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# From balance to imbalance: a shift in the dynamic behaviour of Chhota Shigri Glacier (Western Himalaya, India)

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# Abstract

Mass-balance and dynamic behaviour of Chhota Shigri Glacier have been investigated between 2002 and 2010 and compared to data collected in 1987/1989. During the period 2002/2010, the glacier experienced a negative glacier-wide mass balance of –0.67 ± 0.40 m water equivalent per year (w.e. $yr^{-1}$). Between 2003 and 2010, elevation and ice flow velocities are slowly decreasing in the ablation area leading to a 24 to 37% reduction in ice fluxes, an expected response of the glacier dynamics to its recent negative mass balances. The reduced ice fluxes still remain far larger than the balance fluxes calculated from the year 2002 to 2010 average surface mass balances. Therefore, further slow down, thinning and terminus retreat of Chhota Shigri Glacier are expected over the next years. Conversely, the 2003/2004 ice fluxes are in good agreement with ice fluxes calculated assuming that the glacier-wide mass balance is zero. Given the limited velocity change between 1987/1989 and 2003/2004 and the small terminus change between 1988 and 2010, we suggest that the glacier has experienced a period of near zero or slightly positive mass balance in the 1990s, before shifting to a strong imbalance in the 21st century. This result challenges the generally accepted idea that glaciers of Western Himalaya have been shrinking rapidly for the last decades.

# 1. Introduction

Although Himalayan glaciers have important social and economic impacts (e.g. Barnett and others, 2005), they have never been monitored on a long-term basis and little is known about recent glacier trends or their contribution to local and regional water supplies, even giving rise to a controversial statement in the IPCC 4th assessment report saying that “the likelihood of them disappearing by the year 2035 or perhaps sooner is very high if the Earth keeps warming at the current rate” (Cogley and others, 2010). A general negative mass balance of mountain glaciers on a global level is clearly revealed from recent research (e.g. Cogley, 2009; Zemp and others, 2009) but the effect of global warming in the Himalaya is still under debate (Yadav and others, 2004; Roy and Balling, 2005). Though temperate glacial mass-balance change is one of the best indicators of climate change (Oerlemans, 2001; Vincent and others, 2004; Ohmura and others, 2007) the paucity of mass-balance data in Himalaya makes it difficult to obtain a coherent picture of regional climate change impacts in this region. In the Indian Himalaya the first mass-balance study started on Gara Glacier (Himachal Pradesh) in September 1974 (Raina and others, 1977) and ended in 1983 (Dobhal and others, 2008). According to Dyurgerov and Meier (2005), eight glaciers in the Indian Himalaya were surveyed for mass balance at least for one year during the 1980s. Unfortunately each study was restricted to short periods, not more than one decade (Dobhal and others, 2008). Remote sensing studies were also attempted in this part of Himalaya, but these studies either deal with only surface area changes (e.g. Kulkarni and others, 2007; Bhambri and others, 2011) or cover short periods (Kulkarni, 1992; Berthier and others, 2007).

The present study is based on mass balance and surface ice flow velocity measurements conducted on Chhota Shigri Glacier, Himachal Pradesh, India between 2002 and 2010, and on a comparison to data collected in 1987/1989. In the Indian Himalaya, this is one of the longest continuous field mass-balance dataset. Moreover, in October 2009, a Ground Penetrating Radar (GPR) survey was also conducted to measure ice thickness. Eight years of mass-balance measurements, surface ice velocities and ice thickness data provide an opportunity to study the behaviour of this glacier. The main objectives of this paper are (i) to present the recent mass balance of Chhota Shigri Glacier, (ii) to determine the ice fluxes at five cross sections from thickness and ice velocities, and (iii) to compare these data with the ice fluxes inferred from cumulative surface mass balance upstream of the same cross sections. These results give insights into the mass-balance trend of the glacier over the last two to three decades, and allow us to assess whether it is in equilibrium with the climate of the 21st century.

## 2. Site description and methodology

### 2.1 Site description

Chhota Shigri Glacier (32.2° N and 77.5° E) is a valley-type glacier located in the Chandra-Bhaga River basin of Lahaul and Spiti Valley, Pir Panjal Range, Western Himalaya. This glacier extends from 6263 m to ~4050 m a.s.l., is ~9 km long and covers 15.7 $km^2$ of area. The snout of this glacier is easy to locate from one year to another because it is well defined, lying in a narrow valley and giving birth to a single proglacial stream. The main orientation of this glacier is north except for its tributaries which have a variety of orientations (Fig. 1). The lower ablation area (<4400 m a.s.l.) is partly covered by debris representing ~3.4% of the total surface area. This glacier is located in the monsoon-arid transition zone and is influenced by two atmospheric circulation systems: the Indian monsoon during summer (July–September) and the northern-hemisphere mid-latitude westerlies during winter (January–April) (Singh and others, 1997; Bookhagen and Burbank, 2006; Gardelle and others, 2011).

### 2.2 Mass balance

The first series of mass-balance measurements was performed on Chhota Shigri Glacier between 1987 and 1989 (Nizampurkar and Rao, 1992; Dobhal and others, 1995; Kumar, 1999). The bedrock topography and surface ice velocity were also surveyed over the same period by gravimetric and stake displacement methods respectively (Dobhal and others, 1995; Kumar, 1999). We re-initiated the mass-balance observations in 2002. From that year, annual surface mass-balance measurements have been carried out continuously on Chhota Shigri Glacier at the end of September or beginning of October using the direct glaciological method (Paterson, 1994). Ablation was measured through a network of ~22 stakes distributed between 4300 and 5000 m a.s.l. (Fig. 1) whereas in the accumulation area, the net annual accumulation was obtained at six sites (by drilling cores or pits) between 5100 and 5550 m a.s.l. (Wagnon and others, 2007). In the accumulation area, the number of sampled sites is limited due to the difficulty in access and high elevation. The glacier-wide mass balance, $B_a$ is calculated according to:

$$B_a = \sum b_i \, (s_i / S) \tag{1}$$

Where $b_i$ stands for the mass balance of the altitudinal range (m w.e. $\mathrm{yr}^{-1}$), $i$, of map area $s_i$ and $S$ symbolizes the total glacier area. For each altitudinal range, $b_i$ is obtained from the corresponding stake readings or net accumulation measurements.

## 2.3 Surface velocity

Annual surface ice velocities were measured at the end of each ablation season (September-October) by determining the annual stake displacements (~22 stakes) using a Differential Global Positioning System (DGPS). These geodetic measurements were performed in kinematic mode relative to two fixed reference points outside the glacier on firm rocks. The accuracy in x (easting), y (northing) and z (elevation) of each stake position is estimated at ±0.2 m depending mainly on the size of the hole in which the stake has been set-up. Thus the surface ice velocities measured from stake displacements have an accuracy of ±0.3 m $yr^{-1}$.

### 2.4 Ice thickness

Ground penetrating radar measurements were conducted in October 2009 to determine ice thickness on five transverse cross sections (Fig. 1) between 4400 and 4900 m a.s.l. A pulse radar system (Icefield Instruments, Canada) based on the Narod transmitter (Narod and Clarke, 1994) with separate transmitter and receiver, was used in this study with a frequency centered near 4.2 MHz and antenna length of 10 m. Transmitter and receiver were towed in snow sledges along the transverse profile, separated by a fixed distance of 20 meters, and used to record measurements every 10 m. The positions of the receiver and the transmitter are known through DGPS measurements, within an accuracy of ±0.1 m. The speed of electromagnetic wave propagation in ice has been assumed to be 167 m $\mu s^{-1}$ (Hubbard and Glasser, 2005). The field measurements were performed in such a way as to obtain reflections from the glacier bed located more or less in the vertical plane with the measurement points at the glacier surface, allowing the determination of the glacier bed in two dimensions. The surface of the bedrock was constructed as an envelope of all ellipse functions, which give all the possible reflection positions between sending and receiving antennae. Ice thickness was measured along four transverse profiles (profiles 1-4) on the main glacier trunk and one (profile 5) on a western tributary (Fig. 1).

# 3. Data analysis and results

## 3.1 Glacier-wide mass balance and mass-balance profile

The annual glacier-wide mass balance and cumulative mass balance of Chhota Shigri Glacier between 2002 and 2010 are plotted in Figure 2. The glacier-wide mass balance was negative except for three years (2004/2005; 2008/2009 and 2009/2010). It varies from a minimum value of –1.40 m w.e. in 2002/2003 to a maximum value of +0.33 m w.e. in 2009/2010. The cumulative mass balance of Chhota Shigri is -5.37 m w.e. between 2002 and 2010 while the glacier-wide mass balance averaged over the same period is –0.67 m w.e. $yr^{-1}$.

The quantitative uncertainty associated with the glaciological mass balance requires a distinction between the accumulation zone and the ablation zone. In the accumulation zone, the surface mass-balance measurements were obtained from shallow boreholes (auger). Therefore, they are based on core length and density determination. In the ablation zone, the measurements have been carried out from ablation stakes. The overall error (standard deviation) on point measurements are estimated at 0.30 m w.e. and 0.15 m w.e. in the accumulation zone and in the ablation zone, respectively. The overall error comes from a variance analysis (Thibert and others; 2008) applied to all types of errors (ice/snow density, core length, stake height determination, liquid-water content of the snow, snow height). Although conducted on a glacier in the Alps, the analysis of Thibert and others (2008) can be generalized to other glaciers because it is based on measurement errors which are similar on every glacier when using the glaciological method. However, only 6 sites are sampled in the accumulation zone (11.6 km²), and 22 sites in the ablation zone (4.1 km²). The uncaptured spatial variability of surface mass balance may cause systematic errors on the glacier-wide mass balance. In the accumulation zone, the spatial variability remains unknown and is probably very high as observed for other glaciers (e.g. Machguth and others, 2006). In the ablation zone, stakes set up at the same altitude show similar values except on the terminal tongue which is debris covered (0.54 km²). Consequently, the overall uncertainties on mass-balance profile have been assessed at 0.5 m w.e. in the accumulation zone, 0.25 m w.e. in the white ablation zone and 0.5 m w.e. in the debris covered area of the glacier. Moreover, the surface area estimation also causes systematic error. The uncertainty on the surface area calculated for each altitudinal range is estimated at 5%. Combining these errors at different altitudinal ranges using Equation (1), the uncertainty on the annual glacier-wide mass balance is 0.4 m w.e. yr$^{-1}$. As revealed by other studies (e.g. Vincent, 2002; Thibert and others, 2008; Huss and others, 2009), this estimation confirms that the glaciological method needs to be calibrated by a volumetric method over a long period of monitoring (i.e. >5 years) in order to limit the systematic errors and to improve the accuracy of absolute values of mass balance. Note that the uncertainty of relative changes in mass balance from year to year is smaller than those inherent in annual mass balances, given that the influence of systematic errors can be reduced.

We also calculated the mass-balance profile between 2002 and 2010 (Fig. 3). For each altitudinal range, we computed the average of all available measurements. Figure 3 reveals that melting in the lowest part of the ablation area (below 4400 m a.s.l.) is reduced by about 1 m w.e.

yr$^{-1}$ irrespective to its altitude. This is due to the debris cover (~5 to 10 cm thick debris mixed with isolated rocks) which reduces the melting in this region (Mattson and others, 1993; Wagnon and others, 2007). Moreover the lower part of Chhota Shigri Glacier flows in a north-south oriented deep and narrow valley (Fig. 1), causing the glacier tongue to receive less solar radiation due to the shading effect of the steep valley slopes.

### 3.2 Ice thicknesses and cross section areas

Thanks to clear reflections, ice-bedrock interface was generally easy to determine on all profiles. Figure 4 provides an example of the radargram obtained at cross section 2. A radar wave velocity of 167 m µs$^{-1}$ was used for calculations of ice thickness at all the profiles. The cross sections obtained from GPR measurements reveal a valley shape with maximum ice thickness greater than 250 m (Fig. 5). The centerline ice thickness increases from 124 m at 4400 m a.s.l. (cross section 1 in Fig. 1) to 270 m at 4900 m a.s.l. (cross section 4), which confirms that the thicknesses obtained by gravimetric methods in 1989 (Dobhal and others, 1995), twice lower than the present results, were under-estimated as proposed by Wagnon and others (2007). The cross sectional areas are given in Table 1. The accuracy of the calculated ice thickness is determined, in part, by the accuracy of the measurement of the time delays and the antenna spacing. Additional errors may arise because the smooth envelope of the reflection ellipses is only a minimal profile for a deep valley-shape bed topography, with the result that the ellipse equation will be governed by arrivals from reflectors located toward the side and thus not directly beneath the points of observation. Further errors may be introduced by assuming that all reflection points lie in the plane of the profile rather than on an ellipsoid. No errors associated to radar wave velocity variations between snow and ice have been accounted for because all cross sections were surveyed in the ablation zone or slightly above (with the firn-ice transition depth at the surface or < 2 m deep). Hence, the radar wave velocity for ice (167 m $\mu s^{-1}$) was used to calculate all ice depths. The overall uncertainty in ice thickness is estimated as ±15 m. Given that the uncertainty in ice surface coordinates is low (±0.1 m), the uncertainty on cross section areas mainly arises from the uncertainty in ice thickness. The uncertainties in cross section areas are 16, 9, 10, 10 and 15% for the cross sections 1, 2, 3, 4 and 5 respectively.

### 3.3 Ice velocity

Annual surface ice velocities were also measured between 2002 and 2010. However, some data gaps exist due to discontinuous DGPS signal, or loss of stakes. The ice velocities from 2003/2004 were used in this study because they provided the most complete dataset (Fig. 6). The center line horizontal ice velocities at each cross section were calculated by linear interpolation method along the center line between the velocities measured immediately upstream and downstream of the cross section (ablation stakes visible on Fig. 1). Mean cross section velocities are required to compute the ice fluxes (see section below). A map of the surface ice velocity field has been derived by correlating 2.5-m SPOT5 images acquired on 13 November 2004 and 21 September 2005 (Berthier and others, 2005). Comparison of the satellite-derived velocities with 16 nearly simultaneous DGPS velocity measurements shows a mean difference of 0.2 m $yr^{-1}$ and a standard deviation of 1.6 m $yr^{-1}$. The ratio between the center line horizontal velocity and the mean surface velocity (all extracted from the satellite-derived 2004/2005 velocity field) was found to be 0.80 and 0.78 for cross sections 2 and 3 respectively. Reliable velocity measurements could not be measured from SPOT5 imagery for other cross sections. Using the mean value of

0.79, the mean horizontal velocity has been calculated from the center line velocity for each gate cross section (Table 1).

### 3.4 Ice fluxes from kinematic method

The ice flux $Q$ ($m^3$ of ice per year) through each cross section was calculated using the cross sectional area $S_c$ ($m^2$) and depth-averaged horizontal ice velocity $U$ (m $yr^{-1}$).

$$Q = U\,S_c \qquad (2)$$

The depth-averaged horizontal ice velocity was derived from the mean surface ice velocity calculated in the previous section. Nye (1965) gives ratios of depth-averaged horizontal ice velocity to mean surface ice velocity varying from 0.8 (no sliding) to 1 (maximum sliding). Here, we assume a mean basal sliding, with a constant ratio of 0.9. The calculated ice fluxes and maximum depth at each cross section are given in Table 1. The flux through cross section 3 at 4750 m a.s.l. is higher than the flux through cross section 4 at 4900 m a.s.l. This is due to the ice influx from the western part of the glacier (flux through cross section 5) which contributes to cross section 3 and not to cross section 4 (Fig. 1).

The largest uncertainty on the depth-averaged horizontal ice velocity results from the ratio between the depth velocity and the surface flow velocity. The estimated factor 0.9 and unknown variations in the basal sliding lead to an uncertainty of roughly ±10% in the calculated flux, which lies within the range of uncertainty of the other variables as discussed by Huss and others (2007). Consequently, we can assess that depth-averaged horizontal ice velocity at each cross section is known with an accuracy of 1.0 to 3.0 m $yr^{-1}$ depending on the cross sections. Combining these errors on the cross sectional area and mean velocity, the uncertainties on the ice fluxes are 0.21, 0.69, 0.92, 0.89 and 0.38 x $10^6$ $m^3$ $yr^{-1}$ for the cross sections 1, 2, 3, 4 and 5 respectively. Here we have considered that the errors are systematic, so these uncertainties are probably over-estimated.

### 3.5 Ice fluxes obtained from surface mass balance

We also calculated ice fluxes using annual surface mass balance measured during 2002/2010. Although the dynamic changes are neglected here, this method allows us to estimate the ice fluxes for each section from mass-balance data according to the following equation:

$$Q = \frac{1}{0.9} \sum_{z}^{z\max} b_i s_i \qquad (3)$$

Where $Q$ is the ice flux (converted into $m^3$ of ice per year using an ice density of 900 kg $m^{-3}$, hence the factor 1/0.9) at a given elevation, $z$, and $b_i$ is the annual mass balance of the altitudinal range $i$ of map area $s_i$. The altitudinal ranges taken into account in the calculation are located between $z$ and the highest range of the glacier $z_{\max}$ (highest altitude of the glacier area contributing ice to the cross section). We assume here that on each point of the glacier above this altitude, $z$, the surface elevation has remained unchanged from one year to the next.

The ice fluxes calculated from annual mass-balance data at the 5 cross sections each year are given in Table 3, while the average ice fluxes for the eight years are given in Figure 7. The uncertainties on ice fluxes resulting from surface mass balance are directly derived from the mass-balance uncertainties (see section 3.1) applied to areas contributing to each cross section.

# 4. Discussion

The first and main objective of this section is to discuss the mass-balance change of Chhota Shigri Glacier over the last two to three decades using not only direct mass-balance observations (over the last eight years) but also ice-flux analysis. The second goal is to give insights into the specific dynamics and the future retreat of this glacier that can be expected in relation to its recent surface mass balance (hereafter referred to as SMB).

## 4.1 Null to slightly positive mass balance during the 1990s inferred from ice fluxes

The ice fluxes obtained from the kinematic method using ice thickness and 2003/2004 ice velocities are much higher than the average fluxes derived from the 2002/2010 SMBs, the latter being often negative (Table 2). Thus in this section, to assess the mean state of the glacier corresponding to the ice fluxes obtained by the kinematic method, we compare these measured ice fluxes to theoretical ice fluxes calculated from SMB assuming the glacier to be in steady state. The glacier-wide mass balance obtained by the glaciological method is -0.67 m w.e. $yr^{-1}$ over the 2002/2010 period. Consequently, the SMB needs to be increased by 0.67 m w.e. $yr^{-1}$, for the glacier to be in steady state with the present surface area. For each year (2002/2010), we calculated the theoretical ice flux from SMB at each cross section assuming the glacier was in steady state. For this purpose, every year, a theoretical SMB at each elevation has been calculated by substracting the overall annual specific SMB of the same year. For instance, year 2002/2003 was characterized by a negative annual glacier-wide SMB of -1.40 m w.e. so we calculated a new SMB profile by adding +1.40 m w.e. to the SMB at each elevation. In contrast year 2009/2010 was characterized by a positive annual glacier-wide SMB of +0.33 m w.e. so we calculated a new SMB profile by subtracting 0.33 m w.e. from the SMB at each elevation. The resulting ice fluxes are reported in Table 3, together with the mean ice flux at each cross section over the eight years and the corresponding standard deviations.

These ice fluxes are close to the 2003/2004 ice fluxes obtained by the kinematic method (Fig. 7) indicating that the dynamic behaviour of the glacier in 2003/2004 is representative for steady-state conditions. This suggests that in the years preceding 2003/2004, the glacier-wide mass balance of this glacier has probably been close to zero and that, in 2003/2004, the ice fluxes had not adjusted to previous year negative SMB.

This result is also supported by other observations. First, the ice velocities measured in 1987/1988 (Dobhal and others, 1995) are very close to the 2003/2004 values (Fig. 6) suggesting that the dynamic behaviour of this glacier did not change a lot between 1988 and 2004. Second, the terminus fluctuation measured between 1988 and 2010 show a moderate retreat equal to 155 m, equivalent to only 7 m yr$^{-1}$, in agreement with conditions not far from steady state. Given that Berthier and others (2007) observed a glacier-wide SMB of Chhota Shigri Glacier of approximately -1 m w.e. yr$^{-1}$ during the period 1999 to 2004, the glacier is likely to have experienced a null to slightly positive mass balance between 1988 and the end of the 20th century.

### 4.2 Glacier dynamics starting to adjust to 21st century negative SMB

In theory, the response of ice fluxes to surface mass balance is immediate (Cuffey and Paterson, 2010, p. 468) but observations show a 1-5 year delay (Vincent and others, 2000; Span and Kuhn, 2003; Vincent and others, 2009). For instance, Span and Kuhn (2003) found synchronous decrease in ice velocity between eight glaciers in the Alps, which are driven by the same mass-balance changes (Vincent and others, 2005). Consequently, the recent dynamic behaviour of Chhota Shigri Glacier should be affected by the negative mass balance since 1999. However, the stake network on Chhota Shigri Glacier, originally designed for SMB measurements, is not best suited to accurately compare either the ice velocities or the thickness variations because the measurements have not been performed exactly at the same location every year and they are mainly restricted to the ablation area.

In spite of the above limitation, an attempt has been made to compare ice velocities and elevations from the available stake network. For this purpose, stakes measured at the beginning and at the end of the series have been selected on five short longitudinal cross sections (A, B, C, D and E in Fig.1) along the center line of the glacier where the network is most dense. The elevations in 2003 and 2010 and the ice velocities in 2003/2004 and 2009/2010 have been reported on these longitudinal cross sections to deduce thickness and velocity changes in the ablation area (Fig. 8, Table 4). Although the accuracy of the results is affected by the distance between the point measurements, we can conclude that the part of the glacier below 4750 m a.s.l. is in strong recession. First, the thickness has decreased annually by 0.7 to 1.1 m $yr^{-1}$ over the last seven years. Second, the ice velocities have been reduced by ~7 m $yr^{-1}$ between 2003 and 2010 resulting in a 24 to 37% decrease in the ice fluxes since 2003. Despite an improvable monitoring network, it may be surmised that the ice fluxes have been affected by the negative glacier-wide mass balance during (at least) the last eight years, and the dynamics of this glacier are progressively adjusting to the negative SMB. Consequently, we expect an accelerated terminus retreat in the coming years. If the SMB remained equal to its 2002/2010 average value in the future, the terminus would retreat by 5.6 km to reach an altitude of 4870 m a.s.l. (altitude where the ice flux is equal to zero) (Table 2).

## 5. Conclusion

The Chhota Shigri Glacier experienced negative mass balance over the 2002/2010 period. The glacier-wide mass balance of the glacier is estimated at –0.67 m w.e. yr$^{-1}$ between 2002 and 2010, revealing strong unsteady-state conditions over this period. Conversely, ice fluxes calculated through 5 transverse cross sections by the kinematic method correspond to near steady-state conditions before 2004. Given that ice velocities measured in 2003/2004 are close to those measured in 1988, and that terminus has retreated only 155 m between 1988 and 2010, it seems that the dynamic change was moderate between 1988 and 2004. Therefore, considering that Berthier and others (2007) observed a negative glacier-wide mass balance of about -1 m w.e. yr$^{-1}$ between 1999 and 2004 using satellite images, our analysis suggests that the glacier experienced a period of slightly positive or close to zero mass balance at the end of the 20th century, before starting to shrink. As Chhota Shigri seems to be representative of other glaciers in the Pir Panjal Range (Berthier and others, 2007), it is possible that many Western Himalayan glaciers of northern India experienced growth during the last 10-12 years of the 20th century, before starting to shrink at the beginning of the 21st century.

Since 2003, ice velocities and elevation are decreasing in the ablation area. Our data suggest that the ice fluxes have diminished by 24 to 37% below 4750 m a.s.l. between 2003 and 2010. Even if we account for a 37% decrease in ice fluxes calculated from 2003/2004 ice velocities to obtain present ice fluxes values, it remains a very large imbalance with ice fluxes coming from glacier-wide mass balance of the last eight years. Thus the present dynamics (thickness and ice velocities) of this glacier are far from surface mass balance and climate conditions of the last eight years, even if it is progressively adjusting. Therefore the glacier is likely to undergo accelerated retreat in the near future.

This glacier is almost free of debris and thus its mass-balance variations are closely related to climate changes. This glacier has the longest running series of mass balance measurements in the Himalaya range. In the future, the dynamic behaviour and the mass balance need more detailed investigations, although such field measurements are demanding due to the very high altitude. In order to investigate the annual thickness and the ice velocity changes, we recommend performing elevation and ice velocity measurements on ~12 cross sections including some in the accumulation zone. We also recommend measuring the ice velocities from a dense network of stakes to be set up on longitudinal center lines in order to compare the annual velocity changes at the same points. Finally, we recommend calibrating and checking the mass-balance field measurements from a volumetric method (from photogrammetry or remote sensing techniques).

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## Table and figure captions

Table 1: Calculated ice flux, mean surface ice velocity and maximum ice depth at each cross section. The mean surface horizontal ice velocities are from DGPS measurements performed in 2003/2004. The satellite-derived mean ice velocities are from the correlation of satellite images acquired on 13 November 2004 and 21 September 2005 (NA: Not available).

Table 2: Ice fluxes (in $10^6$ $m^3$ ice $yr^{-1}$), inferred at each cross section from annual mass-balance data.

Table 3: Ice fluxes (in $10^6$ $m^3$ ice $yr^{-1}$), obtained at every cross section, using steady state mass-balance assumption for every surveyed year.

Table 4: Thickness and surface velocity changes between 2003 and 2010 on 5 longitudinal cross sections (NA: Not available).

Figure 1: Map of Chhota Shigri Glacier with the measured transverse cross-sections (lines 1 to 5), the ablation stakes (dots) and the accumulation sites (squares). Also shown are longitudinal sections (lines A-E) used to calculate thickness and ice velocity variations (see section 4.2). The map (contour lines, glacier delineation) was constructed using a stereoscopic pair of SPOT5 (Systeme Pour l'Observation de la Terre) images acquired 12 and 13 November 2004 and 20 and 21 September 2005 (Wagnon and others, 2007). The map coordinates are in the UTM43 (north) WGS84 reference system.

Figure 2: Cumulative (line) and annual glacier-wide mass balances (histograms) of Chhota Shigri Glacier during 2002/2010.

Figure 3: The 2002/2010 average mass-balance profile and the hypsometry of Chhota Shigri Glacier. Altitudinal ranges are of 50 m (for instance, 4400 stands for the range 4400-4450 m, except for 4250 and 5400 which stand for 4050-4300 m and 5400-6250 m respectively).

Figure 4: Radargram of cross section 2: radar signals plotted side by side from west to east in their true spatial relationship to each other (interval between each signal of 10 m). The x-axis gives the amplitude of each signal (50 mV per graduation); the y-axis is the double-time interval (µs).

Figure 5: Ice depth and surface topography of cross-sections 1-5. The horizontal and vertical scales are the same for all cross-sections. All cross sections are oriented from west to east except cross section 5 which is north-south oriented.

Figure 6: Measured ice velocities plotted as a function of the distance from the 2010 terminus position. Measurements were collected along the central flow line.

Figure 7: Ice fluxes at every cross section derived (i) from 2003/2004 ice velocities and section areas (open squares) and (ii) from mass balance method for a glacier-wide SMB = 0 m w.e. (plain squares) or a glacier-wide SMB = -0.67 m w.e. (triangles). The error range for mass balance fluxes calculated from the mass balance method assuming a steady state (±1 standard deviation: hyphenes) is also given.

Figure 8: Elevation (dots) and surface ice velocity (triangles) between 2003 (continuous lines) and 2010 (dashed lines) along the longitudinal sections A, B, C, D and E shown in Figure 1.

| Cross section | Altitude (m a.s.l.) | Area ($10^4$ $m^2$) | Mean surface ice- velocity from field data (m $yr^{-1}$) (central line velocity*0.79) | Satellite-derived mean surface velocities (m $yr^{-1}$) | Ice flux ($10^6$ $m^3$ $yr^{-1}$) | Max. depth at center of cross section (m) |
|---|---|---|---|---|---|---|
| 1 | 4400 | 4.23±0.68 | 20.3 | NA | 0.78 ±0.21 | 124 |
| 2 | 4650 | 12.14±1.09 | 31.2 | 30.7 | 3.41 ±0.69 | 240 |
| 3 | 4750 | 16.49±1.65 | 29.2 | 29.2 | 4.35 ±0.92 | 245 |
| 4 | 4900 | 15.53±1.55 | 30.1 | NA | 4.20 ±0.89 | 270 |
| 5 | 4850 | 6.01±0.90 | 27.1 | 25.5 | 1.47 ±0.38 | 175 |

Table 1

| **Cross section** | **Altitude (m a.s.l)** | **Hydrological years (October – September)** | | | | | | | | **Mean** |
|---|---|---|---|---|---|---|---|---|---|---|
| | | **2002/03** | **2003/04** | **2004/05** | **2005/06** | **2006/07** | **2007/08** | **2008/09** | **2009/10** | **2002/10** |
| Snout | 4050 | -22.26 | -19.28 | 2.27 | -22.21 | -15.59 | -14.65 | 2.06 | 5.24 | -10.55 |
| 1 | 4400 | -22.83 | -19.55 | 3.78 | -22.65 | -15.43 | -14.82 | 3.19 | 6.84 | -10.18 |
| 2 | 4670 | -14.31 | -11.49 | 6.57 | -14.35 | -8.63 | -8.21 | 6.03 | 8.70 | -4.46 |
| 3 | 4735 | -9.88 | -7.68 | 6.89 | -10.04 | -5.45 | -5.46 | 6.16 | 8.43 | -2.13 |
| 4 | 4900 | -1.41 | -1.36 | 4.84 | -2.30 | 0.05 | -0.16 | 4.14 | 5.72 | 1.19 |
| 5 | 4870 | -1.61 | -1.30 | 2.08 | -2.19 | -0.57 | -0.82 | 1.86 | 2.80 | 0.03 |

Table 2

| **Cross section** | **Altitude (m a.s.l)** | **Hydrological years (October – September)** | | | | | | | | **Mean** | **STD*** |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | **2002/03** | **2003/04** | **2004/05** | **2005/06** | **2006/07** | **2007/08** | **2008/09** | **2009/10** | **2002/10** | |
| Snout | 4050 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0 |
| 1 | 4400 | 1.25 | 1.32 | 1.33 | 1.38 | 1.43 | 1.03 | 0.96 | 1.17 | 1.23 | 0.17 |
| 2 | 4670 | 5.07 | 5.30 | 4.59 | 4.99 | 4.94 | 4.55 | 4.24 | 4.14 | 4.73 | 0.41 |
| 3 | 4735 | 6.04 | 6.11 | 5.27 | 5.85 | 5.70 | 5.02 | 4.69 | 4.68 | 5.42 | 0.58 |
| 4 | 4900 | 5.78 | 4.87 | 4.11 | 4.88 | 5.09 | 4.58 | 3.48 | 4.02 | 4.60 | 0.72 |
| 5 | 4870 | 2.76 | 2.48 | 1.63 | 2.17 | 2.49 | 2.06 | 1.46 | 1.77 | 2.10 | 0.46 |

*STD = standard deviation.

Table 3

| Longitudinal section | Elevation change (m) | Velocity change (m yr$^{-1}$) |
|---|---|---|
| A | -5.3 | -6.6 |
| B | -8.6 | -8.8 |
| C | -7.5 | -7.4 |
| D | -2.8 | NA |
| E | -5.6 | +4.8 |

Table 4

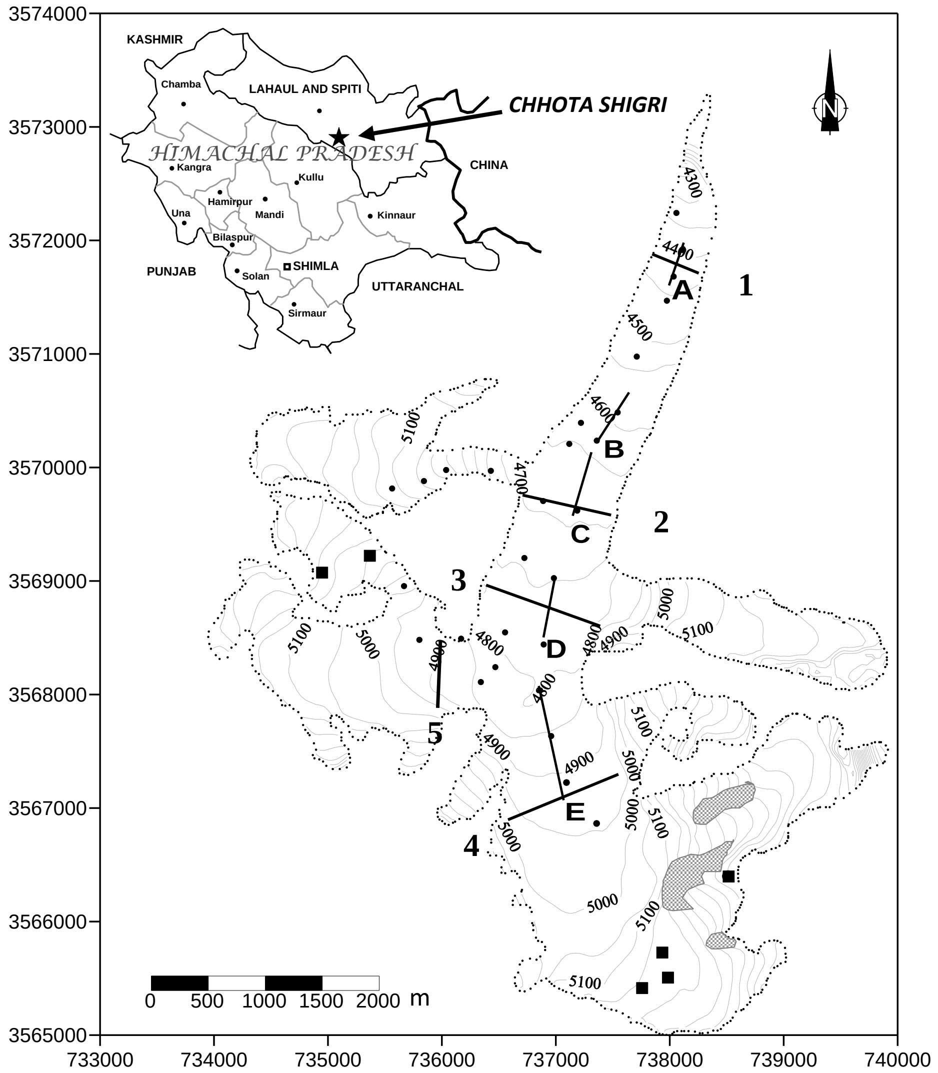

Figure 1

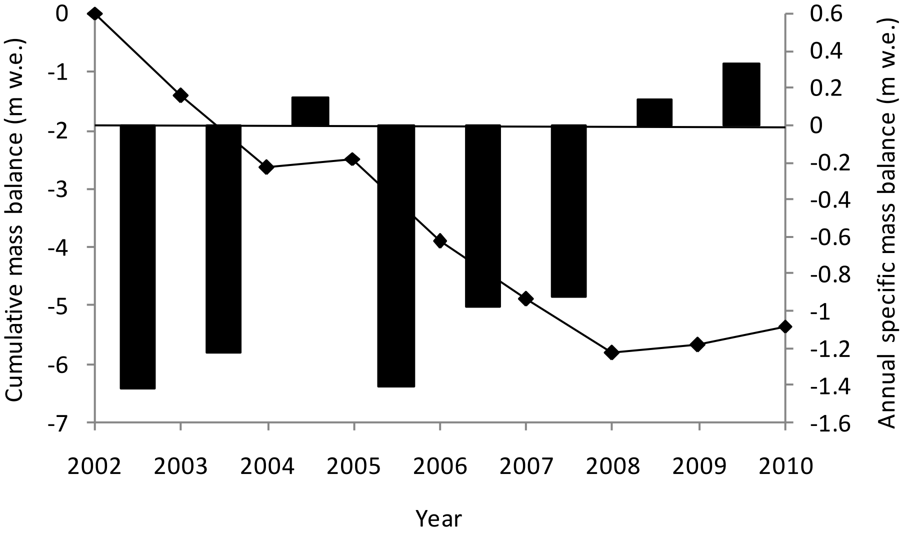

Figure 2

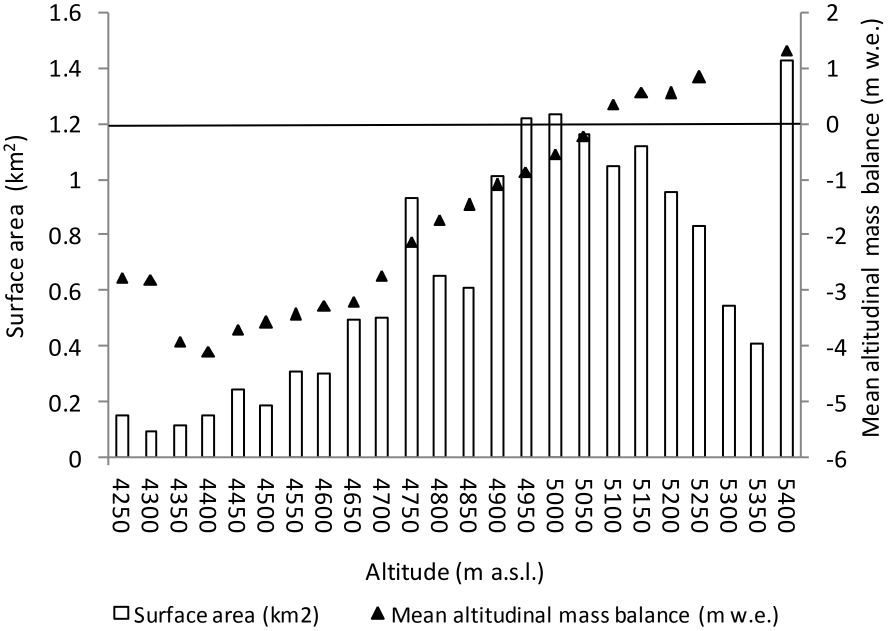

Figure 3

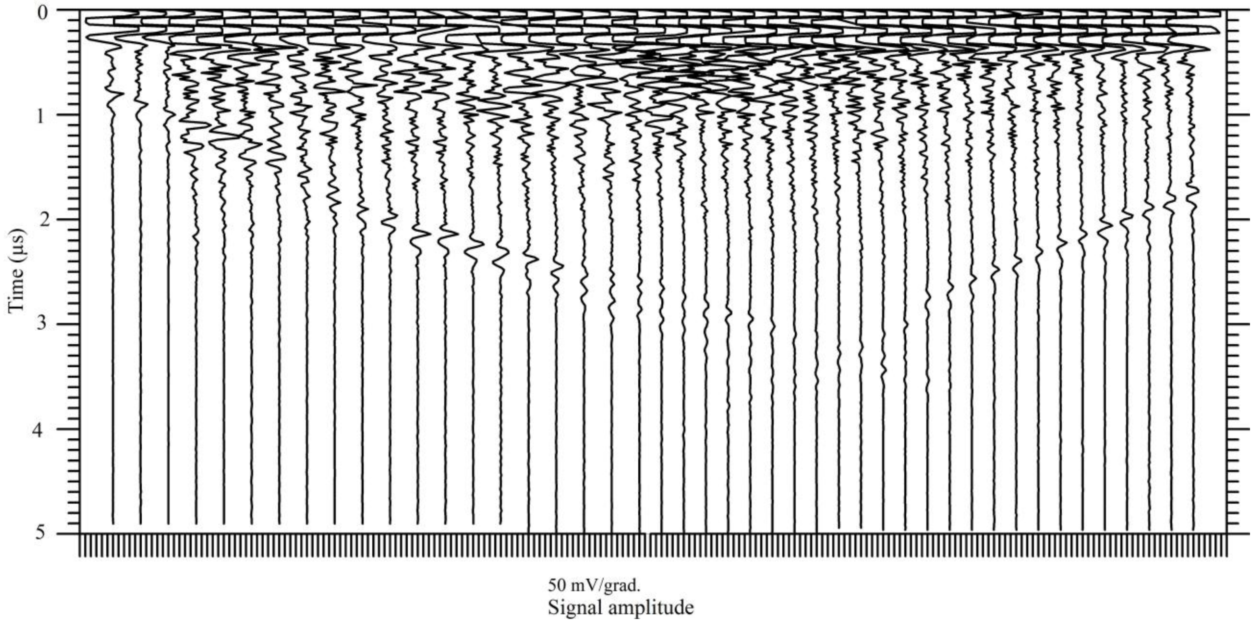

Figure 4

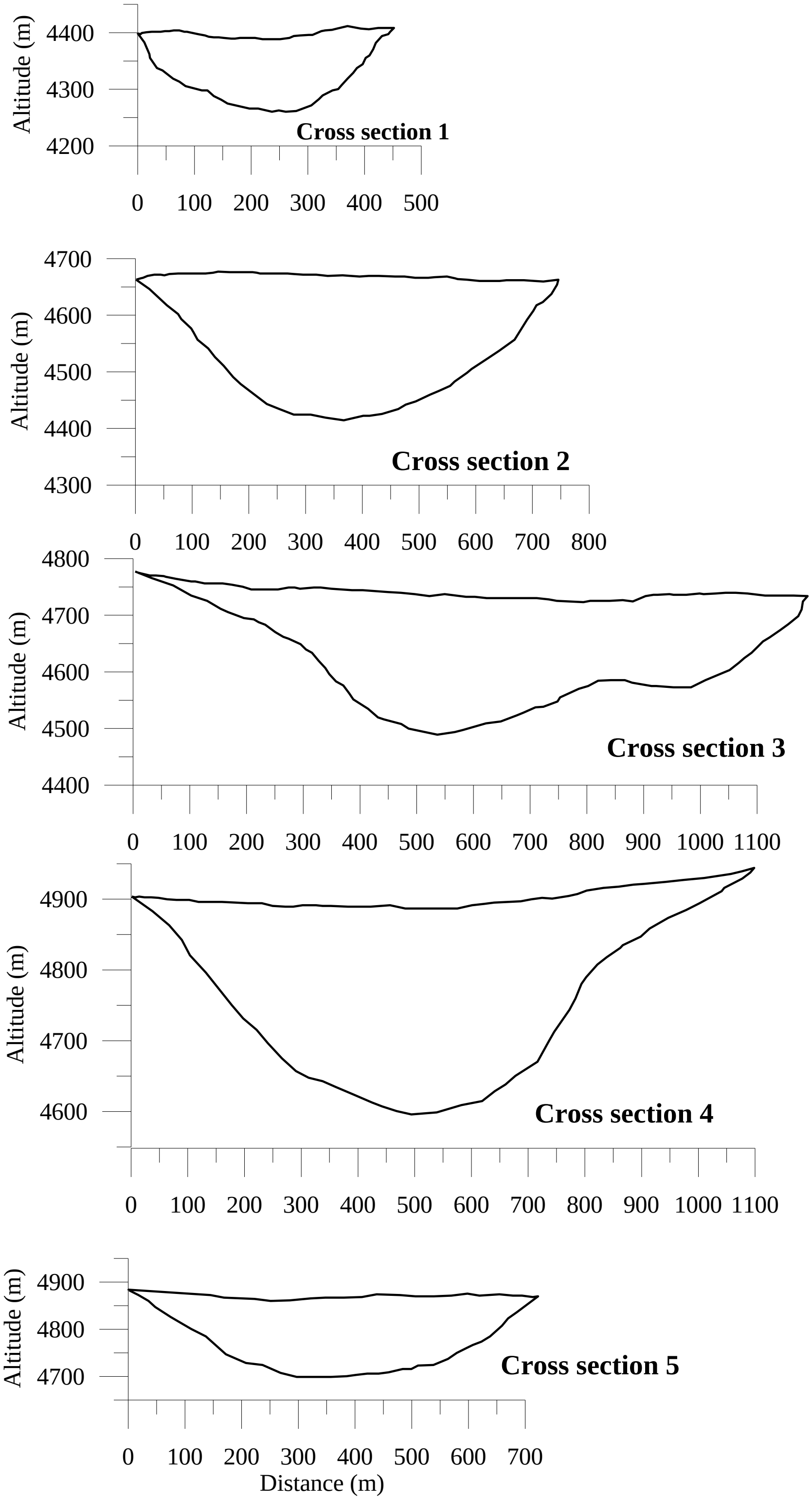
Altitude (m)
4400
4300
4200
Cross section 1
0 100 200 300 400 500
Altitude (m)
4700
4600
4500
4400
4300
Cross section 2
0 100 200 300 400 500 600 700 800
Altitude (m)
4800
4700
4600
4500
4400
Cross section 3
0 100 200 300 400 500 600 700 800 900 1000 1100
Altitude (m)
4900
4800
4700
4600
Cross section 4
0 100 200 300 400 500 600 700 800 900 1000 1100
Altitude (m)
4900
4800
4700
Cross section 5
0 100 200 300 400 500 600 700
Distance (m)

Figure 5

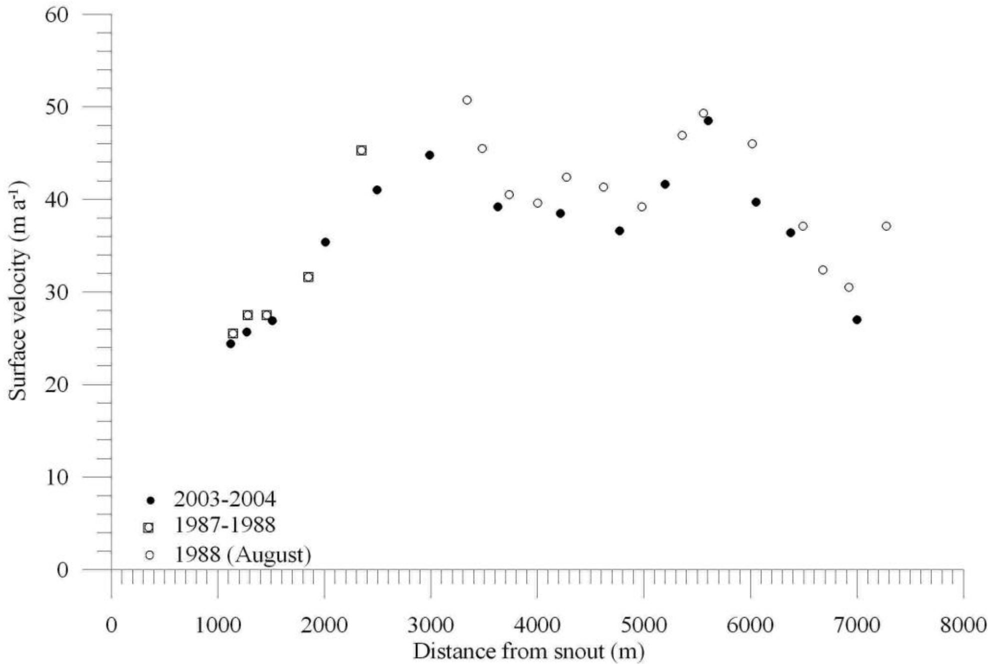

Figure 6

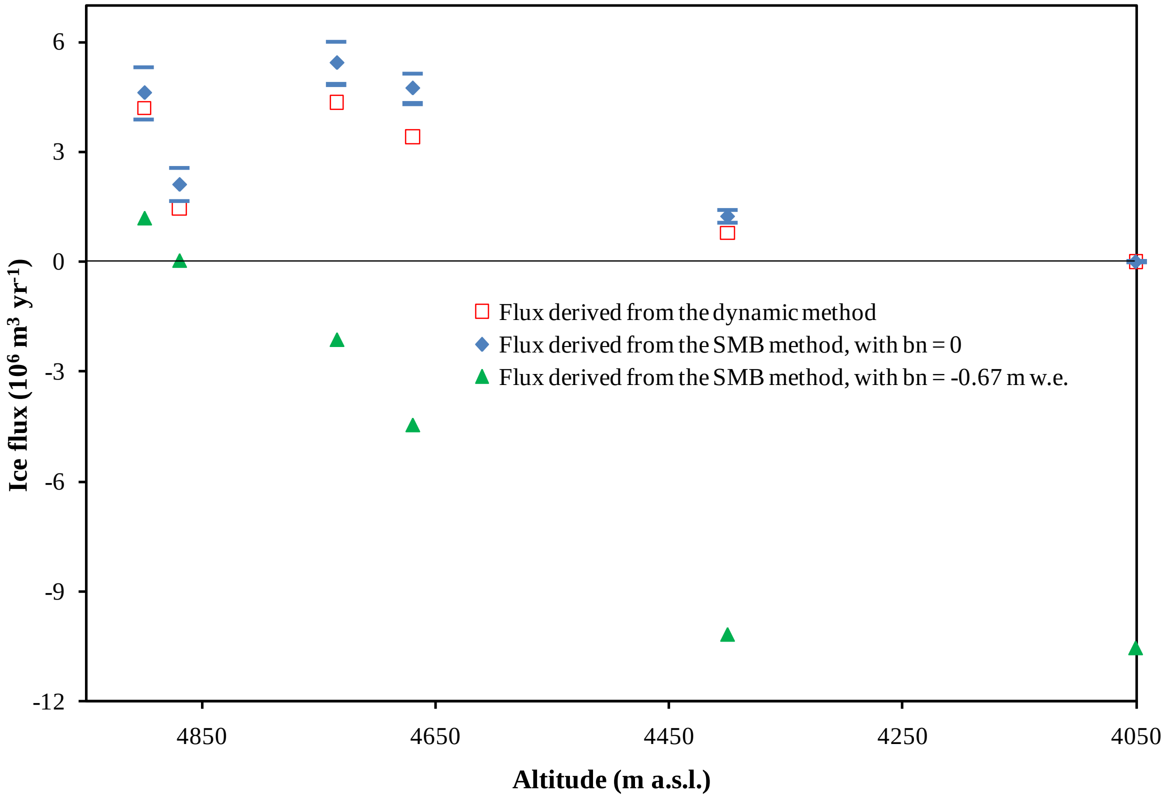

Figure 7

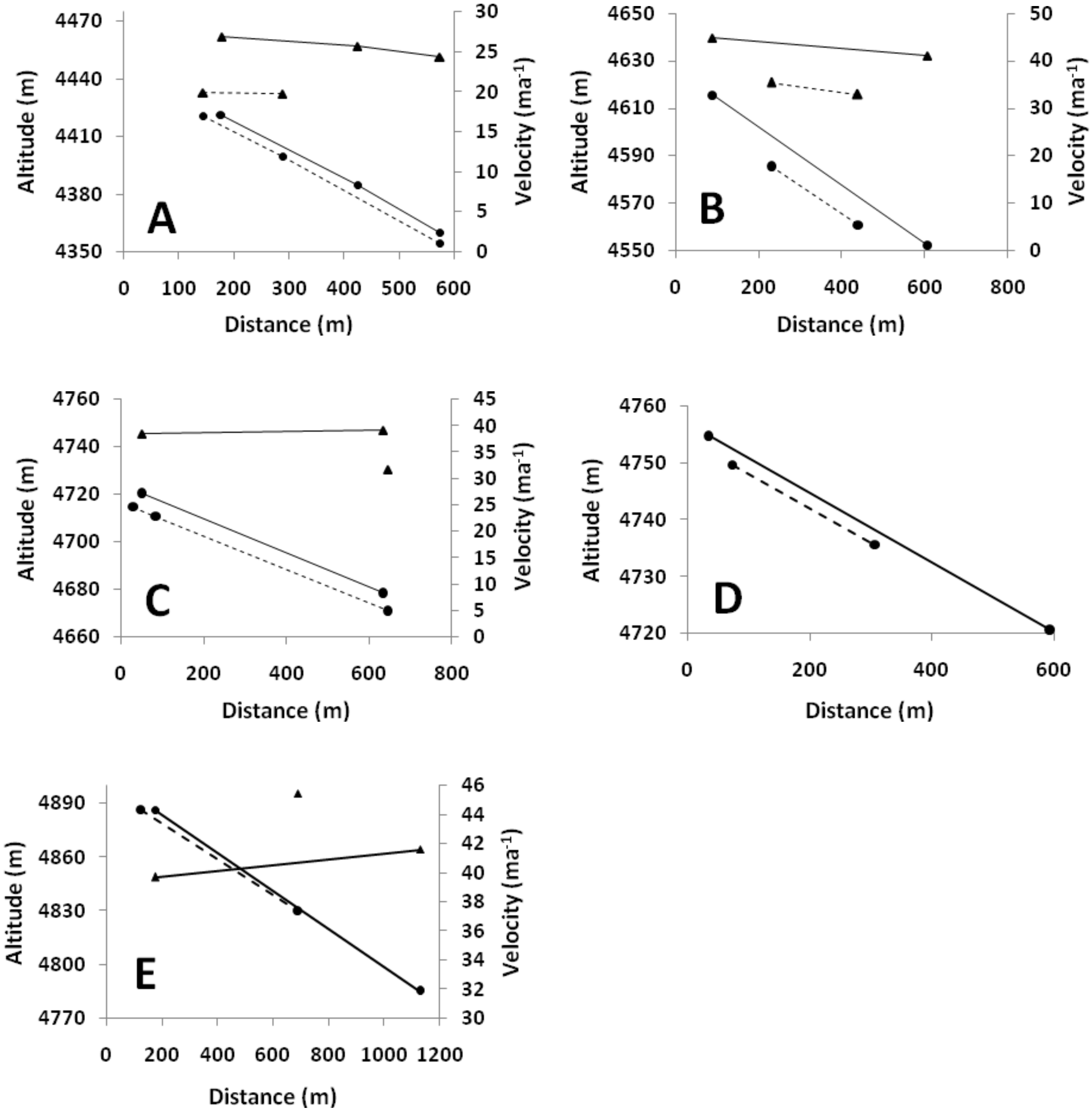
A
Altitude (m)
Velocity (ma$^{-1}$)
Distance (m)
4470
4440
4410
4380
4350
30
25
20
15
10
5
0
0
100
200
300
400
500
600
B
Altitude (m)
Velocity (ma$^{-1}$)
Distance (m)
4650
4630
4610
4590
4570
4550
50
40
30
20
10
0
0
200
400
600
800
C
Altitude (m)
Velocity (ma$^{-1}$)
Distance (m)
4760
4740
4720
4700
4680
4660
45
40
35
30
25
20
15
10
5
0
0
200
400
600
800
D
Altitude (m)
Distance (m)
4760
4750
4740
4730
4720
0
200
400
600
E
Altitude (m)
Velocity (ma$^{-1}$)
Distance (m)
4890
4860
4830
4800
4770
46
44
42
40
38
36
34
32
30
0
200
400
600
800
1000
1200

Figure 8