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A comparison of different methods of evaluating glacier response characteristics: application to glacier AX010, Nepal Himalaya

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Abstract

Himalayan glaciers are considered to be amongst the most sensitive glaciers to climate change. However, the response behaviour of these glaciers is not well understood. Here we use several approaches to estimate characteristic timescales of glacier AX010, a small valley glacier in the Nepal Himalaya, as a measure of glacier sensitivity. Assuming that temperature solely defines the mass budget, glacier AX010 waits for about 8 yr (reaction time) to exhibit its initial terminus response to changing climate. On the other hand, it takes between 29–56 yr (volume response time) and 37–70 yr (length response time) to adjust its volume and length following the changes in mass balance conditions, respectively. A numerical ice-flow model, the only method that yields both length and volume response time, confirms that a glacier takes longer to adjust its length than its volume.

1 Introduction

Glaciers are excellent indicators of climate change. They change their dimensions in response to all climatic variables in the surrounding environment. Several lines of evidence reveal that the interaction between glaciers and climate change is particularly sensitive (e.g. Kaser, 2001). Although this relationship is not completely understood as yet, glacier size/shape, steepness of the bed, hypsometry, and the ratio of annual mass turnover to total mass determine the degree of sensitivity of ice-climate interactions (Oerlemans et al., 1998). Of these factors, a glacier's mass balance, which forms a vital link between the changing atmospheric environment and the dynamic behaviour of a glacier, is the most crucial to its survival. Climate change affects not only accumulation and ablation rates, but also the length of mass balance seasons. In response to a climatic shift, a glacier changes its extent towards a size that makes the net mass balance over the glacier domain zero (e.g. Lemke et al., 2007). Along with changes in its areal extent, a glacier also makes an adjustment in its surface elevation, thereby exhibiting a three-dimensional response.

Glacier thickness responds immediately to perturbations in snow accumulation or snow/ice ablation, but changes in glacier extent lag behind a climate change. This lag in time is generally referred to as reaction time. It provides useful information about how quickly a glacier responds to a shift in its mass balance. However, this characteristic timescale has not drawn much attention from glaciologists. This is probably because it does not reflect a well-defined physical property of a glacier, as it depends on the past evolution of glacier in a non-transparent way (Oerlemans, 2001). Some investigations of reaction time include those of McClung and Armstrong (1993), Pelto and Hedlund (2001), and Calmanti et al. (2007). For a typical valley glacier, these efforts reveal the reaction time to be on the order of a few years. In this paper, we analyze the reaction time of glacier AX010 using a simple time series analysis between local climate and glacier extent.

After the initiation of the terminus response, a glacier obeys an exponential asymptotic path towards a new steady state so that the net mass balance of the glacier becomes zero. The period over which a glacier adjusts its overall geometry to accommodate a change in mass balance is called the response time. Oerlemans (1997b) reports that a glacier carries the mass balance history in its memory over a period equivalent to one or two times the response time. In other words, the response time tells how long a glacier is affected by a sudden change in its mass balance. Based on both field observations and numerical models, several methods have been proposed to estimate the response time. The majority of them reveal that large glaciers with gentle slopes and cold ice, such as polar ice caps, respond more slowly than small valley glaciers (e.g. Lemke et al., 2007). Using a scaling analysis, Bahr et al. (1998) suggest that a small valley glacier responds faster mainly because of its climatic setting (typical mountain climate) and not because of the dynamic characteristics of the glacier arising from its small size, as often assumed. A brief review of some of the widely used methods for response time estimation is summarized in the following paragraph.

The linear kinematic wave theory described in Nye (1960) and a subsequent series of papers gives a simple equation that yields response times on the order of 100 to 1000 yr for valley glaciers. This order of magnitude is probably too long to be representative of real glaciers (Jóhannesson et al., 1989b). A more realistic estimate of response time is obtained using an alternative approach by Jóhannesson et al. (1989a). However, it should be noted that this method does not account for the height-mass balance feedback. This feedback is particularly important for valley glaciers, which typically rest on steep bedrock. A few attempts to include this feedback are those by Harrison et al. (2001), Oerlemans (2001), and Raper and Braithwaite (2009). The potential application of these models to an individual glacier is restricted by the need for extensive annual datasets, so the fidelity of these approaches is not well-assessed yet. Another category of methods is a conventional approach with a numerical ice-flow model, although these have been criticized to yield unreasonably long timescales (Oerlemans, 2007). Inverse modelling (Oerlemans, 2001) provides an alternative approach to estimate response times from the available record of glacier extent. Given the fact that each method has its own strengths and limitations, we consider several of these approaches to calculate and compare the response times of glacier AX010.

2 The study area

Glacier AX010 (27°42′ N, 86°34′ E; Fig. 1), a small debris-free valley glacier located in the eastern part of the Nepal Himalaya, is treated here for the diagnosis of its response characteristics. The glacier head is at the foot of a rock peak at an altitude of 5381 m a.s.l., and the terminus is at a pond at 4952 m a.s.l. as of 1978 AD. The glacier opens eastward in its accumulation zone and its terminus runs down to the south-east. The plan-form geometry of the glacier is typical V-shape, with the accumulation zone about four times wider than the ablation area. It is a monsoon-affected summer-accumulation-type glacier, where summer balance is representative of the annual mass balance (e.g. Kadota and Ageta, 1992).

Glacier AX010 is amongst a few highly-studied glaciers in the Nepal Himalaya and has the densest observations in terms of glacier extent, mass balance, and ice flow (e.g. Fujita et al., 2001). The first detailed study of this glacier was conducted in 1978/1979. In this year, mass balance (Ageta et al., 1980), heat balance (Ohata and Higuchi, 1980), surface velocity (Ikegami and Ageta, 1991), and areal extent data (WGMS, 1998) were collected intensively. Thereafter, glaciological state variables have been monitored intermittently in 1989, 1991, 1995–1999, and 2004. Only the terminus position was recorded in 1989 (Fujita et al., 2001), while in 1991 a topographic map of the whole area of the glacier was completed (WGMS, 1998). Annual monitoring of the glacier was initiated in 1995 and continued until 1999 in order to obtain mass balance, surface velocity, and areal extent data (WGMS, 2005). The ice thickness of the glacier was measured in 1995 (Kadota et al., 1997) by means of radio-echo sounding. The areal extent of the glacier was resurveyed in 2004 (Kayastha and Harrison, 2008).

3 Reaction time: initial terminus response

For any glacier there exists a time lag between the onset of a relatively sudden change in climate and the initiation of a noticeable response of its terminus. This lag in time is generally referred to as reaction time, τ_R. Unlike the response time, this timescale is not a pure physical property of a glacier (e.g. Oerlemans, 2001; Pelto and Hedlund, 2001) because it depends on the glacier and climate history in a non-transparent way. In addition, the term "noticeable response" appearing in the definition does not have a clear meaning. Therefore reaction time is criticized as a loosely-defined timescale (Oerlemans, 2007). Nevertheless, there is intrinsic value to this metric. It is less idealized than the concept of a steady-state response time, and climate-driven changes in glacier length are commonplace and simple to observe. Historical glacier length reconstructions are also possible through dating of moraine sequences, and it would be powerful to be able to invert records of glacier length for the site-specific climate history. A more clearly-defined relationship between climate forcing, glacier history, and reaction time would therefore be of great value. We return to this point later in the paper.

Here we investigate τ_R for glacier AX010 through the conventional method of analyzing historical climate variations and the resulting fluctuations of the glacier terminus.

The first step of this analysis is to define which climatic parameter should be considered or how climatic parameters should be combined to create a representative climate to best describe a glacier's mass balance history. Increased air temperature, rather than reduced precipitation or other climatic variables, plays a key role in the increasing rate of retreat of Himalayan glaciers (e.g. Naito et al., 2001; Kayastha and Harrison, 2008). This is also demonstrated by Adhikari and Huybrechts (2009) using a numerical ice-flow model applied to glacier AX010. We therefore consider temperature as the principal parameter that determines the mass budget of glacier AX010.

Due to the unavailability of a sufficiently long series of temperature recorded on or nearby the glacier, we mainly use data from the Chialsa station (∼20 km southeast of the glacier) as was done by Naito et al. (2001) in their sensitivity study of glacier AX010. This record, however, only covers the period 1976–1996. We therefore need an alternative for the preceding and following periods. Kayastha and Harrison (2008) infer that Kathmandu temperature is highly correlated to regional variations in Equilibrium Line Altitude (ELA) in the eastern part of the Nepal Himalaya. Similarly, Adhikari and Huybrechts (2009) successfully reconstruct the historical terminus position of glacier AX010 by forcing the flow model with temperature anomalies from Kathmandu station (∼150 km to the west). Kathmandu summer temperature is plotted in Fig. 2a, along with the corresponding data from the Chialsa station. The time series of both stations follow a similar trend of variation. A simple linear regression analysis between these datasets for an overlapping period is depicted in Fig. 2b. The relationship is not perfect but the Kathmandu temperature trend is broadly representative of conditions at glacier AX010.

Here we first estimate a qualitative range of τ_R solely based on visual inspection of the temperature trend and the glacier's terminus positions as was done by Pelto and Hedlund (2001). As shown in Fig. 3a, glacier AX010 has been continuously receding since record keeping began in 1978. Therefore the shift of the glacier from either retreating to advancing or advancing to retreating stages cannot be seen in this record. We therefore examine two stages of relatively rapid retreat: one starting in 1989 (at a rate of $-14\,\mathrm{m\,a^{-1}}$ for 2 yr), and another one in 1996 (at an average rate of $-20\,\mathrm{m\,a^{-1}}$ for 8 yr). A coarse inspection of temperature series and length record for the glacier reveals that the first retreat could be the consequence of the temperature increase from 1976. After a few years of decline, temperature rose again in 1988, which could be linked to the latter stage of a rapid retreat. This yields $\tau_R \sim 8$–13 yr. Similarly a period of relative stabilization for the glacier (an average rate of retreat of only $-2.4\,\mathrm{m\,a^{-1}}$ from 1992–1996) in between these two relatively rapid retreating stages could have been caused by a short cooling phase that prevailed from 1984 to 1988. This suggests $\tau_R \sim 8$ yr, which falls in the above estimated range.

Now we make a more rigorous attempt to refine the magnitude of the reaction time. We perform a cross-correlation analysis between time series of climate and glacier extent data as depicted in Fig. 3a (only for an overlapping period, 1978–2004). McClung and Armstrong (1993) use a similar technique to estimate τ_R ("time response" in their words) for the Blue glacier, USA. A similar analysis was also performed by Calmanti et al. (2007) to estimate the "time lag" between the observed climatic trends and ensemble-averaged terminus fluctuations of glacial systems in Piedmont and Val d'Aosta, Italy. A cross-correlation analysis measures the degree of linear relationship between the temperature variation and the resulting fluctuation of the glacier terminus. A time delay of 0 to +20 yr is applied to the length series by shifting its position backward in time. The correlation coefficient r for each year of delay is plotted in Fig. 3b. The figure reveals good correlations for time lags of 7–12 yr, with the peak occurring at 8 yr ($r=-0.67$). This means that it takes about 8 yr for glacier AX010 to initiate its terminus retreat in response to the applied negative mass balance (as a result of a generally warming climate, cf. Fig. 3a). This reaction time is comparable to that of other valley glaciers (Table 1). Note that negative time lags are not applied in this analysis because the glacier behaviour can evidently not be described by the climate of subsequent years.

Although temperature is the key climatic parameter to control the mass balance of glacier AX010, precipitation fluctuations can also be expected to play a role in mass balance variability and the consequent terminus fluctuations of the glacier. Over the period 1978–2004, precipitation data from Kathmandu indicate an increasing trend, at an average rate of ∼0.01 m a^{-1} (annual average=1.16 m). No local precipitation data is available. Assuming that the Kathmandu record is representative of the region, this supports the assertion that precipitation variability has a small influence on mass balance for glacier AX010 (Adhikari and Huybrechts, 2009). If precipitation were a key parameter, the glacier would have been advancing over the last three decades, contrary to observations. However, precipitation fluctuations are likely to play a role in modulating the temperature-driven terminus reaction.

As noted earlier, the concept of terminus reaction time can be ambiguous because it depends on the glacier's state at the time of the climate perturbation (e.g. advancing or retreating) as well as the site-specific glacier geometry. There is therefore an influence from the climate history of a region. We examine this further in Sect. 5.

4 Response time

In the glaciological literature there is no consensus on the definition of characteristic timescales. The majority of definitions are based on the concept of a reference state glacier (Paterson, 1994). A step change in mass balance on the reference state glacier induces a reaction towards a new steady state. The time that a glacier takes to move from the reference state to a new steady state is precisely defined as the equilibrium time (Bahr et al., 1998). Because this timescale can be infinitely long, the response time is generally used as its characteristic measure. It is common to define response time using the e-folding concept (e.g. Jóhannesson et al., 1989b; Oerlemans, 2001). Let us define the initial glacier volume and length to be V_0 and L_0 and the new equilibrium values to be V_f and L_f. The changes in volume and length between the initial and final states are ΔV and ΔL. Over the e-folding response time, glacier adjusts volume by $\left(1-e^{-1}\right)$ ΔV of its volume change and length by $\left(1-e^{-1}\right)$ ΔL of its length change along an exponential, asymptotic path to a new steady state.

An idealized step change in climate causing an evolution from an initial to a final steady state does not occur in nature (Schwitter and Raymond, 1993). This means that the response time cannot be defined directly from field observations. Estimates that rely on field data are based on the assumption that during a period of relatively constant climate after a significant climate shift, the glacier adjusts its geometry to $\left(1-e^{-1}\right)$~63% of the final adjustment.

4.1 Simpler approaches

Based on kinematic wave theory applied to a linearized equation of glacier flow, Nye (1960) proposes a semi-quantitative method to estimate a timescale for glacier adjustment, or glacier memory τ_M. The interval of time τ_M over which a glacier responds to a climate change is in general equivalent to the volume response time τ_V (Jóhannesson et al., 1989b). According to this approach, τ_V is related to the length L and terminus velocity u_T of a reference state glacier so that:

$$\tau_V \sim f \frac{L}{u_T}. \quad (1)$$

The factor f in the above equation is commonly assumed to be about 0.5 (Paterson, 1994). For typical valley glaciers, this model yields τ_V to be on the order of 100 to 1000 yr, which is longer than available observations indicate (e.g. Jóhannesson et al., 1989b). Such a large estimate is probably due to the critical dependence of the timescale on the dynamics at the glacier terminus. Van de Wal and Oerlemans (1995) compare kinematic wave theory to a numerical model of ice dynamics and claim that the latter one explains more accurately the dynamics at the glacier terminus, and hence yields a more realistic estimate of τ_V. Pelto and Hedlund (2001) also criticize the application of this method (Eq. 1) to real glaciers. They report that due to the wide spatial and temporal variability of u_T, this model does not yield a consistently accurate result. Jóhannesson et al. (1989b) argue that Eq. (1) is still valid to estimate τ_V provided that f is a profile shape parameter, defined as the constant of proportionality between the average thickness change of a glacier and the thickness change at its terminus:

$$f(t) = \left[\frac{\overline{\Delta H(t)}}{\Delta H_T(t)} \right]. \quad (2)$$

Here $\Delta H_x(t)$ is the width-averaged thickness change at time t at any section x along the glacier length ($0 \leq x \leq L$), $\overline{\Delta H(t)}$ is the length-wise average of $\Delta H_x(t)$, and H_T denotes the terminus thickness. The parameter f may be interpreted as a measure of the degree to which thickness changes are localized near the terminus ($f \rightarrow 0$) or are spread evenly over the glacier ($f \rightarrow 1$) (Schwitter and Raymond, 1993). According to this, the conventional choice $f=0.5$ corresponds to a near linear variation of thickness change from zero at the head to a maximum value at the terminus. This is equivalent to assuming 100% diffusion in kinematic wave theory (Paterson, 1994).

Now we calculate τ_V for glacier AX010 using the model discussed above. A quick look at the glacier's length record (Fig. 3a) indicates that the glacier was not in an equilibrium or near-equilibrium state over the period 1978–2004. The assumption of a reference state glacier is therefore based on the best available model input. The 1978 AD glacier stand is considered as a reference state glacier whose length and average terminus velocity (observed at L10, Fig. 1) were 1700 m and 4 m a^{-1} respectively (Ikegami and Ageta, 1991). With these data, the conventional value of $f=0.5$ yields $\tau_V \sim 212$ yr.

Next, we estimate τ_V using the parameter f as defined in Eq. (2). To calculate f we use the thickness change data collected from various sources as summarised in Table 2. The glacier-average thickness change during 21 yr (1978–1999) was about −17 m. The thickness change at the glacier terminus is not well documented. Kadota et al. (1993) note that the glacier surface lowered by >30 m around the terminus from 1978–1991. Assuming a wedge-shaped terminus, we conduct a simple geometric analysis to make a rough estimate of thickness change at the terminus (Kadota and Ageta, 1992) for the period 1991–1999. The 1991 topographic map reveals the average surface slope near the terminus to be about 15°. For 102 m of glacier retreat (1991–1999), this yields a 27 m lowering of the glacier surface at the terminus. These data result in $f \sim 0.30$. This value of f falls in the range of 0.1–0.4 (average=0.28) as calculated by Schwitter and Raymond (1993) using the longitudinal profile change data of 15 valley glaciers (Table 3). Considering 1978 stand as a reference state, $f=0.30$ yields $\tau_V \sim 126$ yr. This value for the volume response time is within the range of 30–244 yr estimated by Pelto and Hedlund (2001) for 17 glaciers in the North Cascades, USA, using the same methodology.

Alternatively, Jóhannesson et al. (1989a) propose a simpler approach that provides realistic estimates of τ_V for valley glaciers, on the order of 10 to 100 yr. According to this method, the timescale is obtained by dividing a characteristic ice thickness H by the net annual mass balance (ice-equivalent) at the glacier terminus b_T:

$$\tau_V \sim \frac{H}{-b_T}. \quad (3)$$

A clear definition of H is not available. Agreement between the timescale given by Eq. (3) and other estimates depends on the chosen ice thickness (Oerlemans, 2007). Here we use the maximum ice thickness, as was done by e.g. Schwitter and Raymond (1993) and Naito et al. (2001). Based on radio-echo sounding conducing in June 1995, Kadota et al. (1997) report that the maximum ice thickness of glacier AX010 is 86 m. The net annual mass balance values recorded at the terminus of glacier AX010 from 1996 to 1999 were −1.72, −2.58, −3.60, and −2.91 m w.e. a^{-1} (WGMS, 2005), giving an annual average of −2.97 m a^{-1} (ice equivalent). These values (H=86 m, b_T=−2.97 m a^{-1}) yield $\tau_V \sim 29$ yr. This is compared to the timescales of a few other glaciers as estimated using the same method (Table 4).

A quick comparison of the volume timescales of glacier AX010 obtained from simpler approaches discussed above reveals a broad range of magnitude (29–212 yr). However it is clear that Jóhannesson's estimate based on Nye theory (126 yr) is a better predictor than Nye's estimate (212 yr). The choice between the two estimates following Jóhannesson (126 and 29 yr) is not obvious (see Sects. 4.5 and 6 for a detailed discussion), although the latter one may be more realistic. Despite the appealing simplicity and practical utility of these simpler approaches, they bear a large degree of uncertainty in the sense that none of them accounts for the height-mass balance feedback. Including this feedback is thought to make the response time longer (e.g. Oerlemans, 1997; Harrison et al., 2001).

4.2 Including the height-mass balance feedback

To date, only a few attempts have been made to include the effect of changing surface elevation on mass balance. They include an effort by Harrison et al. (2001) to modify Jóhannesson's estimate (Eq. 3), a similar effort by Oerlemans (2001), and a simple conceptual model of glacier hypsometry (Raper and Braithwaite, 2009). The application of these models is not straightforward as they require extensive annual data. Key inputs for Harrison's estimate are climatic data such as specific mass balance and (exposed) bedrock elevation. The latter two models, on the other hand, require geometric details of the glacier such as annual data of ice volume, ice thickness, glacier area, length, and altitudinal range. Moreover, the fundamental geometry-related assumption in the model by Raper and Braithwaite (symmetric triangular hypsometry) is not satisfied for a glacier like AX010, which has a wide accumulation area and a narrow, steep tongue (see Fig. 1). Therefore in this section we only consider Harrison model to demonstrate the importance of including the height-mass balance feedback.

The Harrison model introduces the idea of a reference glacier geometry, which can be taken as the initially-surveyed volume V_0, area A_0, and surface topography $Z_0(x,y)$ of an ice mass. Over time the glacier evolves to a new geometry $V(t)$, $A(t)$, and $Z(x,y,t)$, with the changes ΔV, ΔA, and $\Delta Z(x,y)$. Based on the concept of a reference surface balance rate (Elsberg et al., 2001), Harrison et al. (2001) modify

Jóhannesson's estimate (Eq. 3) so that:

$$\tau_V \sim \frac{1}{\left(\frac{-\bar{b}_r}{H} - \overline{G}\right)}. \tag{4}$$

Here H is once again the characteristic thickness, b_r is the specific balance rate at the (ice-free) bedrock surface, in m a^{-1}, and $\bar{b}_r$ is the average value of this quantity over area ΔA:

$$\bar{b}_r \sim \frac{\int\limits_{\Delta A} b_r dA}{\int\limits_{\Delta A} dA} = \frac{\int\limits_{\Delta A} b_r dA}{\Delta A}. \tag{5}$$

This accounts for the effect of changes in glacier area ΔA on the balance rate. In Eq. (4), $G = \frac{db}{dz}$ is the vertical gradient in specific mass balance rate, and $\overline{G}$ is its area-weighted average:

$$\bar{G} \sim \frac{\int\limits_{A} G\,(Z - Z_0)\,dA}{\int\limits_{A} (Z - Z_0)\,dA} = \frac{\int\limits_{A} G\,(Z - Z_0)\,dA}{\Delta V}. \tag{6}$$

The $\overline{G}$ term accounts for the effect of the changing surface elevation of a glacier on its balance rate, where $Z - Z_O$ is the difference in elevation at a point on the surface with respect to the reference surface. Elsberg et al. (2001) suggest that in case of glacier retreat, integration of Eq. (6) should be over the reference area A_0 instead of the current area A.

The τ_V defined by Jóhannesson et al. (1989a, Eq. 3) and modified by Harrison et al. (2001, Eq. 4) have the same meaning except that the latter one accounts explicitly for the height-mass balance feedback via $\overline{G}$. Moreover b_T in Eq. (3) characterizes the balance rate at the elevation of the ice surface near the terminus, while $\bar{b}_r$ in Eq. (4)

characterizes the weighted-average of the balance rate at bedrock height over ΔA. It should be noted that Harrison et al. (2001) misinterpret this point, claiming that $\bar{b}_r$ characterizes the balance rate at the elevation of the bed in the vicinity of the terminus, and arrive at the misleading conclusion that $|b_T| < |\bar{b}_r|$. This is not strictly true, although it is likely to be the case if ΔA is concentrated around the glacier terminus and the glacier is in retreat (Fig. 4a). The geometric requirement for this to be strictly true is that the valley side walls are perfectly vertical (Fig. 4b) all along the glacier so that the changes in ice thickness over a time period will neither cover (in case of advancing) nor expose (in case of retreating) the lateral rocks. This rarely happens in nature; for example the exposure of the rocks all along the side walls of glacier AX010 during its retreating stage (1978–1999) can be seen in Fig. 4c. This suggests that ΔA does not necessarily concentrate on the terminus region. Since the balance rate at the bedrock elevation of ΔA in the higher altitudinal range b_r is less negative than that on the glacier terminus b_T, its areal-weighted contribution makes $\bar{b}_r$ less negative. In other words, the magnitude of $\bar{b}_r$ is mainly defined by the altitudinal distribution of ΔA. One should not draw a conclusion concerning the magnitude of $\bar{b}_r$ and b_T without considering this altitudinal distribution.

We calculate τ_V for glacier AX010 using Harrison's estimate (Eq. 4). We take the initial surface (1978 AD, Fig. 1) as the surface at time t=0 years, to be the reference one. The long-term averages of $\bar{b}_r$ and $\overline{G}$ are used in this application. First we derive $\bar{b}_r$, which demands a map of bedrock elevations within ΔA and the specific balance rates b_r at those elevations. However, such details of mass balance data are not available for glacier AX010. We therefore approximate $\bar{b}_r$ based on the long-term specific balance rate at the glacier terminus as suggested by Elsberg et al. (2001). They recommend that 75% of the net balance rate at the glacier terminus is representative of $\bar{b}_r$. As reported in the previous section, the average value of b_T over 1996–1999 was $-2.97\,\mathrm{m\,a^{-1}}$ (ice equivalent), which approximates the average value of $\bar{b}_r$ to be $-2.23\,\mathrm{m\,a^{-1}}$.

Next we estimate the long-term average of the balance rate gradient $\overline{G}$ for glacier

AX010. Although Elsberg et al. (2001) recommend estimation of $\overline{G}$ for each year and to perform a numerical integration, this is not possible for our glacier due to the unavailability of annual hypsometry, mass balance, and thickness change data. We therefore estimate $\overline{G}$ for the period 1978–1991 in a single treatment. We use the hypsometry and thickness change data (WGMS, 1998) recorded at an elevation interval of 20 m. As the balance rate gradient of the conventional surface is equivalent to the reference surface counterpart by the linearity assumption (Elsberg et al., 2001), we use the average annual mass balance data recorded during 1996–1999 at an elevation interval of 50 m. To match the same elevation interval, hypsometric data are organized accordingly (Table 5). This table summarizes the parameters considered to estimate $\overline{G}$ for glacier AX010. The calculation yields $\overline{G}$=0.00355 m a^{-1} m^{-1} (1978–1991 average).

We consider the same thickness scale as used in the previous section, H=86 m. These values (H=86 m, $\overline{G}$=0.00355 m a^{-1} m^{-1}, $\bar{b}_r$=−2.23 m a^{-1}) yield τ_V=45 yr. Comparing this timescale to the one estimated by the original Jóhannesson model in the previous section (29 yr) illustrates the importance of $\overline{G}$, which explicitly accounts for the height-mass balance feedback.

4.3 Calculations with a numerical ice-flow model

In this section we discuss the results of sensitivity tests on glacier AX010 as obtained from a numerical ice-flow model. A detailed description of the model can be found in Adhikari and Huybrechts (2009). The model is based on the vertically-integrated continuity equation that relates the changes in ice thickness to the flux divergence and the net surface mass balance (e.g. Oerlemans, 2001). The sensitivity test is conducted by imposing a step change in mass balance on the reference state of the glacier equivalent to its 1996 AD stand. Assuming that over a sufficiently long period of time the terminus response of a glacier reflects its overall mass balance (Huybrechts et al., 1989), we assess the changes in glacier length for mass balance perturbations in a range between −0.5 and +0.5 m w.e.a^{-1}. The evolution of glacier length and the corresponding e-folding timescales are illustrated in Fig. 5a. The figure reveals that the length response time τ_L for glacier AX010 is in the range of 55–85 yr for both positive and negative perturbations. This is comparable to τ_L of other glaciers as obtained using a similar model (Table 6).

It is worth mentioning that the initial terminus response τ_R revealed by the dynamical flow model ranges from 7 to 21 yr for positive, and 5 to 13 yr for negative perturbations, respectively. The latter range is comparable to the one estimated in Sect. 3 (τ_R~8 to 13 yr), which is also associated with a negative mass balance. However, it should be noted that the estimates from a flow model are based on the assumption that the "noticeable response" of the glacier terminus is equivalent to the grid spacing of the model (in this case a finite difference scheme with 10 m grid spacing). A coarser grid spacing would yield a larger τ_R than one with a finer grid spacing. To analytically and numerically derive τ_R, this timescale should be defined using a clear concept similar to an e-folding concept for defining response times. This however does not seem straightforward. See Sect. 5 for further discussion.

The volume timescale τ_V of glacier AX010 is also investigated using the dynamical ice-flow model (Fig. 5b). The e-folding values of τ_V range between 50 and 62 yr. Comparing this range to the one corresponding to τ_L (55–85 yr) confirms the understanding that a glacier takes a longer time to adjust its length than its volume (e.g. Oerlemans, 2001). This is probably because the ice volume is more directly affected by changes in the specific mass balance (Oerlemans, 1997a).

4.4 Inverse modelling: calculations based on the glacier length records

In this section, we attempt to calculate the response time τ_L of glacier AX010 from its length record. Considering a linear system, for a reference state glacier with length L_0 and ELA E_0 Oerlemans (2001) relates the fluctuation in ELA ΔE to change in glacier length ΔL as:

$$\Delta E\,(t) = \frac{1}{C}\left[\Delta L\,(t) + \tau_L \frac{\mathrm{d}}{\mathrm{d}t}\Delta L\,(t)\right]. \tag{7}$$

Here $L(t)=L_0+\Delta L$ is the glacier length and $E(t)=E_0+\Delta E$ is the ELA at any time t, and $C<0$ is the climate sensitivity that determines how the steady state glacier length is related to the ELA. Although a second- or higher-order formulation for the equilibrium glacier length and ELA would be more accurate, the linear model (Eq. 7) should yield a sufficiently accurate order of magnitude of τ_L. In the cases where an annual record of glacier length is not available, this is circumvented by linear interpolation between available data. A polynomial of degree N is then fitted to the resulting data with a least-square method so that $\Delta L\,(t) = \sum_{n=0}^{N} k_n\, t^n$. Here k_n are the coefficients. The time derivative of length change is then obtained as $\frac{\mathrm{d}}{\mathrm{d}t}\Delta L\,(t) = \sum_{n=1}^{N} n\, k_n\, t^{n-1}$. Once the coefficients k_n and the climate sensitivity C are defined, one can either estimate the response time τ_L or reconstruct the ELA using Eq. (7) provided that one of them is known. Here we use the reconstructed ELA (Kayastha and Harrison, 2008) to estimate τ_L of glacier AX010.

The length of glacier AX010 was recorded intermittently in 1978, 1989, 1991, 1995–1999, and 2004 (Fig. 3a). By interpolating the observed values linearly, we first prepare an annual record of glacier length for the period 1978–2004. The resulting data are plotted and fitted with a polynomial of degree $N=4$ (Fig. 6a). After defining the coefficients k_n, we investigate the climate sensitivity C of the glacier using a flowline model discussed in the previous section. ELA perturbations in a range between −25 and +25 m a.s.l. are imposed on the reference state glacier to obtain the corresponding changes in glacier length. These points are then plotted and fitted with a least-square linear line (Fig. 6b), whose slope determines the climate sensitivity $C=-24.6\,\mathrm{m\,m^{-1}}$ of glacier AX010. The figure reveals a remarkably linear relation between changes in ELA and resulting changes in the glacier length. This means that the linear equation (Eq. 7) should generally yield an accurate estimate of τ_L. Based on the analytical solution, Oerlemans (2001) suggests that for a glacier with accumulation zone roughly four times wider than the ablation part, C can be estimated from the mean slope S

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of a glacier, so that $C = -\frac{2}{S} \times 150\%$. The mean slope of glacier AX010 (1991 AD) is $S \sim 0.21$, which yields $C \sim -15$. This value is smaller than the one obtained from the numerical model.

Kayastha and Harrison (2008) reconstruct the changes in ELA since the Little Ice Age (LIA) in the eastern part of Nepal Himalaya, by using a linear relationship between the ELA and the Toe-to-Head Altitude Ratio (THAR). They suggest the average rate of change of ELA over 1959–1992 in the Khumbu massif to be about $0.87 \pm 1.06\ \mathrm{m\,a^{-1}}$. Using the above estimated values of k_n and C, the model is simulated to reconstruct the ELA by tuning τ_L. The simulations are continued until τ_L is optimized so that the reconstructed ELA best describes the one estimated by Kayastha and Harrison (2008). The reconstruction of the ELA whose linearization yields the same value as obtained by Kayastha and Harrison (2008) is shown in Fig. 6a. Such a reconstruction is associated with $\tau_L \sim 37$ yr. Table 7 summarizes the climate sensitivity C and the length response time τ_L of a few valley glaciers, as derived by inverse modelling.

4.5 Summary

It is interesting to synthesize and compare the response times of glacier AX010 obtained from several approaches (Table 8). The volume response time τ_V is estimated by four, and the length response time τ_L by two different methods. The only method that yields both timescales is a numerical ice-flow model. The outcome of the model confirms that τ_V is shorter than τ_L. It means that the glacier volume is much more directly affected than its length, because a glacier exhibits a three-dimensional response to a climate change.

Amongst the methods considered, the estimates based on the kinematic wave theory yield unrealistically high values of τ_V for a small valley glacier. Nye model yields the largest value, $\tau_V \sim 212$ yr. This is not a surprising result because the consideration of a conventional value for the profile shape parameter $f = 0.5$ is expected to yield a longer timescale. Even a more suitable value of $f \sim 0.30$ (Jóhannesson et al., 1989b) for glacier AX010 does not improve the order of magnitude ($\tau_V \sim 126$ yr). Jóhannesson et al. (1989a) report that the long timescale predicted by the kinematic theory is the consequence of the assumed dynamics near the terminus and the overestimation of the change in ice volume after a sudden change in mass balance. A more realistic timescale (τ_V~29 yr) is obtained using the alternative approach by Jóhannesson and others (1989a). However, these methods do not account for the height-mass balance feedback. Among a few available models that includes this feedback, Harrison's effort to modify the Jóhannesson's estimate was considered here. This model yields τ_V~45 yr. A similar magnitude is obtained using a numerical flow model (τ_V~56 yr). By extracting the thickness and velocity data from a numerical model of glacier AX010, Naito et al. (2001) estimate a similar value, τ_V~53 yr (via Jóhannesson et al., 1989a).

Similarly, the length response time τ_L of glacier AX010 is investigated using two different methods. However the magnitudes of these estimates (τ_L~55–85 yr from a numerical flow model, and τ_L~37 yr from an inverse model) do not overlap. Flow models are known to yield a relatively longer timescales (Oerlemans, 2007), whereas the linearity assumption employed in the inverse model may yield untrustworthy results.

Adhikari and Huybrechts (2009) suggest that the longer response times obtained for glacier AX010 through flow modelling indicate that the flow parameters considered in the model are too low. Based on the volume response, however, it seems that the range of response times obtained from the flow model do not appear excessively large because other methods, especially Harrison's estimate (which includes the height-mass balance feedback), also yield similar values. This may indicate that choice of flow parameters for relatively stiff (cold) ice reflects the real dynamics of glacier AX010.

5 Reaction time revisited

Using the flow model introduced in Sect. 4.3, we revisit the concept of terminus response time to a climate perturbation. To better understand the meaning of this timescale and in an attempt to examine the influence of climate history, we impose sinusoidal temperature (mass balance) perturbations of different amplitudes (±1 to ±3°C) and periods (8 to 40 yr) in 480-year model simulations. The first 240 yr of the simulation are taken as a spinup and we analyze the glacier terminus response in the final 240 yr of each simulation. Figure 7a plots an illustrative time series from the model experiments, for a sinusoidal temperature forcing of ±2°C with a period of 24 yr (solid line). This equates to 12 yr of warming followed by 12 yr of cooling. The dotted line indicates the lagged response of the glacier length to this temperature forcing. The response is strongly correlated with the forcing, with the strongest correlations ($r=-0.87$) found at a lag of 8 yr. There is no apparent difference in the response to cooling or warming trends.

While this meshes with the reaction time estimates for glacier AX010 discussed in Sect. 3, modelled reaction time is strongly sensitive to the period (duration) and magnitude of the forcing. The solid line in Fig. 7b plots the calculated reaction time for temperature forcings of 8 to 40 yr, based on the lag time of the peak negative correlation between the time series of temperature and glacier length. Reaction time is proportional to the duration of the temperature anomaly, increasing roughly linearly from 2 to 18 yr for temperature forcings of 8 to 40 yr. This result may be due to the slower rate of temperature change in the longer-period sinusoidal forcing. With more gentle forcing, the glacier takes longer to respond. To test this, we also calculated the terminus reaction time as a function of the time-integrated (cumulative) warming or cooling. The result is plotted with the dashed line in Fig. 7b. This gives shorter reaction times (from 0 to 8 yr), but the same qualitative relationship of longer reaction times in response to long-duration temperature cycles.

These results can be understood in terms of the "inertia" in the glacier as a result of its climate history. As an example, the 40-yr temperature cycle means that the glacier has experienced 20 yr of cooling at the time the climate reverses and begins to warm. It is in a stable or advancing state and it takes several years before the warm anomaly actually translates to terminus retreat. The specific reaction time for any glacier will be determined by a combination of this climatic history and the glacier's geometric/topographic/dynamic setting. Further exploration is needed to better define (if possible) what is meant by reaction time.

6 Conclusions and recommendations

The response characteristics of glacier AX010 have been investigated using several techniques. First, the initial terminus response or reaction time was estimated. A rough inspection of the variation in temperature trend and resulting fluctuation of glacier terminus reveals a qualitative range of reaction time to be around 8–13 yr. This rough estimate of the reaction time is refined using a cross-correlation analysis, which yields τ_R~8 yr. Glacier modelling with idealized climate perturbations gives reaction times that are consistent with this value, but also highlights the intrinsic ambiguity of this characteristic timescale. The climate history needs to be accounted for to better understand the terminus reaction time of a valley glacier to an instantaneous or short-term climate anomaly.

The volume response time τ_V for glacier AX010 is estimated to be in the broad range of 29–212 yr. The few estimates based on linear kinematic wave theory (212 yr from Nye model, and 126 yr from Jóhannesson model) can be dismissed from this range as many glaciologists (e.g. Jóhannesson et al., 1989b; Van de Wal and Oerlemans, 1995; Oerlemans, 2001) report that kinematic theory yields an unrealistically long timescale. This narrows down the range of τ_V to 29–56 yr. The upper limit of this range is the average estimate of a flow model. The lower limit of the range of timescale comes from a simple approach by Jóhannesson et al. (1989a). As this estimate does not account for the height-mass balance feedback, τ_V is likely to be longer than 29 yr. Harrison's estimate of 45 yr, which includes the height-mass balance feedback, probably best represents the volume response time of glacier AX010. A similar range (37–70 yr) is obtained for the length response time τ_L of glacier AX010. Here also the upper limit is associated with the average estimate from a flow line model. The lower limit of the estimates comes from an inverse model.

A rigorous attempt to quantify the reaction time in terms of (possibly) climatic and

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geometric parameters (of present and past) by using a full-Stokes dynamical glacier model, employing a grid-transformation technique such as finite element, would be an interesting topic for further research. In addition, any efforts at improving the inverse model by considering a sophisticated nonlinear relationship between the local climate and glacier geometry will be useful assets in glaciology.

On a final note, one could of course question the future relevance of dealing with response timescales for a small valley glacier such as AX010 that is predicted to entirely melt down almost in situ for the type of climate warmings projected for the 21st century (Adhikari and Huybrechts, 2009). Under such circumstances, inferences made from past behaviour to understand the future is questionable, but that also applies to many other valley glaciers.

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Table 1. The estimated reaction time τ_R for a few valley glaciers. The range of reaction times tabulated in the first row is based on an analysis of 21 North Cascade Glaciers. Note that McClung and Armstrong (1993) used the term "time response" to their figure, whereas Calmanti et al. (2007) simply used the term "time lag".

Glacier	Country	Reaction time, a	Reference
North Cascade glaciers	USA	4–16	Pelto and Hedlund (2001)
Blue glacier	USA	10	McClung and Armstrong (1993)
Glaciers in Piedmont and Val d'Aosta	Italy	8–10	Calmanti et al. (2007)
Glacier AX010	**Nepal**	**8**	**This paper**

Table 2. A record of average thickness change over the entire glacier AX010 and over its terminus. The question marks indicate the data unavailability.

Time period	Glacier-average ΔH, m	ΔH at terminus, m	Reference
1978–1991	−8.69[a]	−30[b]	WGMS (1998)[a] Kadota et al. (1993)[b]
1991–1996	−5.70	?	Kayastha and Harrison (2008)
1996–1999	−2.47	?	WGMS (2005)

Table 3. Comparison of the profile shape parameter f of a few glaciers. These parameters are calculated using a multi-decadal profile change data.

Glacier	Country	Time period	L at the end of the time period, km	f
Grosser Aletsch[a]	Switzerland	1927–1983	22.50	0.30
South Cascade[a]	USA	1955–1985	3.10	0.30
Glacier AX010	**Nepal**	**1978–1999**	**1.54**	**0.30**
Gries[a]	Switzerland	1923–1979	5.50	0.20
Rhone[a]	Switzerland	1885–1980	10.30	0.13

[a] Schwitter and Raymond (1993)

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Table 4. Comparison of the volume response time τ_V of a few valley glaciers as estimated by Jóhannesson et al. (1989a, Eq. 3).

Glacier	Country	τ_V, a	Reference
Glaciers in temperate maritime climate		15–60	Paterson (1994)
Abramov glacier	Kyrgyzstan	35	Oerlemans (2001)
Rhonegletscher	Switzerland	31	Oerlemans (2001)
Glacier AX010	**Nepal**	**29**	**This paper**
17 North Cascade glaciers	USA	10–25	Pelto and Hedlund (2001)
Blue glacier	USA	20	McClung and Armstrong (1993)
Nigardsbreen	Norway	12	Oerlemans (2001)
Franz Josef glacier	New Zealand	6	Oerlemans (1997a)

Table 5. Summary of various parameters used to calculate $\overline{G}$ for glacier AX010. All data represent the average of 1978–1991 except for the annual mass balance b, which is the 1996–1999 average.

Elevation range, m	Area (A_0) m^2	ΔA m^2	$(Z-Z_0)$ M	b $m\,a^{-1}$	G $m\,m^{-1}\,a^{-1}$	$G\,(Z-Z_0)\,dA$ $m\,m^{-1}\,a^{-1}\,m^3$
5250–5360	174500	−41500	−1.8	0.15	0.0014	104.48
5200–5250	172500	6500	−6.3	−0.12	−0.0023	94.80
5150–5200	68500	−1500	−11.2	−0.71	−0.0142	−238.62
5100–5150	40500	−4500	−15.3	−1.30	−0.0260	−1783.50
5050–5100	61500	−19500	−18.4	−2.32	−0.0465	−16642.03
5000–5050	40500	−2500	−21.2	−2.65	−0.0530	−2815.47
4952–5000	11000	3000	−20.2	−2.97	−0.0619	3743.78
$\int_{A_0} G\,(Z-Z_0)\,dA$, $m\,m^{-1}\,a^{-1}\,m^3$						−17536.55
ΔV, m^3						−4934000
$\bar{G}$, $m\,m^{-1}\,a^{-1}$						**0.00355**

Table 6. The length response time τ_L and volume response time τ_V for some of the valley glaciers obtained from a numerical ice-flow model.

Glacier	Country	τ_L, a	τ_V, a	Reference
Sofiyskiy glacier	Russia	73–114	–	De Smedt and Pattyn (2003)
Glacier AX010	**Nepal**	**55–85**	**50–62**	**This paper**
Nigardsbreen	Norway	63–73	–	Oerlemans (1997b)
Glacier d'Argentière	France	27–45	–	Huybrechts et al. (1989)
Franz Josef glacier	New Zealand	20–27	13–20	Oerlemans (1997a)

Table 7. Comparison of a number of valley glaciers in terms of their climate sensitivity C and their length response time τ_L as estimated by inverse modeling.

Glacier	Country	C, m m^{-1}	τ_L, a
Rhonegletscher[a]	Switzerland	−32.0	58
Unterer Grindelwald[a]	Switzerland	−40.0	40
Glacier AX010	**Nepal**	**−24.6**	**37**
Glacier d'Argentière[a]	France	−35.0	32
Nigardsbreen[b]	Norway	−26.1	35
Briksdalsbreen[b]	Norway	−8.0	5
Vandret da Palu[b]	Switzerland	−9.0	4

[a]Oerlemans (2001),
[b]Oerlemans (2007)

Table 8. Summary of the response times τ_L and τ_V for glacier AX010 as obtained from different methods.

Description of the method	Volume response time τ_V, a	Length response time τ_L, a
Linear kinematic wave theory		
– Nye (1960)	212	–
– Jóhannesson et al. (1989b)	126	–
Simpler approach		
– Jóhannesson et al. (1989a)	29	–
Including height-mass balance feedback		
– Harrison et al. (2001)	45	–
Numerical ice-flow model		
– e.g. Oerlemans (2001)	50–62	55–85
Linear inverse model		
– Oerlemans (2001)	–	37

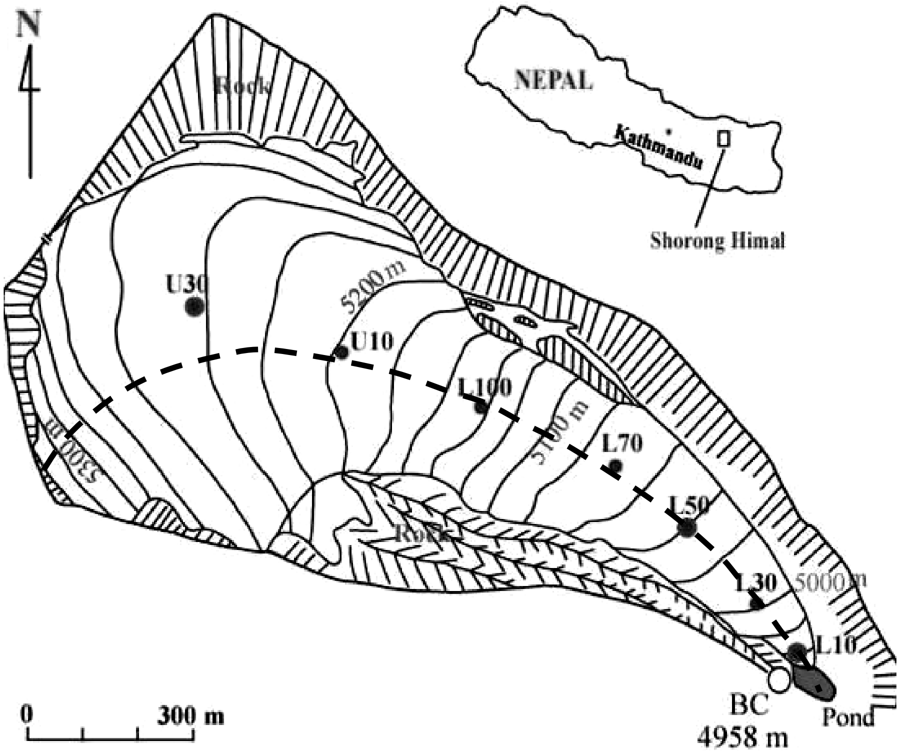

Fig. 1. Location of Shorong Himal and topographic map of glacier AX010 as surveyed in 1979 (Ikegami and Ageta, 1991). Solid circles inside the glacier are positions of mass balance stakes. Solid lines across the glacier are contours of 20 m interval. The dotted line along the glacier is the assumed flow line used for a numerical model.

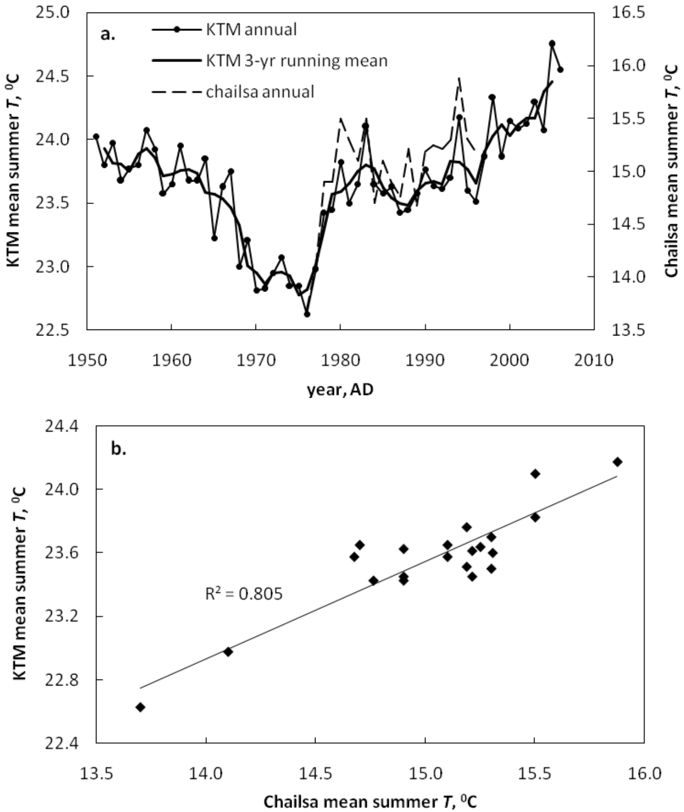

Fig. 2. **(a)** Comparison of mean summer temperature recorded at the Kathmandu (KTM) and Chialsa stations. The black bold line is a 3-yr running mean of KTM data, which is used as the representative temperature for further analysis. **(b)** A regression analysis between KTM and Chialsa temperature using a least-square method.

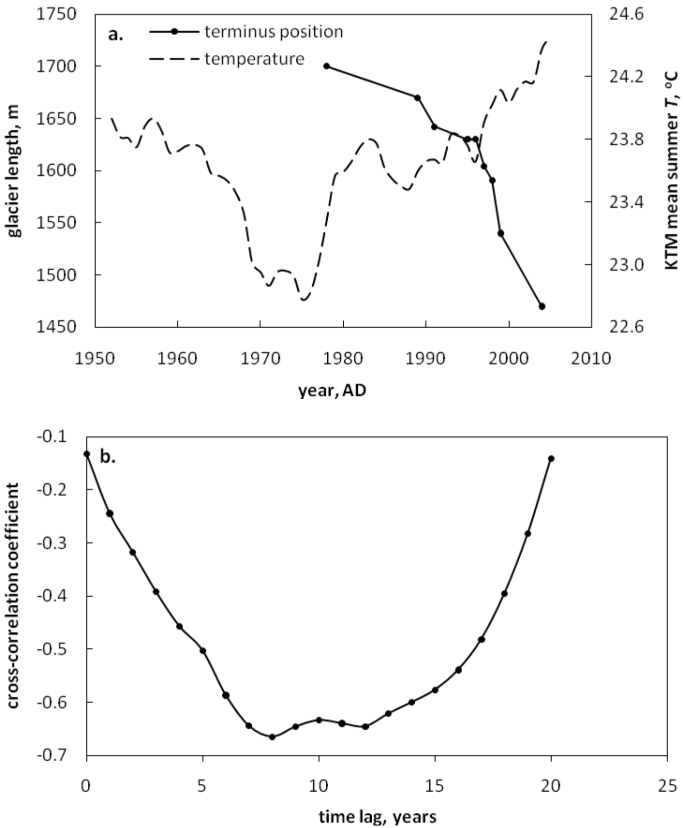

Fig. 3. (a) Length record of glacier AX010 since 1978, along with a 3-yr running mean of KTM temperature. **(b)** A cross-correlation analysis between the temperature variation and the resulting fluctuation of the glacier terminus over the period 1978–2004. A positive lag is applied to the length data.

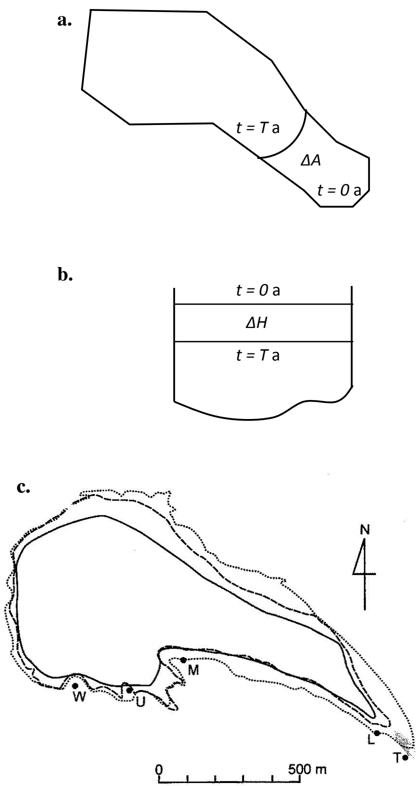

Fig. 4. The glacier geometry **(a** and **b)** that supports the condition $|b_T| < |\bar{b}_r|$. **(a)** The plan of glacier showing the changes in its extent between an initial time t=0 years and any time t=T years. Note that the change in area ΔA is strictly concentrated at the terminus. **(b)** The valley cross-section showing the thinning of glacier ΔH without exposing any lateral rocks. **(c)** The surface boundaries of glacier AX010, as seen in 1978 (outer), 1996 (middle) and 1999 (inner), to illustrate the exposure of lateral rocks all along the glacier during its retreating phase (1978–1999).

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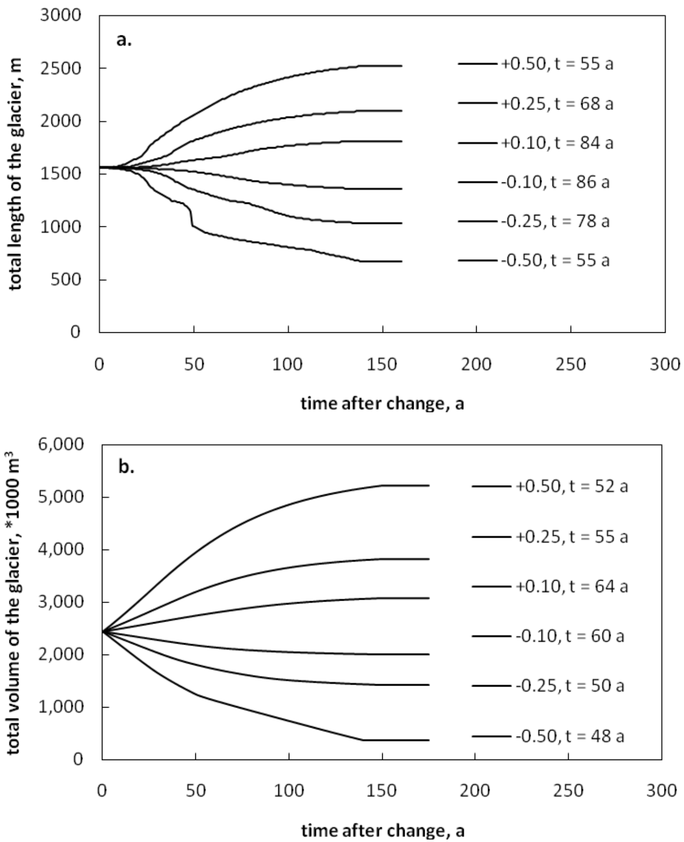

Fig. 5. **(a)** Reaction of the glacier terminus, and **(b)** evolution of glacier volume in response to given changes in mass balance. In both cases, a step change in mass balance in the range between −0.5 and +0.5 m w.e.a^{-1} is imposed on a reference state glacier, whose size corresponds to the 1996 AD glacier stand.

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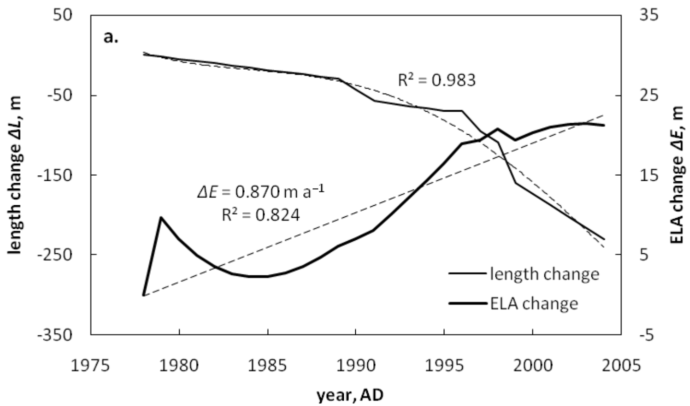
Interactive Discussion



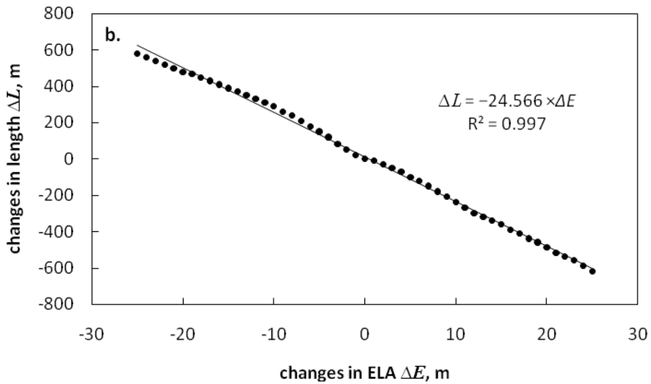


Fig. 6. (a) Glacier length and reconstructed ELA as obtained by inverse modelling. Both curves are plotted with respect to 1978 data, and are accompanied by the corresponding trend lines. The trend line fitted over the length record is a polynomial of degree N=4, while that over the ELA is a linear fit. **(b)** Plot showing the resulting changes in initially steady state glacier length due to applied changes in ELA. The slope of the linear fit reveals the climate sensitivity C of glacier AX010.

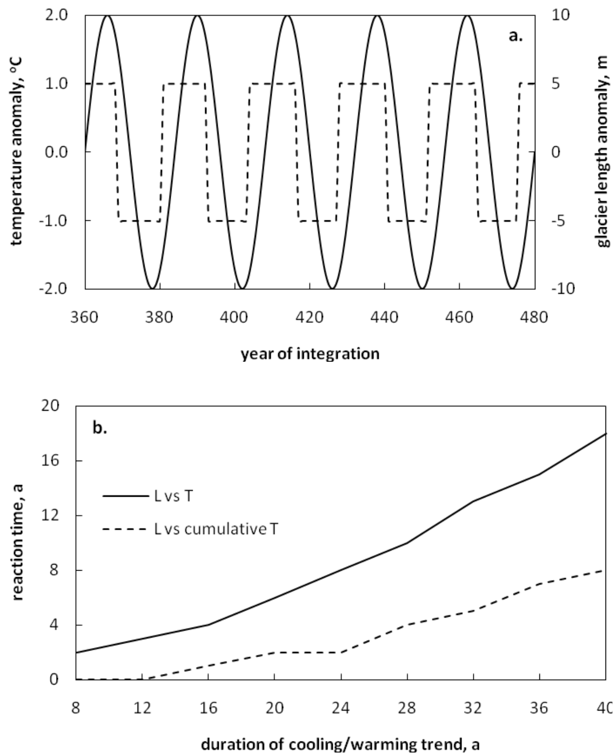


Fig. 7. **(a)** Illustrative numerical experiment testing the glacier length response to sinusoidal temperature perturbations. Plot shows the temperature forcing (solid line) and glacier length anomaly (dotted line) for a 24-yr temperature cycle, model years 360–480 yr. **(b)** Terminus reaction time calculated from the lag time of the peak negative correlation of the glacier length time series with the temperature forcing and the integrated (cumulative) temperature change.