Amplified Last-Glacial-Maximum response of Chandra valley (western Himalaya) glaciers

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This is the EarthARxiv preprint of a manuscript that is under review in Geology.

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ABSTRACT

Geomorphological evidence suggests a subdued response of Himalayan glaciers during the Last Glacial Maximum (LGM), with relatively minor advances (~10 km) reported in several glacierised valleys across the region. This supports the hypothesis that a weakened Indian summer monsoon during the LGM largely counterbalanced the effects of a colder climate on Himalayan glaciers. In contrast, a recently reported major LGM advance (>100 km) along the main trunk of Chandra valley, western Himalaya, led to an alternative hypothesis that Himalayan glaciers did respond strongly to reduced LGM temperatures, in harmony with other glacierised regions in the world. We investigate this distinctive LGM response of Chandra valley glaciers using a two-dimensional ice-flow model, to show that this massive LGM advance was driven by a relatively modest lowering of equilibrium line altitude (ELA) by ~300 m. The vigourous response of Chandra valley glaciers to the ELA perturbations was governed by their high climate sensitivity due to the gentle slope of the main trunk valley. The relatively low value of estimated ELA change in this valley compares favourably with careful estimates reported from other parts of the Himalaya, indicating a prevalent weak climate forcing of glaciers in and around the Himalaya during the LGM.

INTRODUCTION

During the Last Glacial Maximum (LGM), about 20 ka before present (Mix et al., 2001), favourable climatic conditions (Clark et al., 2012) caused equilibrium line altitude (ELA – the elevation separating the accumulation area above it from the ablation zone below) of mountain glaciers to descend by 800–1000 m (Broecker and Denton, 1989) than their present level in several glacierised regions across the globe. The resultant expansion of the accumulation areas induced major (~100 km) glacial advances (Hughes et al., 2013). However, in some of the regions, the LGM extents were not the most extensive local glaciation of the last glacial period, likely due to a possible decline in accumulation in a drier LGM climate which limited the glacial advance (Gillespie and Molnar, 1995; Hughes et al., 2013). The Himalaya may be a prominent example of this effect as the LGM glaciations here were largely restricted to only about 10 km beyond the present glacier termini (Owen, 2011). This relatively weak influence of the global LGM cooling on Himalayan glaciers, which are largely fed by snow from Indian summer monsoon (ISM), has been linked to a weakened ISM during the LGM (Benn and Owen, 1998; Owen et al., 2002; Schäfer et al., 2002). The weakening of ISM during the LGM is well documented in signals extracted from foraminifera, speleothem, ice-core and other natural climate archives (e.g., Duplessy, 1989; Thompson et al., 1997; Herzschuh, 2006; Cheng et al., 2016). This decline of ISM precipitation was possibly driven by a corresponding decline in the northern hemispheric high-latitude solar insolation (e.g., Cheng et al., 2016). Of course, a significant spatial variability of the LGM response in the Himalaya is expected due to the strong variability and an east-west contrast of glacier accumulation regimes in the Himalaya (Maussion et al., 2014), the topographic and/or hypsometric variability (Pratt-sitaula et al., 2011), and the supraglacial debris-cover effects (Vacco et al., 2010; Banerjee and Shankar, 2013). Moreover, the low preservation potential of glacial deposits in the Himalaya and possible uncertainties in their inferred chronology tend to inflate the variability of the reconstructed LGM response of Himalayan glaciers (Eugster et al., 2019).

Along with the possible factors listed above, another intrinsic reason behind variable glacier response between regions, or even between two glaciers in the same region, is the difference in the climate sensitivity of individual glaciers (e.g., Oerlemans, 2001). In simple terms, climate sensitivity is the rate of change of glacier size (i.e., length, or area, or volume) with respect to the variation of ELA. To compare the climate forcing among different regions or different glaciers with significantly different climate sensitivities, the reconstructed glacier advances need to be first translated into the corresponding changes in ELA. There are several methods for estimating paleo-ELA from reconstructed glacier extents (e.g., Benn and Lhemkuhl, 2000; Benn et al., 2005). However, many of these existing methods have inherent limitations due to the simplifying assumptions made, e.g., ignoring the effects of a variable glacier geometry and hypsometry, the ice-elevation feedback, non-linearity of mass-balance profile etc. (Benn et al., 2005). The presence of extensive supraglacial debris and avalanche activities, that are common in the Himalaya, complicates the matter further (Laha et al., 2017). Incidentally, the reported values of ELA changes based on reconstructed LGM glacier extents in the Himalaya varies over a wide range of 100 m to 1000 m (e.g., Owen and Benn, 2005; Heyman, 2014). While part of this reported variability may be due to an inherently inhomogeneous climate forcing over the Himalaya during the LGM, a significant part of it may also have arisen out of the limitations implicit in the methods employed for reconstructing the LGM ELA. Computation of the LGM ELA changes in the Himalaya using Global Circulation Model reconstruction of LGM climate did not prove to be useful due to a large model-to-model variability of such estimates (Rupper and Koppes, 2008). However, based on a careful analysis of the methods used and the accuracy of moraine chronology, Owen and Benn (2005) concluded that estimates of 100m to 300 m of ELA change during LGM as obtained in the Khumbu region (central Himalaya) and Batura glacier (Karakoram) are reliable. These estimates are consistent with the hypothesised weak LGM forcing on Himalayan glaciers discussed at the outset.

Recent evidence from Chandra valley, western Himalaya (Fig. 1) casts doubts on the above picture of a weak LGM forcing of Himalayan glaciers (Eugster et al., 2016). Building upon an existing body of previous work (Owen et al., 1996, 1997, 2001), and using extensive cosmogenic ^{10}Be dating of glacier polished bedrock, Eugster et al. (2016) were able to reconstruct a ~150 km LGM advance of Chandra valley glaciers along the main trunk valley. Based on the evidence, the authors argued in favour of a purely temperature driven LGM response of Chandra valley glaciers in line with other glacierised region in the northern hemisphere, questioning the standard paradigm that the negative feedback of a weakened ISM attenuated the impact of LGM cooling on Himalayan glaciers.

A potentially serious limitation of the above argument by Eugster et al. (2016) is that the authors did not consider the alternative possibility of an exceptionally large climate sensitivity of the Chandra valley glaciers amplifying their response to a prevalent weak ELA forcing during the LGM. For example, a low valley slope is known to induce such amplified glacier response (Eaves et al., 2019). In this paper, we use numerical simulations to test the hypothesis that a high climate sensitivity of Chandra valley glaciers triggered the massive glacial advance along the main-trunk valley by about a couple of hundred kilometers despite a weak ELA forcing during the LGM.

METHODS

We use a vertically integrated two-dimensional shallow-ice approximation (SIA) based ice-flow model (e.g.,Oerlemans, 2001; Le Meur, 2004), that considers only the deformation contribution to flow without any basal slip. An implicit Crank--Nicholson finite-difference scheme is used with spatial and temporal step size of 100 m and 0.01 years, respectively. The ice-free bedrock is derived from Frey et al. (2014). The model is forced by a linear mass-balance profile with a cut-off on maximum accumulation at 1 m/yr. The debris-covered portions of the simulated glaciers are assumed to have a flat ablation rate that equals -2m/yr (Banerjee and Shankar, 2013; Banerjee and Azam, 2016). The present extents of clean and debris-covered ice are obtained from RGI Consortium (2017). To incorporate the large-scale avalanche activity in the region (Laha et al., 2017), we have implemented a scheme for gravitational redistribution of snow/ice along steep slopes (slope > 50%). Simulations are run for up to 3000 years to ensure that a steady state is reached, starting from an ice-free bedrock. The local ELA value is manually tuned to produce steady glaciers with extents similar to the present glacier extent (Fig. 1B and 2B). Subsequently, the model is run with different perturbed values of ELA. More details about the model simulations, and the results of various sensitivity tests are described in the supplementary document.

RESULTS AND DISCUSSIONS

Simulation results show that for a relatively low uniform ELA depression of ~300 m, a 120 km long glacier advance takes place along the presently deglaciated main trunk valley (Fig. 2). The glacier advance could have been even larger in reality due to possible contributions from the tributaries in the Bhaga valley which is just downstream of our simulation domain (Eugster et al., 2016). The modeled steady-state trunk-valley glacier has ice thickness of up to about 1000 m. This modeled state compares well (Fig. 2C) with the reconstructed Chandra valley glaciers during 18-19 ka (Eugster et al., 2016).

The above estimate of ELA change during LGM is dependent on model parameters to some extent. Using different values of mass-balance gradient, accumulation cut-off, constant sub-debris ablation rate, and Glenn’s flow law constant (e.g., Le Meur, 2004), a range of ELA depression between 150 m to 325 m is seen to reproduce steady-state advances similar to the one described above (Supplementary Section S5).

A limitation of our model is that the evolution of the debris-cover extent is not included in it. To investigate the corresponding impact on our results, we have considered two extreme scenarios: a debris free ablation zone during the LGM, and a completely debris-covered ablation zone for the advancing tongue (Supplementary Section S5). The estimated LGM ELA depressions for these two limiting cases are 300 m and 150-200 m, respectively. We note that a nearly debris-free ablation zone during the LGM may be more likely because of the expected exponential decline in debris-production rate in a colder climate (Banerjee and Wani, 2018).

The idealised spatially homogeneous mass-balance profile used in the study is based on the observed mass-balance profiles from Hamtah and Chhota Shigri glaciers in the region (Laha et al., 2016). In reality, there may be considerable spatial variability of glacier mass-balance profiles in the region. However, there are no mass balance profile data available from the glaciers in the upper reaches of the valley. A regional variability of the ELA change during the LGM is also expected due to a strong precipitation gradient across this valley (Ashahi, 2010). We have considered a possible inhomogeneous change in ELA to study such effect. Here, the local ELA perturbation is computed based on the present precipitation distribution (Shea and Immerzeel, 2016). In this experiment, a very similar LGM glacier advance is obtained for a mean ELA change of 290 m across the valley (Section S5), which is consistent with the the range of ELA change mentioned above.

While a lack of accurate data related climate forcing led us to make certain simplifying assumptions as detailed above, the use of a simple ice-flow model based on SIA may be questioned. SIA considers only the horizontal shear stress components and therefore, is inaccurate over steep and/or narrow valleys (Le Meur et al., 2004). However, we note that similar 2-d SIA models have been successfully used for paleo-glaciation studies in other mountainous regions in the world (e.g., Plummer and Phillips, 2003; Kessler et al., 2006; Xu et al., 2013; Eaves et al. 2019). Moreover, the inaccuracies in the estimated ELA changes due to the limitations of the ice-flow model (e.g., use of SIA, neglecting basal sliding etc.) is likely to be relatively insignificant when simulating the long time-scale evolution of large glaciers, with errors due to the uncertainties in the bedrock elevation and that of the mass-balance profile being relatively more important (Greuell, 1992; Leysinger and Gudmundson, 2003). In any case, the

ELA estimates are going to be more reliable than those obtained from the commonly used thumb-rules (Benn and Owen, 2005). In addition, the fact that our targeted LGM state, with an extent much larger than the present glaciers, is reproduced without much fine-tuning gives confidence in the robustness of our ELA estimates.

Results from the simulations and the sensitivity tests, provide strong evidence that the extensive (>100 km) Chandra valley glaciation during 18-19 ka (Eugster et al., 2016) was likely driven by a modest ~300 m ELA change. This estimate is consistent with the reported low values of ELA changes from Batura and Khumbu glaciers (Owen and Benn, 2005). Thus, the extensive LGM glaciation in the Chandra valley is, in fact, consistent with and does not contradict the hypothesis that glaciers in and around the Himalaya experienced a relatively weaker climate forcing due to a partial compensation of the temperature-change effects by a weakening of ISM. The massive LGM response in Chandra valley was the result of a high climate sensitivity of the glaciers in the valley, and does not necessarily imply a strong climate forcing. The exceptionally high climate sensitivity of Chandra valley glaciers can be traced to the gentle slope of about 2% along the main trunk valley, as surface slope is known to be inversely related to climate sensitivity of glaciers (e.g., Oerlemans, 2001; Eaves et al., 2019).

The inherent averaging involved in simulation of all the glaciers in a valley that is larger than 1000 km^2 and contains more than 50 glaciers larger than 1 km, also contributes to the robustness of the estimated ELA. Remarkably, applications of simpler empirical rules to estimate paleo-ELA, when averaged over a large set of glaciers in and around Tibetan Plateau, obtained estimates of 350±200 m of regional ELA depression during the LGM (Heyman, 2014). This estimate is consistent with that obtained here for Chandra valley, or that reported for Khumbu and Hunza valleys (Owen and Benn, 2005).

We note that the above trends are not to be considered signals of a uniform ELA depression during the LGM across and along the strike of the Himalaya. Significantly larger estimated ELA changes were also reported elsewhere in the region (e.g., Ashahi, 2010; Shukla et al., 2018), and that may not entirely be artifacts of the methods used for reconstructing the paleo-ELA. However, our results do contradict the claim of Eugster et al. (2016) that the major LGM advance in Chandra valley was due to the hemisphere-scale temperature forcing alone, notwithstanding any regional influences (Gillespie and Molnar, 1995) such as that of a weakened ISM (Benn and Owen, 1998; Owen et al., 2002; Schäfer et al., 2002). We provide strong evidence that the LGM glacier response in the Chandra valley was driven by a combination of a large climate sensitivity of the main-trunk glacier and a relatively weak ELA forcing, which is consistent with the effect of a weakened ISM. The present study underlines the need to consider the variability of climate sensitivity of glaciers, preferably through a glacier dynamic simulation, while trying to infer about the nature and magnitude of the climate forcing driving any particular paleo-glacial event.

Eugster et al. (2016) emphasised that the post-LGM deglaciation was very rapid in the Chandra valley with a retreat of several tens of km within a few millennia. This is also consistent with our our computed response time for the main-trunk glacier, which is about 110 years. This suggests that the glacier was able to keep pace with the prevalent climate forcing over the millennial time scales.

ACKNOWLEDGMENTS

The development of the numerical model used in this research was supported by MoES grant MoES/PAMC/H&C/80/2016-PC-II, AJ is supported by MoES/PAMC/H&C/79/2016-PC-II, and SP is supported by ESSO-NCPOR grant NCAOR/2018/HiCOM/04 under HiCOM initiative.

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FIGURES

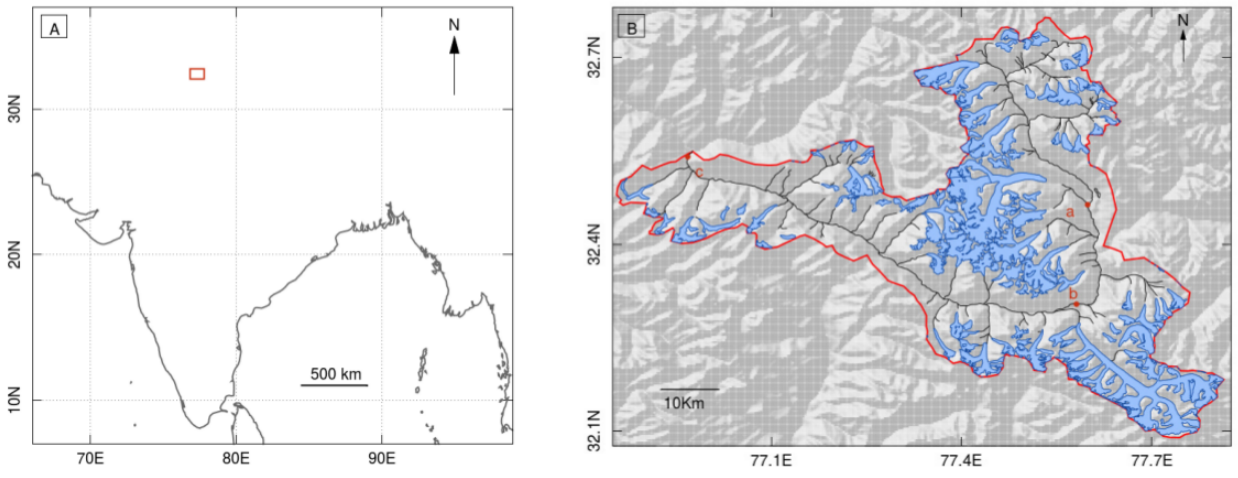

Figure 1. A: The location of the study area is shown with a red rectangle. B: A: A map of Chandra catchment showing the present glacier extents (light blue shaded polygons) and river network (black solid lines). The confluences of the Chandra river with Samuudra Tapu and Bara Shigri Glaciers and Bhaga river are marked with letters a, b and c, respectively.

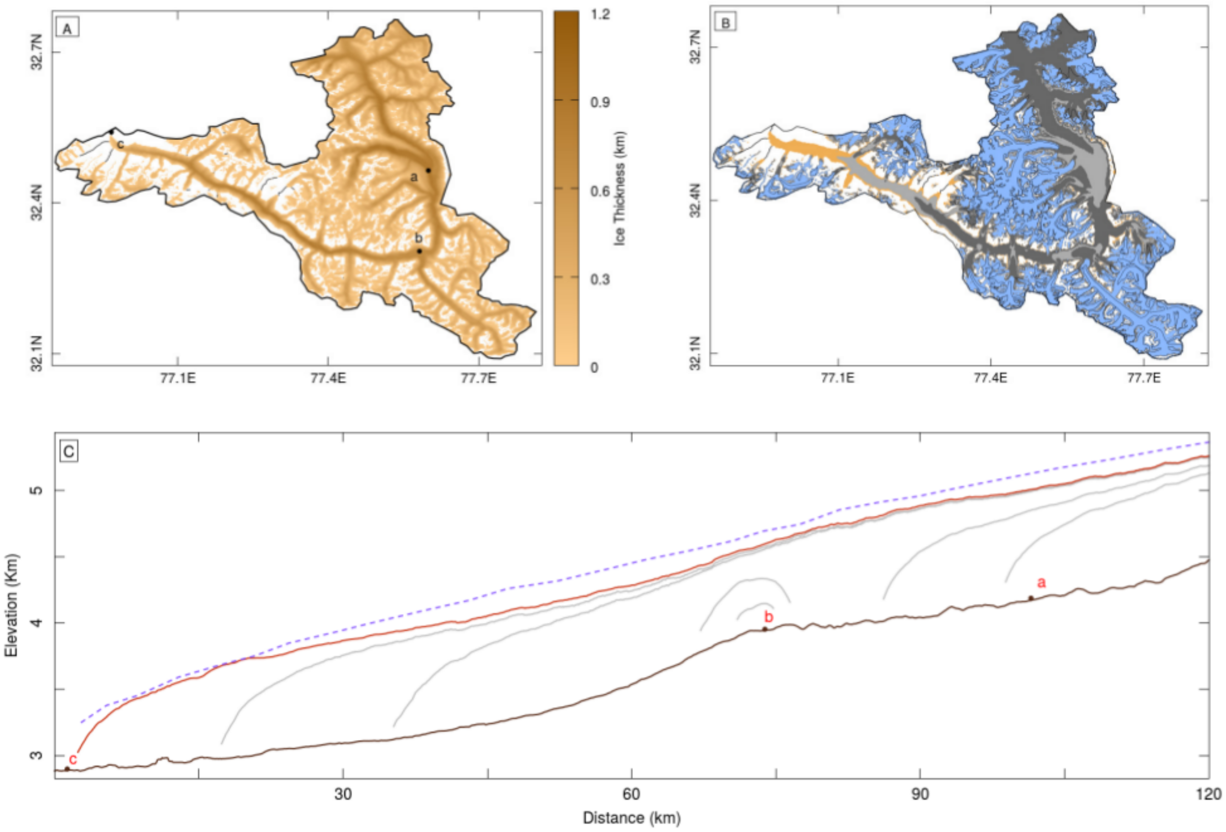

Figure 2. A: The ice-thickness map of the steady-state corresponding to a lowering of ELA by 300 m, resembles the reconstructed glacier extent during 18-19 ka in the valley. B: The simulated steady-state glaciers corresponding to the present extent (light blue shaded polygon) and LGM extent (brown shaded polygon) are shown together with a set of intermediate steady-states corresponding to ELA depression of 100 m , 150 m, 200 m, and 250 m (alternate dark and light gray shaded area). C: The ice thickness profiles for the modeled 18-19 ka state (red solid line), and the states corresponding to ELA depression of 100 m, 150 m, 200 m, and 250 m (gray solid lines) are shown. Purple dashed line denotes the reconstruction of the 18-19 ka advance (Eugster et al. , 2016). The brown line denotes the bedrock profile.