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Glacier surface mass balance modeling in the inner tropics using a positive degree-day approach

L. Maisincho[1,2], V. Favier[3], P. Wagnon[2,4], V. Jomelli[5], R. Basantes Serrano[2,3,8], B. Francou[2], M. Villacis[6], A. Rabatel[3], M. Ménégoz[7], L. Mourre[2], B. Cáceres[1]

[1] INAMHI, Instituto Nacional de Meteorología e Hidrología, Iñaquito N36-14 y Corea, Quito, Ecuador.

[2] IRD, Univ. Grenoble Alpes, CNRS, INPG, Laboratoire d'étude des Transferts en Hydrologie et Environnement (LTHE), UMR5564, F-38000 Grenoble, France.

[3] Univ. Grenoble Alpes, CNRS, Laboratoire de Glaciologie et Géophysique de l'Environnement (LGGE), UMR5183, F-38000 Grenoble, France.

[4] ICIMOD, GPO Box 3226, Kathmandu, Nepal

[5] Université Paris 1 Panthéon-Sorbonne, CNRS, Laboratoire de Géographie Physique (LGP), 92195 Meudon, France.

[6] Escuela Politécnica Nacional (EPN), Ladrón de Guevara E11-253, Quito, Ecuador.

[7] Institut Català de Ciènces del Clima (IC3), Doctor Trueta, 203 3a planta 08005 Barcelona (Spain)

[8] Universidad Regional Amazónica (IKIAM), Km 7, Via Muyuna-Atacapi, Tena, Ecuador.

Correspondance to: Luis Maisincho, INAMHI, Iñaquito N36-14 y Corea, Quito, Ecuador

lmaisincho@inamhi.gob.ec

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Abstract

We present a basic ablation model combining a positive degree-day approach to calculate melting and a simple equation based on wind speed to compute sublimation. The model was calibrated at point scale (4,900 m a.s.l.) on Antizana Glacier 15 (0.28 km^2; 0°28’S, 78°09’W) with data from March 2002 to August 2003 and validated with data from January to November 2005. Cross validation was performed by interchanging the calibration and validation periods. Optimization of the model based on the calculated surface energy balance allowed degree-day factors to be retrieved for snow and ice, and suggests that melting started when daily air temperature was still below 0 °C, because incoming shortwave radiation was intense around noon and resulted in positive temperatures for a few hours a day. The model was then distributed over the glacier and applied to the 2000-2008 period using meteorological inputs measured on the glacier foreland to assess to what extent this approach is suitable for quantifying glacier surface mass balance in Ecuador. Results showed that a model based on temperature, wind speed, and precipitation is able to reproduce a large part of surface mass-balance variability of Antizana Glacier 15 even though the melting factors for snow and ice may vary with time. The model performed well because temperatures were significantly correlated with albedo and net shortwave radiation. Because this relationship disappeared when strong winds result in mixed air in the surface boundary layer, this model should not be extrapolated to other tropical regions where sublimation increases during a pronounced dry season or where glaciers are located above the mean freezing level.

Keywords: Degree-day, melting, surface mass balance, inner tropics, Antizana.

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1 Introduction

Glaciers in Ecuador respond rapidly to climate change, particularly to variations in temperature. The comparison of glacier extents using photogrammetric information available since 1956 (Francou *et al*., 2000) with local variations in temperature suggests that local warming of the atmosphere (about 0.2 °C/decade (Vuille *et al*., 2000)) has played an important role in glacier retreat since the 1950s (Francou *et al*., 2000), with direct consequences for the local water supply to Quito (e.g., Favier *et al*., 2008; Villacis, 2008). The expected warming in the high-elevation Andes over the 21st century (between 4 °C and 5 °C) (Bradley *et al*., 2006; Vuille *et al*., 2008; Urrutia and Vuille, 2009), which is more than estimated warming since the early Holocene (Jomelli *et al*., 2011), could thus have dramatic consequences for glacial retreat in Ecuador. Understanding and producing long-term models of glacial retreat under local warming is thus crucial.

Surface mass balance models for the tropics using minimum inputs have already been built and applied in the outer tropics (Kaser, 2001; Juen *et al*., 2007) but never specifically in the Ecuadorian Andes. To date, only one attempt has been made to link the various energy fluxes to two input variables, monthly precipitation and temperature (Juen *et al*., 2007). Using a similar approach in the inner tropics makes sense because solid precipitation and temperature changes have already been demonstrated to play an important role in the interannual variability of ablation (Francou *et al*., 2004). Even though the interest of the positive degree-day (PDD) model is quite controversial in the tropics, where temperature is generally assumed to have a limited link with the main local ablation processes (Sicart *et al*., 2008), a comprehensive test of such a model has still not been performed in the inner tropics and is timely. Indeed, in the Ecuadorian Andes, air

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temperature is known to be the main variable involved in glacier surface mass balance as it controls the 0 °C level, which oscillates continuously within the ablation zone (Kaser, 2001; Favier *et al*., 2004a&b, Francou *et al*., 2004; Rabatel *et al*., 2013). Thus, slight changes in temperature directly modify the ablation processes at the glacier surface due to the precipitation phase and its impact on surface albedo (e.g., Francou *et al*., 2004; Favier *et al.*, 2004a). As a consequence, during El Niño/La Niña events, atmospheric warming/cooling, or more precisely the rise/drop in the 0 °C level, has major consequences for the precipitation phase over the glacier, leading to high/low melting rates (e.g., Francou *et al.*, 2004; Favier *et al.*, 2004b).

In this study, we developed a basic model based on variations in temperature, precipitation, and wind speed to study the glacier surface mass balance on Antizana Glacier 15 (0.28 km^2; 0^{o}28'S, 78^{o}09'W). Melting was assessed using a typical positive degree-day approach (e.g., Braithwaite, 1995; Hock, 2003), whereas sublimation was calculated using only daily wind speeds. The model was calibrated and tested on data from Antizana Glacier 15 (see GLACIOCLIM observatory: http://www-lgge.ujf-grenoble.fr/ServiceObs/) to judge whether such a simple approach can reasonably quantify local glacier surface mass balance and the transient snowline elevation in Ecuador.

2 Study site, climatic setting and associated glaciological processes

Antizana stratovolcano is one of the main ice covered summits in the Cordillera Oriental of Ecuador (Figure 1). The most recent glacier inventory performed in 2014 showed that glaciers extended over a total surface area of 15 km^2 (Basantes, 2015) distributed in 17 glacier tongues (Hastenrath, 1981). Glaciological and hydrological studies in the area began on Antizana Glacier 15 in 1994. The glacier is located on the north-western

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side of the volcano and is a reference site for long-term observations. The surface of the glacier presently extends from 5,700 m above sea level (a.s.l.) down to 4,850 m a.s.l.

The study area belongs to the inner tropics, which are characterized by very low temperature and moisture seasonality (e.g., Kaser and Osmaston, 2002). The low latitude location yields circadian temperature variations largely exceeding those of the daily mean temperature over one year. Between 2000 and 2008, the annual precipitation recorded at 4,550 m a.s.l. in the catchment of Antizana Glacier 15 ranged from 800 to 1,300 mm a^{-1}. Precipitation was significant every month; monthly variations produced two slight maxima in April and October, and slight minima in July-August and December (Favier *et al*., 2004a). As a consequence of these peculiar climatic settings, accumulation and ablation occur simultaneously and continuously. The mean 0 °C level is generally close to 5,000 m a.s.l., i.e. within the ablation zone. However, this value is subject to year-to-year variability. On the other hand, during the period 2000-2008, wind velocity was subject to pronounced seasonal variations, with intense easterly winds generally occurring between June and October (hereafter referred to as Period 1). Period 2 refers to the period from November to May of the following year, which was associated with marked mass and energy losses through melting (Favier *et al*., 2004a). For instance, between 2000 and 2008, the mean wind speed was 6.1 m s^{-1} (standard deviation of daily values (STD) = 2.9 m s^{-1}) in Period 1 and 3.1 m s^{-1} (STD = 2.1 m s^{-1}) in Period 2.

Finally, most of the local climate variability since the 1970s has been closely linked to the El Niño–Southern Oscillation (ENSO) (Francou *et al*., 2004; Vuille *et al*., 2008). There is a three month delay in the local response of the atmosphere to the ENSO signal. Surface energy balance studies showed that these variations are closely linked with variations in albedo that mirror changes in the precipitation phase at the glacier surface due

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to variations in temperature.

3 Data

3.1 Basic model input data

3.1.1 Daily temperature and precipitation

We used data recorded at five meteorological stations and two tipping bucket rain gauges (Table 1 and Figure 1). We were able to obtain a continuous homogeneous temperature dataset at daily time scale from the temperature sensor of the glacier station located at 4,900 m a.s.l. (hereafter referred to as AWS_{G1}, see Table 1). We filled the data gaps (Table 2) by applying simple correlations between the daily temperature recorded at 4,900 m a.s.l. and at various neighboring stations when the stations were working simultaneously (r^2 always higher than 0.75, Table 3). The stations used to fill the gaps were in order of descending priority, first AWS_{M1} (installed on the lateral moraine of Glacier 15 at 4,900 m a.s.l.), second AWS_{G2} (installed on Glacier 15, at 5,000 m a.s.l.), when AWS_{M1} was not working, third AWS_{G3} (installed on a nearby glacier, Glacier 12 at 4,900 m a.s.l.) and finally AWS_{M2} (installed off-glacier at 4,785 m a.s.l.). Figure 1 and Tables 1 and 2 show the location and provide additional information for each station and explain how the continuous dataset from 2000 to 2008 was obtained. For AWS_{M2} and AWS_{G2} data, lapse rate corrections allowed us to account for the difference in elevation between the two stations and AWS_{G1}. The quality of the temperature data was checked during regular field visits conducted approximately every 10 days to detect any AWS malfunction (i.e. failure of artificial ventilation) and by comparing with data from the closest sensors. When data were considered to be suspicious (1.2% of a total of 3,288 days) they were not used in the present

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study.

Precipitation between 2000 and 2008 came from an automatic tipping bucket HOBO rain gauge referred to as P4 (Figure 1) located on the moorland (*páramo*) at 4,550 m a.s.l. Data from P4 were quality controlled and validated with monthly total precipitation measurements at 4,550 m a.s.l. in the field using a totalizer rain gauge. When daily precipitation was not available at P4 (6% of the total period), we used data from a similar rain gauge hereafter referred to as P2 (Figure 1), located at 4,875 m a.s.l. The determination coefficient of daily precipitation amounts between P2 and P4 was significant ($r^2 = 0.60$, $n = 2{,}378$ days, $p = 0.001$, between 2000 and 2008) even if snow precipitation occurred more frequently at P2 than at P4, and snow melt in the rain gauge was delayed from a few hours to maximum one day because the sensors were not artificially heated.

It is well known that precipitation measurements are subject to large systematic errors, especially when a large proportion of precipitation falls in the form of snow in a windy environment and undercatch prevails (e.g. Immerzeel *et al.*, 2012). This is the case on Antizana Glacier 15. Consequently, based on a detailed analysis of measurements made with a reference gauge suitable for measuring both solid and liquid precipitation (Geonor T-200b, equipped with a weighing device), Wagnon *et al*. (2009) recommended applying a correction factor of +51% to account for this undercatch. Given that P4 systematically collected 16.5% less precipitation than the Geonor rain gauge between 2005 and 2012 (data not shown), the correction factor to apply to P4 measurements was as high as +76% (1.76 = 1.165 * 1.51) Here, we applied this correction to the precipitation at P4 between 2000 and 2008, leading to a mean precipitation of 1,820 mm w.e. a^{-1} at 4,550 m a.s.l. We assumed that precipitation did not vary with elevation due to the small size of the glacier (only 2 km long) (Favier *et al.*, 2008). Nevertheless, we did test the impact of correcting precipitation on calculations of the

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glacier-wide climatic (or surface) mass balance (Section 6.3).

3.1.2 Wind speed used to compute sublimation

Turbulent heat fluxes are known to be very sensitive to wind velocity (Garratt, 1992). Since on Antizana Glacier 15, the latent heat (*LE)* and wind speed are indeed closely correlated at a daily time scale (r = -0.79, n = 530, p = 0.001, see Supplementary Materials, Table S2), like in Favier *et al*. (2008), sublimation amounts for 2000-2008 were computed using surface wind speed recorded at the same stations as temperature (see Section 3.1). When wind speeds were not available (18% of a total of 3,288 days), we used daily wind speed at 600 hPa available from the NCEP-NCAR Reanalysis1 (NCEP1) dataset closest to Antizana volcano (77º W; 0.2º S) (Kalnay *et al.*, 1996) (r = 0.7 for n = 2,685 days with data). A comparison with field data at 4,900 m a.s.l. showed that the reanalyzed wind speed presented a mean bias of 0.3 m s^{-1}, which we assumed to be negligible in our surface mass balance computation.

3.2 Data used for model calibration and validation

3.2.1 Data used for modeling the surface energy balance

The surface energy balance (SEB) was computed at 4,900 m a.s.l. for 530 days between March 14, 2002 and August 31, 2003. Continuous data were available at the AWS_{G1}, except between May 2 and May 6, 2002. The sensors installed on the AWS_{G1} and the available data are described in Favier *et al*. (2004a&b). A second data set was used to compute the surface mass balance from January 1, 2005 to November 30, 2005. Except for incoming long-wave radiation, which was available only at AWS_{M1}, all the meteorological variables came from AWS_{G1}. These results allowed us to calculate daily ablation, which was then used as reference data to calibrate and validate the basic model. The basic model was first calibrated using data from the period March 2002 to August 2003 and validated using data from January to November 2005. Cross validation was then performed by interchanging the calibration and validation periods.

Finally, the albedo measurements collected at AWS_{G1} and AWS_{G3} were used to get information on surface state between 2000 and 2008 (Table 1&4). The albedo data used in 2006 were collected at 4,900 m a.s.l. on Antizana Glacier 12 located on the south-western flank of the volcano, 3 km from Antizana Glacier 15. The slope and the aspect of the two glaciers are similar and albedo was measured at the same elevation. The sensors used for radiation measurements on the latter AWS were the same as those installed on AWS_{G1}.

3.2.2 Glaciological data used for model validation

We used the following data for model validation (Table 4):

1) Daily melting amounts for 43 days in 2002-2003 were obtained using "melting boxes" similar to those described in Wagnon *et al*. (1999). The values used in this study are those of Favier *et al.* (2004a). Melting box accuracy is hard to assess, but the comparison of melting amounts from melting boxes and from surface energy balance data (see Section 4.2) suggests that measured melting is generally lower, likely because initial liquid water is retained by/between the small ice blocks due to capillarity. The uncertainty of daily melting measured by melting boxes cannot be assessed with accuracy but Favier *et al.* (2004a) observed a 30% difference between measured and computed melting, suggesting the error range is likely in this order of magnitude. Finally, based on the application of the surface energy balance model (Section 4.2), we calculated that melting occurring below 20 cm under the surface represented 1.6 % of

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total melt and was thus negligible.

2) The monthly mass balance and snow accumulation at 4,900 m a.s.l. from 2000 to 2008. These data allowed us to assess the thickness of the snow cover and the changes in the level of the surface due to ablation. Data from 2000 to 2003 were already used and described by Francou *et al.* (2004).

3) The annual Antizana Glacier 15 climatic mass-balance profiles (hereafter referred to as b(z)) obtained in the ablation zone from field measurements and the ELA from 2000 to 2008 (available on the GLACIOCLIM and WGMS databases). Above 5000 m a.s.l., Basantes Serrano et al. (2016) have shown that using Lliboutry's approach to interpolate the measured data is more accurate by significantly reducing the discrepancy between the glaciological and geodetic balance. All details regarding the glaciological measurements and methods are described in Francou *et al*. (2004) and Basantes Serrano et al. (2016).

4) The Antizana Glacier 15 glacier-wide climatic annual mass balances (B_a) from 2000 to 2008. Here we present data from Basantes Serrano *et al.* (2016), in which the glacier-wide annual mass balance of Antizana Glacier 15 computed using the glaciological method was recalculated using an updated delineation of the glacier and adjusted with the geodetic method based on photogrammetric restitution of aerial photographs taken in 1997 and 2009.

5) The annually updated hypsometry and glacier surface area of Antizana Glacier 15 computed by Basantes Serrano *et al.,* (2016). The glacier-wide annual mass balance calculated from the basic model accounted for this revised hypsometry and area.

6) Intermittent observations and terrestrial photographs (Table 4) of the glacier surface that were made to estimate the elevation of the transient snowline on the glacier during field

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trips between 2004 and 2008. The daily transient snowline elevation was estimated from photographs obtained with a low resolution automatic camera (Fujifilm FinePix 1400) installed on the frontal moraine at 4,785 m a.s.l. These photographs were taken from the location labelled "Photo" in Figure 1, and were georeferenced (Corripio, 2004). A total of 712 good quality daily photographs allowed us to almost continuously monitor the transient snowline elevation over time with an accuracy of ±10 m.

4 Methods

4.1 Statistical test to assess model performance

To test model performance, we used the efficiency statistical test (E) proposed by Nash and Sutcliffe (1970):

$$E = 1 - (RMSE/s)^2 \tag{1}$$

where s is the standard deviation of the observations and RMSE is the root mean squared error of the simulated variable (perfect agreement for E=1). The correlation between measurements and model was also analyzed. Except when the p value is mentioned, correlations are considered as significant if p is 0.001 or less.

4.2 Surface energy balance computation

At point scale (4,900 m a.s.l.), daily melting was calculated for a horizontal surface by applying the classical SEB approach described in Favier *et al.* (2011), which includes subglacial processes (see Supplementary Materials for details) not originally accounted for in Favier *et al.* (2004a&b). Daily melting was quantified from March 14, 2003 to August 31, 2003 and from January 1, 2005 to November 30, 2005. Calculations were validated using melting amounts measured with the melting boxes and with point mass balance measured on

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stakes in the vicinity of the AWS_{G1} (see Supplementary Materials, Figure S2). Results using Favier *et al.* (2011) approach agreed with measured melting amounts better than results in Favier *et al.* (2004a&b), with a correlation coefficient of r = 0.91 (instead of 0.86). The regression line is also closer to the 1:1 line (slope of 1.01 instead of 0.89). Over one year, the heat storage below the surface is zero, and the energy excess at the surface (i.e. $Q_{surface}$) is used to melt the snow/ice at the surface or below. As a consequence, the mean annual computed melting rates only differed by 0.4% (see Supplementary Materials, Figure S1) from those given by Favier *et al*. (2004a&b), suggesting that heat conduction (G) into the ice/snow over one year can be disregarded. However, daily differences between the results of the present study and the calculations in Favier *et al*. (2004a&b) are significant (reaching 20 mm w.e d^{-1} with a standard deviation of 5 mm w.e. d^{-1}) because refreezing may occur in particular when sublimation is high (in Period 1) demonstrating that the use of a computation scheme including G and solar radiation penetration into the ice is necessary to study ablation processes at a daily timescale (e.g., Mölg *et al*., 2008, 2009).

4.3 The basic model

4.3.1 The positive degree-day model

The positive degree-day model enables calculation of daily snow or ice melt $m_j(z)$ (in mm w.e.) at a given elevation z (in m a.s.l.), and at time step j (in days) (Braithwaite, 1995; Hock, 2003):

$$m_j(z) = F\,(T_j(z_{ref}) + LR\,(z - z_{ref}) - T_{threshold}) \quad \text{if} \quad T_j(z_{ref}) + LR\,(z - z_{ref}) > T_{threshold} \quad (2),$$

$$m_j(z) = 0 \quad \text{if} \quad T_j(z_{ref}) + LR\,(z - z_{ref}) \leq T_{threshold} \quad (3),$$

where F is the degree-day factor (in mm w.e. °C^{-1} d^{-1}), T_j(z) (in °C) is the mean daily temperature, z_{ref} = 4,900 m a.s.l. and z (in m a.s.l.) is the reference elevation and the given elevation respectively, $T_{threshold}$ (in °C) is a threshold temperature above which melting begins, and LR is the lapse rate in the atmosphere (in °C m^{-1}, hereafter expressed in °C km^{-1} for better readability). The PDD model generally assumes that $T_{threshold}$ = 0 °C (van den Broeke *et al.*, 2010). However, during short periods in the daytime, melting may occur when daily mean is below 0 °C (e.g. Van den Broeke *et al.*, 2010). Here, we used $T_{threshold}$ as a calibration parameter of the model (See section 5.1).

The model can be run using different F values depending on the presence or absence of snow at the glacier surface at the previous time step, where $S_{j-1}(z)$ is the amount of snow in mm w.e. at the time step j-1:

$$F = F_{snow} \quad \text{if} \quad S_{j-1}(z) > 0 \qquad \text{(in mm w.e. °C}^{-1}\text{ d}^{-1}\text{)} \quad (4).$$

$$F = F_{ice} \quad \text{if} \quad S_{j-1}(z) = 0 \qquad \text{(in mm w.e. °C}^{-1}\text{ d}^{-1}\text{)} \quad (5).$$

Snow cover is the difference between ablation and snow accumulation at a given elevation z. In ablation computations, sublimation was assessed using a simple relationship based on the regression line between wind speed and sublimation (see Equation 6, Section 4.3.2) like in Favier *et al*. (2008). Solid precipitation is assumed if the air temperature is below a threshold ($T_{snow/rain}$ = 1 °C (Wagnon *et al.*, 2009)), otherwise solid precipitation is zero. This threshold was obtained from field measurements and from direct observations of the precipitation phase in the Andes, which showed that, below this temperature, more than 70% of precipitation is solid (e.g., L'hôte *et al.*, 2005). Temperature at a specific elevation z was computed assuming a constant lapse rate (LR) between the reference elevation z_{ref}, where meteorological data are available, and z. Half-hourly field temperature measurements performed in artificially ventilated shelters at three different elevations on Antizana Glacier 12 (3 km south of the Antizana Glacier 15) suggested a mean LR of -8.5 °C km^{-1} (standard

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deviation of 3.0 °C km^{-1} on half hourly values, for 18,685 values) (data not shown). This vertical temperature gradient is steeper than the moist adiabatic gradient because Antizana Glacier 12 and Antizana Glacier 15 are located on the leeward side of the volcano, where there is a strong foehn effect whose consequence is to steepen *LR* (e.g., Favier *et al.*, 2004a). The *LR* values may present a seasonal cycle, which can strongly impact the modeled glacier-wide mass balance. Over one year (July 2012-July 2013), this gradient was steeper in July-August (around -9.2 °C km^{-1}) when the wind was stronger than in the rest of the year (-8.2 °C km^{-1}). We used -8.5 °C km^{-1} in the present paper, and a model sensitivity test against this parameter is presented in section 6 to quantify to what extent our results depend on *LR* seasonality.

The elevation of the transient snowline $z_{SL,j}$, i.e. the elevation above which daily snow accumulation was positive, was an output of the model and was then compared with field observations. Finally, the modeled ELA is the altitude at which the annual surface mass balance $b_j(z)$ is zero.

4.3.2 Incorporating sublimation in the basic model

In the tropics, sublimation is known to be an important ablation process (Winkler *et al.*, 2009) which is worth including in a basic model. The regression line between the daily wind speed and turbulent latent heat flux (see Supplementary Materials) provides the equation needed to compute daily sublimation:

$$Sublimation = LE*24*3600 / L_s = -5.73\ u \quad (6)$$

where u is daily mean wind speed (in m.s^{-1}) and L_s is latent heat of sublimation (L_s = 2.834 10^6 J kg^{-1}).

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Between March 2002 and August 2003, sublimation represented 3.7% of the total ablation at 4,900 m a.s.l. on Antizana Glacier 15 (Favier *et al*., 2004a). This rate may increase with elevation as melting amounts decrease and wind speed increases. However, because sublimation decreases with a drop in air temperature (e.g., Bergeron *et al*., 2006), sublimation is still limited at high elevations due to colder temperatures. Moreover, the frequent presence of lenticular clouds on the summit of Antizana Glacier 15 suggests that water condensation or re-sublimation takes place at the summit (as confirmed by frequent frost deposition) and sublimation at the glacier snout likely results from the notable effect of the foehn (Favier *et al*., 2004a). The sublimation gradient is thus unclear. Because the mean sublimation at 4,900 m a.s.l. was of the same magnitude (-300 mm w.e. a^{-1}) as the mean 2000-2008 glacier-wide mass balance (-240 mm w.e. a^{-1}), i.e. 15% of the glacier-wide ablation (2060 mm w.e. a^{-1}), assuming that sublimation is constant or decreases rapidly with elevation has important consequences for the final modeled glacier-wide surface mass balance. Since the gradient of sublimation as a function of altitude is not yet available from SEB modelling, we developed rough hypotheses for its distribution with elevation. In this study, the sublimation gradient was assumed to be equal to 0 (constant sublimation), but a sensitivity test was performed (Section 6.3) using a linear decrease until zero was reached at the summit where frost may result in insignificant sublimation.

5 Model calibration and validation at point scale

5.1 Calibration over the 2002-2003 period

When running the model at point scale (4,900 m a.s.l., i.e. the elevation of the input data), *LR* can be discarded and only three parameters F_{snow}, F_{ice} and $T_{threshold}$ need to be calibrated. $T_{threshold}$ was first obtained as the zero melting intercept given by the regression line between daily temperature and daily melting from March 14, 2002 to August 31, 2003 ($T_{threshold}$ = -2.05°C). The basic model was trained over 2002-2003 period using ablation computed from the SEB approach (hereafter referred to as 'SEB ablation'). We distinguished days with snow at the surface from days without (bare ice) using a separation according to a threshold ($a_{threshold}$) applied on measured surface albedo. Optimization of this threshold allowed us to calibrate the F_{snow} value only in the presence of snow cover and F_{ice} only in the case of bare ice. This value is not a parameter of the PDD model, since it is not used when the model computes the surface state (snow or ice).

We then multiplied the mean daily temperature by the corresponding F value and added the daily sublimation computed with Equation (7). The resulting ablation is hereafter referred to as 'T/ablation'.

Model calibration was performed using a Monte-Carlo approach based on 1,000,000 simulations to obtain the best calibration parameters, i.e. the degree-day factors (F_{ice} and F_{snow}) in equation (2) and the albedo threshold ($a_{threshold}$). The basic model was optimized at a daily time scale and the best score (r = 0.81; RMSE = 6.5 mm w.e. d^{-1}, E = 0.64) was obtained for F_{snow} = 5.68 mm w.e. °C^{-1} d^{-1}, F_{ice} = 10.53 mm w.e. °C^{-1} d^{-1} and $a_{threshold}$ = 0.49. The latter threshold is consistent with field observations (Figure 2).

Logically, the cumulative ablation obtained with the basic model (11.0 m w.e) is similar but slightly lower than the value given by the full energy balance model (11.3 m w.e.) and the melting obtained with the basic model was indeed highly correlated with that derived from the SEB equation. More instructively, the annual ablation cycle was accurately reproduced (Figure 3a – red line) with reduced ablation during windy periods and increased ablation when the wind speed is low. As a consequence, once the respective degree-day

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factors were accurately calibrated, the model was able to correctly reproduce the seasonal variability of ablation.

5.2 Validation using the year 2005

To validate the model, we applied it to the year 2005 period using parameters optimized over the 2002-2003 period (Figure 3b – red line). Even though the cumulative ablation obtained with the basic model (6.7 m w.e from January 1 to November 30 2005) was slightly overestimated compared with that from the full energy balance model (5.8 m w.e.), the scores (r = 0.81, n = 334, p = 0.001, RMSE = 6.5 mm w.e. d^{-1}, E=0.57) were acceptable, which gave us confidence in the ability of the model to reproduce melting and ablation.

5.3 Cross validation of the model

To assess the impact of the choice of the calibration period on the accuracy of model parameters and in turn, on model results, the periods 2002-03 and 2005 were interchanged and used as validation and calibration periods, respectively (Figure 3 – blue lines). This time, the zero melting intercept is obtained with $T_{threshold}$ = -2.14 °C, and the best score (r = 0.84; RMSE = 5.5 mm w.e d^{-1}, E = 0.69) was obtained for F_{snow} = 4.24 mm w.e. °C^{-1} d^{-1}, F_{ice} = 9.45 mm w.e. °C^{-1} d^{-1} and $a_{threshold}$ = 0.56, which is not very different from the original parameters. This time, the cumulative ablation obtained for the validation period with the basic model (10.4 m w.e from March 14, 2002 to August 31, 2003) was slightly underestimated compared with that from the full energy balance model (11.3 m w.e.). Nevertheless, the scores (r = 0.8, n = 334, p = 0.001, RMSE = 6.8 mm w.e. d^{-1}, E=0.62) remained acceptable which confirmed our confidence in the model.

5.4 Validation with ablation stakes

An in-depth analysis of the model was performed at 4,900 m a.s.l. The model was run using the mean daily temperature and wind speed recorded from 2000 to 2008 at 4,900 m a.s.l. The separation between snow and ice was not based on albedo values, but directly from the computed presence of snow at the surface. The results were consequently independent of $a_{threshold}$. Surface ablation was computed using the calibration described in section 5.1 between March 14, 2002 and August 31, 2003, whereas the calibration described in Section 5.3 was preferred in 2005 (January 1 to November 30, 2005). For the other periods, the calibrated parameters in 2002-03 and 2005 were averaged (F_{snow} = 4.96 mm w.e. °C^{-1} d^{-1}, F_{ice} = 9.99 mm w.e. °C^{-1} d^{-1}, $T_{threshold}$ = -2.09°C).

The results showed that the model accurately reproduced well the cumulative glacier mass balance at 4,900 m a.s.l. (Figure 4). In particular, the moderate ablation from 2000-2001 and 2008 was clearly reproduced. This suggests that the model accurately distinguished the surface states and accurately computed accumulation and ablation. Finally, using the mean calibration described in the previous paragraph, we observed that the model worked perfectly for the period 2000-2008.

5.5 Validation of mean coefficients with melting boxes

The model was applied using the set of mean parameters described in section 5.4, and the resulting daily melting values were compared with the melting amounts collected by the melting boxes over a period of 43 days. The surface states used to calculate melting were those identified in the field, giving a more accurate validation of melting amounts according to the real surface state. The correlation between modeled and measured daily melting was

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significant ($r = 0.8$; $n = 43$; $p = 0.001$), but the mean modeled melting rate was higher (22.9 mm w.e. d^{-1}) than observations (15.6 mm w.e. d^{-1}), because the slope of the regression line between observed and modeled melting was 1.05. This discrepancy is likely explained by the water retained in the melting boxes that leads to underestimation of the actual melting amounts. Indeed, the mean melting rate computed from the SEB approach (19.4 mm w.e. d^{-1}) was closer to the results of the basic model.

5.6 Final validation and model parameterization used in this study

The calibrated parameters in 2002-03 and 2005 were averaged (F_{snow} = 4.96 mm w.e. °C^{-1} d^{-1}, F_{ice} = 9.99 mm w.e. °C^{-1} d^{-1}, $T_{threshold}$ = -2.09°C). Averaging albedo threshold values ($a_{threshold}$ = 0.525), also allowed us to assess the uncertainty of the model compared to the surface energy balance model in 2002-03 and 2005. We observed that the model keeps a good score while using these averaged parameters (Table 5 and Figure 3 – green lines). The cumulative ablation obtained with the basic model was slightly underestimated for the period 2002-2003 (10.7 m w.e.) and slightly overestimated for 2005 (6.4 m w.e), but logically, the biases were then reduced for both periods.

In the following sections, we describe how the model was applied using this final set of averaged parameters, and how model uncertainty was tested using the parameters obtained by each calibration separately.

6 Model validation at glacier scale

The model was applied at 25 m intervals in the elevation range using the mean daily temperature and wind speed recorded from 2000 to 2008 at 4,900 m a.s.l. To assess the

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accuracy of the model and review its parameterizations, a sensitivity test of the computed surface mass balance was performed on the main model parameters (Table 6): the temperature threshold, the degree-day factors for ice and snow, and the temperature lapse rate (*LR*). We tested the uncertainty of the optimal degree-day factors and $T_{threshold}$ linked to the choice of $a_{threshold}$. The assumptions concerning sublimation and precipitation distributions with elevation were also tested.

6.1 Modeling the distributed surface mass balance over the period 2000-2008

We checked whether our basic model was able to properly reproduce 1) the temporal and 2) spatial variability of the surface mass balance of Antizana Glacier 15. The model was run using mean daily temperature, wind speed, and precipitation recorded at 4,900 m a.s.l. on the glacier from 2000 to 2008, using the parameter set described in section 5.6. We assumed that sublimation was constant with elevation. The resulting mass balances were compared with the measurements of the surface mass balance (b_a) made on the glacier.

Overall, simulated and measured vertical mass balance b(z) agreed fairly well in the ablation zone (Figure 5), even though in 2002-2003, the mass balance gradient with elevation was too steep between 4900 m a.s.l and 5000 m a.s.l.. However, in the upper part of the glacier, the point mass balance (accumulation) was generally overestimated (see Section 7.4).

Compared with the other years, the performance of the model was rather weak in 2002-2003 and in 2005. One peculiarity of this 2002-03 hydrological year was that the albedo was particularly low over both snow covered surfaces and bare ice (Figure 6). This suggests that the snow and ice were frequently dirty. Indeed, albedo measurements made in

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the ablation zone of Antizana Glaciers 15 and Glacier 12 between 2000 and 2008 were often below 0.3 for the ice and rarely above 0.56 for snow (Figure 6). We consequently decided to re-run the model using the parameters described in section 5.1 for 2002-03 (respectively described in section 5.3 for 2005). The performance of the model was improved in the ablation zone suggesting that the parameters described in section 5.1 are suitable for years with dirty ice (i.e. 2002 and 2003) but are likely too high for years when the ice is cleaner. Conversely, the parameters described in section 5.3 are suitable for years with clean ice.

We also compared our modeled glacier-wide climatic mass balance with mass balance estimates in Basantes Serrano *et al.* (2016) (Figure 7). Results were in good agreement with Basantes Serrano's estimates, and even better (r^2 = 0.87, p = 0.001, RMSE = 0.29 m w.e. a^{-1}) when we used the optimized parameters obtained in section 5.1 (and in section 5.3) for the year 2002-2003 (and for 2005, respectively). In this case, the mean modeled mass balance between 2000 and 2008 (0.06 m w.e. a^{-1}) was slightly more positive than the mean observed geodetic mass balance (-0.12 m w.e. a^{-1}) for 2000-2008 (Figure 7a). However, when we only used the parameters given in section 5.6 the mean modeled mass balance between 2000 and 2008 (0.08 m w.e. a^{-1}) was still closer to field observations (r^2 = 0.78, p = 0.001).

6.2 Modeling the transient snowline and ELA variations

To further validate the model, we compared the modeled *vs.* measured annual ELA, and the modeled *vs.* measured transient snowline at a daily time step. The modeled and measured transient snowline elevations were averaged over 15 days to reduce the impact of precipitation uncertainty on model results and to improve the readability of the figure. Indeed, because the tipping bucket rain gauges are not artificially heated, the snow can accumulate inside the funnel and only melt several hours or even a day after the precipitation occurred. This can lead to some shifts in the modeled daily transient snowline time series.

The modeled snowline was in good agreement with the observed measured transient snowline (r = 0.69, n = 96, p = 0.001 for 15-day periods between 2004 and 2008 and r = 0.70, n = 712, p = 0.001, based on daily values) demonstrating that the model was able to reproduce the altitudinal distribution of accumulation and ablation at a short time scale (Figure 8a&b). The difference between the modeled and the measured transient snowline was small (45 m (standard deviation STD = 47 m) for 15-day average snowlines and 30 m (STD = 59 m) for the daily snowlines (712 observations)). The modeled annual ELA also matched the measured ELA well (r^2 = 0.83, n = 9 years, p = 0.003, RMSE = 19 m), as a direct consequence of the good agreement between the modeled and the observed transient snowlines.

6.3 Model sensitivity

A sensitivity test was performed on every model parameter (F_{ice}, F_{snow}, $T_{threshold}$ and LR), on the gradient of sublimation as a function of altitude as well as of precipitation amounts (Table 6). We also tested the way the degree-day factor and threshold temperature were impacted by the choice of $a_{threshold}$ (Table 5), showing that the model parameters described in Section 5.6 were close to those obtained with the best calibration of $a_{threshold}$ for both 2002-2003 and 2005. This suggests that calibration is not very sensitive to $a_{threshold}$. As is always the case with PDD models (e.g. Azam *et al.*, 2012), the results are sensitive to LR, degree-day factors, and temperature threshold. These parameters are actually inter-dependent and different parameter sets could thus provide similar results. The results are

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also very sensitive to the amount and distribution of precipitation over the glacier area. This analysis showed that without applying a +76% correction for precipitation, as suggested by Wagnon et *al*. (2009), the agreement between simulated and measured mass balance would have been much worse. In conclusion, this basic model is able to properly simulate the mass change and the melting of Antizana Glacier 15 provided that it is thoroughly calibrated using a substantial dataset, which is a prerequisite for such modeling. This study suggests that in-situ measurements tend to underestimate precipitation amounts (strong undercatch of snow, especially when the weather is windy), and a significant correction is needed to assess real precipitation.

7 Discussion

7.1 On the existing relationship between T and energy fluxes

To understand which physical processes are responsible for the good performance of this basic model, we compared the basic model melting amounts (hereafter referred to as T/melting) with the different energy fluxes recorded at AWS_{G1}. A significant correlation was found between the T/melting and the net shortwave radiation *S* (r = 0.71, n = 530 days, p = 0.001). A moving correlation coefficient (r) between *S* and the *T*/melting over 30 days revealed variations over the annual cycle but the coefficient was generally 0.8 when temperatures underwent significant variations over a period of one month (data not shown). However, the correlation decreased when there was no variation in temperature over a longer period.

An in-depth analysis of correlations between daily energy fluxes and temperature (see Supplementary Materials) revealed moderate but significant (at p = 0.001) correlations

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between air temperature, *S* or albedo, and incoming shortwave radiation but only during periods with low speed winds. Since melting amounts during those periods (Period 2 and Period 1 with $u < 4$ m s^{-1}) represented more than 73% of the total melting amounts over the study period (i.e. 11.3 m w.e. between March 14, 2002 and August 31, 2003), the relationship between T and *S* likely largely explains the link between ablation and T.

The constant temperate conditions are always close to melting, and any slight increase in the incoming energy will enhance melting. As a consequence, any small change in *T* may have important consequences for precipitation phase, albedo, *S* and finally for melting. However, the relationship with *S*↓ only exists when the wind speed is low. Consequently, the model performance is likely to decrease when the wind becomes stronger. In our case, this had limited effects on melting on Antizana Glacier 15, since the windy periods were also low-melting periods, and as a consequence, had no significant impact on total melting amounts.

7.2 On the accuracy of model parameterization

7.2.1 Glacier slope and aspect

Several studies have shown that degree-day factors vary according to the slope and aspect of a glacier (e.g., Vincent *and Six.*, 2013). Here, the basic model calibration was performed using results from surface energy balance calculations for a horizontal surface whereas the glacier ablation area presents a mean slope of 28° and is oriented NW. Based on the characteristics of the ablation zone, the best score of the basic model calibration would decrease (r = 0.56, p = 0.001, E = 0.31 for 2002-2003) if we account for a 28° slope facing NW, suggesting that these calibration values at a daily time scale are only suitable for a

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horizontal surface. However, the impact is more limited at monthly and annual time scales, because the glacier is located at the latitude of 0° and there are fewer seasonal variations in melting caused by changes in the solar zenith angle than at other latitudes. Thus, the difference between annual ablation for a horizontal surface and for the mean slope and aspect of the ablation zone was less than 7% over 18 months. Nevertheless, it may be preferable to use the basic model to assess glacier ablation for horizontal surfaces.

7.2.2 $T_{threshold}$

Model parameterization suggests that melting began when the daily temperature was below 0 °C. First, using results from the surface energy balance model, we analyzed the frequency of melting events that occurred when the mean daily temperature was negative. We found that, (except for 4 days), melting was always significant, even when daily temperature was equal to -1.7°C, but nil at -2.1°C, which was the lowest daily temperature recorded in 2002-2003 and 2005. Observations made with melting boxes also showed that out of the 43 days of direct field observations, melting amounts were always significant, even if the mean daily air temperature was below 0 °C on nine days. For example, a daily melt of 3.8 mm w.e. d^{-1} on July 31, 2002 was measured when the mean daily air temperature was -1.3 °C.

The same situation has already been observed in Greenland (Van den Broeke *et al.*, 2010), where a -5 °C threshold was necessary to remove modeling biases caused by the occurrence of short periods of melting when significant nocturnal refreezing occurred. Indeed, these periods were characterized by mean daily air temperatures below 0 °C due in particular to unbalanced longwave budgets at night, but also by major incoming shortwave radiation leading to diurnal melting.

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7.2.3 Albedo threshold

The optimal albedo threshold between ice and snow surfaces was rather low compared to values reported in the literature (e.g., Oerlemans *et al*., 2009). The snow cover was generally thin because permanent snow is very rare at 4,900 m a.s.l. on Antizana (Wagnon *et al.*, 2009). This suggests that the ice below the surface snow cover may impact albedo measurements. The patchy distribution of snow on the surface of the glacier caused by the high winds on Antizana (Wagnon *et al.*, 2009) may also explain the low values. Indeed, even when thin snow still covers the surface of the glacier, snow may not be present everywhere. In such a case, the exposed ice surfaces may impact the mean albedo values.

7.3 Degree-day factors *F*

When we compared the *F* values from other regions, we found that our calibrations (F_{snow} = 4.96 mm w.e. °C^{-1} d^{-1}, F_{ice} = 9.99 mm w.e. °C^{-1} d^{-1} from section 5.6) were close to those obtained in the sub-tropical zone, for instance at Chhota Shigri Glacier (32.28°N, 77.58°E, F_{snow} = 5.28 mm w.e. °C^{-1} d^{-1}, F_{ice} = 8.63 mm w.e. °C^{-1} d^{-1}) (Azam *et al.*, 2014). However, they are also similar to those observed Storbreen in Norway (61.57°N 8.13°E, F_{snow} = 4.9 mm w.e. °C^{-1} d^{-1}, F_{ice} = 8.5 mm w.e. °C^{-1} d^{-1}) or Svartisheibreen in Norway (66.58°N, 13.75°E, F_{snow} = 6.0 mm w.e. °C^{-1} d^{-1}, F_{ice} = 9.8 mm w.e. °C^{-1} d^{-1}) (Radic and Hock, 2011).

7.4 Model accuracy in the accumulation zone

Accumulation data on Antizana Glacier 15 are currently poorly reliable. Using the precipitation correction proposed by Wagnon *et al*. (2009), we observed that the simulated and measured vertical mass balance *b*(z) agreed at low elevations, but not in the

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accumulation zone (Figure 5).

Nevertheless, Basantes Serrano *et al.* (2016) adjusted the mass balance series of glacier 15 with the 1997-2009 geodetic mass balance. The two matched only if the original accumulation measurements were systematically underestimated by a factor of 60%, due to the difficulty in recognizing a year-to-year reference level inside the snow during field observations, leading to sometimes erroneous in-situ accumulation measurements. This suggests that a correction factor should be applied to accumulation measurements given by Francou *et al*. (2004). Except for 2000, where this assumption yields a peculiar shape of modeled accumulation above 5000 m a.s.l., and for 2002 where accumulation was still underestimated, this assumption yields better agreement between modeled and observed mass balance at any elevation (Figure 5). This confirms that accumulation was largely underestimated by field measurements.

7.5 Temporal variation in degree-day factors

We observed that the quality of the model in the ablation zone was improved if the model was applied with different parameters in 2002-2003 and in 2005 than the parameters used in other years. This suggests that our optimized parameters may vary depending on the period of time (e.g., Huss and Bauder, 2009) reflecting variations in albedo (since degree-day factors for ice differ with the state of the surface).

For past or future climate reconstructions, given that degree-day factors may vary as a function of time, the uncertainty range of F values should always be taken into consideration when assessing the final uncertainty of the results. El Niño events are characterized by enhanced melting, partly due to low-albedo conditions (Francou *et al*., 2004), whereas the opposite situation is observed during La Niña events. Consequently, using different F values for the two events is highly recommended, irrespective of whether the goal of the study is to reconstruct past ablation or to make future projections.

7.6 Accuracy of the modeled transient snowline

Overall, there was a good agreement between the modeled and measured transient snowline, suggesting that even though the model is not physically based, it is able to broadly simulate most of the important physical processes controlling the surface mass balance of the glacier, or the transient snowline, likely because the 0 °C level, which has a direct impact on the precipitation phase (snow or rain) is a key variable governing the mass change of this glacier (e.g., Favier *et al*., 2004a&b, Francou *et al*., 2004). Nevertheless, in 2008, the differences between the 15-day average of the modeled and observed transient snowlines were larger than during the rest of the study period. These differences are likely due to either inaccurate observations of the snowlines due to some failures of the automatic camera, or the exceptional variability of the snowline in 2008 (Figure 8b). Indeed, during camera breakdowns, the 15-day snow line elevation was obtained from photographs taken during field trips. But such trips were conducted once or twice every 15 days, and as a consequence, the 15-day average only corresponded to 1 or 2 observations that were possibly not representative of the 15-day period. In addition, the simulated snowline sometimes varied considerably from day to day, which was less visible for the observed snowline (Figure 8b). This marked variability is probably due to recurrent small snow falls over an icy surface, a situation in which the model is very sensitive to precipitation uncertainties. Indeed, if solid precipitation is underestimated, snowfalls are not large enough to durably cover the glacier surface, leading to large day-to-day variability of the snowline elevation, although in reality, the glacier is mostly snow covered. On the contrary, if solid

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precipitation is overestimated, some simulated snowfalls may artificially shift the snowline to lower elevations than in reality while the glacier may be mostly free of snow.

8 Conclusion

The good agreement between temperature and glacier ablation or mass balance is not fortuitous but based on similar relationships as those found at other latitudes. Despite the limited variation in annual temperature (less than 3.5 °C, based on daily means), our study revealed a significant correlation between daily temperature and melting if a distinction between ice and snow was made, and provided that the model parameters (F_{ice}, F_{snow}, $T_{threshold}$, LR) were correctly calibrated. The comparison between daily temperature and the energy fluxes demonstrated that both air temperature and surface melting were closely linked to the net shortwave radiation budget through the impact of the albedo, which is mainly controlled by the precipitation phase (Favier *et al*, 2004a). However, we observed that the relationship between temperature and incoming shortwave radiation disappeared when the wind speed was high.

Moreover, despite the often weak correlations between incoming heat fluxes and temperature, a basic model including simple sublimation estimation, applied to local data gave accurate results on Antizana Glacier 15. The model is also suitable for the estimation of the transient snowline and ELA.

Because the correlation between temperature and melting is less significant with high speed winds, this type of model should not be used in the case of high sublimation (i.e. windy periods). However, in the case of Antizana Glacier 15, the consequences were limited for the mean monthly ablation because high sublimation events generally occurred when the temperature and melting (and in turn ablation since melting is the main ablation process

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here) on the glacier were low. This study showed that variations in the annual mass balance were well reproduced when the temperature was accurately assessed and when the model enabled correct estimation of the surface state (i.e. indirect estimation of surface albedo). However, full SEB computation reproduces measured ablation better, demonstrating that a complete surface energy balance model is preferable when accurate incoming fluxes are available (see Supplementary Materials, Figure S2). Several results also suggest that melting began when the daily mean air temperature 2 m above the surface of the glacier was still below 0 °C. If a threshold below 0 °C for temperature is not accounted for, a new calibration of the degree-day factors and the temperature lapse rate would be needed, which would lead to higher degree-day factors and/or a steeper temperature gradient.

In spite of the fairly good results we obtained, the model should be used with caution at high elevations where ablation is reduced, and when the wind speed is high. However, this study goes one step further in demonstrating the high sensitivity of glaciers to temperature changes in Ecuador. The Antizana glaciers have lost more than 30% of their area since 1950 (Francou *et al*., 2000; Rabatel *et al*., 2013), and temperatures in the tropical Andes have increased by up to 0.68 °C since 1939 (Vuille *et al*., 2008). Because several studies suggest that atmospheric warming will accelerate in the future and may reach 5 °C at the end of the 21st century (e.g., Vuille *et al*., 2008; Urrutia and Vuille, 2009), the ELA may rise 600 m and reach almost the elevation of the summit of Antizana. In these conditions, Antizana glaciers might drastically shrink or even disappear, which will have major consequences for local water supplies. Knowing the exact range of expected future temperature changes is thus crucial to assess its impact on local water resources.

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Table 1: Equipment used. Sensors installed at AWS_{G1} station (for SEB calculations) and at AWS_{M1} at 4,900 m a.s.l. and their specifications; thermometers at AWS_{G2} (5,000 m a.s.l.) and at AWS_{M2} station (4,785 m a.s.l.) and rain gauge characteristics (4,550 m a.s.l.), and albedometer at AWS_{G3} (4,900 m a.s.l. on Glacier 12).

Data measured[1]	Type of sensor	Station name/elevation[2]/type of surface	Period with data	Accuracy[3]
Air temperature, °C	Vaisala HMP 45, aspirated[4]	AWS_{G1} / 4,900 / Glacier	2000-2005, 2007-2008	±0.2°C
	Vaisala HMP 45, aspirated[4]	AWS_{G2} / 5,000 / Glacier	2003-2004	±0.2°C
	Vaisala HMP 45, aspirated[4]	AWS_{M1} / 4,900 / Moraine	2005-2008	±0.2°C
Air temperature, °C	Automatic HoboPro	AWS_{M2} / 4,785 / Moraine	2000-2008	± 0.2°C
Relative humidity, %	Vaisala HMP 45, aspirated[4]	AWS_{G1} / 4,900 / Glacier	2000-2005	±2 %
Wind speed, m s^{-1}	Young 05103	AWS_{G1} / 4,900 / Glacier	2000-2005	±0.3 m s^{-1}
Wind direction, deg	Young 05103	AWS_{G1} / 4,900 / Glacier	2000-2005	±3 deg
Incident short-wave radiation,W m^{-2}	Kipp&Zonen CM3, $0.305<\lambda<2.8\mu m$	AWS_{G1} / 4,900 / Glacier	1999-2005,2007-2008	±3 %
	Kipp&Zonen CM3	AWS_{G3} / 4,900 / Glacier 12	2006	±3 %
Reflected short-wave radiation, W m^{-2}	Kipp&Zonen CM3, $0.305<\lambda<2.8\mu m$	AWS_{G1} / 4,900 / Glacier	1999-2005,2007-2008	±3 %
	Kipp&Zonen CM3	AWS_{G3} / 4,900 / Glacier 12	2006	±3 %
Incoming long-wave radiation, W m^{-2}	Kipp&Zonen CG3, $5<\lambda<50$ µm	AWS_{G1} / 4,900 / Glacier	2002-2004	±3 %
	Kipp&Zonen CG3	AWS_{M1} / 4,900 / Moraine	2005	±3 %
Outgoing long-wave radiation, W m^{-2}	Kipp&Zonen CG3, $5<\lambda<50\mu m$	AWS_{G1} / 4,900 / Glacier	2002-2004	±3 %
Daily precipitation, mm	Automatic Hobo Rain Gauge[5]	P2 / 4,785 / Moraine	2000-2008	Opening: 200 cm^2 Height: 100 cm
	Automatic Hobo Rain Gauge[5]	P4 / 4,785 / Moorland	2000-2008	Opening: 200 cm2 Height: 100 cm

[1]Quantities are half-hourly means of measurements made at 15-s intervals except for wind direction, which are instantaneous values measured at 30-minute intervals.
[2] m a.s.l.
[3]according to the manufacturer
[4]artificially aspirated to prevent over heating due to radiation.
[5]Tipping bucket rain gauge, measured precipitation 0.214 mm by tipping.

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Table 2: Data used in this study

Modeling	Period	Reference input data for modeling	Data gaps[1]	Data used to fill gaps[2]	Model validation
PDD/Mass balance/Transient snowline	2000-2008	2000-2001: AWS_{G1} 2002-2004: AWS_{G1} 2005: AWS_{G1} 2006: AWS_{M1} 2007-2008: AWS_{G1}	2000-2001: 41% 2002-2004 : 14% 2005: 21% 2006: 13% 2007-2008: 8%	AWS_{M2} AWS_{G2} (otherwise: AWS_{M2}) AWS_{M1} (otherwise: AWS_{M2}) AWS_{M2} AWS_{M1} (otherwise: AWS_{M2})	Measured mass balance & Melting boxes

[1]Number of days with missing data (percent)
[2]for wind speed, when data were missing at every station, we used NCEP1 reanalysis output

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Table 3: Description of daily temperature data used and correlations with Antizana Glacier 15 data at 4,900 m a.s.l. All the correlation coefficients are significant at p = 0.001.

Station	AWS_{G1}	AWS_{M1}	AWS_{M2}
Determination coefficient (r^2) with AWS_{G1} during each period[1]		Not available	0.80 (2002-2004)
Determination coefficient (r^2) with AWS_{G2} during each period[1]	0.90 & 0.85 (2002 & 2003)	0.87 & 0.89 (2005 & 2008)	0.75 & 0.86 (2003 & 2007-2008)
Determination coefficient (r^2) with AWS_{M2} during each period[1]	0.80 (2002-2004)	0.89 (2005-2008)	

[1]Periods are in parentheses.

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Table 4: Data used for basic model calibration and validation

Data	Model/method	Period	Location
Daily melting	SEB calculation	March 14, 2002-August 31, 2003 Jan – Nov 2005	AWS_{G1}
Measured melting	Melting boxes	March 12, 2002-June 11, 2003 (43 days): Antizana, (4,900 m a.s.l.)	AWS_{G1}
Terrestrial photographs	Automatic camera	2004–2008: 712 daily photos	Antizana Glacier 15 frontal moraine (4,875 m a.s.l.)
Daily albedo	Kipp & Zonen CM3 pyranometer	1999-2005: at AWS_{G1} 2006: on Antizana Glacier 12 (4,900 m a.s.l.) 2007-2008: at AWS_{G1}	On Antizana Glaciers 15 and 12

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Table 5: Summary of the optimized sets of parameters with their respective modeling scores, and sensitivity of parameters to $a_{threshold}$ variations in 2002-2003 and 2005.

Parameter	F_{ice}[1]	F_{snow}[1]	$T_{threshold}$[2]	$a_{threshold}$	Measured ablation[3]	Modeled ablation[3]	R	RMSE[1]	E
Calibration on 2002-03	10.53	5.68	-2.05	0.49	11.3	11.0	0.81	6.5	0.64
Validation on 2005	10.53	5.68	-2.05	0.49	5.8	6.7	0.81	6.5	0.57
Calibration on 2005	9.45	4.24	-2.14	0.56	5.8	6.2	0.84	5.5	0.69
Validation on 2002-03	9.45	4.24	-2.14	0.56	11.3	10.4	0.8	6.8	0.62
Validation of mean parameters on 2002-03	9.99	4.96	-2.09	0.52	11.3	10.7	0.81	6.6	0.63
Validation of mean parameters on 2005	9.99	4.96	-2.09	0.52	5.8	6.4	0.83	5.9	0.64
Sensitivity in 2002-03	9.91	4.87	-2.14	0.56	11.3	11.1	0.79	6.3	0.63
Sensitivity in 2005	9.38	4.38	-2.05	0.49	5.8	5.7	0.82	5.7	0.67

[1]in mm w.e. °C^{-1} d^{-1}
[2]in °C
[3]in m w.e.

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Table 6: Model sensitivity tests. Values (in m w.e. a^{-1}) are the differences between the "original" mass balance (0.08 m w.e. a^{-1}) and the mass balance resulting from the sensitivity test over the 2000-2008 period. Here, the "original" mass balance refers to the mass balance obtained with final parameters given in section 5.6, and the sensitivity test mass balance results from computations with the parameter value given in the same cell.

$T_{Threshold}$		$\boldsymbol{F}_{ice}$		$\boldsymbol{F}_{snow}$		*Lapse Rate*		*Sublimation*		*Precipitation*	
Value (°C)	difference	Value (mm w.e. °C^{-1} d^{-1})	difference	Value (mm w.e. °C^{-1} d^{-1})	difference	Value (°C km^{-1})	difference	Value (mm w.e. a^{-1})	difference	Value	difference
-2.14	-0.06	**10.53.**	-0.02	**5.68.**	-0.21	**-9.2**	0.08	**Constant with elevation**	0	**-76%**	-1.13
-2.09	0	**9.99**	0	**4.96**	0	**-8.5**	0				
-2.05	0.05	**9.45.**	0.02	**4.24.**	0.21	**-8.2**	-0.04	**Linear decrease to 0 (summit)**	0.13	**0%**	0

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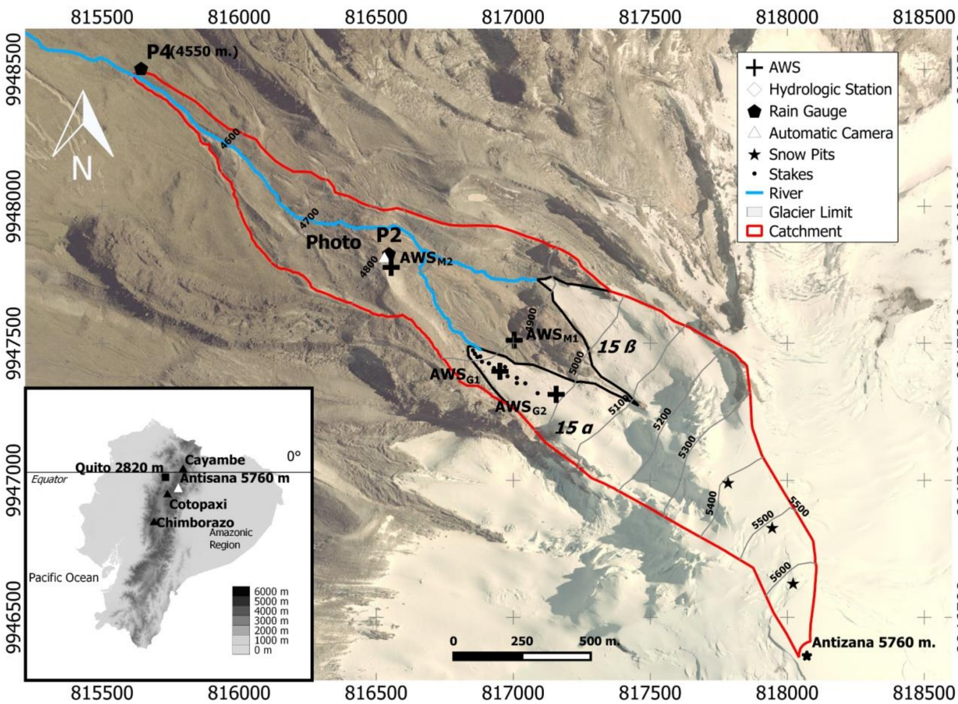

Figure 1: Orientation map of Antizana Glacier 15 showing location of monitoring equipment. Projection is on UTM zone 17, coordinate system is WGS84. Inset is a physical map of Ecuador. In this figure, AWS_{G3} is not shown. The background image is an orthoimage of Antizana Volcano taken in 2010 by the *Instituto Geográfico Militar* of Ecuador: *http://www.geoportaligm.gob.ec/portal/index.php/catalogo-de-datos/)*

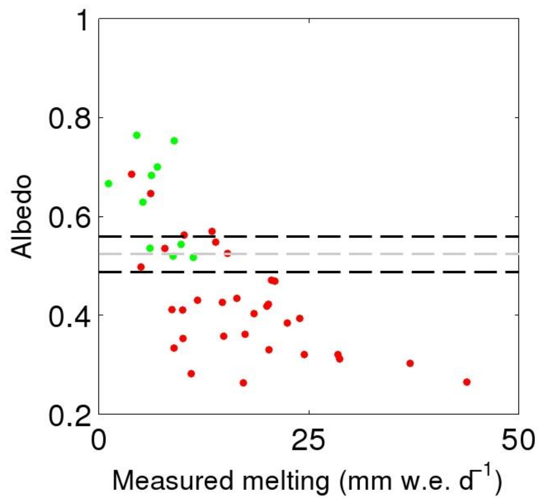

Figure 2: Comparison between measured melting rates in melt boxes and mean daily albedo for snow (green) and ice (red). Dashed horizontal black lines are optimized thresholds between snow, and ice for 2002-2003 ($a_{threshold}$ = 0.49) and 2005 ($a_{threshold}$ = 0.56). The dashed gray line is the mean albedo threshold ($a_{threshold}$ = 0.525).

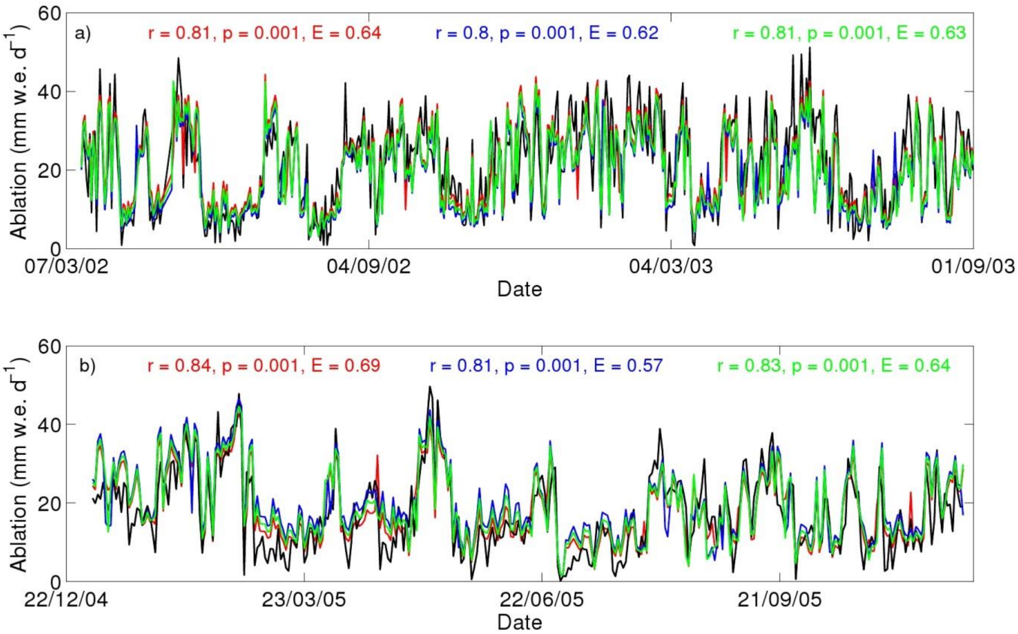

Figure 3: (a) Comparison between computed daily ablation rates obtained with the SEB model (black) and with the basic model. The red curve is the optimized modeling for 2002-2003, i.e. using parameters (F_{ice}, F_{snow}, $T_{threshold}$ and $a_{threshold}$) optimized on 2002-2003. The blue curve is the validation curve using parameters optimized using 2005. The green curve is the same as the red curve but accounts for the mean parameters given in section 5.4. (b) Same as (a) but for 2005. The colors of the correlation coefficients correspond to those of the curves.

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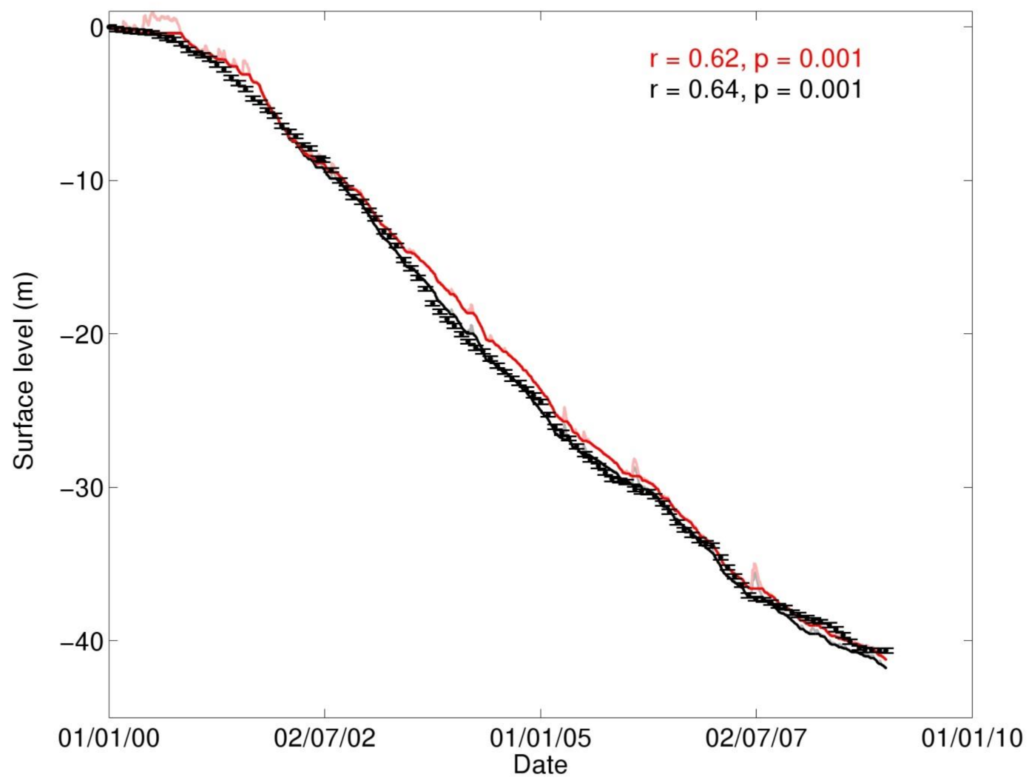

Figure 4: Comparison between modeled (black and red lines) and measured (dots with error bars) surface mass balance (in m of ice) at 4,900 m a.s.l. between 2000 and 2008. The gray and pink lines represent the level of snow assuming a density of 200 kg m^{-3}. The red line is the results of the basic model using the mean parameters given in section 5.6. The black line shows the results using mean parameters given in section 5.6, except between March 15, 2002 and August 31, 2003, when the parameters given in section 5.1 were preferred, and between January 1, 2005 and November 30, 2005 when the parameters given in section 5.3 were preferred. The correlation coefficients are between observed and modeled monthly mass balance. The colors of the determination coefficients correspond to the colors of the symbols.

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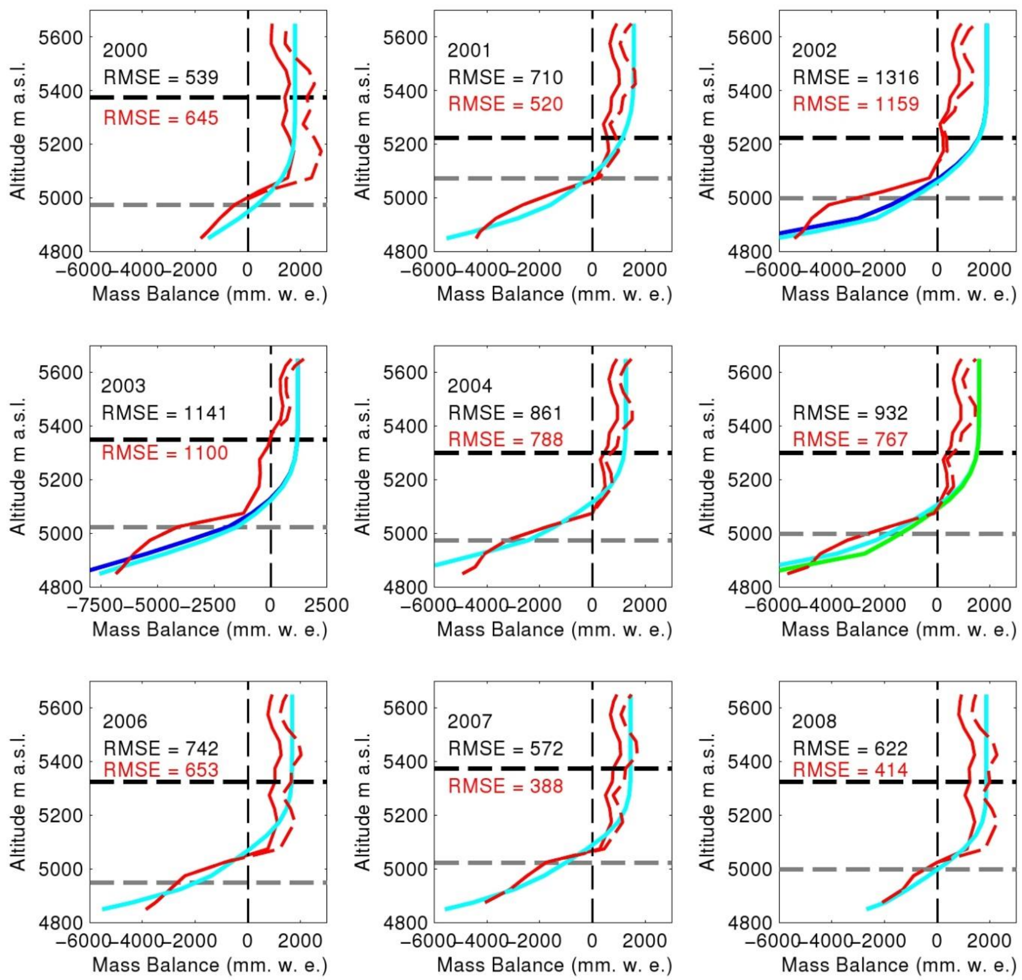

Figure 5: Variations in the point mass balance versus elevation for each year between 2000 and 2008 assuming sublimation remains constant with elevation. The study year is given in the upper left corner of each panel. The continuous red curve is the measured mass balance; the dashed red line is accumulation multiplied by 1.6 as suggested by Basantes Serrano et al (2016); the light (dark) blue lines are modeled mass balance assuming model parameters are the mean coefficient given in section 5.6 (in section 5.1, respectively). The green line is the modeled mass balance assuming model parameters are the coefficient given in section 5.3. The horizontal dashed lines represent the elevation of the lowest accumulation measurement

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(black) and of the highest ablation stake (gray) surveyed during the corresponding year. RMSE is computed between observed and modeled mass balance at each elevation range. Values in black are for b(z) values given by Basantes Serrano et al (2016), values in red are those obtained when accumulation is multiplied by 1.6.

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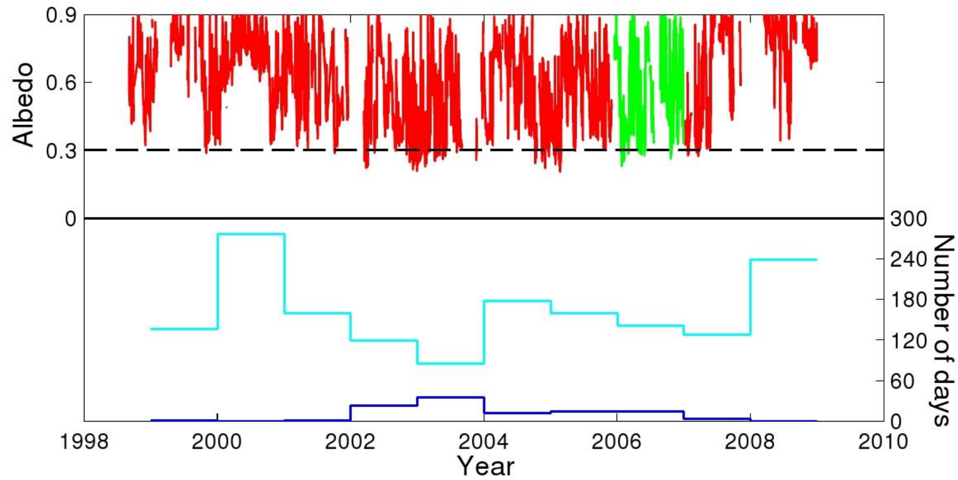

Figure 6: Daily albedo (upper panel) at 4,900 m a.s.l. on Antizana Glacier 15 (from 1999 to 2005 and in 2008, in red) and at 4,900 m a.s.l. on Antizana Glacier 12 (2006, green). Missing data between December 17, 2001 and March 14, 2002 are not accounted for, otherwise it would have increased the number of occurrences in 2002. Lower panel: The dark blue line shows the number of days with albedo values below 0.3, the light blue line shows the days with albedo values above 0.56.

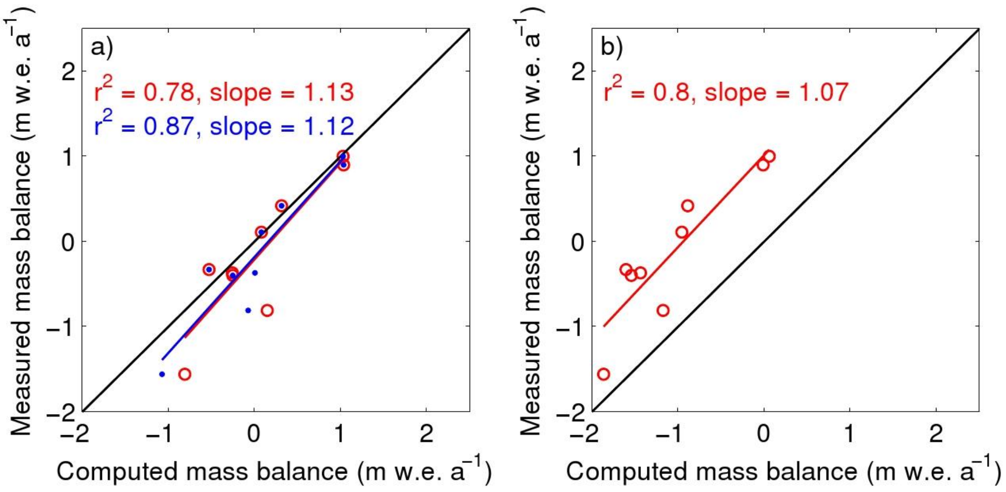

Figure 7: Comparison between the computed and the measured glacier-wide annual mass balance of Antizana Glacier 15. (a) Modeled data are forced with temperature and precipitation data from Antizana Glacier 15 catchment. Blue dots indicate the results using the mean calibration parameters given in section 5.6. Red circles are the results using optimized parameters given in section 5.6 except for 2002-03 and 2005 where parameters come from section 5.1 and section 5.3 respectively. (b) Same as (a) but modeled mass balance do not account for precipitation correction of 76% suggested by Wagnon *et al.* (2009). The colors of the determination coefficient and slope values correspond to the colors of the symbols. The 1:1 line is also shown in black.

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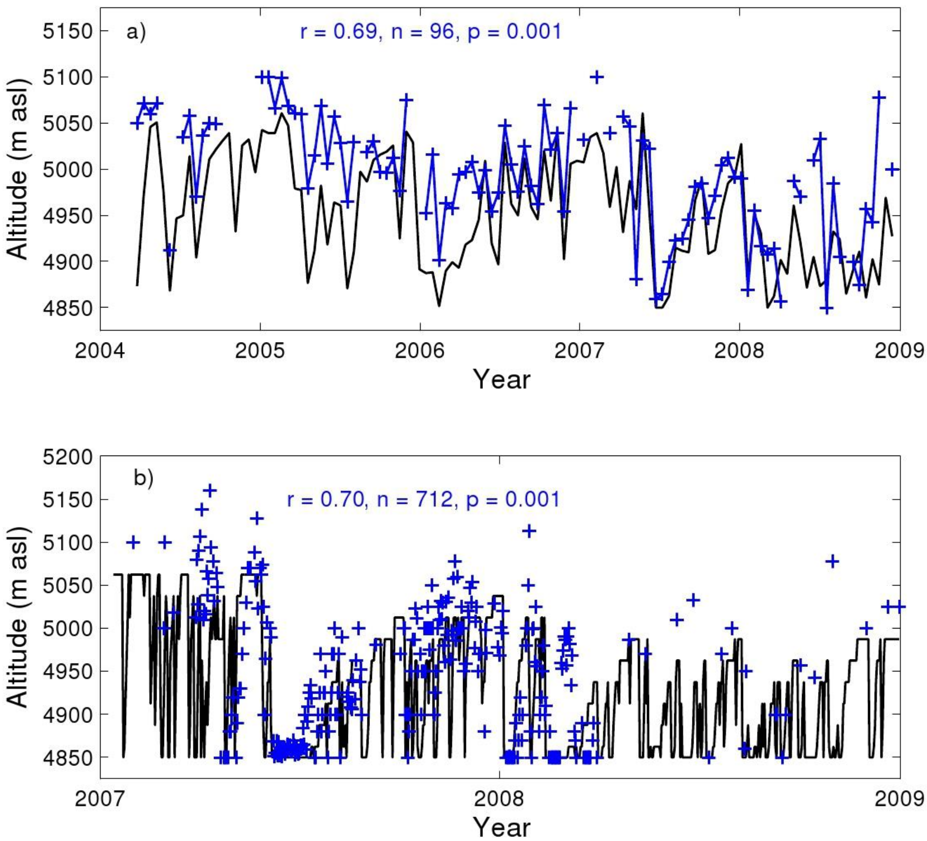

934

935 Figure 8: Comparison between the observed (blue) and modeled (black) transient snowline
936 elevations accounting for a 76% increase in precipitation compared with measurements, at
937 (a) a 15-day time scale (15-day averages) over the period 2004-2008, and at (b) a daily time
938 scale over the period 2007-2008.