Increasing precipitation due to climate change could partially offset the impact of warming on glacier loss in the monsoon-influenced Himalaya until 2100 CE

Anya M. Schlich-Davies[1*], Ann V. Rowan[2*], Andrew N. Ross[1], Duncan J. Quincey[3], Vivi K. Pedersen[4]

[1]Priestley International Centre for Climate, School of Earth and Environment, University of Leeds, UK
[2]Department of Earth Science, University of Bergen and Bjerknes Centre for Climate Research, Bergen, Norway
[3]School of Geography, University of Leeds, UK
[4]Department of Geoscience, Aarhus University, Aarhus C, Denmark

*These authors contributed equally to this work

Correspondence to: Ann V. Rowan (ann.rowan@uib.no)

Abstract. Glacier mass in the Himalaya is projected to shrink by 53–70% due to climate change by 2100 CE. However, the impact of changes in precipitation amount and distribution on future glacier change remains uncertain because mesoscale meteorology is not represented in current glacier models. We explored the combined effects of past and future changes in air temperature and precipitation amount and distribution on the evolution of Khumbu Glacier in the Everest region of Nepal. We used a climate-glacier modelling approach that forced an ice-dynamical glacier evolution model with surface mass balance forcings that included mesoscale meteorological variables derived from downscaling of Regional Climate Model results. Our simulations show that historical warming has committed Khumbu Glacier to future mass loss of 10–23% during this century, and that under an intermediate future emissions scenario (RCP4.5), Khumbu Glacier could lose 70% mass by 2100 CE due to warming. However, the projected increase in precipitation in tandem with warming could offset about half of the projected glacier loss, such that the total decrease in glacier mass by 2100 CE compared to the present day would be reduced to 34%. Under a high future emissions scenario (RCP8.5) glacier loss due to warming will not be compensated by changes in precipitation, but will instead result in substantial ablation above 6,000 m with devastating consequences for one of the highest glaciers on Earth.

1. Introduction

Projecting glacier mass change in response to climate change is important for determining the impact of anthropogenic warming on regional water availability (Pritchard, 2019). High Mountain Asia is projected to lose 34 ± 19% of glacier mass by 2100 CE if warming is limited to 1.5°C to meet the ambitious Paris Agreement target (Kraaijenbrink et al., 2017). Less ambitious projections give 53 ± 23% glacier mass loss by 2100 CE under the intermediate emissions scenario RCP4.5, and 69 ± 20% under the high emissions scenario RCP8.5 (Kraaijenbrink et al., 2017; Marzeion et al., 2020; Rounce et al., 2023). However, such projections are challenging to make because accumulation and ablation processes in mountain environments are driven by orographic feedbacks between high-relief topography and atmospheric circulation systems such as the South Asian Summer Monsoon (Bookhagen and Burbank, 2006). Furthermore, large uncertainties arise from the challenge of simulating the interactions between the mass balance regimes of monsoon-influenced glaciers, where accumulation and ablation both occur during the monsoon season, and the dynamics of glaciers flowing through high-relief topography that includes processes such as the development of supraglacial debris layers that modify surface melting (Dehecq et al., 2019; Miles et al., 2018b; Salerno et al., 2023). Variability in the extent and intensity of the Indian Summer Monsoon during the Last Glacial Maximum affected glacier expansion in the monsoon-influenced Himalaya through changes in snowfall distribution (Benn and Owen, 1998; Owen et al., 2009). Future Indian Summer Monsoon precipitation

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and variability projected in Global Circulation Models (GCMs) will increase with current global warming (Katzenberger et al., 2021), but the effect of projected changes in precipitation amount, timing, and phase (rain/snow) on Himalayan glaciers remain poorly constrained (Immerzeel et al., 2012; Mölg et al., 2014; Ragettli et al., 2016; Shaw et al., 2022; Shea et al., 2015).

Supraglacial debris covers 4–7% of glacier surfaces globally and 30% of glacier ablation areas in the Himalaya, and modifies the response of glaciers to climate change relative to regional trends (Herreid and Pellicciotti, 2020; Kraaijenbrink et al., 2017; Rounce et al., 2023; Rowan et al., 2015). Satellite observations show that the rate of glacier mass loss across the Himalaya has accelerated over the last 40 years for both clean-ice glaciers and debris-covered glaciers (Maurer et al., 2019). Observations and modelling studies indicate that thick supraglacial debris caused historical mass loss from debris-covered glaciers to lag that of clean-ice glaciers, such that debris-covered glaciers are currently larger than would otherwise be the case (King et al., 2020; Rounce et al., 2023; Rowan et al., 2021). However, the dampening effect of supraglacial debris on glacier mass loss is overturned by the development of extensive supraglacial ponds and ice cliffs within debris layers (Miles et al., 2018a; Strickland et al., 2023) and the stagnation and detachment of debris-covered tongues from the upper and more active sections of glaciers (Rowan et al., 2021). Quantifying the impact of feedbacks set up by the formation and expansion of supraglacial debris layers at a regional scale requires exploring these processes at scales that can be resolved in ice-dynamical glacier evolution models (Rowan et al., 2015; Nicholson et al., 2021; Compagno et al., 2022). These processes can be considered in 2-D (along the glacier flowline) either considering stochastic debris delivery to the glacier (Vacco et al., 2010; Wirbel et al., 2018) or continuous debris delivery, which can result in the over-accumulation of debris at the terminus (Anderson and Anderson, 2016; Ferguson and Vieli, 2020; Jouvet et al., 2011), or in 3-D (using the horizontal and vertical ice flow fields), which allows the lateral transport and deposition of debris to the margins of the ablation area (Rowan et al., 2015).

While recent rapid warming resulted in a rise in regional equilibrium line altitude (ELA), that caused recession and collapse of glacier termini for both clean-ice glaciers and debris-covered glaciers, the decay of the former ablation areas of debris-covered glaciers is relatively delayed by supraglacial debris, such that the terminus of the actively flowing glacier can remain in contact with the detached ice tongue rather than separating (Maurer et al., 2019; Pellicciotti et al., 2015; Rowan et al., 2021). In common with most large debris-covered Himalayan glaciers, Khumbu Glacier in the Everest region of Nepal is in greater imbalance with climate than a climatically equivalent clean-ice glacier, and has maintained a more extensive ice volume than would be possible without supraglacial debris (Rowan et al., 2021). However, as a result of reduced ice flux from the accumulation area, the debris-covered tongue no longer receives much (or any) input of ice, and has dynamically detached from the active glacier (Fig. 1c); this observation is confirmed by the rapid reduction in ice flow and the peak in glacier surface lowering below the Khumbu Icefall where the debris layer is thinnest (King et al., 2020; Quincey et al., 2009). Therefore, the active glacier and the stagnant debris-covered tongue will evolve along different trajectories, and only the part of Khumbu Glacier above the terminus of the active glacier can be considered dynamic (Fig. 1). Projections of future glacier evolution should therefore discount the heavily debris-covered former tongue, which is decaying *in situ* without any input of new ice from the accumulation area, while considering the development of supraglacial debris across the ablation area of the active glacier.

We applied a novel glacier-climate modelling approach to Khumbu Glacier to test the hypothesis that changes in precipitation in response to climate change will reduce the impact of warming on glacier mass loss. Khumbu Glacier is a benchmark debris-covered glacier in the monsoon-influenced Himalaya flowing from 7,981 m above sea level (a.s.l.) to 4,879 m a.s.l. that is representative of the majority of glaciers in this region (Fig. 1). We used a 3-D ice-flow model forced by mass balance calculated from mesoscale meteorological variables to simulate the evolution of Khumbu Glacier from the late Holocene (~1 ka) through the present day (2015 CE) until 2100 CE using results from three downscaled Regional Climate Models (RCMs) under two Relative Concentration Pathways (RCPs). This approach represents an advance in the use of such models to understand the evolution of Himalayan glaciers; for the first time mesoscale meteorological forcing is used with a climate-glacier model that represents the

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Deleted: Models of debris-covered glacier evolution represent the dynamic feedback between debris transport, mass balance and ice flow that differentiates the evolution of glaciers with a substantial supraglacial debris layer from climatically-equivalent clean-ice glaciers (Zekollari et al., 2022). Such models require numerical representation of the processes controlling debris delivery to glacier surfaces from hillslope erosion, englacial transport of debris through the glacier to accumulate at the ice surface in the ablation area, and the impact of an evolving supraglacial debris layer on surface melting (Nicholson et al., 2021).

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processes of sublimation, snow avalanching, and debris transport, all of which are important controls on the mass balance of Himalayan glaciers. Simulations start from the late Holocene when Khumbu Glacier was last in dynamic equilibrium with the local climate, as evidenced by large ice-marginal moraines dated to 1.3 ± 0.1 ka surrounding the present-day glacier (Hornsey et al., 2022), and when the glacier surface was free of debris (Rowan et al., 2015). The focus of our experiments is to simulate glacier evolution to the end of the 21st Century. However, the centennial dynamic response time of a large debris-covered glacier such as Khumbu Glacier means that the glacier continues to evolve beyond this time scale, and we continued the simulations until 2300 CE to explore longer-term glacier evolution, albeit with greater uncertainties associated with results beyond 2100 CE.

2. Climate-glacier modelling of Khumbu Glacier

The climate-glacier glacier model experiments used mesoscale meteorological variables at an appropriate scale to calculate surface mass balance for the Khumbu Glacier catchment in combination with a debris-covered glacier evolution model to represent the surface processes that modify mass balance. Our approach produced a total of six simulations of Khumbu Glacier from three CORDEX South Asia region RCMs (NOAA, CCCma, IPSL; Lutz et al., 2016) and two RCPs (RCP4.5 and RCP8.5; Collins et al., 2013) to explore the impacts of possible variability in future precipitation amount and distribution in tandem with warming on glacier evolution. Before we use the RCMs to force the future climate scenarios, we evaluated their capabilities against observations of present-day weather and climate. The experimental design represents an advance compared with previous climate-glacier modelling efforts by including robust representations of: (1) mesoscale meteorological phenomena, including sublimation, (2) the redistribution of surface mass balance by snow avalanching, and (3) the feedbacks between debris transport, ice flow and mass balance. This section describes the experimental design for the climate-glacier modelling workflow (Fig. 2); a description of the study site, downscaling of the present-day RCMs using meteorological data from automatic weather stations in the Khumbu Valley, downscaling of the future RCMs for both RCPs, the surface energy and mass balance calculations using COSIPY, and the debris-covered glacier evolution modelling using iSOSIA.

Khumbu Glacier (RGI2000-v7.0-G-15-08331) is 16.0 km long with an area of 26.4 km^2. The median glacier elevation is 6,025 m a.s.l. from the terminus at 4,879 m a.s.l. to the headwall at 7,981 m a.s.l. (RGI 7.0 Consortium, 2023). The stagnant debris-covered tongue has an area of 6.2 km^2 (23% of the total glacier). The 'Little Ice Age' (LIA) maximum of Khumbu Glacier occurred about 500 years before present, which is consistent with ages produced for moraines elsewhere in the central Himalaya (Hornsey et al., 2022; Rowan, 2017). Khumbu Glacier was slightly larger than today during the late Holocene, transitioning from a clean-ice glacier with high velocities and efficient export of debris to an debris-covered glacier with lower velocities during the LIA; this change was initiated by the reduction in ice flux promoted by a rise in ELA and thickening supraglacial debris (Rowan et al., 2015). Observations and modelling of the dynamics and structure of Khumbu Glacier show that the lowest five km (25% of the total length, 20% of total ice volume) is stagnant and dynamically detached from the active glacier in the last century (Miles et al., 2021; Quincey et al., 2009; Rowan et al., 2021). Basal ice at the glacier surface indicates that the active terminus overrides the stagnant glacier tongue (Miles et al., 2021) and measurements of surface displacement show no longitudinal flow through the detached debris-covered tongue, which is collapsing laterally at a rate of about 3 m a^{-1} (Watson et al., 2017).

2.1 Glacier-climate model experimental design

As a starting point for our transient simulations of Khumbu Glacier, we reconstructed the late Holocene glacier from an ice-free domain using an ELA of 5,325 m a.s.l. and an atmospheric lapse rate of 4.0°C km^{-1} in a 5 kyr simulation. The simulation continued through the LIA forced by forced by a step change in mean annual air temperature (MAAT) equivalent to 1.5°C colder than the present day over 500 years following Rowan et al. (2015, 2021). Ice-marginal moraines denoting the late Holocene (1.3 ± 0.1 ka) glacier extent and thickness (Hornsey et al., 2022) were used to constrain the spin-up simulation. The ice-free domain was found by subtracting the estimated ice thickness (Farinotti et al., 2019) from a 30-m digital elevation model (DEM) acquired from the Shuttle Radar Topography Mission (Farr et al., 2007). The ice-free model domain incorporated the full hydrological catchment including the steep hillslopes in the Western Cwm that provide snow to the glacier surface by avalanching. We simulated

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The simulations continued from the present day to 2100 CE forced by the distributed glacier surface mass balances calculated for each of the three RCMs and two RCPs using COSIPY. The three RCMs and two future RCPs represent a range of possible future climates including distinctly different precipitation trends, equivalent to dry, moderate and wet scenarios under warming of 1.4–2.2°C under RCP4.5 and 3.8–4.1°C under RCP8.5 (Table 1; Section 2.2). We used climate time slices representing the present day (2015–2020 CE) and the end of the 21st Century (2095–2100 CE) to calculate surface mass balance. The five-year time slices were chosen to reduce the computational expense of the modelling (~24 hours per simulation), and the preceding decade was used to evaluate these time slices Section 3.3). We used a step forcing rather than interpolating mass balance over time, whereby the future mass balance was imposed and the glacier adjusted to this from the start of the century in question. Thus we arrived at the present-day simulation from the LIA simulation by forcing the LIA glacier with the 2015–2020 CE mass balance for 200 years. We use the output from the present-day simulation with the 2095–2100 CE mass balance to force the model to 2100 CE for a period of 80 years. We then use the result from this simulation as the starting point for the 2200 CE simulation forced by the 2195–2200 CE mass balance for 100 years, and the same approach for the 2300 CE using the 2295–2300 CE mass balance. Beyond 2100 CE less detailed climate projections are available; given the absence of regional climate projections, globally projected temperature changes were used to extend the end-of-century mass balances. These gave a further increase in temperature of 0.5°C by 2200 CE and 0.7°C by 2300 CE under RCP4.5, and 2.8°C by 2200 CE and 4.1°C by 2300 CE under RCP8.5 (Table 1; Collins et al., 2013). No precipitation changes were applied to the post-2100 CE climates due to the absence of projections for precipitation in the CORDEX RCMs and the high uncertainties associated with projections of global precipitation changes for this period.

We tested a range of lapse rates from 3.0°C km^{-1} to 6.0°C km^{-1} based on the range of monthly values calculated from regression of NASA MODIS land surface temperature data for the Central Himalaya while maintaining the same ELA, which resulted in a difference in ice volume of 0.4 x 10^{9} m^{3} and no change in glacier length at the present day. We examined the uncertainty in accumulation resulting from the application of a calculation to move snowfall from slopes susceptible to avalanching. Observations of high-elevation Himalayan glaciers, including Khumbu Glacier, indicate that up to 75% of accumulation occurs by avalanching rather than direct snowfall (Benn and Lehmkuhl, 2000). Avalanching affects Khumbu Glacier in two ways; by moving snow from steep hillslopes onto the glacier surface thus increasing accumulation from that calculated from direct snowfall onto the glacier surface, and by redistributing snow across steep sections of the glacier surface. If avalanching was not considered in the glacier model then accumulation of snow calculated using COSIPY within the catchment but outside the glacier outline would have no impact on mass balance, resulting in an underestimation of ice volume. When avalanching was not simulated, accumulation occurred at a uniform rate of 2.0 w.e. m a^{-1} across the Western Cwm, and the resulting glacier had a similar extent but a volume more than double that of the glacier simulated with avalanche redistribution of snow, because mass was not redistributed effectively across steep sections of the glacier surface.

2.2 Present-day RCM downscaling using meteorological observations

Six RCMs were assessed on their fidelity to present-day climate using hindcasting (Biemans et al., 2013) with emphasis on temperature seasonality and seasonal precipitation dynamics, given the importance of these variables for glacier mass balance. RCMs from the Coordinated Regional Downscaling Experiment (CORDEX) South Asia domain dynamically downscaled from CMIP5 GCMs

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by the Indian Institute of Tropical Meteorology to a 50 km spatial resolution (Lutz et al., 2016) were downloaded for the grid box containing Khumbu Glacier (27.9065056°N, 86.4352951°E). Three of the six CORDEX South Asia RCMs (NOAA, CCCma, IPSL) spanning a range of possible future precipitation conditions (Table 1) were selected as discrete scenarios for the glacier surface energy and mass balance calculations. The three remaining RCMs were discounted due to being intermediate to those selected for our experiments (i.e. close to the future precipitation scenario represented by CCCma) or particularly poor at reproducing seasonal temperature and precipitation cycles. For example, despite the annual precipitation sums from the CSIRO RCM being closest to observed values and having the potential to be the 'driest' scenario examined, analysis of precipitation seasonality indicated that the monsoon signal was completely absent, with this RCM instead showing a strong dominance of winter precipitation.

The present-day RCM results were downscaled using quantile mapping, also known as "distribution mapping", using 14 years of observations from three automatic weather stations (AWS; Fig. 1c and Appendix A) collected between January 2006 and November 2019 with gaps filled using interpolated data from neighbouring stations if required (Fig. 2). Parametric quantile mapping (Piani et al., 2010) was used, whereby a statistical relationship between the raw climate model outputs and observations was formed by substituting the RCM results with observations at a cumulative density function of the prescribed distribution (e.g., a gaussian distribution for temperature; Luo et al., 2018; a gamma distribution for precipitation; Piani et al., 2010). This correction was applied to the raw RCM outputs to produce a third downscaled dataset with an improved the fit to the observations (Maraun et al., 2016). The quantile mapping approach was chosen because this is effective for downscaling precipitation and reduces errors in the standard deviation, the coefficient of variation, and the skewness of distributed values relative to other methods (Lafon et al., 2013; Reiter et al., 2018). The AWS data were also used to disaggregate the resultant daily downscaled present-day and end-of-century climate model outputs to an hourly resolution for energy balance modelling. All meteorological variables, excluding precipitation, were downscaled using the MELODIST Python tool (Förster et al., 2016). Seasonal means were applied for precipitation to reproduce the 'nocturnal peak' seen during the monsoon that MELODIST was unable to replicate. Further information on RCM downscaling and AWS data analysis are provided in Appendix A.

2.3 Future RCM downscaling

Two future emission scenarios (RCP4.5 and RCP8.5) were available from CORDEX South Asia, which represent only intermediate and high emissions by 2100 CE relative to the present day. These two emissions scenarios are frequently used in climate impact studies enabling the comparison of our results with studies that use other climate/glacier model projections. The two future emissions scenarios were analysed for each of the three CORDEX RCMs to account for the inherently high uncertainties in future precipitation trends associated with climate models and the interplay of changing precipitation with atmospheric warming. The same statistical downscaling approach and disaggregation used for the three present-day RCMs (described in Section 2.2) was applied to the raw CORDEX RCM daily outputs for the three future RCM time slices under RCP4.5 and RCP8.5. The temperature change between the present day and the future time slices was preserved and there was no evidence of any imposed strengthening in the monsoon resulting from this downscaling. An increase in the frequency of days per year outside of the monsoon season with high precipitation amounts (defined here as over 15 mm of daily precipitation) accounted in large part for the higher annual precipitation amounts relative to the present day that were found in four out of the six RCMs. However, the total future annual precipitation increase was on average 8.8% greater in the downscaled climates relative to the raw RCMs, suggesting that this positive trend was inflated by downscaling. The downscaled climates reduced the frequency of precipitation, although, as in present day observations, monsoon precipitation occurred frequently and could be characterised as predominantly drizzle in the future.

2.4 COSIPY surface energy balance modelling

COSIPY is a glacier surface energy and mass balance model that integrates a surface energy and mass balance model with a multi-layer snow and ice model (Weidemann et al., 2018; Sauter et al., 2020). COSIPY was chosen as it is currently considered a leading open-source method for estimating glacier

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mass balance and has previously been applied to glaciers in High Mountain Asia. COSIPY includes a calculation of sublimation, which is an important ablation process for high-elevation glaciers because ablation can still occur if the latent heat flux is negative through sublimation, even in instances where surface temperature and/or air temperature are well below the melting point (Bonekamp et al., 2021; Brun et al., 2023; Huintjes et al., 2015). COSIPY integrates a surface energy balance model with a multi-layer snow and ice model and thereby resolves all energy fluxes (F) at the ice surface that contribute to surface melt ($Qmelt$):

$$F = SWin \cdot (1-\alpha) + LWin + LWout + Qsens + Qlat + Qg + Qliq \qquad \text{Eq.(1)}$$

Where $SWin$ is incoming shortwave radiation, α is albedo, $LWin$ and $LWout$ are incoming and outgoing longwave radiation, and $Qsens$, $Qlat$, and Qg are the sensible, latent, and ground heat fluxes (Oerlemans et al., 2001) and and $Qliq$ is the heat flux from liquid precipitation; the latter variable is often neglected in ablation calculations (Cuffey and Paterson, 2010) but is of particular importance here as the Indian Summer Monsoon brings a significant amount of liquid precipitation to the lower reaches of Khumbu Glacier. The resulting F is equal to the energy available for surface melt ($Qmelt$) when surface temperature (T_s) is at melting point (0°C). T_s is used to calculate $LWout$, $Qsens$, $Qlat$, QG and to partition solid and liquid precipitation. When T_s exceeds the melting point it is reset to 0°C (273.15 K) and the residual F fluxes equal $Qmelt$. In this instance, subsurface melt is triggered when the energy fluxes, for example, penetrating $SWin$ warm the ice layer so that T_s exceeds the melting point of ice (Sauter et al., 2020).

The COSIPY model domain was taken from the 30-m DEM that was resampled to 200-m grid spacing following sensitivity analyses that revealed minimal impact on the results whilst greatly reducing computational expense (Fig. 4). CORDEX RCM daily climate variables (temperature, precipitation, the radiation components, wind speed, relative humidity and atmospheric pressure) were used to force COSIPY. Snowfall measurements can be used as an input to COSIPY, but precipitation was partitioned into rainfall and snowfall using the snow transfer scheme within COSIPY given the paucity of observations and high uncertainties associated with AWS observations, climate reanalysis, and modelled snowfall products for this region (Sauter et al., 2020). COSIPY was forced using hourly meteorology with nine variables to calculate the energy balance and mass balance components at an hourly time step from the sum of accumulation by solid precipitation, deposition, and refreezing of melt water percolation, and ablation by melt and sublimation. The exchange processes at the surface, including energy release and consumption with phase changes, control temperature distribution and phase changes within the glacier (comprised of horizontal ice and snow layers). The coupling of the surface energy balance component with a multi-layer subsurface snow and ice model accounts for meltwater refreeze and percolation with the meltwater produced from the surface melt calculations acting as an input. The mass balance was calculated at an hourly resolution, with accumulation resulting from the accumulation of solid precipitation on the ice surface, refreeze of meltwater, and deposition of water vapour, and ablation resulting from subsurface and surface melt and sublimation (Sauter et al., 2020). The impacts of supraglacial debris on ablation and of snow avalanching on accumulation were handled in the glacier model.

2.5 Glacier evolution modelling

The second-order shallow ice approximation model (iSOSIA) is a 3-D higher-order ice-dynamical glacier evolution model that solves for the flow of ice including longitudinal and transverse stress gradients that are imposed on ice flow through high-relief topography (Egholm et al., 2011). This glacier model simulates the evolution of debris-covered glaciers by incorporating the feedbacks between debris transport, mass balance and ice flow (Rowan et al., 2015) and includes two processes that are important for many Himalayan glaciers—the redistribution of snow by avalanching that is estimated to account for 75% of accumulation, and the formation of a supraglacial debris layer that insulates the ice surface to modify ablation (Rowan et al., 2015). While previous versions of this glacier model used depth-integrated ice flow, this version simulates the evolution of Khumbu Glacier in 3-D as the ice thickness is divided into 20 vertical layers to calculate englacial debris transport (Rowan et al., 2015). The glacier

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The Coupled Snowpack and Ice-surface Energy and Mass Balance model in Python (COSIPY) was used to calculate surface energy balance (Sauter et al., 2020). COSIPY is developed and modularised in Python and integrates a surface energy balance model with a multi-layer snow and ice model and thereby resolves all energy fluxes at the ice surface that contribute to surface melt.

Moved up [13]: COSIPY was chosen as it is currently considered a leading open-source method for estimating glacier mass balance and has previously been applied to glaciers in High Mountain Asia.

Moved up [16]: CORDEX RCM daily climate variables (temperature, precipitation, the radiation components, wind speed, relative humidity and atmospheric pressure) were used to force COSIPY. Snowfall measurements can be used as an input to COSIPY, but precipitation was partitioned into rainfall and snowfall using the snow transfer scheme within COSIPY given the paucity of observations and high uncertainties associated with AWS observations, climate reanalysis, and modelled snowfall products for this region (Sauter et al., 2020). COSIPY was forced using hourly meteorology with nine variables to calculate the energy balance and mass balance components at an hourly time step from the sum of accumulation by solid precipitation, deposition, and refreezing of melt water percolation, and ablation by melt and sublimation

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model has a variable time step that can adjust up to a maximum of 0.1 years to allow greater computational efficiency.

The distributed surface mass balances calculated using COSIPY using the downscaled RCMs for the periods 2015–2020 CE and 2095–2100 CE were used as inputs to the glacier model with no change in forcing applied between time steps. Surface processes within the glacier model modified the distribution of accumulation and ablation but this was not updated into the surface topography used in COSIPY. Simulated accumulation was the result of the total snowfall in each cell and avalanching of snow imposed for the accumulated snowpack from hillslopes by removing snow and ice from hillslopes greater than 28° and redistributing this mass across less steep surfaces using a non-linear hillslope flux model (Roering et al., 1999). The avalanching routine was found to be sufficient to prevent snow and ice accumulation on slopes that are observed to be free of glacier ice such as the southwest face of Sagarmatha (Mt. Everest) while allowing accumulation on steep sections of the glacier (Rowan et al., 2015), resulting in accumulation rates at the glacier surface in line with the limited available observations for Himalayan glaciers of 2 m water equivalent (w.e.) per year (Benn and Lehmkuhl, 2000).

Rock avalanching is responsible for much of the debris accumulation on the glacier surface but there is little information about the magnitude and frequency of these events, so headwall erosion was assumed to be uniform at 1 mm a^{-1} (Rowan et al., 2021). Debris produced by headwall erosion was delivered to the glacier surface using a similar non-linear hillslope flux model to snow avalanching. The reduction in ablation beneath supraglacial debris from clean-ice values was represented as a reciprocal function that scaled clean-ice ablation (b_{clean}) to give sub-debris melt (b_{debris}) as a function of debris thickness (h):

$$b_{debris} = b_{clean} \times \frac{h_0}{h + h_0} \qquad \text{Eq. (2)}$$

where h_0 is a constant representing the characteristic debris thickness at which the reduction in ablation due to insulation by supraglacial debris is 50% of the value for an equivalent clean-ice surface (Anderson and Anderson, 2016; Rowan et al., 2021). The observed heterogeneity of surface ablation required a parameterisation of sub-debris melt representing the effects of differential ablation, which was represented in Equation (2) using a value for h_0 of 0.8 m (Bartlett et al., 2021; Rowan et al., 2021; Strickland et al., 2023).

3. Results

3.1 COSIPY parameter perturbations

The sensitivity of glacier mass balance to individual meteorological variables (MAAT, radiative fluxes, relative humidity, lapse rate, precipitation amount, precipitation phase, glacier surface roughness) was calculated in experiments that perturbed these variables individually. Perturbations were made within the range of the possible uncertainties for each variable that arise from a combination of the choice of observations or climate models, the downscaling approach used, and the distribution of meteorological variables. The values used for perturbations of MAAT and precipitation amount were similar to those expected for possible future climate forcings.

The spatially averaged mass balance was most sensitive to changes in MAAT (perturbed by ±1.5°C, 2.0°C and 3.0°C), *LWin* and *SWin* (±10% and 20%). Perturbations of relative humidity (±10% and 20%) had the least impact on mass balance. The use of a seasonal lapse rate of 5.38°C km^{-1} yielded a spatially averaged mass balance that was 5.6% less than the reference calculation value, while a diurnal lapse rate gave a mass balance that was only 0.45% lower because the reference lapse rate was close to the mean of the day/night lapse rates, whereas the environmental lapse rate (6.50°C km^{-1}) gave a mass balance that was 1.24% higher than the reference value. The relatively small difference in mass balance due to the choice of lapse rate is due to the extremely high elevation of Khumbu Glacier, which means that MAAT is below 0°C in the accumulation area for much of the year and a higher lapse rate does not affect rain/snow partitioning. The largest difference in mass balance due to the choice of lapse rate was

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Moved up [4]: dynamically detached from the active glacier in the last century (Miles et al., 2021; Quincey et al., 2009; Rowan et al., 2021). Basal ice at the glacier surface indicates that the active terminus overrides the stagnant glacier tongue (Miles et al., 2021) and measurements of surface displacement show no longitudinal flow through the detached debris-covered tongue, which is collapsing laterally at a rate of about 3 m a^{-1} (Watson et al., 2017).

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Deleted: 2.5 Glacier model experimental design¶
The late Holocene (~1 ka) glacier was reconstructed using a 5000-year equilibrium simulation starting from an ice-free domain and used as the starting point for three transient simulations through the ‘Little Ice Age’ maximum forced by a step change in mean annual air temperature (MAAT) equivalent to 1.5°C colder than the present day (Appendix B). The simulation was then forced to present-day conditions using the three surface mass balances (one for each RCM) calculated using COSIPY. The simulations continued to 2100 CE forced by the distributed surface mass balances calculated for each of the three RCMs and two RCPs using COSIPY. Khumbu Glacier is surrounded by ice-marginal moraines denoting the late Holocene (1.3 ± 0.1 ka) extent and ice thickness (Hornsey et al., 2022), which are used to constrain the spin-up simulation. Observations and modelling of the dynamics and structure of Khumbu Glacier show that ... [10]

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of ±24% just below the ELA. The installation of the National Geographic automatic weather stations on Mt. Everest in 2019 provided an opportunity to examine lapse rates at the highest elevations. For the period April–November 2019, the observed lapse rate was 4.68°C km^{-1} between Phortse (3,810 m a.s.l.) and Everest Base Camp (5,315 m. a.s.l.) and 5.36°C km^{-1} between Camp II and South Col, similar to the value used in this study. The lapse rate above 8,000 m a.s.l. was about 1.2°C km^{-1} greater than that below 5,600 m a.s.l. between the two highest AWS (South Col and Balcony) indicating that in the highest-elevation sections of the catchment, lapse rates may be best represented by values considered suitable for the free atmosphere. The glacier ice surface roughness (z_0) value was 1.7 mm (Table 2), which is a reasonable estimate for clean-ice glaciers (Mölg et al., 2012). The z_0 values reported in the existing literature vary widely even for clean-ice glaciers and do not consider debris-covered glacier surfaces, and so two substantially different z_0 values were tested as endmembers of the likely range in z_0 values; 0.1 mm from Midtre Lovénbreen in Svalbard (Irvine-Fynn et al., 2014) and August-One Glacier in China (Guo et al., 2018), and 6.9 mm for the clean-ice section of Haut Glacier D'Arolla (Brock et al., 2006). Adjusting z_0 had minimal impact on mass balance, although a higher (lower) z_0 did result in slightly increased (decreased) mass balance.

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Coupled parameter testing involved perturbing precipitation and MAAT simultaneously. The most significant change in spatially averaged mass balance followed a 3°C increase in MAAT and 20% decrease in precipitation amount. The change in ablation following an increase in temperature of 1.5°C was compensated by accumulation resulting from 20% higher precipitation. The impact on mass balance of two precipitation phase (rain/snow) partitioning schemes was investigated and compared with the default snow transfer function in COSIPY; (1) using threshold temperatures of 0.5°C, 2.0°C, and 3.5°C, and (2) using a calculation that smoothly scaled rain/snow partitioning from 100% solid precipitation at –1°C to 0% solid precipitation at 4°C. The height of the 0°C isotherm during months that experienced significant ablation (May–September) fluctuated around 5,125–6,250 m a.s.l., which correlated with the elevations that experienced the greatest mass balance change with lapse rate. While the lapse rate used to distribute MAAT did not have a significant impact on glacier-wide mass balance, the elevation of the 0°C isotherm from the pre-monsoon until the end of the monsoon was sensitive to the air temperature distribution.

3.2 Evaluation of COSIPY surface energy and mass balance results

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Turbulent fluxes and energy balance components across Khumbu Glacier were explored across a three-year period to assess the performance of COSIPY and understand their relative spatial importance. A reference simulation was made using COSIPY for comparison with observed glacier mass balance and tested across a range of grid spacings (30 m to 1 km). The reference simulation represented the period 2013–2015 CE and was forced by AWS data (Appendix A) using the model parameters in Table 2.

COSIPY was used to calculate clean-ice surface mass balance from the downscaled RCMs, and the insulating effects of supraglacial debris were calculated in the glacier model. The glacier-wide clean-ice mass balance for the three-year reference period was –3.4 m w.e, which equates to –1.13 m w.e a^{-1}. Maximum ablation was up to 16.2 m w.e. over three years (Fig. 4). High precipitation events were observed to offset some ablation if they occurred outside the core monsoon season (e.g., in October 2013 and May 2014) but did not influence monsoon season ablation when high air temperatures and strong incoming radiative fluxes rapidly remove snow cover and drive melting. Higher minimum temperatures in winter 2013–2014 CE relative to the other winters did not significantly influence accumulation rates, which remained similar to those in 2014–2015 CE. Low precipitation amounts during the 2015 monsoon (286 mm in 2015, compared to 330.8 mm in 2013, and 333.9 mm in 2014) resulted in lower accumulation in the upper reaches of the glacier. The precipitation gradient was calibrated to 1 x 10^{-5} % m^{-1} to match observed accumulation rates. However, this gradient largely arises from avalanching (Benn and Lehmkuhl, 2000) which is challenging to represent in COSIPY and was instead handled in the glacier model (Section 2.5).

The energy available for ablation peaked in the pre-monsoon and monsoon, bringing higher rates of sublimation and subsurface melt in April–June (Fig. 5). Simulated sublimation occurred at all elevations, with the highest cumulative loss near the South Col (EB7910) where sublimation dominated mass balance and only slightly slowed from December until May. Sublimation rates were increasingly tied to seasonality with distance down-glacier, with rates on the lower section of the tongue (EB4980) increasing from April until the start of the monsoon in July. Calculated subsurface melt was negligible at or above the ELA (5,950 m a.s.l.) whereas at lower elevations sub-surface melt dominated mass balance with a stronger seasonal cycle related to surface temperature. The interannual variability in subsurface melt was linked to surface temperature, although low simulated subsurface melt rates in the first year of the reference simulation were largely due to persistence of the initial snow cover that shielded the subsurface from surface temperature until the subsurface adapted to local conditions. Refreezing occurred across the entire glacier, with a staggered onset due to increased elevation, and the absolute values were low. The higher latent heat flux during the monsoon resulted in higher deposition of snow to the glacier at the lower elevations, with negligible rates at higher elevation. Similar absolute values and patterns are seen for condensation.

Calculated incoming shortwave radiation matched well with observations from Ev-K2-CNR, GlacioClim, and high-elevation AWS networks (Matthews et al., 2020) indicating that the radiation model in COSIPY performed well across the extreme relief of the Khumbu Glacier catchment (Fig. 6). Net shortwave radiation contributed the largest energy input to the glacier surface at lower elevations, correlating most strongly with the energy available for melt with a mean correlation coefficient of 0.79. There was high temporal variability related to variable cloud cover exhibited in the hourly incoming shortwave radiation forcing and fluctuating albedo during the warmer months with the melting of the snowpack. The high incoming shortwave radiation the upper reaches of the glacier indicate that low net shortwave radiation is not due to topographic shading. Net shortwave radiation was correlated with albedo ($r = 0.86$), and the persistence of snow throughout much of the year reduced the energy available for melt. Net longwave radiation also contributed to the energy available for melt as the pattern of both fluxes corresponded. Between 5,900–7,900 m a.s.l., net longwave radiation sometimes exceeded zero during the monsoon, most likely due to heavy cloud cover and increased temperatures relative to the glacier surface. The latent heat flux was almost zero at the lower elevation sites as the arrival of the monsoon resulted in higher relative humidity, and this pattern was similar, but dampened, at higher elevation. At the South Col (EB7910) the energy available for melt correlated exactly ($r = 1$) with the sensible heat flux (Fig. 6).

Grid spacings for the climate-glacier model of 30 m, 50 m, 200 m and 1 km were tested to ensure that that the COSIPY calculations captured orographic effects without unnecessary computational expense. The simulated maximum accumulation rate did not change significantly with grid spacing, giving accumulation rates of 2.1–3.9 m w.e. at 6,500–7,000 m a.s.l. in the reference simulation. The 1 km grid spacing contained only 27 glacier points, and gave a similar spatial mean mass balance to the finer-resolution calculations (Fig. 4), but there were large gaps in mass balance calculated across the glacier that affected the height of the ELA and significantly reduced the calculated maximum accumulation value. The 30-m and 50-m grid spacings captured greater spatial variability in mass balance relative to the 200 m resolution calculation, particularly at elevations between 5200–5400 m a.s.l.. However, as the ELA, and the calculated maximum and minimum mass balances were not significantly different between these finer-resolution calculations, the 200 m grid spacing was used throughout to benefit from the much reduced computational expense.

3.3 Evaluation of the RCM downscaling

The downscaled climate variables from the three RCMs for the present-day time slices (2015–2020 CE) were evaluated against 14 years of observations from three AWS to assess the representation of means, seasonality, diurnal cycles, day-to-day variability, and interannual variability. All three downscaled RCMs showed good agreement between mean annual air temperature (–2.15 ± 0.05°C) and observed air temperature from the Pyramid AWS (Appendix A). The representation of the monsoon was greatly improved by the RCM downscaling; temperature seasonality was well resolved following quantile mapping and the monthly mean and minimum air temperatures were similar to observations across the present-day time slices (Fig. A1). The monsoon stabilised air temperatures and reduced the range between minimum and maximum temperatures in the downscaled RCMs, which was in better agreement with AWS observations, but was not present in the raw RCMs prior to downscaling. We note

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that the downscaled maximum air temperature was at times higher than observations amongst all RCMs during the post-monsoon and winter (Fig. A1), but that the distribution of downscaled air temperatures was similar to observed values (Fig. A2). Gamma distribution quantile mapping substantially improved the absolute precipitation values relative to the AWS observations compared to those in the raw RCMs; the overestimation of winter precipitation and relative underestimation of monsoon precipitation amounts in the raw RCMs was reduced and downscaled results show a clearer monsoon signal (Fig. A3). When compared with AWS observations, RCM downscaling slightly over-corrected the seasonal precipitation pattern with a slight underestimation of winter precipitation for the most extreme winter events. Across the three present-day simulations, the surface mass balance calculated using the NOAA RCM was more positive than for the ISPL and CCCma RCMs and most similar to the mass balance calculated from meteorological observations.

3.4 Evaluation of the glacier model results

The simulated glacier geometry and dynamics (Fig. 7) were compared with remotely sensed observations of velocity, surface elevation change, and debris cover extent for the present-day glacier. The distributed surface mass balances calculated using COSIPY were most similar to observed values after the calculated surface mass balances were integrated with the glacier model to include accumulation by snow avalanching and the reduction in surface melting beneath supraglacial debris; the glacier extent was underestimated if supraglacial debris is not simulated (Fig. 8). The supraglacial debris-mass balance feedback in the glacier model reproduced the observed reversed mass balance gradient and peak in ablation below the Khumbu Icefall (Fig. 1). The simulated glacier area was 7.8 km^2 and similar to that obtained from structural mapping in 1979 CE (Nakawo, 1986). Radio-echo sounding in 1999 CE obtained ice thickness estimates close to the active terminus of ~160 m (Gades et al., 2000) and simulated ice thickness at the terminus was 130 m. The simulated thickness at the active glacier terminus thickness was approximately 175 m in 1999 CE, which agreed well with observations from DEMs of difference that show thinning here of up to 55 m between 1984–2018 CE (King et al., 2020). Simulated surface elevation change in the ablation area was –30 m over 20 years to the present day and similar to values derived from satellite observations for 1984–2015 CE (King et al., 2020). Simulated present-day glacier velocities (Fig. 9) show a similar pattern and magnitude to glacier surface velocities observed using remote sensing observations, which reach a maximum of 220 m a^{-1} in the Khumbu Icefall (Altena and Kääb, 2020) and up to 20 m a^{-1} in the ablation area (Quincey et al., 2009; Dehecq et al. 2019). The simulated present-day velocities in this study were a better fit to remote sensing observations than those from previous simulations that used an elevation-dependent mass balance forcing (Rowan et al., 2015, 2021) where the maximum simulated velocities were 118 m a^{-1}.

3.5 Climate change and glacier evolution from the present day until 2100 CE

Khumbu Glacier is responding to historical climate change and will continue to shrink even if warming ceases today. Indeed, if we allow the spin-up experiment to reach equilibrium with the present-day NOAA RCM mass balance, the glacier terminus will recede by 2.1 km and the maximum ice thickness will decrease from 246 m to 206 m by 2100 CE without any additional warming (Fig. 10a). In this simulation, a supraglacial debris layer up to 1.3 m thick extends 1 km up-glacier from the terminus and partially dampens the committed volume loss by sustaining 13% more ice volume than would be the possible for a clean-ice glacier surface with the same mass balance. The committed glacier volume loss due to historical warming in the absence of any further climate forcing is 10–23% of the present-day glacier mass (Fig. 9b) with the associated uncertainty represented by this range of values arising from the parameterisation of the impact of supraglacial debris evolution on surface melting.

Now considering the effects of additional warming under the RCP scenarios for the NOAA experiment, we find that greater warming occurs in winter than in summer under both RCPs (Sanjay et al., 2017) and results in an increase in annual precipitation amount of about 15% made up of a greater increase in winter precipitation than summer precipitation. The climate forcing from the downscaled NOAA RCM under RCP4.5 is 1.4°C warmer than the present day (–0.75°C in 2095–2100 CE compared with –2.15°C in 2015–2020 CE) and annual precipitation increases by 14.8% from 581.4 mm at present day to 664.8 mm a^{-1} by 2100 CE with summer (June–September) precipitation increasing by 5.4% and winter (December–February) precipitation increasing by 14.1% (Fig. 2). Under RCP8.5, the downscaled

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climate forcing is projected to be 3.8°C warmer than present day (1.65°C in 2095–2100 CE) with an increase in annual precipitation of 14.9% by 2100 CE, with summer precipitation increasing by 9.8% and winter precipitation increasing by 19.4%. In the NOAA RCM RCP4.5 experiment, the spatially averaged cumulative glacier mass balance is –0.14 m w.e. a^{-1} in 2100 CE, which is slightly more positive than the present-day value of –0.21 m w.e. a^{-1}. Glacier volume decreases by 36% between the present day and 2100 CE (Fig. 10). While significant, this end-of-century glacier loss is partially offset by the concurrent increase in precipitation. In comparison, an equivalent simulation forced only by warming and without any change in precipitation results in a more linear trajectory of glacier change and 70% loss of glacier volume by 2100 CE (light blue line in Fig. 10d) demonstrating that 34% of potential glacier loss from warming could be compensated by the increase in precipitation that occurs as a result of warming.

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The CCCma and IPSL RCMs projected greater warming from the present day by 2100 CE than the NOAA RCM under RCP4.5 with a value of 1.6°C (+0.2°C compared with the NOAA RCM) in the IPSL RCM experiment and 2.2°C (+0.8°C) in the CCCma RCM experiment. These two RCMs also projected slightly greater warming by 2100 CE under RCP8.5, with a value of 3.9°C (+0.1°C compared with the NOAA RCM) for the IPSL RCM experiment and 4.1°C (+0.3°C) for the CCCma RCM experiment. The projected increase in precipitation amount across the three RCMs is similar between RCPs with annual totals above 600 mm by 2100 CE. The CCCma RCM gives the greatest increase in annual precipitation amount of 100 mm by 2100 CE. There is no evidence of change in the intensity of the Indian Summer Monsoon, as the seasonal split in precipitation remains similar to the present day, but the frequency of days with high precipitation (over 15 mm per day) increases by 2100 CE, giving twice as many days in the NOAA RCM experiment and up to seven times as many days in the IPSL RCM experiment. Under RCP8.5, all experiments showed similar results for mass balance by 2100 CE with only a 10% difference in glacier volume between the three RCMs (Fig. 10d). The CCCma RCM experiment has only a 1% difference in volume loss between RCP4.5 and RCP8.5 by 2100 CE despite a 1.9°C difference in MAAT—this is a surprising result given the significant temperature difference, which can be attributed to the greater number of high-magnitude precipitation events that occur under RCP8.5 in combination with the small difference in winter temperatures between the two RCPs. Indeed, in the CCCma RCM experiment under RCP4.5, the maximum winter temperature is 1.7°C higher than for the other RCMs, resulting in ablation and rainfall (rather than snowfall) during the winter.

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3.6 Climate change and glacier evolution from 2100 CE until 2300 CE

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Projections of climate change beyond 2100 CE are more uncertain than those for this century and the uncertainties associated with the model experimental design become larger, but these results do indicate a future trajectory for Khumbu Glacier. As there are no regional temperature projections beyond 2100 CE we used global values to continue the simulations into the next century (Table 1). However, there are no global projections of precipitation beyond 2100 CE and to avoid introducing potentially significant uncertainties to our results by estimating these values, no changes in precipitation were applied beyond 2100 CE.

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In all the RCP4.5 experiments, there is little change in glacier volume between 2200 CE and 2300 CE compared with 2100 CE regardless of the RCM forcing used (Table 1 and Fig. 10b). In the NOAA RCP4.5 experiment, the Khumbu Icefall is maintained until 2300 CE and ice continues to flow from the Western Cwm to below 6,000 m so that the glacier remains in contact with the dynamically detached tongue. Therefore, keeping warming within the limit of RCP4.5 could restrict future volume loss to only 26% beyond that already committed to by historical climate change, and Khumbu Glacier would reach a new dynamic equilibrium that maintains a sufficient ice thickness to survive for at least two centuries. However, in all the RCP8.5 experiments, substantial glacier loss occurs after 2100 CE and Khumbu Glacier completely decays before 2300 CE.

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Physical detachment of the debris-covered tongue from the active glacier, whereby this area contained no active glacier ice, occurs around 2140 CE in the NOAA experiment (Fig. 9) and around 2070 CE in the CCCma and IPSL experiments. We defined the glacier to be stagnant when the maximum rate of ice flow is less than 10 m a^{-1}; a conservative estimate of the uncertainty associated with observations

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of glacier velocities (Dehecq et al., 2019). Accordingly, we consider Khumbu Glacier to no longer be a viable glacier system at the point where there is no ice flow above this value in the entire glacier since there is minimal throughput of ice mass. In the NOAA RCP8.5 experiment, the glacier area is 1.2 km^2 and the mean velocity reduces to 10 m a^{-1} by 2260 CE, such that the glacier is no longer viable. Glacier breakdown occurs earlier for the CCCma and IPSL RCM experiments (Fig. 10b) by 2170–2180 CE because loss of ice volume due to warming is not compensated to the same magnitude by the increase in precipitation projected under RCP8.5 in the NOAA RCM experiment.

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4. Discussion

4.1 Uncertainties associated with the climate-glacier modelling approach

Sources of uncertainty in our results arose from each step of our climate-glacier modelling workflow, and we considered how the experiments could be designed to reduce these uncertainties. In this section we discuss the potential sources of uncertainty associated with the choice of RCMs, the downscaling of the RCMs, the use of time slices rather than continuous mass balance calculations, the representation of future precipitation in the RCMs, the use of GCM forcing beyond 2100 CE, and the representation of avalanching in the glacier model.

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A single RCM was not considered sufficient to represent both present-day climate and potential future climatic extremes, but the climate-mass balance forcing ensemble was limited in size due to the small number of RCMs available. The use of three RCMs allowed the implications of uncertainties in understanding of local climate for glacier evolution to be evaluated. A multi-model mean approach using all the CORDEX South Asia RCMs (widely used elsewhere) was not considered sufficient to represent present-day and future climate conditions in the Khumbu Valley, as this approach gives equal weighting to models irrespective of their performance (Pierce et al., 2009) and does not enable intercomparison of results for future climate conditions. Five-year downscaled RCM time slices were chosen to reduce computational expense associated with COSIPY and the integration with the glacier model. To ensure that the five-year periods selected were representative, the preceding decade was used for comparison with the time-slice results (Appendix A). The use of quantile mapping with 14 years of AWS data as the downscaling method limited the influence of any natural variability by ensuring that the period did not reflect an extreme phase of natural climate oscillation. This comparison was particularly important for the future time slices, where large uncertainties arise between RCMs and observational data cannot be used for evaluation of the downscaled climate or the resulting mass balance.

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The time slice approach resulted in a stepped response in terms of ice volume change over time where the forcing was changed (Fig. 10d). The glacier continued to evolve through each century rather than equilibrate with the mass balance forcing, but we note that this stepped approach could be improved by interpolating the mass balance over time and coupling the COSIPY and iSOSIA models such that mass balance was calculated dynamically for the evolving ice surface; however this was beyond the scope of our experiments. The experiments were repeated using mid-century (2045–2050 CE) mass balance forcings to simulate glacier evolution until 2100 CE to investigate if this produced a different end-of-century result. These experiments produced near-identical results in 2100 CE to the experiments with no mid-century forcing, because the response time of the simulated glaciers was longer than the 40-year period between the present-day and future time slices. Thus, a mid-century surface mass balance forcing was not considered necessary in our experiments and instead we used a step forcing for mass balance rather than interpolation between mass balance calculations in the glacier model. The reacceleration in glacier mass loss after 2100 CE was in part due to the stepped forcing approach and the uncertainties associated with GCM projections, which increase with time after 2100 CE particularly under RCP8.5. For example, forecasts of global warming for 2281–2300 CE relative to 1986–2005 CE under RCP8.5 range from 3.0°C to 12.6°C (Collins et al., 2013).

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The differences in simulated glacier change and response time that resulted from the RCMs were at times greater than those resulting from the RCPs due to differences in projections of precipitation. Whilst the three selected RCMs performed relatively well in representing annual precipitation cycles from the six available CORDEX RCMs, we note that this representation was still fairly poor, although

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substantially improved by quantile mapping. The poor representation of monsoon dynamics in the present-day RCMs highlights an additional uncertainty associated with future precipitation scenarios and that these results should be treated as a set of possible scenarios. The CORDEX CMIP5 and CMIP6 projects only produced dynamically downscaled RCMs for two future emissions scenarios (RCP4.5 and RCP8.5) and as such the implications of other RCPs for glacier evolution could not be assessed. The downscaled future climates were compared with those from other studies using CORDEX results, and showed similar annual and seasonal regional temperature trends strongly linked to the choice of RCP, and similar positive precipitation trends with poor agreement between RCMs (Kaini et al., 2019; Sanjay et al., 2017). The relationship between precipitation and warming in the two future emissions scenarios was less clear than that for air temperatures because the monsoon-influenced Himalaya has particularly poor RCM consensus and high levels of uncertainty in future precipitation trends with warming relative to other regions in High Mountain Asia (Sanjay et al., 2017). The transition in the glacier model between the downscaled RCM and the GCM forcing could be improved by homogenising the climate model results across 2090–2100 CE, however, as noted above, the computational expense of forcing COSIPY with downscaled RCM outputs to create inputs to the glacier model required the use of a time-slice approach that was limited to five-year periods, and integration of the RCM results with GCM results was beyond the scope of this study. In the absence of RCMs that can project changes in precipitation after 2100 CE, precipitation was maintained at the same level for the glacier surface energy and mass balance model simulations beyond 2100 CE. The end-of-century precipitation amount is therefore unlikely to be reflective of the more distant future, and more realistic precipitation projections are required to explore whether the active glacier can be sustained further into the future or will lose ice more rapidly than is found in this study. However, while future precipitation changes may be important for glacier volume change under RCP4.5, we do not expect a sufficient increase in precipitation beyond 2100 CE to compensate for the warming projected under RCP8.5.

A large uncertainty in the glacier model arose from the parameterisation of avalanching, which resulted in increased accumulation along the glacier surface in the Western Cwm and improved the agreement between simulated and observed accumulation rates and distribution (Benn and Lehmkuhl, 2000). Future work to resolve the impact of low frequency–high magnitude avalanche events on accumulation rates would be useful to refine this calculation but the contribution of avalanches to glacier accumulation over decadal time scales remains challenging to measure. Our study addresses fine-scale temporal (hourly) and spatial (100 m) glacier surface processes, including avalanching and sublimation, that affect glacier surface mass balance across the elevation range of Khumbu Glacier, but further observations of meteorological and glaciological conditions at the highest elevations would be beneficial, and are needed if micro-scale processes are to be included in glacier models (Brun et al., 2023; Khadka et al., 2021; Mölg et al., 2014; Shaw et al., 2022).

4.2 Comparison with outcomes under RCP8.5

Current global greenhouse gas emissions are following the trajectory of the intermediate emissions scenario RCP4.5, while the high emissions scenario RCP8.5 could be described as 'low possibility but high impact' (Pedersen et al., 2020). However, as represented in the RCMs used in our study, mountain regions are warming more rapidly than the global mean such that a global temperature rise of 1.5°C will lead to 2.1 ± 0.1°C of warming in High Mountain Asia (Kraaijenbrink et al., 2017; Pepin et al., 2022) although the occurrence of elevation-dependent warming above 5,000 m a.s.l. is debated (Gao et al., 2018). The high temperatures projected under RCP8.5 could potentially be offset partly by increased precipitation, given that high-magnitude precipitation events from winter Westerly disturbances increased by a factor of seven between present day and 2100 CE in the IPSL RCM under RCP8.5. However, we found no evidence of future increases in precipitation offsetting RCP8.5 warming; net glacier mass balance was strongly negative in all RCP8.5 experiments and insufficient to maintain any actively flowing glacier. Under RCP8.5, glacier mass balance in the monsoon-influenced Himalaya may therefore shift from being driven by accumulation during the monsoon to predominantly during winter. Monsoon precipitation would only result in snow accumulation at the very highest elevations and would be insufficient to maintain flowing glaciers. This outcome is avoidable by limiting anthropogenic warming to within RCP4.5, which, due to the associated increase in precipitation, could sustain nearly

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The uncertainties associated with GCM projections increase with time after 2100 CE, particularly under RCP8.5.

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A recent global glacier modelling study forced by an ensemble of 10 GCMs projected mass loss of 64% for Khumbu Glacier by 2100 CE (Rounce et al., 2023). In comparison, our experiments project less severe rates of decline, resulting in 30% less mass loss under the RCP4.5 future climate scenario than in the global study (Fig. 10d). One difference between these results is that rather than using the global glacier inventory outline to define the glacier margins we consider only the actively flowing glacier and so exclude 20% of the starting glacier volume in the detached tongue. We would expect the two sections of the glacier to evolve along different paths; while the active glacier responds to climate change as projected in our experiments, thick supraglacial debris mantling the detached tongue could allow this ice mass to survive and slowly decay *in situ* for many decades beyond the present day. The decay of the detached tongue may however increase due to erosion of the surface by ice cliffs and supraglacial water bodies that are expanding across the former glacier surface (King et al., 2020). Our experiments only consider the rapidly changing active glacier, and we expect that the debris-covered tongue would melt more slowly than projected in the global modelling study, but as we do not consider the stagnant tongue to be part of the present-day glacier the ice volume simulated at the start of our experiments is smaller than that represented by Rounce et al. (2023) and other studies based on the RGI glacier inventory. The dynamically detached debris-covered tongue represents 20% of the present-day glacier volume and contains ice estimated as up to 360 m thick. The mean present-day ablation rate across this section of the glacier simulated in Rowan et al. (2021) is –0.54 m w.e. a^{-1} which can be used to estimate the life expectancy of the debris-covered tongue assuming no input of ice from the active glacier and no change in ablation rate due to thickening of supraglacial debris or the development of ice cliffs and supraglacial ponds. While the thickest part of the detached tongue may survive for ~600 years, the mean life expectancy of this ice mass is 176 ± 148 years from the present day meaning that the former debris-covered tongue will vanish by 2200 CE.

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4.3 Impacts of microscale meteorology on glacier change

Sublimation simulated in our study occurred at all elevations with the highest rate of ice loss due to sublimation (–0.12 m w.e. a^{-1}) in the upper reaches of the Khumbu Glacier catchment near to South Col (about 7,495 m a.s.l.) where sublimation dominated ablation with only minor seasonality. Whilst this amount of ice loss by sublimation is not negligible, it is almost half that found in the point-based calculations after adjusting for the different time periods represented by our studies (Matthews et al., 2020), which is likely due to the assumed uniformity of wind speed across the model domain in COSIPY. Future work to improve the calculation of sublimation in distributed surface mass balance calculations for high-elevation glaciers would be valuable. While we have considered the effects of mesoscale meteorology on glacier mass balance, smaller-scale processes operating close to the land surface could also be important. Katabatic winds were suggested to explain a local 15-year decrease in maximum air temperatures and precipitation over glaciers while minimum air temperatures continued to rise (Salerno et al., 2023). However, the impact of micro-scale near-surface cooling on the duration and extent of mesoscale precipitation and accumulation is likely to be minimal and therefore unlikely to significantly affect glacier-wide mass balance (Mott et al., 2020; Shaw et al., 2024). Observations from an AWS on Khumbu Glacier (6,464 m a.s.l.) indicate that surface energy fluxes may be sufficient to cause non-negligible melting of glacier surfaces despite freezing air temperatures (Matthews et al., 2020). Results from an ice core from South Col Glacier (>8,000 m a.s.l.) combined with COSIPY experiments suggested that ablation may also take place at even at the highest elevations (Potocki et al., 2022). However, a subsequent study found no evidence of change, and identified large uncertainties associated with simulating mass balance at these extreme elevations where sub-daily air temperature gradients and the duration of snow cover strongly affect ablation and accumulation (Brun et al., 2023). Future work is needed to reduce these uncertainties, as very few observations of accumulation processes and the upper limit of ablation processes exist for high-elevation Himalayan glaciers.

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4.4 The response of large debris-covered glaciers to climate change

The dynamic response time of large glaciers to climate change is of the order of centuries; for this reason, we start our simulations from the late Holocene (~1 ka) moraine extent when Khumbu Glacier was last considered dynamically stable (Hornsey et al., 2022; Rowan et al., 2015). The relationship between glacier response time and mass balance becomes less important after 2100 CE when the glacier is so small that ice flow has little impact on glacier volume change. Global and regional glacier modelling studies typically start their simulations in the current century (e.g., 2000–2007 CE in Marzeion et al. (2020); 2015 CE in Rounce et al. (2023)) and a further complication arises from the use of global glacier inventories as a starting point for glacier modelling studies as such inventories cannot capture the current dynamic state of glaciers that are imbalanced, and include all ice-covered areas rather than identifying only actively flowing ice. However, satellite-derived velocity products could be used identify where ice flow within glacier outlines declines to negligible rates (Dehecq et al., 2019).

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The RGI 7.0 inventory for Khumbu Glacier is based on imagery from 1999 CE (RGI 7.0 Consortium, 2023) where the detached debris-covered tongue represents 20% of the glacier volume contained within this outline (Fig. 1c). Simulations that integrated the stagnant tongue into the model domain rather than as part of the flowing ice improved the representation of simulated ice flow compared to observed values, supporting our conclusion that the debris-covered tongue has been dynamically detached from the active glacier for 50–100 years (Rowan et al., 2021). Field observations support the concept of active and stagnant sections co-existing in contact with each other as englacial optical televiewing indicated that thrusting occurs at several sites, denoted by skewed internal debris layers and of basal ice that has been thrust to the glacier surface, near to the active glacier terminus (Fig. 1c) from the direction of Khumbu Icefall (Miles et al., 2021). Our simulations show that development of supraglacial debris at the active glacier terminus reduced net volume loss by 13% (Fig. 8). Dynamic detachment of debris-covered tongues could allow these glaciers to move closer to equilibrium with a rapidly changing climate, the local mass balance gradient is a more important control on glacier change for both clean-ice glaciers and debris-covered Himalayan glaciers.

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5. Conclusions

In the monsoon-influenced Himalaya, 85% of the glacier area is located higher than 5,000 m above sea level and 21% is above 6000 m. Despite these high elevations, Himalayan glaciers are rapidly losing ice in response to recent warming and are projected to shrink by 53% to 70% during this century. However, the impact of future changes in precipitation on glacier loss remains uncertain, because mesoscale meteorology is not often represented in glacier model projections. We explored the effects of future warming in tandem with changes in precipitation by simulating the evolution of Khumbu Glacier in the Everest region of Nepal—a benchmark glacier in the monsoon-influenced Nepal Himalaya—using mesoscale glacier surface energy and mass balance modelling forced by downscaled Regional Climate Model outputs. Historical warming commits Khumbu Glacier to loss of 10–23% of the total ice volume by 2100 CE. While warming due to intermediate future greenhouse gas emissions (RCP4.5) will lead to glacier volume loss of 70% by 2100 CE, the projected concurrent increase in precipitation amount will offset 34% of this and so reduce glacier loss by about a half. However, high future emissions (RCP8.5) will not be compensated by changes in precipitation amount but will instead result in substantial ablation above 6,000 m with devastating consequences for one of the highest glaciers on Earth. Our results indicate that the net mass balance of Khumbu Glacier could be close to zero in 2100 CE under RCP4.5 and therefore, if climate change is limited to this intermediate emissions scenario, Khumbu Glacier will recede to the base of the icefall with insignificant further change in glacier volume beyond this point. In this scenario, Khumbu Glacier has a similar extent in 2100 CE to the active section of the present-day glacier, and represents at least one example of how monsoon-influenced Himalayan glaciers could persist into the future if global efforts are sufficient to mitigate anthropogenic climate change.

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Code availability

The COSIPY version 1.3 glacier surface energy and mass balance model is available from the original publication describing this model (Sauter et al., 2020) and Zenodo (Arndt et al., 2020). The iSOSIA version spm-3.3.3r glacier model is available from Zenodo (Rowan and Pedersen, 2024).

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Data availability

Daily data from the Coordinated Regional Downscaling Experiment (CORDEX) South Asia domain were downloaded from the Indian Institute of Tropical Meteorology website (http://cccr.tropmet.res.in/home/cordexsa_about.jsp) for the grid box nearest to Khumbu Glacier (27.9065°N, 86.4353°E). Incoming shortwave and longwave radiation components were downloaded from the ESGF portal (https://esgf-ui.ceda.ac.uk/cog/projects/cordex-ceda/). 14 years of meteorological observations were derived from the two Pyramid AWS at 5,050 m a.s.l. and at 5,035 m a.s.l. (SHARE network Ev-K2-CNR; https://www.evk2cnr.org) and the West Changri Nup glacier AWS at 5,363 m a.s.l. (GlacioClim: https://glacioclim.osug.fr/).

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Author contributions

Conceptualisation: DJQ, ANR, AVR
Data curation: ASD, ANR, AVR
Formal analysis: ASD, ANR, AVR
Funding acquisition: DJQ, ANR, AVR
Investigation: ASD
Methodology: ASD, ANR, AVR, VKP
Project administration: DJQ, ANR
Resources: DJQ, ANR
Software: AVR, VKP
Supervision: DJQ, ANR, AVR
Validation: ASD, AVR
Visualisation: ASD, AVR
Writing – original draft preparation: ASD, AVR, DJQ, ANR, VKP
Writing – review and editing: ASD, AVR, DJQ, ANR, VKP

Competing interests

The authors declare that they have no conflict of interest.

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Tables and captions

Table 1. Regional Climate Models (RCMs) chosen for this study and details of the Global Circulation Models (GCMs) from which these are derived. The NOAA RCM that was considered most representative of conditions in the Everest region. The temperature forcings used to project climate change beyond 2100 CE are global values (they are simply included against each of the RCMs for ease comparing with 2100 temperature change) and include no change in precipitation after 2100 CE.

CORDEX South Asia regional climate model	Driving CMIP5 global climate model	CMIP5 modelling centre	RCM name in this study	Future precipitation scenario (qualitative)	2100 CE mean temperature change from present day (°C)		2200 CE mean temperature change from 2100 CE (°C)		2300 CE mean temperature change from 2100 CE (°C) */2200CE*	
					RCP 4.5	RCP 8.5	RCP 4.5	RCP 8.5	RCP 4.5	RCP 8.5
ITTM-RegCM4	NOAA-GFDL-GFDL-ESM2M	National Oceanic and Atmospheric Administration (NOAA), USA	NOAA	Wet	1.4	3.8	0.5	2.8	0.7 */0.2*	4.1 */1.3*
IITM-RegCM4	CCCma-CanESM 2	Canadian Centre for Climate Modelling and Analysis (CCCma), Canada	CCCma	Moderate	2.2	4.1	0.5	2.8	0.7	4.1
IITM-RegCM4	IPSL-CMSA-LR	Institue Pierre-Simon Laplace (IPSL), France	IPSL	Dry	1.6	3.8	0.5	2.8	0.7	4.1

Table 2: Model parameters, where α is albedo (of fresh snow, firn and ice), *t** decay time from snow to firn , *d** the constant for the effect of snow depth on albedo, and z_0 surface roughness length. The albedo values are widely used within the literature for clean. z_0 values are less well parameterised with high spatial variability across glacier surfaces. Testing revealed highest sensitivity of mass balance to $\alpha_{freshsnow}$ relative to the other parameters.

Parameter	**Value**	**Unit**	**Source**
$\alpha_{freshsnow}$	0.85	-	Mölg et al. 2012; Wagnon et al., 2009
α_{firn}	0.6	-	Knap and Oerlemans, 1996; Mölg et al. 2012
α_{ice}	0.3	-	Mölg et al. 2012
t*	20	days	Mölg et al. 2012
d*	1.0	cm	Mölg et al. 2012
Z_{0snow}	0.24	mm	Gromke et al., 2011
Z_{0firn}	4.0	mm	Brock et al., 2006
Z_{0ice}	1.7	mm	Brock et al., 2006
Z_0 ageing length (linearly from Z_{0snow} to Z_{0firn})	60	days	Mölg et al. 2012

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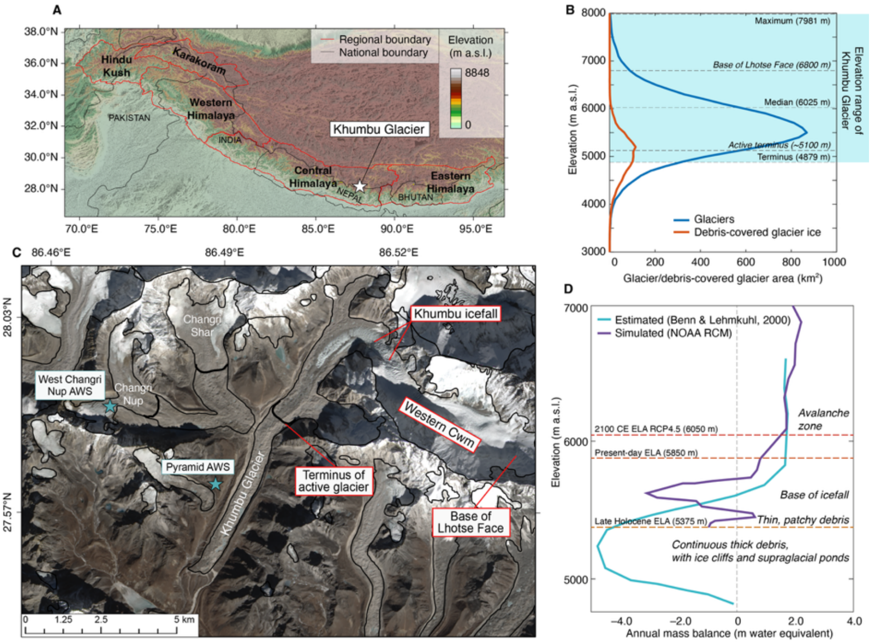

Figure 1: Khumbu Glacier location and context. (a) Location map of High Mountain Asia showing the location of the monsoon-influenced Central and Eastern Himalaya and Khumbu Glacier. (b) hypsometry of glaciers and debris-covered glacier ice in the Central and Eastern Himalaya compared with the elevations of Khumbu Glacier. (c) Satellite image of Khumbu Glacier showing the glacier outline from the RGI database (black line) that is equivalent to the late Holocene (~1 ka) glacier extent identified from ice-marginal moraines, the extent of supraglacial debris, location of the Khumbu Icefall, the extent of active ice flow inferred from observations of glacier velocity, and location of the automatic weather stations used for RCM downscaling (blue stars). (d) Estimated mass balance gradient for debris-covered glaciers in the Everest region (Benn and Lehmkuhl, 2000) compared with the glacier mass balance gradient simulated using the NOAA RCM and showing change in the equilibrium line altitude (ELA) of Khumbu Glacier in the historical and future simulations for the NOAA RCM RCP4.5 experiment.

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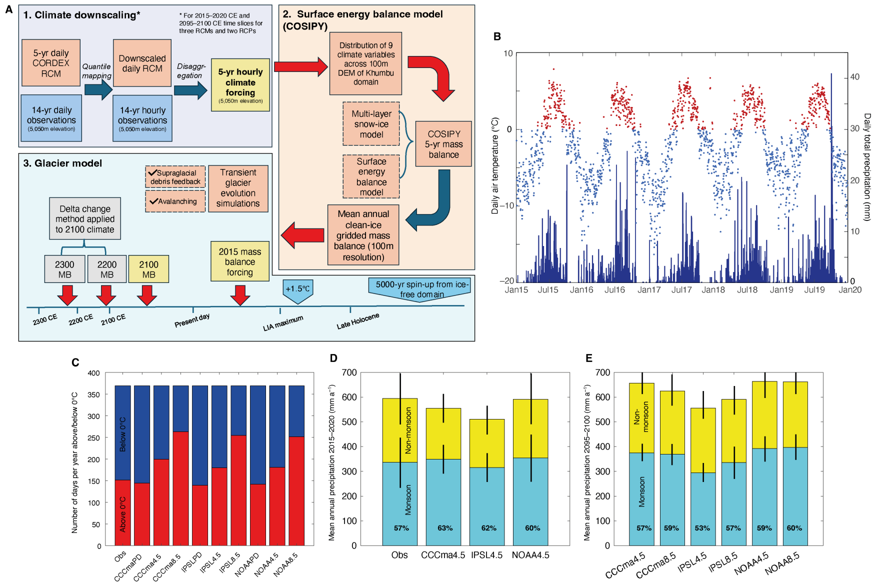

Figure 2: Glacier model experimental design and evaluation of RCM downscaling. (a) Schematic diagram of the glacier modelling approach showing the methods used for downscaling through quantile mapping and disaggregation of climate data. Note that this process does not apply to the post-2100 CE climate forcings which are subject to delta change. Surface energy balance modelling using COSIPY includes the pre-processing stage of meteorological distribution across the Khumbu domain, which is repeated for each RCM in the 2015–2020 CE climates and for the three RCMs and two RCPs for the 2095–2100 CE climates. The simulated mass balance is then used to force the glacier evolution model. (b) Daily mean temperature and daily total precipitation from the NOAA RCM for the present day (2015–2020 CE) following downscaling using quantile mapping with air temperature categorised into above freezing (red) and below freezing (blue). (c) Proportion of air temperatures above and below freezing for the present day for each RCM and RCP for the downscaled daily data compared with observations. (d) Annual precipitation totals for non-monsoon and monsoon with standard deviation between selected years shown by black bars for the downscaled daily data compared with observations. (e) Future (2095–2100 CE) time-slice annual precipitation totals for non-monsoon and monsoon months with standard deviation between selected years shown by black bars. In (d) and (e) the percentage of the total annual precipitation occurring during the monsoon is indicated by the value in bold text. (Obs = meteorological observations from AWS).

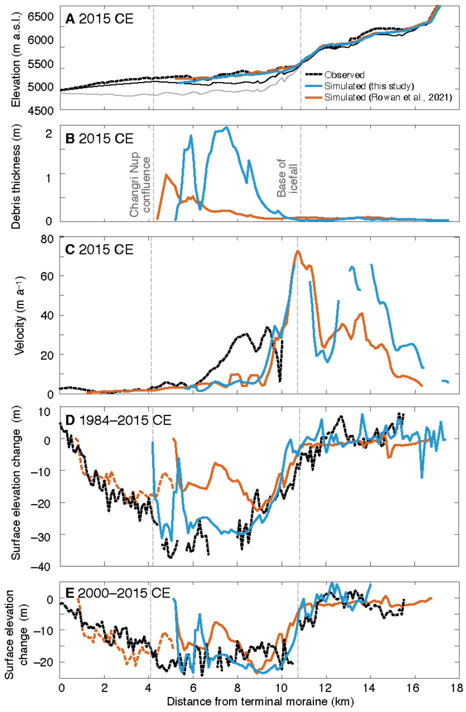

Figure 3. Evaluation of present-day simulation showing; (a) mean simulated glacier surface elevation and bed elevation calculated from a 500 m-wide swath profile along the central flowline of the glacier. Subglacial topography including the dynamically detached debris-covered tongue is shown by the solid black line and subglacial topography used in the entire glacier simulations in Rowan et al. (2015) is shown for comparison by the lowermost grey solid line. The estimated present-day ice surface elevation (Farinotti et al., 2019) is shown by the dashed black line. (b) mean simulated debris thickness, (c) simulated and observed velocities from the NASA MEaSUREs ITS_LIVE project (Dehecq et al., 2019), and simulated and observed mean surface elevation change between (d) 1984–2015 CE and (e) 2000–2015 CE compared with results from the simulations in this study and those in Rowan et al. (2021) where further information about the model evaluation can be found.

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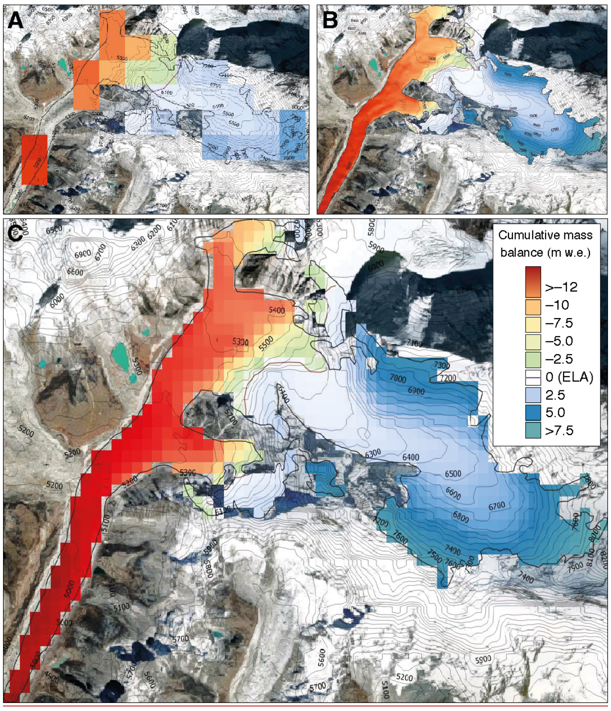

Figure 4. COSIPY reference calculation of present-day mass balance for Khumbu Glacier for the period 2013–2015 CE showing the results from calculations using different grid spacings using (a) a 1-km grid, (b) a 30-m grid, (c) the 200-m grid spacing used throughout the experiments in this study.

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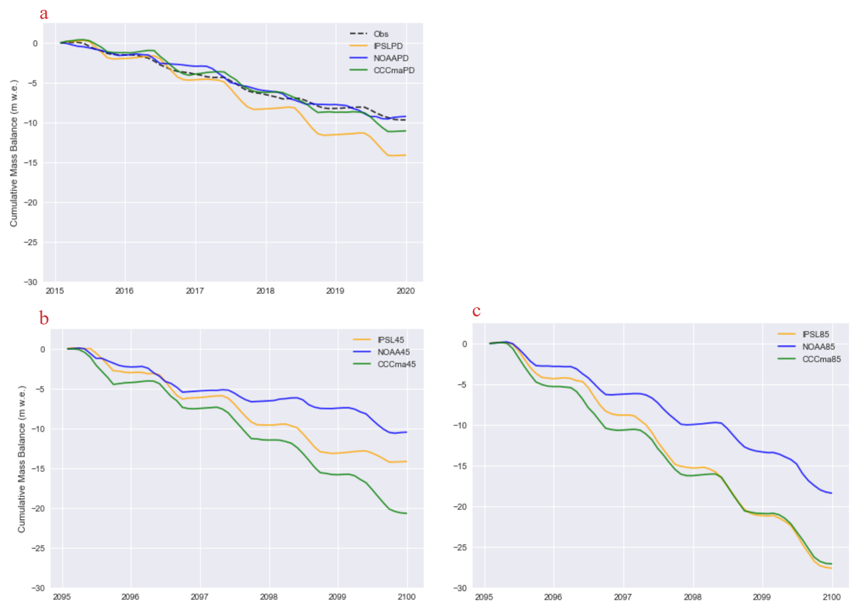

Figure 5. Spatially averaged cumulative clean-ice mass balance with clear seasonality for (a) the present day time-slice including the mass balance forced by the observations used for downscaling, and the end-of-century time-slice under (b) RCP4.5 and (c) RCP8.5. The low annual glacier-wide mass balance values shown here are the result of the extent of the model domain used to force the glacier model that includes the larger catchment beyond the glacier margins and therefore contains a higher proportion of lower elevations than those of the glacier itself. However the similar mass balance results for simulations forced by NOAA RCM and observations can be clearly seen (a), and the differences between the three RCMs is apparent in all time-slices (a-c).

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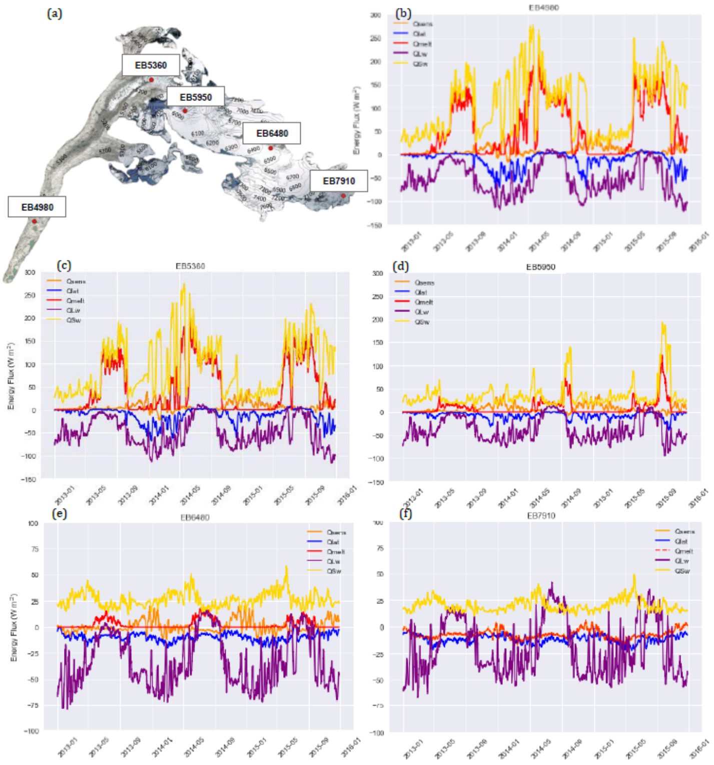

Figure 6. (a) Locations of energy balance (EB) points used for energy flux and melt components analysis (named after corresponding altitude e.g., EB6480) and (b–f) 5-day average of energy fluxes across study period for each site. Note that scales are different for (e) and (f) compared to (b)–(d) due to the marked difference in absolute values.

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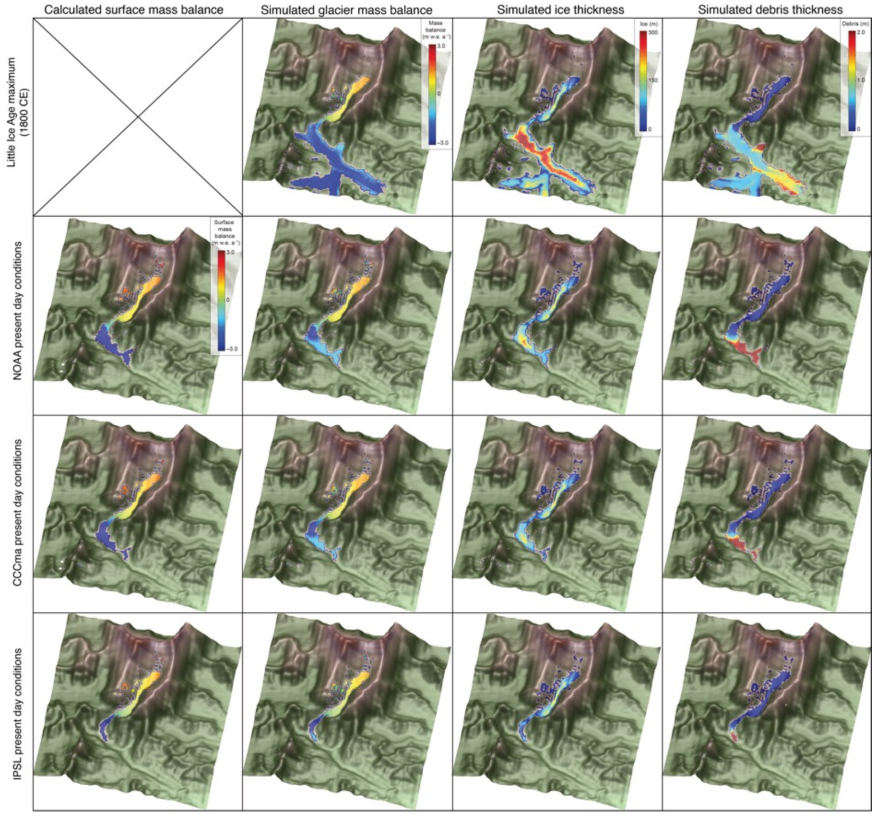

Figure 7. Glacier model sensitivity to surface energy and mass balance forcing, showing Little Ice Age (~1800 CE) glacier mass balance, ice thickness and debris thickness. Present-day results for surface mass balance calculated using each RCM with COSIPY showing glacier mass balance calculated using the same climate forcing following integration with the glacier model, simulated ice thickness, and simulated debris thickness.

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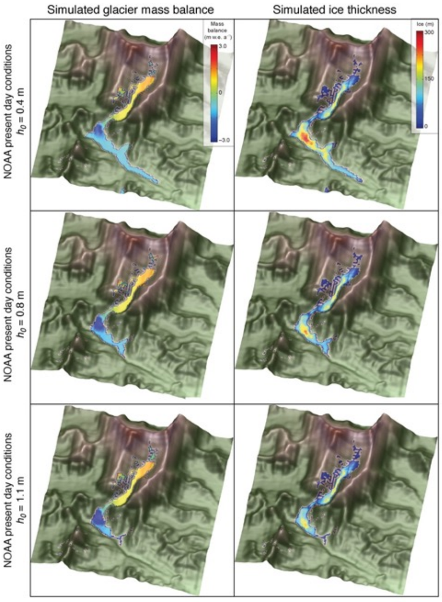

Figure 8. Glacier mass balance and ice thickness simulated using the NOAA RCM climate forcing and the resulting simulated ice thickness for h_0 values of 0.4 m, 0.8 m, and 1.1 m where h_0 is a constant in Equation(1) representing the characteristic debris thickness at which the reduction in ablation due to insulation by supraglacial debris is 50% of the value for an equivalent clean-ice surface (Anderson and Anderson, 2016; Rowan et al., 2021).

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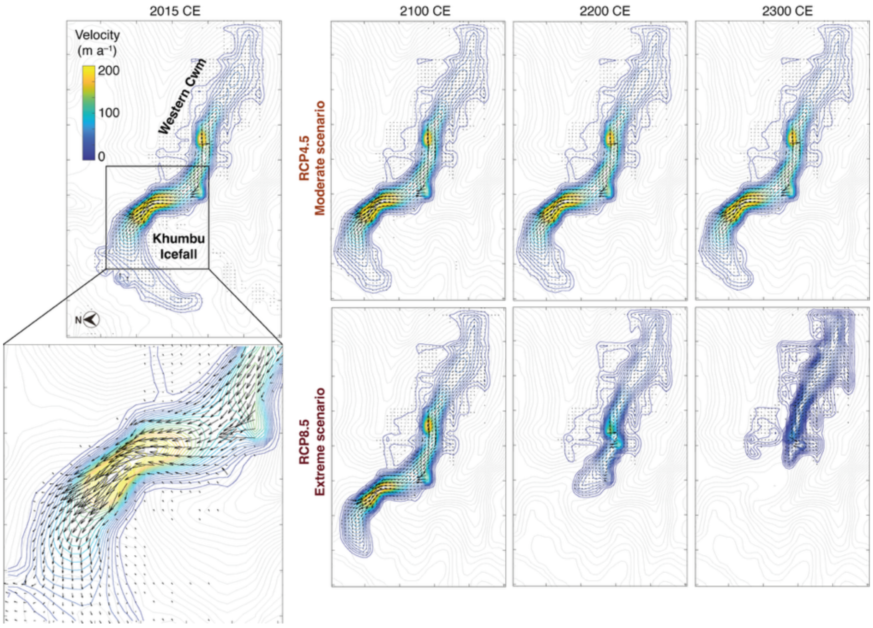

Figure 9. Simulated ice flow for Khumbu Glacier. Velocity-vector maps showing simulated ice flow magnitude and direction from the present day (2015–2020 CE) until 2300 CE under RCP4.5 and RCP8.5 using the downscaled NOAA climate forcing. Simulated ice flow speed is shown as colour shading with blue contours, and the bed topography is shown by grey contours. The outermost contour in each plot represents the slowest ice flow close to the glacier margins with depth-integrated velocities of 5–10 m a^{-1}. Note that rapid flow across the Western Cwm indicated by one arrow shows the effects of avalanching rather than sustained glacier flow.

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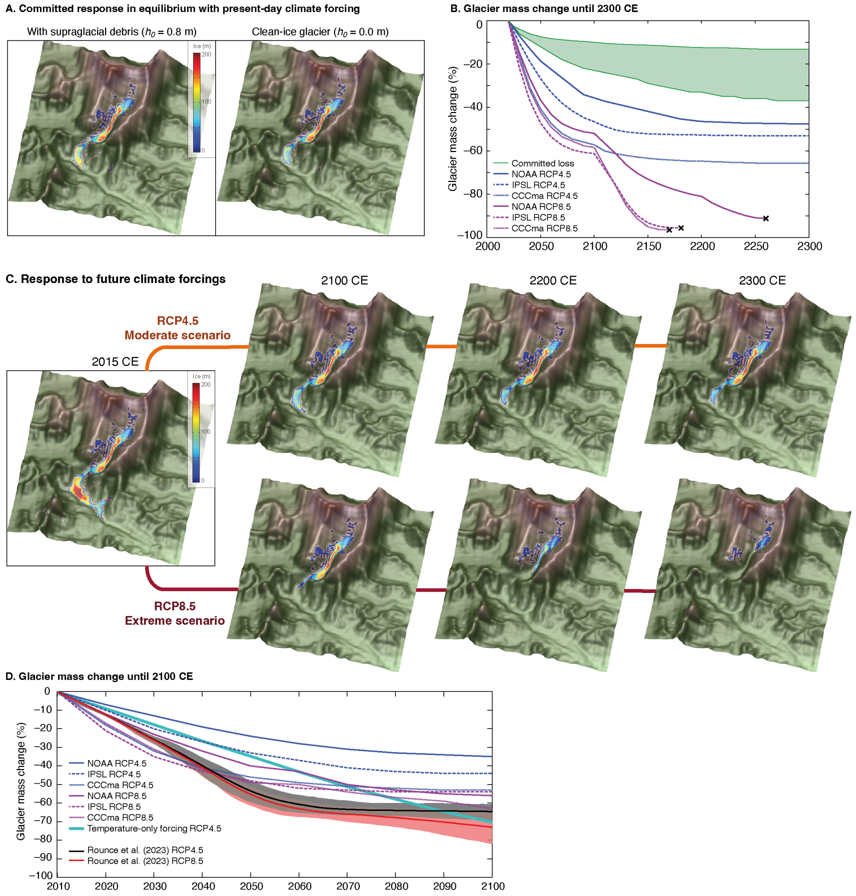

Figure 10. Future glacier volume change projections. (a) Equilibrium ice thickness accounting for the committed response to recent climate change using the downscaled NOAA RCM climate forcing with and without the effect of supraglacial debris on mass balance. (b) Simulated glacier volume change from the present day (2015–2020 CE) until 2300 CE under RCP4.5 and RCP8.5 for the three downscaled RCMs. The black crosses mark when ice flow has declined sufficiently that the glacier is considered almost absent or no longer viable. The green shading shows the range of the committed volume loss due to historical warming. (c) Simulated ice thickness under RCP4.5 and RCP8.5 for 2100 CE, 2200 CE and 2300 CE using the downscaled NOAA RCM climate forcing. (d) Comparison of projected shrinkage of Khumbu Glacier by 2100 CE from this study with those from Rounce et al. (2023) showing results from each of the six experiments in this study with results from RCP4.5 and RCP8.5 from Rounce et al. (2023), the equivalent result for a simulation using a change in MAAT equivalent to the NOAA RCP4.5 forcing where precipitation does not change from the present-day value (cyan line).

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Moved up [29]: Simulated ice flow speed is shown as colour shading with blue contours, and the bed topography is shown by grey contours. The outermost contour in each plot represents the slowest ice flow close to the glacier margins with depth-integrated velocities of 5–10 m a^{-1}. Note that rapid flow across the Western Cwm indicated by one arrow shows the effects of avalanching rather than sustained glacier flow.¶
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Moved up [30]: (b) Simulated glacier volume change from the present day (2015–2020 CE) until 2300 CE under RCP4.5 and RCP8.5 for the three downscaled RCMs. The black crosses mark when ice flow has declined sufficiently that the glacier is considered almost absent or no longer viable. The green shading shows the range of the committed volume loss

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Appendix A: Climate Model Downscaling

This section contains further details on the RCM downscaling, evaluation of the present-day downscaled RCM results using meteorological data, and the distribution of these results across the model domain.

A1. Meteorological data collection and analysis

14 years of meteorological observations were collected from two AWS at the Pyramid Observatory at 5,050 m a.s.l. and 5,035 m a.s.l. and the West Changri Nup Glacier AWS at 5,363 m a.s.l.. Missing data were replaced through interpolation with an alternative AWS from this group. Precipitation was measured at 15-minute intervals using a Geonor T-200BM sensor mounted 1.8 m above the surface. Evaporation from the bucket is supposedly blocked by a layer of oil but some does occur as evidenced by precipitation values below 0 mm. Noise from wind and evaporation were corrected for by compensating any negative change over the 15-minute time step with the neighbouring positive value such that accumulated precipitation was unchanged. Periods with prolonged evaporation were set to zero. Undercatch of snowfall by rainfall gauges was corrected through precipitation phase partitioning using wind speed observations (Wagnon et al., 2009). For interpolation of air temperature, hourly lapse rates were used that averaged 5.54 °C km^{-1} to adjust to the height of the reference point at 5,050 m a.s.l.. Where possible, precipitation data taken from the Pyramid AWS at 5,035 m a.s.l. because this precipitation gauge provides a longer period of continuous observations than the other gauges and avoids errors due to low precipitation amounts measured by tipping bucket gauges, which are known to systematically underestimate snowfall particularly during high winds (Sherpa et al., 2017).

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A1. Downscaled climate model results

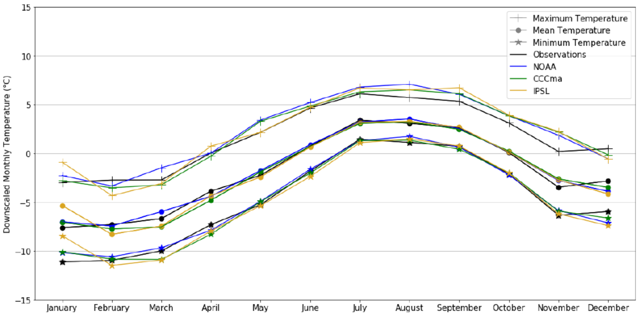

Figure A1: Downscaled monthly mean, maximum, and minimum temperature calculated for the present day time slice.

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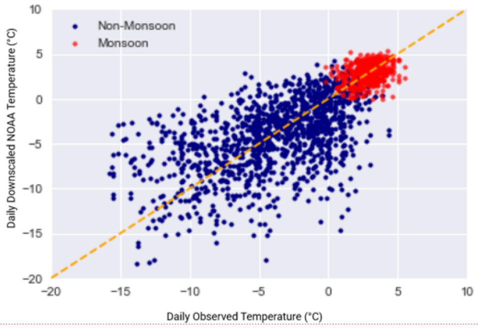

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Figure A2: Daily downscaled temperature from the NOAA RCM against observations, split by monsoon/non-monsoon with a 1:1 line to aid analysis of the temperature distributions (dashed orange line).

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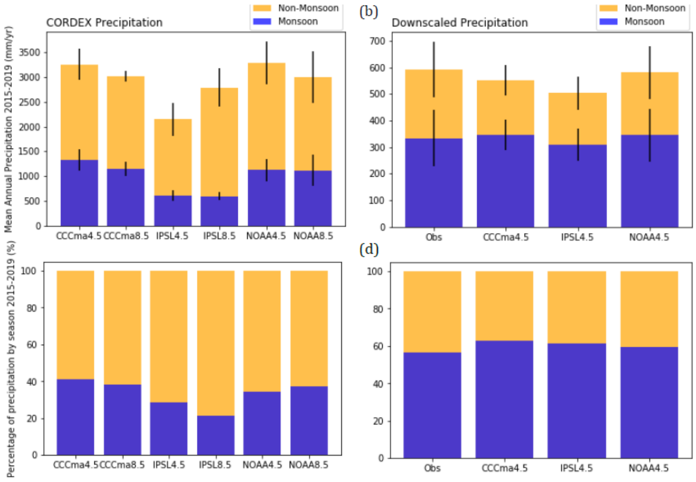

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Figure A3. Annual precipitation totals for non-monsoon and monsoon months before and after downscaling with standard deviation between selected years shown by black bars (a and b) and as their seasonal percentages (c and d). The annual precipitation matches measurements in the southern Dudh Koshi catchment for the gridbox nearest to Khumbu Glacier is located at 27.9065056°N, 86.4352951°E which is 2,100 m a.s.l..

A2. Regional Climate Model analysis and selection

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Three of the six available CORDEX South Asia RCMs (NOAA, CCCma, IPSL) were selected as discrete scenarios that span the range of possible future precipitation conditions (Table 1); either wet, moderate, or dry climate in 2080–2100 CE (Figure A4). The raw RCMs significantly overestimate annual total precipitation by at least a factor of five for the selected gridpoint which is corrected for by downscaling these results using AWS data.

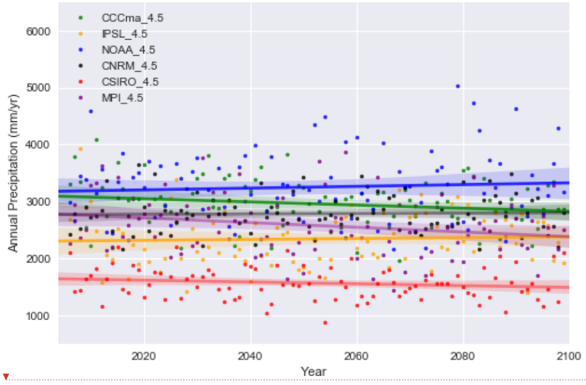

Figure A4: Annual precipitation sums (dots) with fitted trend line from the start of the RCP experiments (2006) until the end (2100) for each of the six Indian Institute for Tropical Meteorology CORDEX models for RCP4.5.

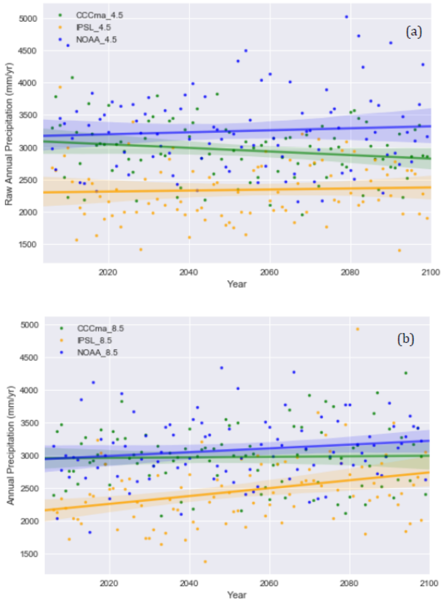

Figure A5: Annual precipitation sums (dots) with fitted trend line from the start of the RCP experiments (2006) until the end (2100) for the three selected of the six CORDEX models for RCP4.5 (a) and 8.5 (b).

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A3. Downscaling parameters and method

Though minimum and maximum temperature are not required to force COSIPY, these were downloaded and statistically downscaled using QM with normal distribution to aid disaggregation to an hourly time step using MELODIST. Quantile mapping for the CORDEX wind speed data was found to be ineffective when analysing the time series output against observations, both for the absolute wind speed as well as the reduced day-on-day variability seen during the monsoon. Therefore, GARD was used instead. This is a simple statistical analogue regression downscaling method appropriate for pointwise downscaling.

Table A1: RCM-derived parameters and the method used for downscaling or bias correction.

RCM-derived parameters	**Downscaling/bias correction method**	**Parametric distribution model (for QM)**	**References**
Precipitation (kg per m^2 per s, converted to mm day^{-1})	Quantile mapping (QM)	Gamma	Vrac et al., 2007; Piani et al., 2010
Mean temperature (K) Minimum temperature (K) Maximum temperature (K)	QM	Normal / Gaussian	Li et al., 2010, Gupta et al., 2016; Luo et al., 2018
Incoming shortwave (W m^{-2}) Incoming longwave (W m^{-2}) Relative humidity (%)	QM	Beta	Ruane et al., 2015
Pressure (hPa)	Bias correction	N/A	N/A
Wind speed (m s^{-1})	Regression downscaling	N/A	Gutmann et al., 2022

A4. Meteorological distribution across the model domain

Temperature and precipitation were interpolated across the 100 m DEM using a linear relationships:

$$V_{interp} = (V) + (Z_{pixel} - Z_{station}) \cdot LR$$

where V_{interp} is the variable to be interpolated, V is the hourly variable in question (e.g. temperature or precipitation), Z_{pixel} is the elevation (m a.s.l.) of the target pixel in the domain, $Z_{station}$ the elevation (m) of the station, and LR is the lapse rate for temperature or gradient for precipitation calculated from meteorological observations, as described below. Note that the distribution is from a prescribed elevation, not from the exact location of the AWS.

The mass balance sensitivity to lapse rate that differed depending on season or time of day were examined, showing a lesser impact on glacier-wide mass balance than in other studies due to the large elevation range of Khumbu Glacier where a smaller fraction of the glacier relative to total area is located along the zero degree isotherm (*cf.* Yala Glacier; Immerzeel et al., 2014). Lapse rates that changed depending on season and time of day (given marked monsoonal and nocturnal lowering of lapse rates) averaging 5.54 °C km^{-1} were calculated and, following integration with the glacier model, produced glacier-wide mass balance and spatial calculations that were closest to those from geodetic observations, including maximum rates of surface lowering in the upper ablation area where the debris layer is thinnest (King et al., 2020) (Fig. 1d).

Analysis of meteorological observations made between 2,600 m and 5,600 m a.s.l. from the Ev-K2-CNR and Glacioclim networks indicated that precipitation gradients were weak, slightly negative or absent across the Dudh Koshi catchment, confirming the observations of Salerno et al. (2015) and Yang et al. (2017). Given the high incidence of missing precipitation data from high-elevation AWS, the undercatch of snow associated with tipping bucket rain gauges, and the absence of precipitation measurements above 5,600 m a.s.l. precipitation was not varied with elevation. To test the sensitivity of precipitation to elevation, COSIPY was forced by a gridded climate distributed using weak negative, weak positive, and no precipitation gradients. The results of these experiments were used to force the

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glacier model, and the simulated historical glacier evolution was similar, with only a 10 m difference in the maximum ice thickness between simulations with different precipitation gradients.

Direct solar radiation across the model domain was corrected by the slope, azimuth, and shadowing potential of each pixel (Wohlfahrt et al., 2016; Sauter et al., 2020). A footprint-weighted correction was also applied to horizontal measurements of net radiation. The fraction of diffuse incoming shortwave radiation was estimated by using the ratio of total shortwave (global) radiation and potential shortwave radiation to define a clearness index (Wohlfahrt et al., 2016). This clearness index was used to calculate diffuse radiation, which is calibrated with data from Neustift, an eddy covariance station in the Austrian Alps (Wohlfahrt et al., 2008). The distributed radiative fluxes were compared with high-elevation stations for 2019 to assess the efficacy of this method across the domain. Pressure was distributed across the domain by first calculating sea level pressure (*cf.* Lente and Osz, 2020) and then interpolated with the barometric equation. The relative humidity gradient was calculated as –0.002 % m^{-1} from Ev-K2-CNR and Glacioclim networks and validated with National Geographic network to capture trends at higher elevations (Matthews et al., 2020). Wind speed was assumed to be uniform across the domain.

Additional references for Appendix A

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balance model that integrates a surface energy and mass balance model with a multi-layer snow and ice model (Weidemann et al.,

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Where *SWin* is incoming shortwave radiation, *α* is albedo, *LWin* and *LWout* are incoming and

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