

# An investigation of the thermo-mechanical features of Laohugou Glacier No.12 in Mt. Qilian Shan, western China, using a two-dimensional first-order flowband ice flow model

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**Abstract.** By combining *in situ* measurements and a two-dimensional thermo-mechanically coupled ice flow model, we investigate the thermo-mechanical features of the largest valley glacier (Laohugou Glacier No.12; LHG12) in Mt. Qilian Shan located in the arid region of western China. Our model results suggest that LHG12, previously considered as fully cold, is probably polythermal, with a lower temperate ice layer overlain by an upper layer of cold ice over a large region of the ablation area. Modeled ice surface velocities match well with the *in situ* observations in the east branch (~~mainstream~~main branch) but clearly underestimate ~~the ice surface velocities~~those near the glacier terminus possibly because the convergent flow is ignored and the basal sliding beneath the confluence area is underestimated. The modeled ice temperatures are in very good agreement with the *in situ* measurements from a deep borehole (110 m deep) in the upper ablation area. The model results are sensitive to surface thermal boundary conditions, for example, surface air temperature and near-surface ice temperature. In this study, we use a Dirichlet surface thermal condition constrained by 20 m borehole temperatures and annual surface air temperatures. Like many other alpine glaciers, strain heating is ~~an important parameter~~important in controlling the englacial thermal structure ~~in of~~ LHG12. Our transient simulations indicate that the accumulation zone becomes colder during the last two decades as a response to the elevated equilibrium-line altitude and the rising summer air temperatures. We suggest that the extent of accumulation basin (the amount of refreezing latent heat from meltwater) of LHG12 ~~may have~~has a considerable impact on the englacial thermal status.

## 1 Introduction

The storage of water in glaciers is an important component of the hydrological cycle at different time scales (Jansson et al., 2003; Huss et al., 2010), especially in arid and semi-arid regions such as northwestern China, where many glaciers are currently retreating and disappearing (Yao et al., 2012; Neckel et al., 2014; Tian et al., 2014). Located on the northeastern edge of the

Tibetan Plateau (36 – 39 °N, 94 – 104 °E), Mt. Qilian Shan (MQS) develops 2051 glaciers covering an area of approximately 1057 km<sup>2</sup> with a total ice volume of approximately ~~50.5~~ 51 km<sup>3</sup> (Guo et al., 2014, 2015). Meltwater from MQS glaciers is a very important water resource for the agricultural irrigation and socio-economic development of the oasis cities in northwestern China. Thus, the changes in the MQS glaciers that occur as the climate becomes warmer in the near future are of concern.

5 Due to logistic difficulties, few MQS glaciers have been investigated in previous decades. However, Laohugou Glacier No.12 (hereafter referred to as LHG12), the largest valley glacier of MQS, has been investigated. Comprised of two branches (east and west), LHG12 is located on the north slope of the western MQS (39°27' N, 96°32' E; Fig. 1), with a length of approximately ~~9.8~~ 10 km, an area of approximately ~~20.4~~ 20 km<sup>2</sup>, and an elevation range of 4260 – 5481 m a.s.l. (Liu et al., 2011). LHG12 was first studied by a Chinese expedition ~~from~~ during 1958 – 1962 and was considered again in short-term field campaigns in  
10 the 1970s and 1980s that were aimed at monitoring glacier changes (Du et al., 2008). Since 2008, the Chinese Academy of Sciences has operated a field station for obtaining meteorological and glaciological measurements of LHG12.

The temperature distribution of a glacier primarily controls the ice rheology, englacial hydrology, and basal sliding conditions (Blatter and Hutter, 1991; Irvine-Fynn et al., 2011; Schäfer et al., 2014). A good understanding of the glacier thermal conditions is important for predicting glacier response to climate change (Wilson et al., 2013; Gilbert et al., 2015), improving glacier  
15 hazard analysis (Gilbert et al., 2012), and reconstructing past climate histories (Vincent et al., 2007; Gilbert et al., 2010). ~~The thermal regime of a glacier is mainly controlled by the surface thermal boundary conditions. For example, Different energy inputs at the glacier surface determine the~~ near-surface ~~warming from refreezing meltwater and cooling from the cold air of crevasses influence the thermal regimes of glaciers~~ ice temperature. In particular, the latent heat released by meltwater refreezing in the percolation zone can largely warm the near-surface area (e.g., Müller, 1976; Huang et al., 1982; Blatter, 1987; Blatter and Kappenberger, 1988; Gilbert et al., 2014b). The importance of surface thermal boundary condition in controlling the thermal regime of a glacier has been recognized by recent thermo-mechanical glacier model studies (e.g., Wilson and Flowers, 2013; Gilbert et al., 2014a; Meierbachtol et al., 2015). Using both *in situ* measurements and numerical models, Meierbachtol et al. (2015) argued that shallow borehole ice temperatures served as better boundary constraints than surface air temperatures in Greenland. However, for the east Rongbuk glacier on Mt. Everest, which is considered polythermal, Zhang et al.  
20 (2013) found that the modeled ice temperatures agreed well with the *in situ* shallow borehole observations when using surface air temperatures as the surface thermal boundary condition. Therefore, careful investigation of the upper thermal boundary condition is highly necessary for glaciers in different regions under different climate conditions.

LHG12 is widely considered as an extremely continental type (cold) glacier and is characterized by low temperatures and precipitation (Huang, 1990; Shi and Liu, 2000). However, in recent years, we have observed extensive and widespread melt-  
30 water on the ice surface and at the glacier terminus. In addition, percolation of snow meltwater consistently occurs in the accumulation basin during the summer. Therefore, we address the following two pressing questions in this study: (i) What is the present thermal status of LHG12? and (ii) How do different surface thermal boundary conditions impact the modeled ice temperature and flow fields? Because temperate ice can assist basal ~~slip~~ motion and accelerate glacier retreat, understanding the current thermal status of LHG12 is very important for predicting its future dynamic behaviour.

To answer these questions, we conduct diagnostic and transient simulations for LHG12 by using a thermo-mechanically coupled first-order flowband ice flow model. This paper is organized as follows: First, we provide a detailed description of the glaciological datasets of LHG12. Then, we briefly review the numerical ice flow model and the surface mass balance model used in this study. Next, we perform both diagnostic and transient simulations. In the diagnostic simulations, we first investigate the sensitivities of ice flow model parameters, and we then ~~compare the diagnostic results with measured ice surface velocities and ice temperature profile obtained from a deep borehole, and we also~~ explore the impacts of different surface thermal boundary conditions and assess the contributions of heat advection, strain heating, and basal sliding to the temperature field of LHG12. Transient simulations are performed during the period ~~1961–2013. The~~ 1961–2011. We then compare the transient results with measured ice surface velocities and ice temperature profile obtained from a deep borehole. We also investigate the evolution of the temperature profile in the deep borehole ~~is then investigated~~ and the impacts of initial equilibrium line altitude (ELA) on the thermal regime of the glacier. Finally, we discuss the limitations of our model and present the important conclusions that resulted from this study.

## 2 Data

Most *in situ* observations, e.g., borehole ice temperatures, surface air temperatures and ice surface velocities, have been made on the east branch (~~mainstream~~ main branch) of LHG12 (Fig. 1). Measurements on the west branch are sparse and temporally discontinuous. Thus, we only consider the *in situ* data from the east tributary when building our numerical ice flow model.

### 2.1 Glacier geometry

In July – August 2009 and 2014, two ground-penetrating radar (GPR) surveys were conducted on LHG12 using a pulseEKKO PRO system with center frequencies of 100 MHz (2009) and 50 MHz (2014) (Fig. 1b). Wang et al. (2016) have presented details regarding the GPR data collection and post-processing.

As shown in Fig. 2, the east branch of LHG12 has a mean ice thickness of approximately 190 m. We observed the thickest ice layer (approximately 261 m) at 4864 m a.s.l. Generally, the ice surface of LHG12 is gently undulating, with a mean slope of  $0.08^\circ$ , and the bed of LHG12 shows significant overdeepening in the middle of the center flowline (CL) (Fig. 2). To account for the lateral effects exerted by glacier valley walls in our 2D ice flow model, we parameterize the lateral drag using the glacier half widths. Based on the GPR measurements on LHG12, we parameterize the glacier cross-sections by a power law function  $z = aW(z)^b$ , where  $z$  and  $W(z)$  are the vertical and horizontal distances from the lowest point of the profile, and  $a$  and  $b$  are constants representing the flatness and steepness of the glacier valley, respectively (Svensson, 1959). The  $b$  values for LHG12 range from 0.8 to 1.6, indicating that the valley is approximately “V”-shaped (Wang et al., 2016). As an input for the flowband ice flow model, the glacier width,  $W$ , was also calculated by ignoring all tributaries (including the west branch) (Fig. 1b and 2).

## 2.2 Ice surface velocities

The ~~ice~~ surface velocities of ~~the ice in~~ LHG12 were determined from repeated surveys of stakes drilled into the ice surface. All stakes were located in the distance between km 0.6 – 7.9 along the CL (Fig. 3), spanning an elevation range of 4355 – 4990 m.a.s.l. (Fig. 1b). We measured the stake positions using a real-time kinematic (RTK) fixed solution by a South Lingrui S82 GPS system (Liu et al., 2011). The accuracy of the GPS positioning is ~~an~~ on the order of a few centimeters and the uncertainty of the calculated ice surface velocities is estimated to be less than  $1 \text{ m a}^{-1}$ . Because it is difficult to conduct fieldwork on LHG12 (due to, e.g., crevasses and supra-glacial streams), it was nearly impossible to measure all stakes each observational year. Thus, the current dataset includes annual ice surface velocities from 2008 – 2009 and 2009 – 2010, summer measurements from June 17 – August 30, 2008, and winter measurements from February 1 – May 28, 2010.

The *in situ* ice surface velocities shown in Fig. 3 are all from stakes near the CL (Fig. 1b). Small ice surface velocities ( $< 17 \text{ m a}^{-1}$ ) are clearly visible in the upper accumulation (km 0 – 1.2) and lower ablation areas (km 6.5 – 9.0) (Fig. 3). Fast ice flow ( $> 30 \text{ m a}^{-1}$ ) can be observed between elevations of 4700 – 4775 m a.s.l. (km 4.0 – 5.0), where the ice surface velocities during the summer are approximately  $6 \text{ m a}^{-1}$  greater than the annual mean velocity ( $< 40 \text{ m a}^{-1}$ ). Measurements of winter ice surface velocities ( $< 10 \text{ m a}^{-1}$ ) are only available near the glacier terminus showing a clear inter-annual variation of the ice flow speeds.

## 2.3 Borehole ice temperature

In August 2009 and 2010, we drilled three 25 m deep shallow boreholes on LHG12 (Fig. 1b). One borehole was drilled in the upper ablation area (site 2, approximately 4900 m a.s.l.) and two boreholes were drilled at the AWS locations (sites 1 and 3). The snow/ice temperatures were measured in the boreholes during the period of October 1, 2010 – September 30, 2011. The seasonal variations of the snow/ice temperatures in the shallow boreholes are presented in Figs. 4a, b and c. Our measurements show very little fluctuation ( $\pm 0.4 \text{ K}$ ) in the ice temperatures over the depth range of 20 – 25 m. Below the 3 m depth, the annual mean temperature profiles for sites 1 and 2 show a linearly increase in temperature with depth, while the annual mean temperature profile for site 3 is convex upward. The mean annual 20 m ice temperatures ( $T_{20\text{m}}$ ) at sites 1, 2, and 3 are 5.5 K, 3.0 K, and 9.5 K higher than the mean annual air temperatures ( $T_{\text{air}}$ ), respectively. Despite its higher elevation, the near-surface snow/ice temperatures below a depth of 5 m at site 3 are greater than ~~the near-surface snow/ice temperatures~~ those in the ablation area (sites 1 and 2), largely due to the latent heat released as the meltwater entrapped in the surface snow layers refreezes.

To determine the englacial thermal conditions of LHG12, we drilled ~~a deep ice core (an ice core to a depth of 167 m)~~ in the upper ablation area of LHG12 (~~approximately~~ 4971 m a.s.l., Fig. 1b). In October 2011, the ice temperature were measured to a depth of approximately 110 m using a thermistor string after 20 days of the drilling, as shown in Fig. 4d. The string consists of 50 temperature sensors with a vertical spacing of 0.5 m and 10 m at the ice depths of 0 – 20 m and 20 – 110 m, respectively. The accuracy of the temperature sensor is around  $\pm 0.05 \text{ K}$  (Liu et al., 2009). From Fig. 4d we can see that the temperature profile is close to linear with a temperature gradient of around  $0.1 \text{ K m}^{-1}$  at the depths of 9 – 30 m. Below the depth of 30

m, the ice temperature demonstrates a linear relationship with depth as well but with a smaller temperature gradient of around  $0.034 \text{ K m}^{-1}$ .

## 2.4 Meteorological data

Two automatic weather stations (AWS) were deployed on LHG12, one in the ablation area at 4550 m a.s.l. (site 1, Fig. 1b), and the other in the accumulation area at 5040 m a.s.l. (site 3). During the period of 2010 – 2013, the mean annual air temperatures (2 m above the ice surface) at sites 1 and 3 were  $-9.2^\circ\text{C}$  and  $-12.2^\circ\text{C}$ , respectively, indicating a lapse rate of  ~~$-0.0061$~~   $-0.0061 \text{ K m}^{-1}$ .

The historical knowledge of the surface air temperature and the ELA of LHG12 is a necessity for running the transient model. We reconstruct the past air temperature on LHG12 based on the observations of the Tuole meteorological station (3367 m a.s.l.), which is approximately 175 km southeast to LHG12. We get the precipitation on LHG12 by downscaling the CAPD (China Alpine Region Month Precipitation Dataset) in Qilian Shan. CAPD provides the monthly sum of precipitation during the period of 1960 – 2013 with a grid spacing of 1 km. We calculate the precipitation on LHG12 from its surrounding 91 grids in CAPD based on the relationship between precipitation and geometric parameters. More details about the reconstruction of both air temperature and precipitation for LHG12 can be found in the Supplement.

## 3 ~~Thermomechanical~~ Thermo-mechanical ice flow model

In this study, we ~~used~~ use the same two-dimensional (2D), thermo-mechanically coupled, first-order, flowband ice flow model as in Zhang et al. (2013). Therefore, we only address a very brief review of the model here.

### 3.1 Ice flow model

We define  $x$ ,  $y$ , and  $z$  as the horizontal along-flow, horizontal across-flow and vertical coordinates, respectively. By assuming the vertical normal stress as hydrostatic and neglecting the bridging effects (Blatter, 1995; Pattyn, 2002), the equation for momentum balance is given as

$$\frac{\partial}{\partial x}(2\sigma'_{xx} + \sigma'_{yy}) + \frac{\partial \sigma'_{xy}}{\partial y} + \frac{\partial \sigma'_{xz}}{\partial z} = \rho g \frac{\partial s}{\partial x}, \quad (1)$$

where  $\sigma'_{ij}$  is the deviatoric stress tensor,  $\rho$  is the ice density,  $g$  is the gravitational acceleration, and  $s$  is the ice surface elevation. The parameters used in this study are given in Table 1.

The constitutive relationship of ice dynamics is described by the Glen's flow law (Cuffey and Paterson, 2010)

$$\sigma'_{ij} = 2\eta \dot{\epsilon}_{ij}, \quad \eta = \frac{1}{2} A^{-1/n} (\dot{\epsilon}_e + \dot{\epsilon}_0)^{(1-n)/n}, \quad (2)$$

where  $\eta$  is the ice viscosity,  $\dot{\epsilon}_{ij}$  is the strain rate,  $n$  is the flow law exponent,  $A$  is the flow rate factor,  ~~$\dot{\epsilon}_e$~~   $\dot{\epsilon}_e$  is the effective strain rate, and  $\dot{\epsilon}_0$  is a small number used to avoid singularity. The flow rate factor is parameterized using the Arrhenius relationship

as

$$A(T) = A_0 \exp\left(-\frac{Q}{RT}\right), \quad (3)$$

where  $A_0$  is the pre-exponential constant,  $Q$  is the activation energy for creep,  $R$  is the universal gas constant, and  $T$  is the ice temperature. The effective strain rate  $\dot{\epsilon}_e$  is related to the velocity gradient by

$$\dot{\epsilon}_e^2 \simeq \left(\frac{\partial u}{\partial x}\right)^2 + \left(\frac{\partial v}{\partial y}\right)^2 + \frac{\partial u}{\partial x} \frac{\partial v}{\partial y} + \frac{1}{4} \left(\frac{\partial u}{\partial y}\right)^2 + \frac{1}{4} \left(\frac{\partial u}{\partial z}\right)^2, \quad (4)$$

where  $u$  and  $v$  are the velocity components along the  $x$  and  $y$  direction, respectively. By assuming  $\partial v / \partial y = (u/W)(\partial W / \partial x)$ , we parameterize the lateral drag,  $\sigma'_{xy}$ , as a function of the flowband half width,  $W$ , following Flowers et al. (2011)

$$\sigma'_{xy} = -\frac{\eta u}{W}. \quad (5)$$

For an easy numerical implementation, we reformulate the momentum balance equation (1) as

$$\begin{aligned} & \frac{u}{W} \left\{ 2 \frac{\partial \eta}{\partial x} \frac{\partial W}{\partial x} + 2 \eta \left[ \frac{\partial^2 W}{\partial x^2} - \frac{1}{W} \left( \frac{\partial W}{\partial x} \right)^2 \right] - \frac{\eta}{W} \right\} \\ 10 \quad & + \frac{\partial u}{\partial x} \left( 4 \frac{\partial \eta}{\partial x} + \frac{2 \eta}{W} \frac{\partial W}{\partial x} \right) + \frac{\partial u}{\partial z} \frac{\partial \eta}{\partial z} + 4 \eta \frac{\partial^2 u}{\partial x^2} + \eta \frac{\partial^2 u}{\partial z^2} = \rho g \frac{\partial s}{\partial x}, \end{aligned} \quad (6)$$

where the ice viscosity is defined-expressed as

$$\eta = \frac{1}{2} A^{-1/n} \left[ \left( \frac{\partial u}{\partial x} \right)^2 + \left( \frac{u}{W} \frac{\partial W}{\partial x} \right)^2 + \frac{u}{W} \frac{\partial u}{\partial x} \frac{\partial W}{\partial x} + \frac{1}{4} \left( \frac{\partial u}{\partial z} \right)^2 + \frac{1}{4} \left( \frac{u}{W} \right)^2 + \dot{\epsilon}_0^2 \right]^{(1-n)/2n}. \quad (7)$$

### 3.2 Ice temperature model

The ice temperature field can be calculated using a 2D heat transfer equation (Pattyn, 2002),

$$15 \quad \frac{\partial T}{\partial t} = \frac{k}{\rho c_p} \left( \frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial z^2} \right) - \left( u \frac{\partial T}{\partial x} + w \frac{\partial T}{\partial z} \right) + \frac{4 \eta \dot{\epsilon}_e^2}{\rho c_p} \frac{4 \eta \dot{\epsilon}_e^2}{\rho c_p}, \quad (8)$$

where  $w$  is the vertical ice velocity,  $k$  and  $c_p$  are the thermal conductivity and heat capacity of the ice, respectively.

The pressure melting point of the ice,  $T_{\text{mp}}$ , is described by the Clausius-Clapeyron relationship

$$T_{\text{mp}} = T_0 - \beta(s - z), \quad (9)$$

where  $T_0$  is the triple-point temperature of water and  $\beta$  is the Clausius-Clapeyron constant. Following Zhang et al. (2013), we determined the position of the cold-temperate ice transition surface (CTS) by considering the following two cases: (1) melting condition, i.e., cold ice flows downward into the temperate ice zone, and (2) freezing condition, i.e., temperate ice flows upward into the cold ice zone (Blatter and Hutter, 1991; Blatter and Greve, 2015). For the melting case, the ice temperature

profile at the CTS simply follows a Clausius-Clapeyron gradient ( $\beta$ ). However, for the freezing case, the latent heat,  $Q_r$ , that is released when the water contained in the temperate zone refreezes is determined as (Funk et al., 1994)

$$Q_r = \mu w \rho_w L, \quad (10)$$

where  $\mu$  is the fractional water content of the temperate ice,  $\rho_w$  is the water density and  $L$  is the latent heat of freezing. In this case, following Funk et al. (1994), the ice temperature gradient at the CTS can be described as

$$\frac{\partial T}{\partial z} = -\frac{Q_r}{k} + \beta. \quad (11)$$

### 3.3 Free surface

The free surface evolution follows the kinematic boundary equation,

$$\frac{\partial s}{\partial t} = b_n + w_s - u_s \frac{\partial s}{\partial x}, \quad (12)$$

where  $s(x, t)$  is the free surface elevation,  $u_s$  and  $w_s$  are the surface velocity components in  $x$  and  $z$ , and  $b_n$  is the surface mass balance.

### 3.4 Boundary conditions

#### 3.4.1 Boundary conditions for ice flow model

For the ice flow model, we assume a stress-free condition for the glacier surface, and use the Coulomb friction law to describe the ice-bedrock interface where the ice slips (Schoof, 2005),

$$\tau_b = \Gamma \left( \frac{u_b}{u_b + \Gamma^n N^n \Lambda} \right)^{1/n} N, \quad \Lambda = \frac{\lambda_{\max} A}{m_{\max}}, \quad (13)$$

where  $\tau_b$  and  $u_b$  are the basal drag and velocity, respectively,  $N$  is the basal effective pressure,  $\lambda_{\max}$  is the dominant wavelength of the bed bumps,  $m_{\max}$  is the maximum slope of the bed bumps, and  $\Gamma$  and  $\Lambda$  are geometrical parameters (Gagliardini et al., 2007). Here we take  $\Gamma = 0.84 m_{\max}$  following ~~and~~ Gagliardini et al. (2007) and Flowers et al. (2011). The basal effective pressure in the friction law,  $N$ , is defined as the difference between the ice overburden pressure and the basal water pressure (Gagliardini et al., 2007; Flowers et al., 2011),

$$N = \rho g H - P_w = \phi \rho g H, \quad (14)$$

where  $H$  and  $P_w$  are the ice thickness and basal water pressure, respectively, and  $\phi$  implies the ratio of basal effective pressure to the ice overburden pressure. The basal drag is defined as the sum of all resistive forces (Van der Veen and Whillans, 1989; Pattyn, 2002). It should be noted that the basal sliding is only permitted when basal ice temperature reaches the local pressure-melting point.

### 3.4.2 Boundary conditions for ice temperature model

We apply a Dirichlet temperature constraint ( $T_{\text{sbc}}$ ) on the ice surface in the temperature model. In some studies,  $T_{\text{sbc}} = T_{\text{air}}$  is used (e.g. Zhang et al., 2013), which, as suggested by recent studies, could result in ~~lower-velocity-values-and~~ cold bias in ice temperature simulations (Meierbachtol et al., 2015). By contrast, Meierbachtol et al. (2015) recommended using the  
5 near-surface temperature  $T_{\text{dep}}$  at depth where inter-annual variations of air temperatures are damped (15 – 20 m deep) (~~as~~ a proxy for the annual mean ice surface temperature). One advantage of using  $T_{\text{dep}}$  is that the effects of refreezing meltwater and the thermal insulation of winter snow can be included in the model (Huang et al., 1982; Cuffey and Paterson, 2010). In fact, the condition  $T_{\text{dep}} = T_{\text{air}}$  is acceptable only in dry and cold snow zones (Cuffey and Paterson, 2010); however,  $T_{\text{dep}} > T_{\text{air}}$  is often observed in zones where meltwater is refreezing in glaciers, such as the LHG12 (Fig. 4). In this study, we set  $T_{\text{sbc}}$  in  
10 the accumulation zone to the glacier near-surface temperature  $T_{\text{dep}}$ , while  $T_{\text{sbc}}$  in the ablation area is prescribed by a simple parameterization (Lüthi and Funk, 2001; Gilbert et al., 2010) as

$$T_{\text{sbc}} = \begin{cases} T_{\text{dep}}, & s > ELA, \\ T_{\text{air}} + c, & s \leq ELA, \end{cases} \quad (15)$$

where ELA is the equilibrium-line altitude, and  $c$  is a tuning parameter implicitly accounting for effects of the surface energy budget and the lapse rate of air temperature (Gilbert et al., 2010). We ~~denote-use~~ Eq. (15) as the surface thermal boundary  
15 condition of the reference experiment (~~E-ref~~) after comparing another numerical experiment by setting  $T_{\text{sbc}} = T_{\text{air}}$  (E-air) (see Sect. 6.1.2 for details).

At the ice-bedrock interface, we apply the following Neumann-type boundary condition in the temperature model,

$$\frac{\partial T}{\partial z} = -\frac{G}{k}, \quad (16)$$

where  $G$  is the geothermal heat flux. We here use a constant geothermal heat flux,  $40 \text{ mW m}^{-2}$ , an *in situ* measurement from  
20 the Dundee ice cap in the western MQS (Huang, 1999), over the entire model domain.

### 3.5 Numerical solution

In our model we use a finite difference discretization method and a terrain-following coordinate transformation. The numerical mesh we use contains 61 grid points in  $x$  and 41 layers in  $z$ . The ice flow model (Eq. (6)) is discretized with a second-order centered difference scheme while the ice temperature model (Eq. (8)) employs a first-order upstream difference scheme  
25 for the horizontal heat advection term and a node-centered difference scheme for the vertical heat advection term and the heat diffusion terms. The velocity and temperature fields are iteratively solved by a relaxed Picard subspace iteration scheme (De Smedt et al., 2010) in Matlab. The free surface evolution (Eq. (12)) is solved using a Crank-Nicolson scheme. More details are given in Zhang et al. (2013).



## 4 Surface mass balance model

In our parameterization of the surface thermal boundary condition, a transient ELA is important in controlling the extent of accumulation zone which can be largely warmed by the refreezing of meltwater. We estimate the annual surface mass balance  $b_n$  of LHG12 ~~between 1960 and 2013, and find the location during 1960–2013, and determine the position~~ of ELA by  $b_n = 0$ .

- 5 The daily ablation at elevation  $z$ ,  ~~$a(z)m_a(z)$~~ , is calculated based on a degree-day method (Braithwaite and Zhang, 2000):

$$\underline{am_a}(z) = f_m \text{PDD}(z), \quad (17)$$

where  $f_m$  is the degree-day factor, and PDD is the daily positive degree-day sum for glacier surface melt,

$$\text{PDD} = T_{\text{mean}} - T_m, \quad (18)$$

- where  $T_{\text{mean}}$  is the daily mean air temperature, and  $T_m$  is the threshold temperature when melt occurs. Note that surface melt  
10 may happen even if  $T_{\text{mean}} < 0^\circ\text{C}$ , due to the positive air temperature during the day, indicating a negative  $T_m$  in such cases.

As the CAPD dataset only provides monthly precipitation sums, we estimate the daily precipitation on LHG12 by assuming a uniformly distributed precipitation over a month. The daily accumulation at elevation  $z$ ,  ~~$e(z)m_c(z)$~~ , is calculated as follows,

$$\underline{cm_c}(z) = \begin{cases} f_P \cdot P_{\text{total}}, & T_{\text{mean}} < T_{\text{crit1}}, \\ f_P \cdot \frac{T_{\text{crit2}} - T_{\text{mean}}}{T_{\text{crit2}} - T_{\text{crit1}}} \cdot P_{\text{total}}, & T_{\text{crit1}} \leq T_{\text{mean}} \leq T_{\text{crit2}}, \\ 0, & T_{\text{mean}} > T_{\text{crit2}}, \end{cases} \quad (19)$$

- 15 where  $P_{\text{total}}$  is the daily precipitation,  $T_{\text{crit1}}$  and  $T_{\text{crit2}}$  are the threshold temperatures for the snow and rain transition, and  $f_P$  is a tuning parameter for the precipitation to account for the uncertainties of the gridded CAPD data and the downscaling method. The model is well calibrated by investigating the sensitivities of model parameters and by comparing the simulated results with the observed mass balance ~~in during~~ 2010–12 (~~See details in the see~~ Figs. S5–8 in the Supplement).

## 5 Simulation strategies

### 20 5.1 Surface relaxation

In order to remove the uncertainties remained in the model initial conditions (including initial topography and model parameters) (Zwinger and Moore, 2009; Seroussi et al., 2013), we allow the free surface to relax for a period of 3 years ~~assuming zero~~  
~~under a constant present-day~~ surface mass balance and ~~zero~~ basal sliding (~~See see~~ details in the Supplement). The time step for the relaxation experiment is set to 0.1 year. We apply surface relaxation before running all of the diagnostic and transient  
25 simulations in this ~~paperstudy~~.

### 5.2 Diagnostic simulations

In our diagnostic simulations, we assume a thermal steady-state condition and use the relaxed present-day geometry of LHG12. First, we explore the model sensitivities to geometrical bed parameters ( $\lambda_{\max}$  and  $m_{\max}$ ), ratio of basal effective pressure ( $\phi$ ), water content ( $\mu$ ), geothermal heat flux ( $G$ ), and the valley shape index ( $b$ ). Next, we tune the using diagnostic simulations with relaxed present-day geometry of LHG12. Then, we calibrate the parameters in surface thermal boundary condition parameters, i.e., ELA, ( $ELA_0$ ,  $T_{\text{dep}}$  and  $c$ ) to fit the modeled steady-state temperature profile to observations in the deep borehole (Fig. 1). We then investigate the modeled horizontal velocity field and thermal structure of LHG12, where  $ELA_0$  is the initial ELA, by first running a diagnostic simulation with a given set of parameters ( $ELA_0$ ,  $T_{\text{dep}}$  and  $c$ ) and compare the diagnostic simulation results with observations. We also c) and then performing a transient simulation using the diagnostic run as an initial condition with time-dependent ELA and surface air temperature (see Sect. 5.3). This numerical process of finding the optimal parameter set of  $ELA_0$ ,  $T_{\text{dep}}$  and  $c$  is repeated until we find a good agreement between the modeled and measured deep borehole temperatures in 2011. Finally, we perform experiments to investigate sensitivity of the thermomechanical the sensitivities of the thermo-mechanical model to different surface thermal boundary conditions (E-ref-E-ref-D and E-air). Finally, we We also perform four experiments (E-advZ, E-advX, E-strain, and E-slip) to explore the effects of heat advection, strain heating, and basal sliding on the thermal distribution and flow dynamic behaviours of LHG12.

### 5.3 Transient simulations

To investigate the impacts of historical climate conditions on the thermal regime of LHG12, time-dependent simulations are performed from 1961 to 2011. The diagnostic run with the calibrated surface thermal parameters is used as the initial condition of the transient simulations. In our simulations, we assume that the surface temperature ( $T_{\text{dep}}$ ) in the accumulation zone is constant in time and space. Due to a lack of in-situ in situ observations (e.g. firn thickness, firn densities) and coupled heat-water transfer model (e.g. Gilbert, 2012; Wilson, 2013), we do not simulate the complex processes of heat exchange in the accumulation zone. To understand how the thermal status varies over time in the deep borehole, we design three (cold, warm and reference) transient simulations transient simulations (E-cold, E-warm, E-ref-T) by setting  $T_{\text{dep}}$  to  $-5^\circ\text{C}$ ,  $-1^\circ\text{C}$  and  $-2.1^\circ\text{C}$   $-1.8^\circ\text{C}$  (as calibrated in our steady-state diagnostic simulations; see Sect. 6.1.1), respectively. We also design two experiments (E-high and E-low) to explore the impacts of the initial ELA ( $ELA_0$ ) on the thermal regime of LHG12.

In the transient model, the ice surface temperature ( $T_{\text{src}}$  mean annual air temperature ( $T_{\text{air}}$ ) and ELA are allowed to vary in time. However, we keep the glacier geometry fixed in the transient simulations for the following two reasons: two main reasons: (1) The tributaries of LHG12, which our flowband model neglects, may have non-negligible inflow ice fluxes that impact the mass continuity equation; (2) The mean surface elevation change above 4600 m a.s.l. (the confluence area) over 1957–2014 is close to negligible (approximately  $-10.4$  m, around 4.4% of the mean ice thickness) (See Fig. S10 in the Supplement). Before we run the transient experiments, we perform a 10-year spin-up using the mean values of surface air temperatures and ELAs during 1961–1970.

## 6 Simulation results and discussions

### 6.1 Diagnostic simulations

#### 6.1.1 Parameter sensitivity

As shown in Figs. 5 and 6, we conduct a series of sensitivity experiments to investigate the relative importance of different model parameters ( $\lambda_{\max}$ ,  $m_{\max}$ ,  $\phi$ ,  $\mu$ ,  $G$ ,  $b$ ) on ice flow speeds and temperate ice zone (TIZ) sizes by varying the value of one parameter while holding the others fixed. We set ELA to 4980 m a.s.l., as observed in the 2011 remote-sensing-Landsat image of LHG12.  $T_{\text{dep}}$  ~~and  $c$  are~~ is set to  $-2.7^{\circ}\text{C}$  ~~and which is the 20 m deep ice temperature at site 3.  $c$  is set to  $-4.3^{\circ}\text{C}$ , which are calculated from the~~ is the mean differences between the 20 m deep ice temperatures and mean annual air temperatures for the two shallow boreholes in the accumulation basin, respectively, at sites 1 and 2.

The friction law parameters,  $\lambda_{\max}$  and  $m_{\max}$ , which describe the geometries of bedrock obstacles (Gagliardini et al., 2007; Flowers et al., 2011), have non-negligible impacts on the model results. As shown in Figs. 5a and b, the modeled velocities and TIZ sizes increase as  $\lambda_{\max}$  increases and  $m_{\max}$  decreases, similar to the results observed by Flowers et al. (2011) and Zhang et al. (2013). A large increase in the modeled basal sliding velocity occurs when  $m_{\max}$  is lower than 0.2. The ratio,  $\phi$ , is an insensitive parameter in our model when it is larger than 0.3 (Fig. 5c). Although the water content,  $\mu$ , in the ice does not directly impact the ice velocity simulations (the flow rate factor  $A$  is assumed independent of the water content in ice), ~~the water content~~ it can affect the temperature field ~~and, consequently,~~ and consequently influence  $A$  and the ice velocities (Fig. 5d). Fig. 6d shows that increasing the water content may result in larger TIZ ~~sizes~~ size. In addition, we test the sensitivity of the model to different geothermal heat flux values. A larger geothermal heat flux can result in ~~larger TIZs~~ a larger TIZ but has a limited impact on ~~modeling modeled~~ ice velocity (Figs. 5e and 6e). As shown by Zhang et al. (2013), our model results are mainly controlled by the shape of the glacial valley, specifically the  $b$  index (Sect. 2.1). A large value of  $b$  indicates a flat glacial valley and suggests that a small lateral drag ~~was~~ is exerted on the ice flow (Figs. 5f and 6f).

Based on the sensitivity experiments described above, we adopt a parameter set of  $\lambda_{\max} = 4$  m,  $m_{\max} = 0.3$ ,  $\phi = 1$  (no basal water pressure),  $\mu = 3\%$ ,  $G = 40$  ~~mW m<sup>-2</sup>~~ mW m<sup>-2</sup>, and  $b = 1.2$  as a diagnostic reference in our modeling experiment (~~E-ref~~). ~~For a better fitting to the observed ice temperature in the deep borehole, we further tune the parameters ELA,  $T_{\text{dep}}$  and  $c$  by~~ E-ref-D). To calibrate the surface thermal parameters, we perform a series of model runs ~~using the above parameter set by~~ varying ELA<sub>0</sub>,  $T_{\text{dep}}$  and  $c$  in the range of 4940–5010 m a.s.l.,  $-3.3$ – $-1.5^{\circ}\text{C}$  and  $1$ – $6^{\circ}\text{C}$ , respectively. The performance of different parameter combinations is evaluated by comparing the ~~root-mean-squares~~ root-mean-squares (RMS) of the differences between the measured and modeled ~~temperature profiles for the~~ temperatures below the 40 m depth of the deep borehole (Fig. 7). We find the ~~RMS sufficiently small when ELA = 4990~~ RMSs are sufficiently small ( $< 0.11$  K) when ELA<sub>0</sub> = 4940 m a.s.l.,  $T_{\text{dep}} = -2.1^{\circ}\text{C}$ , and  $c = 4^{\circ}\text{C}$ , close to observations. By doing this we argue that the transient climate information of the past is partly included in the tuned parameters in our diagnostic simulations.  $T_{\text{dep}} = -1.8^{\circ}\text{C}$  and  $c = 3^{\circ}\text{C}$  (Fig. 7a–f and h). The sensitivity experiments of ELA<sub>0</sub> and their impacts on the thermal regime of LHG12 are discussed in Sect. 6.2.3.

### 6.1.2 Comparison with *in situ* observations

In the reference experiment (E-ref), we diagnostically simulate the distributions of horizontal ice velocities and temperatures (Figs. ??a and c). Next, the model results are compared to the measured ice surface velocities and the ice temperature profile in the deep borehole (Figs. ??b and d). Generally, the modeled ice surface velocities are in good agreement with *in situ* observations from the glacier head to km 4.8 along the CL (Fig. ??b). However, from km 4.8 to the glacier terminus, our model generally underestimates the ice surface velocities as shown in all simulations in Fig. 5. There are three possible reasons for this underestimation. First, the model neglects the convergent ice fluxes from the west branch. Second, an enhanced basal sliding zone may exist at the confluence area, which is not captured by the model. In addition, the diagnostic model using fixed topography cannot capture the time-dependent glacier changes which can largely influence the ice flow dynamics. Here we verify the first two hypotheses by conducting two other experiments, E-W and E-WS. In E-W the glacier widths are increased by 450 m at km 5.8–7.3 as a proxy of including the impact of the convergent flow from the west branch (Fig. 10). In E-WS, except for the same glacier width increasing as in E-W, we also increase  $\lambda_{\max}$  by 100 and decrease  $m_{\max}$  by 33 for accelerating the basal sliding at km 5.8–7.3 (Fig. 10). We can clearly find that while both factors have a non-negligible contribution to the model results, the basal sliding may play a bit more important role in the confluence area. This indicates a need of considering glacier flow branches and spatially variable sliding law parameters in real glacier modeling studies.

The model predicts a TIZ overlain by cold ice over a horizontal distance of km 0.6–7.4 (Fig. ??e). In addition, we further compare our model results with the *in situ* 110-m deep ice temperature measurements (Fig. ??d). Modeled and measured borehole temperature profiles show a close match within a root-mean-square difference of 0.3 K below the 10-m depth. Because *in situ* ice temperature data from below 110 m have not been obtained, we were unable to compare the modeled and measured ice temperatures at the ice-bedrock interface. Glacier thermal regime is largely influenced by the climate history and firm thickness. However, our diagnostic simulation assumes a thermal steady-state and applies a simple thermal boundary condition. The good agreement between modeled and measured temperature profiles is likely due that we select a snow line altitude representing a steady state for the glacier mass balance, and the surface thermal boundary condition includes the effect of refreezing meltwater in the accumulation zone.

### 6.1.2 Choice of surface thermal boundary condition

To investigate the impacts of different surface thermal boundary conditions on the thermo-mechanical fields of LHG12, we perform an experiment (E-air) in which we set  $T_{\text{sbc}} = T_{\text{air}}$ , and compare its results with those of E-ref.

Fig. ?? E-ref-D, Fig. 8a shows that the ice temperatures along the CL are highly sensitive to  $T_{\text{sbc}}$ . From E-air, it is observed that LHG12 becomes fully cold (Fig. 8b), with an average field temperature 6.5 K colder 6.3 K lower than that of E-ref E-ref-D (Fig. ??8a), which decreases the ice surface velocity by approximately 10.0 11 m a<sup>-1</sup> (Fig. ??b). In contrast to the measured deep borehole temperature profile at site 3, E-air results in an unreliable temperature profile that is much colder than the actual temperatures at all depths. 8c).

As noted above, the dynamics of LHG12 can be strongly influenced by the choices of different surface thermal boundary conditions. For LHG12, most accumulation and ablation events overlap ~~during the summer season in summer~~ (Sun et al., 2012). The meltwater entrapped in snow and moulins ~~during the summer season can release can refreeze at a depth where the temperature is below the melting point. Refreezing of meltwater releases~~ large amounts of ~~heat due to refreezing when the~~ ~~temperature decreases, which may significantly increase the ice temperatures in the latent heat, which can significantly raise the~~ near-surface snow/ice ~~layers (Fig. 4) temperatures~~ and result in the warm bias of  $T_{20m}$  ~~(compared with the mean annual surface air temperature compared to  $T_{air}$  (Fig. 4))~~. Therefore, compared with E-air, the experiment ~~E-ref better incorporate E-ref-D better incorporates~~ the effects of meltwater refreezing in the accumulation basin into the prescribed surface thermal boundary constraints resulting in more accurate simulations of ice temperature and flow fields.

### 6.1.3 Roles of heat advection, strain heating and basal sliding

To assess the relative contributions of heat advection and strain heating to the thermo-mechanical field of LHG12, we ~~conducted~~ ~~conduct~~ three experiments (E-advZ, E-advX, and E-strain), in which the vertical advection, horizontal along-flow advection and strain heating were neglected, respectively. In addition, to investigate the effects of basal sliding predicted by the Coulomb friction law on the thermal state and flow dynamics of LHG12, we ~~performed perform~~ an experiment (E-slip) with  $u_b = 0$ .

~~Figs. ?? and ?? compare Fig. 8 compares~~ the ice velocity and temperature results of E-advZ, E-advX, E-strain and E-slip with those of ~~E-ref E-ref-D~~. If the vertical advection is neglected (E-advZ~~;~~), cold ice at the glacier surface cannot be transported downwards into the interior of LHG12~~;~~, and therefore LHG12 becomes warmer (~~Figs. ??a and e Fig. 8a and b~~) and flows faster relative to other experiments (Fig. ~~??b~~). ~~As described for the discontinuous surface thermal boundary conditions across the ELA (a straightforward result from the refreezing meltwater in the accumulation basin), a discontinuous 8c). Fig. 8a shows~~ a transition of the mean column ice temperature ~~is observed along the CL at km 1.3 at km 1.8~~ (the horizontal position of ELA) in E-advX (~~Fig. ??a~~). ~~This transition arises from the discontinuous surface thermal boundary conditions across ELA~~. Compared with ~~E-ref, the E-advX experiment predicts colder the reference experiment, E-advX predicts lower~~ field temperatures (by ~~approximately 2.0 K 2.3 K on average, Fig. 8a~~) and much smaller surface ice velocities ~~-(by 7.2 m a<sup>-1</sup> on average, Fig. 8c)~~. Because the accumulation basin of LHG12 is relatively warm, E-advX, which neglects the horizontal transport of ice from upstream ~~and to~~ downstream, predicts much colder conditions for LHG12 ~~, i.e., and~~ the modeled temperate ice only appears ~~at two discontinuous grid points in the accumulation zone~~ (Fig. ~~??e~~). ~~As described by, we observed 8d). We observe~~ that strain heating contributes greatly to the thermal configuration of LHG12. If we leave away the strain heating (~~E-strain~~), LHG12 ~~becomes fully cold, with a, the modeled~~ mean ice temperature field ~~is~~ lower than that of ~~E-ref by approximately 0.5 K, E-ref-D by 0.4 K on average (Fig. 8a)~~. The TIZ becomes much thinner compared to E-ref-D, indicating the importance of strain heating ~~in the formation of basal temperate ice (Fig. 8d)~~. Previous studies have suggested that basal sliding can significantly influence the thermal structures and velocity fields of glaciers (e.g. Wilson et al., 2013; Zhang et al., 2015). However, in this study the neglect of basal sliding (E-slip) results in a temperature field very similar to that of ~~E-ref E-ref-D~~. We attribute this difference to the relatively small modeled basal sliding values for LHG12. ~~The observed and modeled ice temperature profiles of E-advZ,~~

E-advX, E-strain, E-slip, and E-ref are also compared for the deep ice borehole in Figs. ??d and ??d. The differences of the profiles can also be explained by our above explanations.

## 6.2 Transient simulations

As can be clearly seen from Figure ??a,

### 5 6.2.1 Comparison with *in situ* observations

In the reference experiment (E-ref-T), we simulate the distributions of horizontal ice velocities and temperatures in a transient manner (Figs. 9a and c). Next, the model results are compared to the measured ice surface velocities and the ice temperature profile in the deep borehole (Figs. 9b and d). Generally, the modeled ice surface velocities are in good agreement with *in situ* observations from the glacier head to km 4.8 along the CL (Fig. 9b). However, from km 4.8 to the glacier terminus, our model  
10 generally underestimates the ice surface velocities as shown in all simulations in Fig. 5. There are three possible reasons for this underestimation. First, the model neglects the convergent ice fluxes from the west branch. Second, an enhanced basal sliding zone may exist at the confluence area, which is not captured by the model. Third, the transient model uses a fixed topography, which may fail to capture the time-dependent glacier changes that can largely influence the ice flow dynamics. Here we verify the first and second hypotheses by conducting two other experiments, i.e. E-W and E-WS. In E-W the glacier widths are  
15 increased by 450 m at km 5.8 – 7.3 as a proxy of including the impact of the convergent flow from the west branch (Fig. 10). In E-WS, except for the same glacier width increasing as in E-W, we also increase  $\lambda_{\max}$  by 100% and decrease  $m_{\max}$  by 33% for accelerating the basal sliding at km 5.8 – 7.3 (Fig. 10). We can clearly find that while both factors have a non-negligible contribution to the model results, the basal sliding may play a bit more important role in the confluence area. This indicates a need of considering glacier flow branches and spatially variable sliding law parameters in real glacier modeling studies.

20 The model predicts a TIZ overlain by cold ice over a horizontal distance of km 0.6 – 6.5 (Fig. 9c). In addition, we further compare our model results with the *in situ* 110 m deep ice temperature measurements (Fig. 9d). Modeled and measured borehole temperature profiles show a close match within a root-mean-square difference of 0.2 K below the 20 m depth. Because *in situ* ice temperature data below the 110 m depth have not been obtained, we are unable to compare the modeled and measured ice temperatures at the ice-bedrock interface.

### 25 6.2.2 Evolution of the borehole temperature profile

The transient temperature profiles in our reference experiment (E-ref-T) show a cooling trend in the upper surface (0–50 m deep) particularly during the period 1991–2011 (Fig. 11a). This surface cooling trend is also suggested by the temperature profile in the deep borehole can be remarkably determined by two factors, the thermal condition ( $T_{\text{dep}}$ ) and the extent (ELA) of the accumulation basin. experiments E-cold and E-warm in which the thermal parameter  $T_{\text{dep}}$  directly impacts the upstream  
30 surface thermal boundary condition of LHG12, while ELA controls the heat conduction near the deep borehole by varying the extent of accumulation basin and thus the amount of refreezing heat of melt water. In Figure ??a, the cold (warm) case is set

to  $-5^{\circ}\text{C}$  and  $-1^{\circ}\text{C}$ , respectively. However, E-cold (E-warm) underestimates (overestimates) the ice temperature in the deep borehole, compared with the observation. The reference case From temperature profile modeled by the reference diagnostic experiment (E-ref-D; Fig. 11a), we can see that the shape of the upper temperature profile (0–30 m) is still difficult to simulate.

5 In the surface thermal boundary condition, the parameter  $T_{\text{dep}}$  simply represent a mean thermal status of the near-surface region in the accumulation zone. The reference transient experiment, which adopts the tuned ~~vaule-value~~ of  $T_{\text{dep}}$  ( $-2.1^{\circ}\text{C}$ )~~for the diagnostic experiment~~ ( $-1.8^{\circ}\text{C}$ ), shows a good agreement between the modeled and observed temperature profile, suggesting a possibly supportive evidence that the calibrated  $T_{\text{dep}}$  ~~for the diagnostic simulation might may~~ indeed contain part of the historical climate information of LHG12. In Figure ~~??~~11b, we can see that both the summer (June, July and August)  
 10 air temperature and ELA appear a slight (large) increase during 1971–1991 (1991–2011), which can explain the small (large) decrease of ice temperatures for all model ~~eases-experiments~~ over the same time period, indicating that LHG12 (or similar type glaciers) may not accordingly become cold under a cold climate scenario, when a relatively large ~~aeemulaiton~~ accumulation basin grows and more refreezing latent heat from meltwater is released. ~~In addition, a comparison of vertically averaged ice velocity and temperature field between the diagnostic and transient (reference case) simulations is presented in Figure ??c and~~  
 15 ~~d. In general, the pattern of~~

### 6.2.3 Impacts of the initial ELA on the glacier thermal regime

For LHG12, ELA determines the size of accumulation zone and the ~~velocity profiles shows the opposite to that of the temperature profiles~~. For example, the diagnostic-velocity (temperature) are larger (smaller) than the transient result over the horizontal distances of km 0–1.8 and km 8–9.8. The transient model appears to bring more heat from the accumulation basin  
 20 ~~to downstream over the past decades~~ amount of latent heat released by meltwater refreezing. In the above transient simulations (see Sect. 6.2.1 and 6.2.2), the initial conditions are generated by diagnostic runs in which the  $\text{ELA}_0$  is set to 4940 m a.s.l. To investigate the impacts of initial ELA on the thermal regime of LHG12, we perform two additional experiments with different initial ELAs, i.e. E-high ( $\text{ELA}_0 = 4800$  m a.s.l.) and E-low ( $\text{ELA}_0 = 5000$  m a.s.l.).

The initial differences of the column mean temperature between the sensitivity experiments (E-high and E-low) and the  
 25 reference experiment (E-ref-T) are significant near the downstream of the  $\text{ELA}_0$ , as can be seen in Fig. 12. After 50-year runs, however, they are largely reduced because we set the same surface thermal boundary conditions for E-high/E-low and E-ref-T during the transient simulations. But we can still clearly observe two phase shifts of temperature difference in space between the model initialization and the year 2011. By heat advection the temperature field of the downstream and the deep part of LHG12 can still be remarkably impacted by the thermal status of the upstream after dozens of years, which suggests that the  
 30 reconstruction of past climate changes on the glacier is crucial to better estimate the change of englacial thermal condition (e.g. Gilbert et al., 2014a).



### 6.3 Model limitations

Although our 2D, first-order, flowband model can account for part of the three-dimensional nature of LHG12 by parameterizing the lateral drag with glacier width variations, it cannot fully describe the ice flow along the  $y$  direction, and are not able to account for the confluence of glacier tributaries. The shape of the LHG12 glacier valley is described using a constant value for index  $b$  (1.2; approximately “V” type cross-sections), which was determined from several traverse GPR profiles (Fig. 1). However, for real glaciers, the cross-sectional geometry profiles are generally complex, resulting in an inevitable bias when we idealize the glacier cross-sectional profiles by using power law functions across the entire LHG12. Although the regularized Coulomb friction law provides a physical relationship between the basal drag and sliding velocities, several parameters (e.g.,  $\lambda_{\max}$ ,  $m_{\max}$ ) still must be prescribed based on surface velocity observations. Another uncertainty could be from the spatially uniform geothermal heat flux that we assume in the model, as it may have a great spatial variation due to mountain topography (Lüthi and Funk, 2001). In addition, we can also improve our model ability by linking the water content in the temperate ice layer to a physical thermo-hydrological process in the future.

Due to the limitations of *in situ* shallow borehole ice temperature measurements, the surface thermal boundary condition in our temperature model is determined using a simple parameterization based on observations at three elevations (Fig. 1b). In addition, the parameterized surface thermal boundary condition only provides a rough estimate of the overall contributions of the heat from refreezing meltwater and ice flow advection. At this stage, we cannot simulate the actual physical process involved in the transport of near-surface heat from refreezing, which has been suggested by ~~Because of the uncertainties of the knowledge of past climate and glacier geometry, we here do not include the complete transient simulations of LHG12, but instead we carefully calibrate our diagnostic model parameters as a proxy of implicitly representing the impacts of past climate~~ Gilbert et al. (2012) and Wilson and Flowers (2013). In our transient simulations, the glacier geometry and surface thermal parameters are assumed constant in time, which may also lead to some unphysical model outputs. ~~Because of the uncertainties of~~ Due to a lack of long-term historical climate and geometry inputs, our transient simulations are only performed during 1961–2011 initialized with according diagnostic runs. However, the transient simulations in this paper study are mainly for seeking some possible reasons for the formation of the current temperature profile of the deep borehole. ~~We~~ Therefore, we do not expect accurate model results from the transient experiments.

## 7 Conclusions

For the first time, we investigate the thermo-mechanical features of a typical valley glacier, Laohugou Glacier No.12 (LHG12), in Mt. Qilian Shan, which is an important fresh water source for the arid regions in western China. We assess the thermo-mechanical features of LHG12 using a two-dimensional thermo-mechanically coupled first-order flowband model based on available *in situ* measurements of glacier geometries, borehole ice temperatures, and surface meteorological and velocity observations.

Similar to other alpine land-terminating glaciers, the mean annual horizontal ice flow speeds of LHG12 are relatively low (less than  $40 \text{ m a}^{-1}$ ). However, we observed large inter-annual variations in the ice surface velocity during the summer and



winter seasons. Due to the release of heat from refreezing meltwater, the observed ice temperatures for the shallow ice borehole in the accumulation basin (site 3; Fig. 1) are higher than ~~for the temperatures those~~ at sites 1 and 2 at lower elevations, indicating the existence of meltwater refreezing, ~~as observed in our field expeditions~~. Thus, we parameterize the surface thermal boundary condition by accounting for the 20 m deep temperature instead of only the surface air temperatures. We observed that LHG12 has a polythermal structure with a temperate ice zone that is overlain by cold ice near the glacier base throughout a large region of the ablation area. Time-dependent simulations reveal that the englacial temperature becomes colder in recent two decades as a consequence of the shrink of accumulation area and rising surface air temperature.

Horizontal heat advection is important on LHG12 for bringing the relatively warm ice in the accumulation basin (due to the heat from refreezing meltwater) to the downstream ablation zone. In addition, vertical heat advection is important for transporting the near-surface cold ice downwards into the glacier interior, which “cools down” the ice temperature. Furthermore, we argue that the strain heating of LHG12 also plays an important role in controlling the englacial thermal status, as suggested by Zhang et al. (2015). However, we also observed that simulated basal sliding ~~contributes little to the thermal-mechanical configuration of LHG12~~ (very small;  $< 4 \text{ m a}^{-1}$ ) contributes little to the thermal-mechanical configuration of LHG12.

The mean annual surface air temperature could serve as a good approximation for the temperatures of shallow ice, where seasonal climate variations are damped at cold and dry locations (Cuffey and Paterson, 2010). However, for LHG12, using the mean annual surface air temperature as the thermal boundary condition at the ice surface would predict an entirely cold glacier with very small ice flow speeds. For LHG12, a decline of ELA under a cold climate may assist an increase of the amount of refreezing latent heat in the accumulation basin, and therefore possibly raise the englacial temperature. Because warming is occurring on alpine glaciers in, for example, Mt. Himalayas and Qilian Shan, further studies of supra-glacial and near-surface heat transport are very important because they will affect the surface thermal conditions and, eventually, the dynamical behaviours of the glacier.

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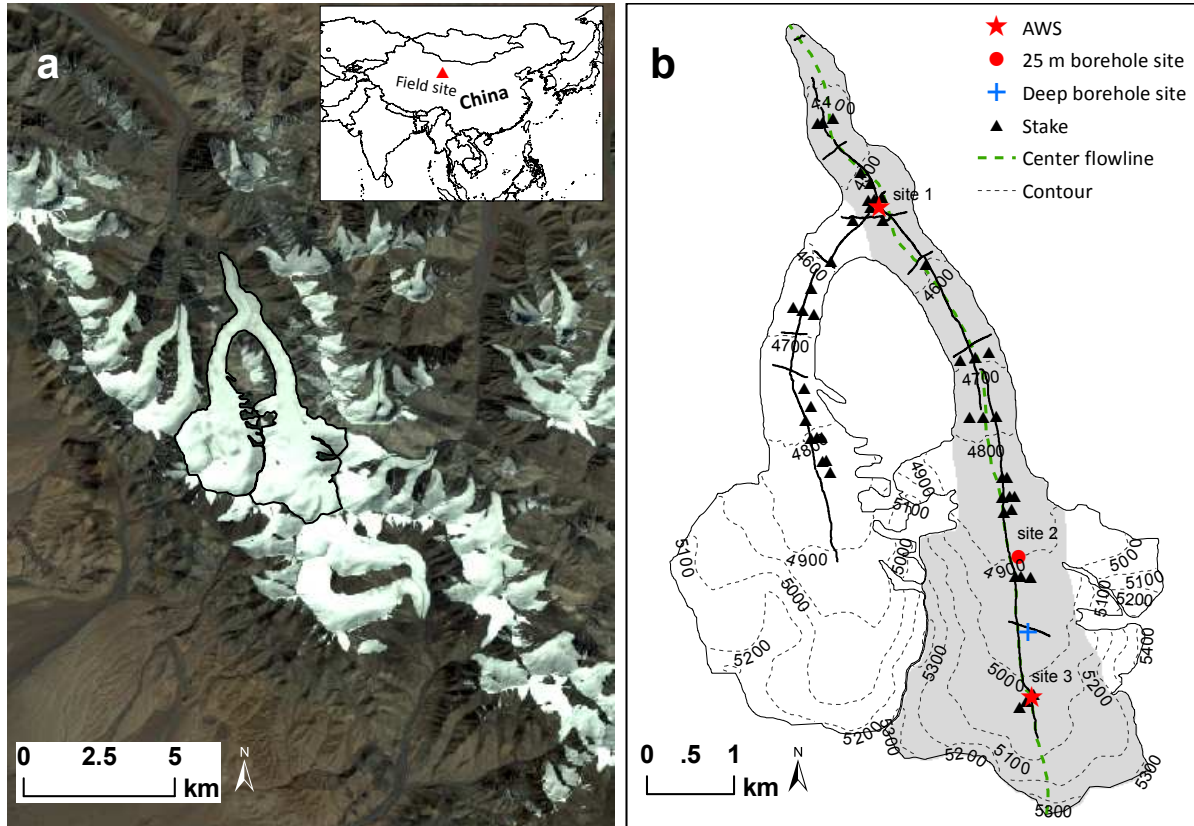
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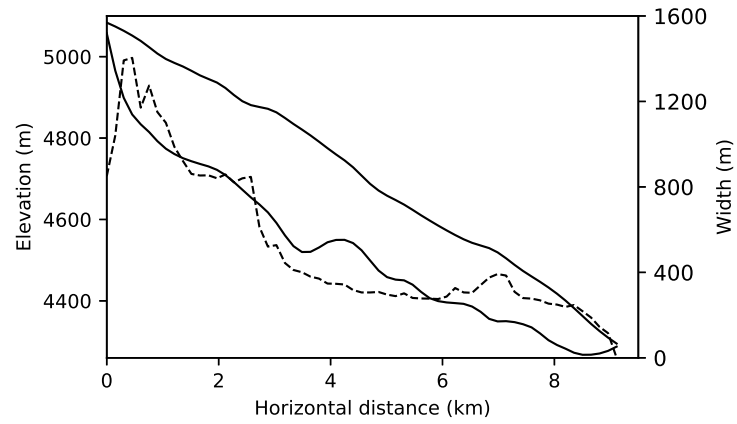
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**Table 1.** Parameters used in this study

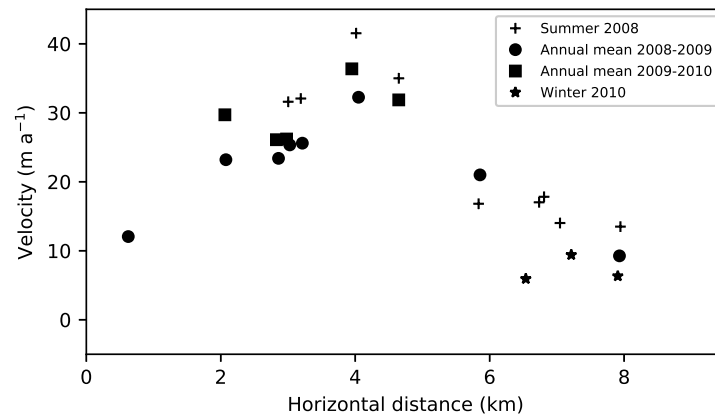
Symbol	Description	Value	Unit
$\beta$	Clausius-Clapeyron constant	$8.7 \times 10^{-4}$	$\text{K m}^{-1}$
$g$	Gravitational acceleration	9.81	$\text{m s}^{-2}$
$\rho$	Ice density	910	$\text{kg m}^{-3}$
$\rho_w$	Water density	1000	$\text{kg m}^{-3}$
$n$	Exponent in Glen's flow law	3	-
$\dot{\epsilon}_0$	viscosity regularization	$10^{-30}$	$\text{a}^{-1}$
$A_0$	Flow law parameter		
	when $T \leq 263.15 \text{ K}$	$3.985 \times 10^{-13}$	$\text{Pa}^{-3} \text{ s}^{-1}$
	when $T > 263.15 \text{ K}$	$1.916 \times 10^3$	$\text{Pa}^{-3} \text{ s}^{-1}$
$Q$	Creep activation energy		
	when $T \leq 263.15 \text{ K}$	60	$\text{kJ mol}^{-1}$
	when $T > 263.15 \text{ K}$	139	$\text{kJ mol}^{-1}$
$R$	Universal gas constant	8.31	$\text{J mol}^{-1} \text{ K}^{-1}$
$k$	Thermal conductivity	2.1	$\text{W m}^{-1} \text{ K}^{-1}$
$c_p$	Heat capacity of ice	2009	$\text{J kg}^{-1} \text{ K}^{-1}$
$L$	Latent heat of fusion of ice	$3.35 \times 10^{-5}$	$\text{J kg}^{-1}$
$T_0$	Triple-point temperature of water	273.16	K



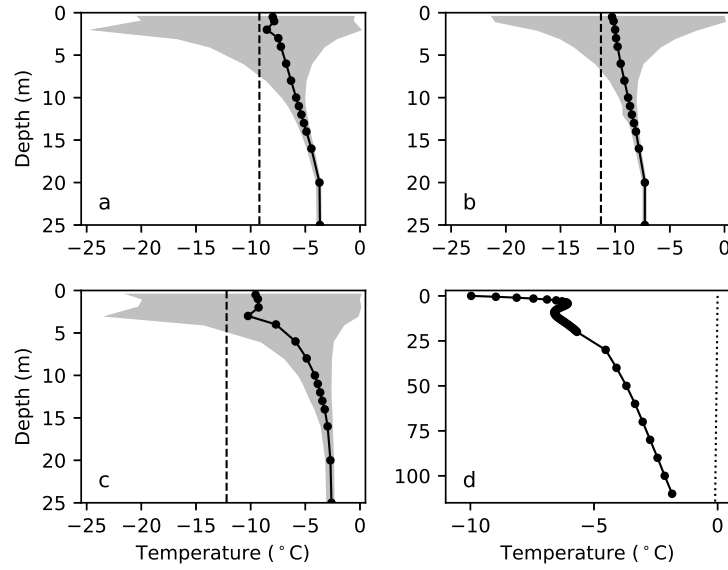
**Figure 1.** (a) The location of Laohugou Glacier No.12 (LHG12) in the west Mt. Qilian Shan, China. Underlain is a Landsat 5 TM image acquired on 22 September, 2011. (b) The solid and thick black lines indicate the ground-penetrating radar (GPR) survey lines. The shaded area denotes the mainstream-main branch of LHG12, which only includes the east branch and neglects the west branch and all small tributaries. The green dashed line represents the center flowline (CL). Red stars indicate the locations of the automatic weather stations (AWSs) and the 25 m deep shallow boreholes (sites 1 and 3). A solid red circle represents the location of the shallow borehole at site 2, and a blue cross represents the location of the deep ice borehole. Black triangles show the positions of the stakes used for ice surface velocity measurements. The black contours are generated from SRTM DEM in 2000.



**Figure 2.** The glacier geometry of LHG12 along the center-flowline. Solid lines show the glacier surface and bed elevations, while the dashed line shows the variation of glacier half widths along the flowline.

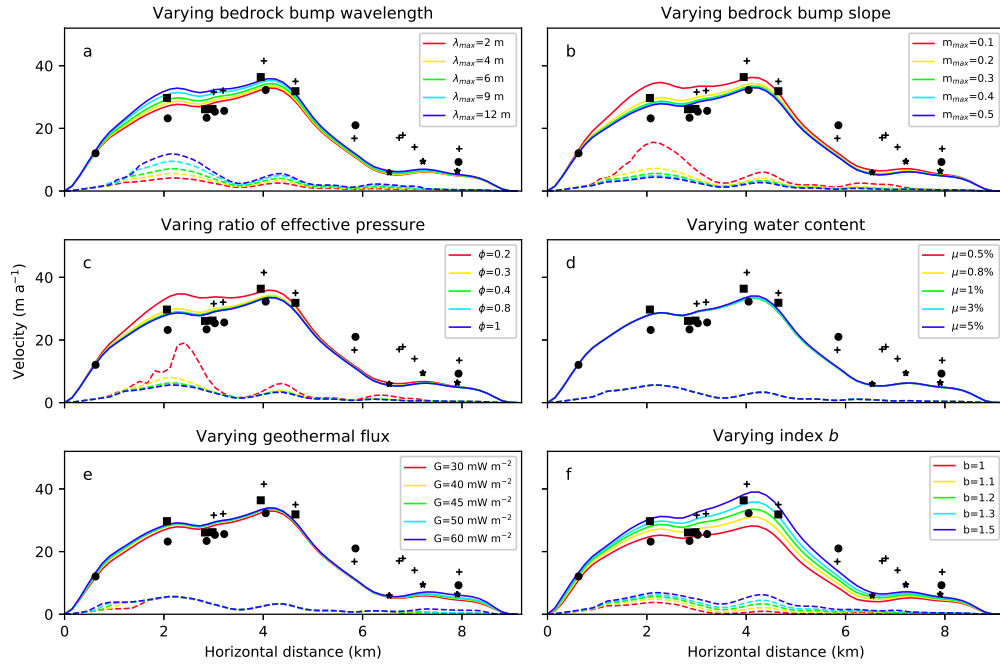


**Figure 3.** Measured ice surface velocities along the center-flowline.

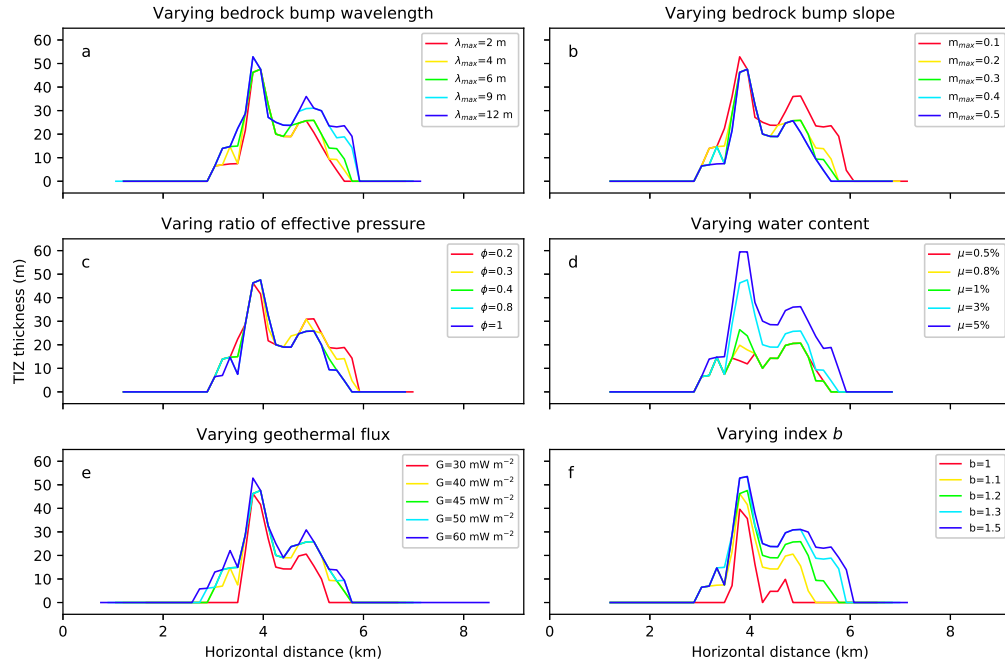


**Figure 4.** Ice temperature measurements from four ice boreholes. (a, b, c) Ice temperature measurements from the 25 m deep boreholes at site 1, 2, and 3, respectively. The black dots show the mean annual ice temperatures over the period of 2010 – 2011. The shaded areas show the yearly fluctuation range of the ice temperature. The dashed lines indicate the mean annual air temperature. (d) Measured ice temperatures from the deep borehole. The dotted line indicates the pressure-melting point (PMP) as a function of ice depth.

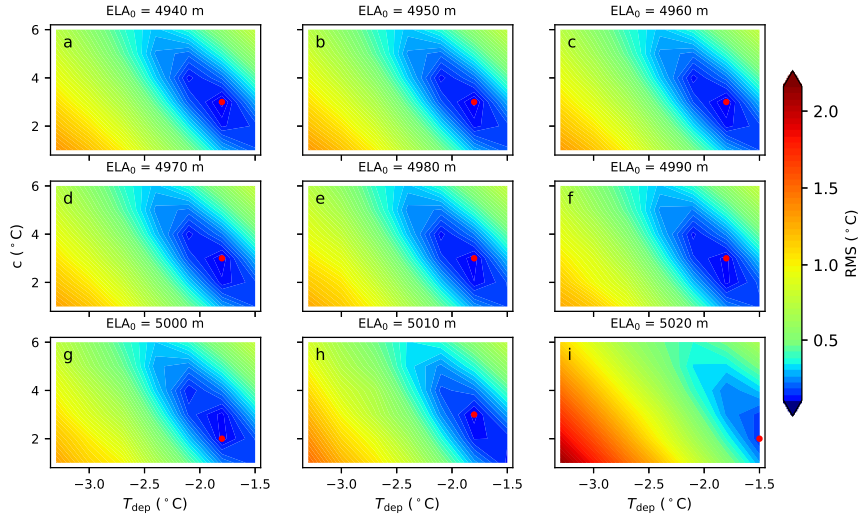




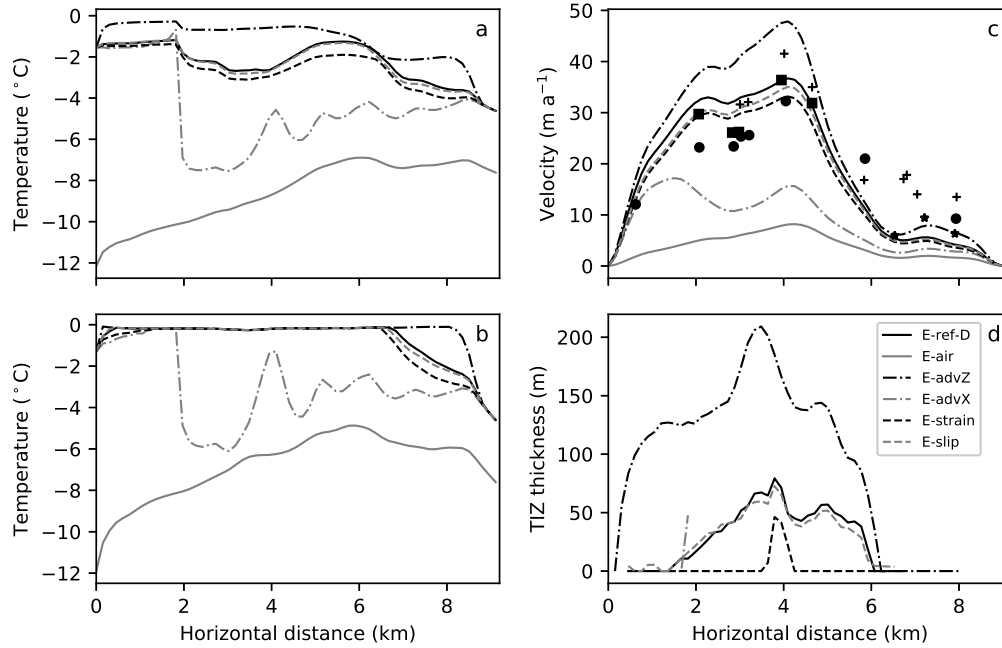
**Figure 5.** Sensitivity of modeled ice flow speeds to parameters along the CL. The solid and dashed lines indicate the modeled surface and basal sliding velocities, respectively. ~~Symbols show~~ Markers indicate the measured ice surface velocities ~~(same as those shown in Fig. 3).~~



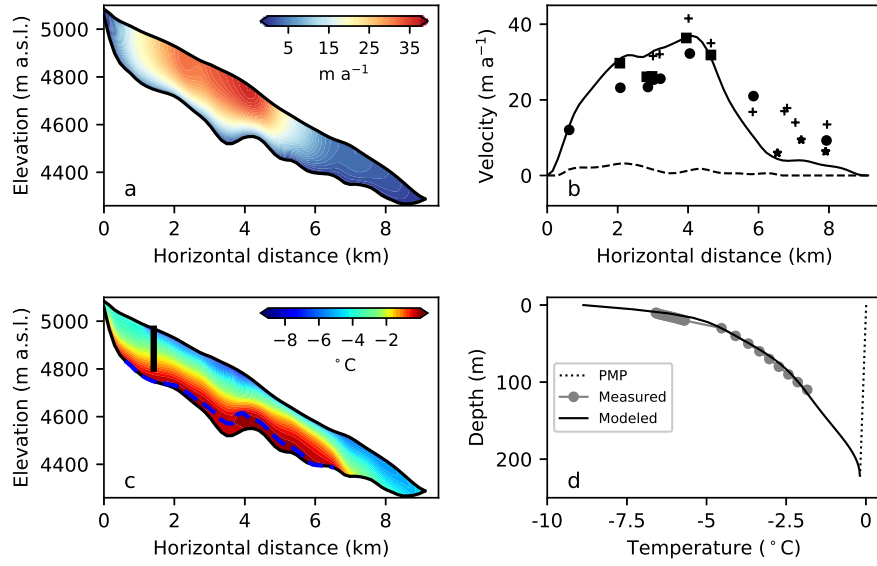
**Figure 6.** Sensitivity of modeled temperate ice thicknesses to parameters along the CL. The parameter settings are the same as described in Fig. 5.



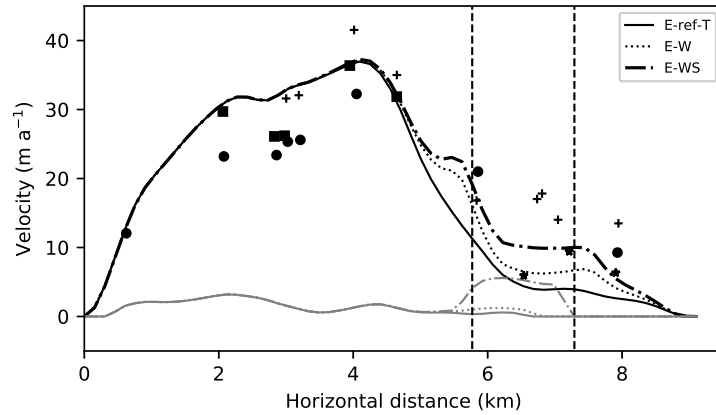
**Figure 7.** Root mean squares (RMS) of differences between measured and modeled temperature profiles in the deep borehole. The red circle indicates the minimum of RMS. The parameter  $T_{\text{dep}}$  is varied from  $-3.3^{\circ}\text{C}$  to  $-1.5^{\circ}\text{C}$  with a step-size of  $0.3^{\circ}\text{C}$ , while  $c$  is varied in the range of  $1\text{---}6^{\circ}\text{C}$  with a step-size of  $1^{\circ}\text{C}$ . The equilibrium line altitude (ELA) is fixed in each panel.



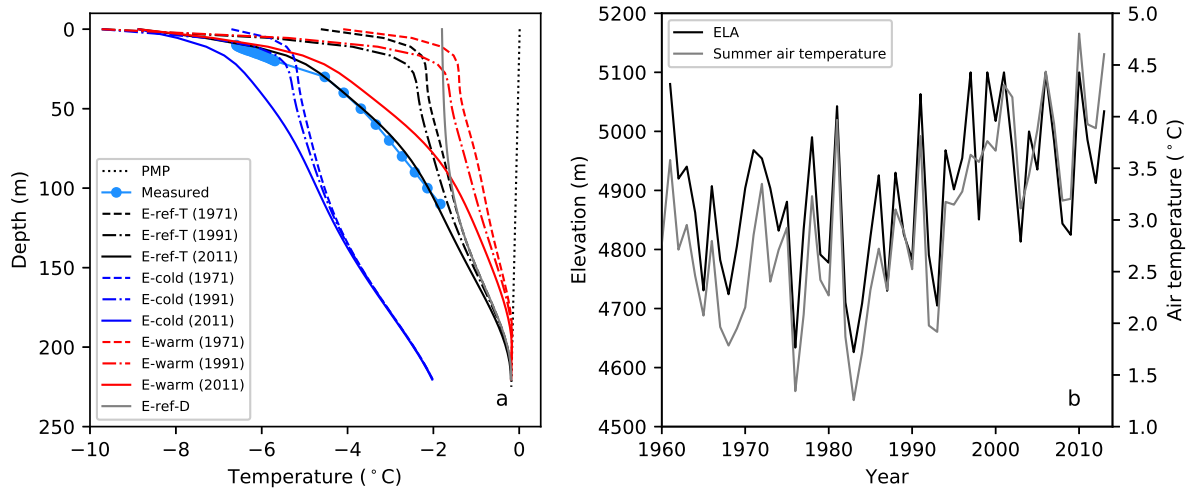
**Figure 8.** Modeled ice temperatures, velocities and TIZ thicknesses along the CL for diagnostic experiments E-ref-D (black solid line), E-air (gray solid line), E-advZ (black dash-dotted line), E-advX (gray dash-dotted line), E-strain (black dashed line), and E-slip (gray dashed line). (a) Modeled column mean ice temperatures. (b) Modeled basal ice temperatures. (c) Modeled surface horizontal velocities. (d) Modeled TIZ thickness.



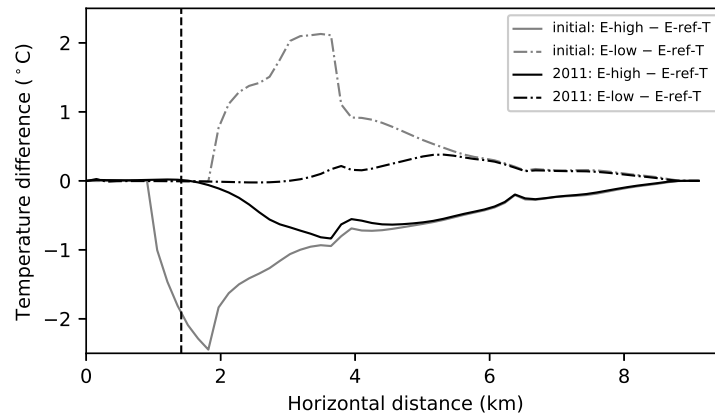
**Figure 9.** Comparison of measured and transiently modeled horizontal velocities and ice temperatures of LHG12. (a) Modeled distribution of horizontal ice velocity. (b) Measured (markers) and modeled surface (solid line) and basal (dashed line) horizontal velocities along the CL. (c) Modeled distribution of ice temperature. The blue dashed line indicates the CTS position, and the black bar indicates the location of the deep ice borehole. (d) Measured (dots) and modeled (solid line) ice temperature profiles in the deep borehole. Pressure-melting point is indicated by the dotted line.



**Figure 10.** Modeled ice surface (black lines) and basal (gray lines) horizontal velocities along the CL for experiments E-ref-E-ref-T (blue solid line), E-W (red dotted line), and E-WS (green dash-dotted line). The glacier widths in the zone of km 5.8 – 7.3 (bounded by the vertical dashed lines) are increased by 450 m for E-W and E-WS. In E-WS, we also include a basal sliding enhancement between km 5.8 – 7.3.



**Figure 11.** (a) Modeled temperature profiles in the deep borehole for experiments E-ref-T (black line), E-cold (blue line), E-warm (red line), and E-ref-D (gray line). Dots indicate the measured ice temperature profile. Pressure-melting point is indicated by the dotted line. (b) The variations of ELA and summer air temperature at 4200 m a.s.l. during 1961–2011.



**Figure 12.** Differences of the column mean temperatures along the CL between the sensitivity experiments (E-high and E-low) and the reference experiment (E-ref-T). Gray lines indicate the initial temperature differences, whereas black lines indicate the temperature differences in 2011.

# Response to Anonymous Referee #1

Yuzhe Wang, Tong Zhang

We are grateful to the reviewer for providing insightful comments and constructive suggestions, which substantially improved our manuscript and helped us develop numerical simulations of the glacier thermal regime. Our response to all the comments is given below.

## General comments

*The authors do not specify the initial condition in the transient run which is crucial for such short simulation. Even with the 10 years spin-up, results are still dependent of the initial condition for a glacier of this size. It would make more sense to initialize the transient simulation from a diagnostic run.*

*The authors still calibrate the diagnostic run by fitting the complete measured deep profile. Again, this is not the good approach! Your work on mass balance and ELA reconstruction show a significant change in ELA elevation leading to significant cooling in the deep borehole during the recent years. This cooling very likely did not reach steady state yet, you cannot constrain your diagnostic run on this. However, you have now everything to do it properly:*

*Find  $(ELA, T_{dep}, c)_{steady}$  in a way that your model fits the measurement AFTER running the transient model using the diagnostic run as initial condition.*

*So:*

- 1- Run the diagnostic model with  $(ELA, T_{dep}, c)_{steady}$*
- 2- Run the transient model until 2011 with the diagnostic model as an initial condition*
- 3- Compare the result of the transient model with measurement to adjust  $(ELA, T_{dep}, c)_{steady}$*

We thank the reviewer for giving us constructive advices, which helps us model the thermal regime of LHG12 in a proper way. Following the reviewer's suggestions, we calibrate the parameters in surface thermal boundary condition ( $ELA_0$ ,  $T_{dep}$  and  $c$ ), where  $ELA_0$  is the initial ELA, by first running a diagnostic simulation with a given set of parameters ( $ELA_0$ ,  $T_{dep}$  and  $c$ ) and then performing a transient simulation using the diagnostic run as an initial condition with time-dependent ELA and surface air temperature. This numerical process of finding the optimal parameter set of  $ELA_0$ ,  $T_{dep}$  and  $c$  is repeated until we find a good agreement between the modeled and measured deep borehole

temperatures in 2011.

In this version of the manuscript, the structure is re-organized. The subsection “Comparison with *in situ* observations” is moved under the section “Transient simulations”. In the discussions of the transient simulations, we also investigated the evolution of the borehole temperature profile during 1971–2011, and the impacts of the initial  $ELA_0$  on the glacier thermal regime. Please find more details in the manuscript.

## Specific comments

*Line 17 p7 I would remove “result in lower velocity values (Sugiyama et al., 2014)”, this is an indirect consequence of the cold bias. Better to focus on temperature only here.*

Agreed. This has been removed.

*Line 26 p7 Use other symbol ( $c$  is already used in eq. 15)*

Fixed. We now use the symbol  $m_c$  to represent the daily accumulation. In order to maintain a consistent style, we represent the daily ablation as  $m_a$ .

*Line 11 p13 “The transient model appears to bring more heat from the accumulation basin to downstream over the past decades.” Unclear plus typo (downstream). Also, this is because of the 10 years spin up with the 1960-1970 climate condition that give wider accumulation area... this not link to the transient climate that actually tend to cool the glacier...*

Thanks for pointing out the typo. In this version of the manuscript, we performed transient simulations initialized with diagnostic runs. We also added a subsection to discuss the impacts of the initial condition on the simulation of the thermal regime.



# Response to Referee Andy Aschwanden

Yuzhe Wang, Tong Zhang

We are grateful to the reviewer for providing constructive comments and positive feedbacks, which helped to improve the manuscript. Our response to all the comments is given below.

## General comments

*My main concern is the use of  $n = 3$  as the exponent of both the flow law (Eq. 2) and as the exponent in the sliding law (Eq. 13). While there is relatively good evidence that supports  $n = 3$  in the flow law, this is not the case for the sliding law. Simulated flow speeds are quite sensitive to the choice of  $n$ . For a Greenland example, see Supplementary Fig. 1 in Aschwanden et al. (2016). In my experience this is one of the most crucial parameters in ice dynamics so the authors would need to convince me that their conclusions remain robust.*

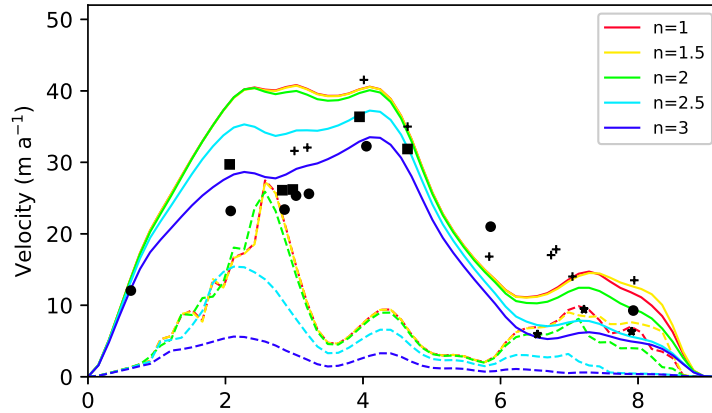


Figure 1: Sensitivity of the modeled surface horizontal velocities and sliding velocities to the exponent  $n$  in the Coulomb friction law.

We investigate the sensitivity of the modeled surface horizontal velocities and sliding velocities to the exponent  $n$  in the Coulomb friction law by varying it from 1 to 3 with a step size of 0.5 (Fig. 1). It can be seen that the mod-

eled surface velocities for  $n < 2.5$  are much larger than those of the reference experiment (i.e.,  $n = 3$ ) due to largely increased basal sliding velocities. The largest basal sliding velocity at the km 2.6 accounts for approximately 70% of the surface velocity for  $n < 2.5$ , which is likely unrealistic. The modeled surface velocities for  $n = 2.5$  is close to the results of the reference experiment. But it also predicts a larger basal sliding velocities compared to the reference result with a maximum value of about  $15 \text{ m a}^{-1}$  at the km 2.0. In this study, we prefer to use  $n = 3$  due to the agreements between the modeled and measured surface velocities. This setting would not affect our main conclusions about the thermo-mechanical features of LHG12. However, we argue that a varying  $n$  in the Coulomb friction law may be possible due to the changing basal shear stresses along the flowline (Cuffey and Paterson, 2010).

## Specific comments

*p. 1, l. 19 as well as later. “. . . of approximately  $50.5 \text{ km}^2$ ”. In combination with the word approximately, I would only reduce the number of significant digits to approximately  $50 \text{ km}^2$ .*

Agreed. We now write “. . . with a total ice volume of approximately  $51 \text{ km}^3$ ”. Following this way of expression, the numbers combining with the word “approximately” in other places are also changed.

*p. 2, l. 15 It seems appropriate to cite the oldest relevant manuscript. The importance of the surface thermal boundary has been discussed by Blatter and Kappenberger (1988). There is also a bunch of older literature on Storglaciären, northern Sweden that talks about the refreezing of snow melt in the firn area. My memory is a bit more shaky there, I may cite some papers in Aschwanden and Blatter (2005).*

Agreed. We now cite relevant studies of the thermal regimes of glaciers in Arctic (Müller, 1976; Blatter, 1987; Blatter and Kappenberger, 1988) and in China (Huang et al., 1982). They have discussed the variable energy inputs in glacier zones that determine the near-surface temperature. The sentence is now reformulated as “Different energy inputs at the glacier surface determine the near-surface ice temperature. In particular, the latent heat released by meltwater refreezing in the percolation zone can largely warm the near-surface area (e.g., Müller, 1976; Huang et al., 1982; Blatter, 1987; Blatter and Kappenberger, 1988; Gilbert et al., 2014b). The importance of surface thermal boundary condition in controlling the thermal regime of a glacier has been recognized by recent thermo-mechanical glacier model studies (e.g., Wilson and Flowers, 2013; Gilbert et al., 2014a; Meierbachtol et al., 2015).”.

*p. 2, l. 28 change “basal slip” to “basal motion” so you can avoid talking about the actual mechanism by which the motion occurs.*

Changed.

*p. 3, l. 7 change “mainstream” to “main branch”*  
Changed.

*p. 3, l. 27 change “is an order” to “is on the order”*  
Fixed.

*p. 5, l. 9 The first order model was first described by Blatter (1995) but Pattyn (2002) did a nicer job explaining the equations. I suggest using both citations.*  
Agreed. This citation has been added.

*p. 6, l. 16 ... in the temperate zone refreezes*  
Changed.

*p. 7, l. 28 change “We denote” to “We use”*  
Changed.

*p. 9, l. 89 Are you really using a “zero surface mass balance” for the relaxation? Maybe I mis-interpret this, but to me, this means that the surface mass balance is zero at every grid point at the surface. That would be an odd choice. Commonly, one uses a constant-in-time surface mass.*

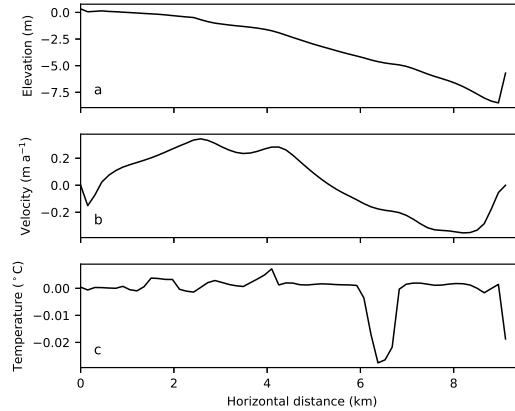


Figure 2: Differences between the two relaxation experiments with and without present-day surface mass balance. (a) Differences of the ice thicknesses after the relaxation. (b) Differences of the vertically averaged horizontal velocities after the relaxation. (c) Differences of the vertically averaged ice temperatures after the relaxation.

It is correct that we used the zero surface mass balance at each grid point. We did this because there was not enough measurements of surface mass balance covering the whole flowline. Now we perform a relaxation experiment using

the mean surface mass balance as modeled during 2009–2013. Then we compare the differences between the two relaxation experiments with and without present-day surface mass balance (Fig. 2). It shows that the modeled mean ice thickness after the relaxation using present-day surface mass balance is lower than that of using zero mass balance by about 3 m (Fig. 2a). The modeled vertically averaged horizontal velocities and temperatures along the flowline show little differences (Fig. 2b and c). In this version of the manuscript, we relax the free-surface using a constant present-day surface mass balance as suggested by the reviewer.

*p. 10–11, Parameter sensitivity I am not so familiar with the temperature model used here but it appears that the authors try to simulate a polythermal with non-polythermal physics, which may question some of the conclusions. Please correct me if I have misread the manuscript. For example, the authors state that “a larger geothermal heat flux can result in larger TIZs but has limited impact on modeling ice velocity” (I think it should read: “a larger geothermal heat flux can result in a larger TIZ but has a limited impact on simulated ice velocities.”). This is not a surprise as the model used here does not soften the ice as a function of the water content. Nonetheless, I agree with the conclusion. Using a polythermal model that includes this feedback does not change the shape of the TIZ and the simulated velocities much (viscosity is a linear function of water content). This has been discussed by Greve (1997a,b); Aschwanden et al. (2012) and recently by Hewitt and Schoof (2017).*

We agree with this comment. Our model is currently not fully polythermal, as it doesn’t include a mechanism that copes well with the water content in the temperate ice layer. We discuss it in the second paragraph of Section 6.1.1. Ideally, a warmer and larger TIZ may lead to softer basal ice, which may contribute more of the increase of the ice velocity field. To improve this, an englacial hydrology model should be coupled for calculating the water content and basal effective pressure. At this point, we are also trying to add the enthalpy scheme for the thermal model following Aschwanden et al. (2012), which we believe will eventually make some progress on this issue. We thank the reviewer for pointing it out.

*p. 12., l. 29–31 This needs to be reformulated. From Fig. 12a, I cannot see why the temperature profile is determined by the thermal boundary condition and by the ELA. “can be remarkably determined” is not proper English. Also it needs to read ... , while the ELA controls...*

In this version of the manuscript, this section has been largely changed. Please find more details in Section 6.2.2 of the revised manuscript.

*p. 14, l. 18 change “which “cools down” the ice temperature to “which cools down” the ice. One can warm and cool ice, but temperature can only be lowered or raised.*

Agreed. This has been fixed.

*Figure 1a The map looks nice but please state what the background image is.*

*Also, it's pretty dark so it is hard to see any features. Maybe use a pan-sharpened RGB composite instead?*

We now replace the black-white background image with a true-color Landsat 5 satellite image acquired on 22 September, 2011.

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