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

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Abstract

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

1. Introduction

Glaciers in the Himalaya are rapidly losing mass (Bolch et al., 2012). However, data describing past and present glacier volumes are scarce, resulting in varying predictions of future glacier volumes (Cogley, 2011; Kääb et al., 2012). To improve predictions of how Himalayan glaciers will decline through the 21st Century and the impact on Asian water resources, we need to quantify the processes that drive glacier change (e.g. Immerzeel et al., 2013; Pellicciotti et al., 2015; Ragettli et al., 2015; Shea et al., 2015). Changes in glacier volume are driven by climate variations, particularly changes in atmospheric temperature and precipitation amount, and modified by ice flow (Bolch et al., 2012; Kääb et al., 2012). The lower portions of clean-ice glaciers lose mass rapidly during periods of warming. As glaciers recede to higher elevations, a new equilibrium state between this smaller glacier and the warmer climate may be established. Numerical modelling is required to understand the processes that cause glaciers to change because we cannot rely simply on the extrapolation of present-day trends. Previous studies of Himalayan glaciers using models designed for clean-ice glaciers resulted in predictions of widespread rapid deglaciation (e.g. Shea et al., 2015). However, debris-covered glaciers respond differently to warming because debris insulates the ice surface (Jouvet et al., 2011; Kirkbride and Deline, 2013; Pellicciotti et al., 2015; Østrem, 1959). Debris-covered glaciers lose most mass by surface lowering rather than terminus recession (Hambrey et al., 2008). Debris-covered glaciers can persist at lower elevations than would be possible for an equivalent clean-ice glacier even when dramatically out of equilibrium with climate (Anderson, 2000; Benn et al., 2012). As glaciers lose mass preferentially from areas of clean ice and mass loss results in the melt-out of englacial debris, debris coverage will increase as a glacier shrinks (Bolch et al., 2008; Kirkbride and Deline, 2013; Thakuri et al., 2014). Therefore, predicting the future of the Himalayan cryosphere and water resources depends on understanding the impacts of climate change on debris-covered glaciers.

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

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

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

2. Khumbu Glacier, Nepal

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

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

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

3. Methods

3.1 Bed topography

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

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

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

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

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

3.2 Glacier topography

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

3.3 Glacier dynamics

Glacier velocities (i.e. surface displacements) were calculated using the panchromatic bands of multi-temporal Landsat Operational Land Imager imagery and a Fourier-based cross-correlation feature tracking method (Luckman et al., 2007). The images were first co-registered with sub-pixel accuracy using large feature (128 x 128 pixels; 1920 m square) and search (256 x 256 pixels; 3840 m square) windows focusing on non-glacierised areas. Glacier displacements were then calculated using finer feature and search windows of 48 x 48 pixels (720 m square) and 64 x 64 pixels (960 m square). Sufficiently robust correlations were accepted on the strength of their signal-to-noise ratio and matches above an extreme threshold of 100 m a^{-1} were removed as blunders. The remaining displacements were converted to annual velocities assuming no seasonal variability in flow. Errors in the velocity data comprise mismatches associated with changing surface features between images, and any inaccuracy in the image co-registration. Given that the glacier is slow-flowing (and thus features do not change rapidly), and that the images were co-registered to a fraction of a pixel, we estimate a maximum theoretical error of one pixel per year (i.e. 15 m). Empirically measured displacements in stationary areas adjacent to the glacier suggest the real error is around half this (i.e. 7–8 m a^{-1}).

3.4 Numerical modelling

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

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

$$\bar{\boldsymbol{u}} = \bar{\boldsymbol{u}}_d + \boldsymbol{u}_b$$

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

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

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

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

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

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

where $\boldsymbol{u}$ is the three-dimensional ice velocity vector. The equation is based on the assumption that debris is transported passively with the ice, and hence that any change in debris concentration in a point is controlled by the flux of debris and ice to and from that point. For example, at the surface in the ablation zone, debris concentration generally increases over time because melting of ice causes the total influx of ice by flow to be positive. Debris may also accumulate along the base of the ice, because basal melting, controlled by the excess heat at the glacier bed (Egholm et al., 2012), drives ice towards the bed. However, most debris follows a concave path from the ice surface in the accumulation zone, down to some depth within the glacier, and then back to the glacier surface in the ablation zone. As a boundary condition to the above equation, we assumed that debris is fed to the surface of the glacier in the accumulation zone and that c_{sa}=0.001 (the concentration of debris at the ice surface) is constant across the accumulation area. Debris in the high parts of Khumbu Glacier is likely transported to the glacier by avalanches, and the high energy of the avalanches can spread snow and debris across wide areas of the glacier surface. Without detailed knowledge of the distribution and frequency of avalanches, we used a constant surface debris concentration in the accumulation zone as the simplest possible boundary condition. We note, however, that because localised quantities of debris in the accumulation zone have a tendency of diffusing during transport in the glacier, the wide-spread distribution of debris near the terminus of the glacier is largely insensitive to variations in the debris input distribution of the accumulation zone. The order of c_{sa} was roughly estimated by considering the total area of the surrounding ice-free hillslopes and assuming that the mean erosion rate is about 1 mm a^{-1}. The total hillslope sediment production was then uniformly distributed across the area of the ice accumulation zone. We note that sediment production from these hillslopes varies through time in response to variations in rock uplift and climate change (Scherler et al., 2011). However, our model experiments focus on the spatial patterns of debris distribution and disregard any temporal evolution of debris production. The rate of debris input used here should consequently only be regarded as a first-order estimate.

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

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

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

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

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

$$\dot{m}_b = \frac{q_b + u_b \cdot \tau_b - q_c}{\rho L}$$

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

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

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

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

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

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

3.5 Experimental design

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

3.5.1 Initial Late Holocene simulation

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

3.5.2 Simulation from the LIA to the present day

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

3.5.3 Simulation from the present day to AD2200

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

3.6 Mass balance sensitivity

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

3.7 Comparison with simulations that do not transport debris

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

4. Results

4.1 Glacier morphology

Reconstruction of Khumbu Glacier using moraine crests showed that, since the LIA, glacier area has decreased from 28.1 km^2 to 26.5 km^2 (a reduction of 6%). If the glacier is considered only in terms of active ice, then area has declined to 20.3 km^2 (a reduction of 28%) (Fig. 4). These values exclude the change in area attributed to the dislocation of the Changri Nup and Changri Shar tributaries (Fig. 4). The volume of the active glacier is 1.7×10^9 m^3 (50% of the LIA volume). The lack of dynamic behaviour in the tongue can be observed from the relict landslide material on the true left of the glacier that has not moved between 2003 and 2014 (Fig. 2a). Comparison of swath topographic profiles of the glacier surface and the LIA lateral moraine crests (Fig. 2c) indicate mean surface lowering across the debris-covered tongue of 25.5 ± 10.6 m, or 0.05 ± 0.02 m a^{-1} since the LIA. Glacier volume decreased from 3.4×10^9 m^3 to 2.3×10^9 m^3 (66% of the LIA volume), a loss of 1.2×10^9 m^3 and equivalent to 2.3×10^6 m^3 a^{-1}. Mean surface lowering observed between 1970 and 2007 across the ablation area was 13.9 ± 2.5 m (Bolch et al., 2011) suggesting that rates of mass loss have accelerated over the last 50 years compared to the last 500 years, and consistent with the observed decrease in the active glacier area (Quincey et al., 2009).

4.2 Glacier modelling

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

4.2.1 The Little Ice Age to the present day

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

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

4.2.2 Present day to AD2200

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

5. Discussion

5.1 Validation of model simulations

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

The decrease in the elevation of the simulated glacier surface over the 40 years prior to the present day was close to zero at the terminus and increased to 8–10 m in the upper part of the ablation area. The simulated surface lowering shows good agreement both in terms of the absolute values and the distribution of surface lowering to that observed for a similar period (1970–2007) which gave an elevation difference of –13.9 ± 2.5 m across the ablation area (Bolch et al., 2011). Integrated mass balance for the simulated present-day glacier was –0.22 m w.e. a^{-1}, slightly less negative than but not dissimilar to geodetic mass balance values estimated between 1970 and 2007 as of –0.27 ± 0.08 m w.e. a^{-1} (Bolch et al., 2011) and between 1992 and 2008 as –0.45 ± 0.52 m w.e. a^{-1} (Nuimura et al., 2012).

5.2 Equilibrium Line Altitude

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

5.3 Sources of uncertainty associated with modelling debris-covered glaciers

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

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

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

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

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

6. Conclusions

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

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

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Table caption

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

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

Figure captions

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

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

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

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

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

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

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

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

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