (c) during the Little Ice Age, and (d) at the present day.

780

781 Figure 9. Simulations of Khumbu Glacier in AD2100 and AD2200. (a) Simulated ice
782 thickness in AD2100 under the maximum IPCC CMIP5 RCP 4.5 warming scenario
783 equivalent to an increase in temperature of 1.6°C from the present day, and (b) assuming the
784 same warming scenario from the present day, the ice thickness in AD2200 after the active
785 glacier detached from the debris-covered tongue. Differences in ice thickness from the
786 present day simulation in (c) AD2100 and (d) AD2200 [Note the different scales for
787 difference in ice thickness]. Debris thickness in (e) AD2100 and (f) AD2200. The Little Ice
788 Age lateral moraine crest, present day glacier surface and simulated glacier surface for (g) the
789 AD2100 and (h) the AD2200 simulations.