Modelling melt, runoff, and mass balance of a tropical glacier in the Bolivian Andes using an enhanced temperature-index model

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Abstract:

This paper evaluates the feasibility of applying a coupled melt, runoff, and mass balance model to the tropical Zongo glacier (Cordillera Real, Bolivia) during two hydrological years. Melt rate was estimated using the standard degree-day method (DDM) and an enhanced temperature-index model (ETI). The latter was run with values of parameters obtained for Haut Glacier d’Arolla and a recalibrated parameter set for Zongo glacier. Glacier mass balance was calculated using snowfall inputs and modelled melt and sublimation. Estimated monthly mass balance and discharge were compared with observations from a stake network in the ablation zone and data from a hydrometric station. We concluded that ETI model agrees very well with the reference runoff and mass balance. Net mass balance over the whole glacier was predicted accurately in the ablation zone, but the model overestimated mass balance in the accumulation zone owing to the absence of observations at higher elevations; the equilibrium line altitude and accumulation area ratio were predicted within reasonable limits. The results demonstrate that ETI model is applicable in tropical conditions, provided that the parameters are recalibrated for the climatic settings of this region.

KEYWORDS Zongo glacier; glacier hydrology; mass balance; temperature-index model; tropical glaciers; Bolivia

INTRODUCTION

Low latitude glaciers are found in the Andes mountain range of South America, the Rwenzori Mountains and high volcanoes of East Africa, and at mountain summits in the Irian Jaya region of New Guinea, where they are confined to elevations of 4500 m a.s.l. (above sea level) and higher. The tropical Andes host more than 99% of the total area of tropical glaciers worldwide; 71% are in Peru, 20% in Bolivia, 4% in Ecuador, and 4% in Colombia and Venezuela (Kaser, 1999). Using an extended World Glacier Inventory, Rádíc and Hock (2010) estimated that tropical glaciers in South America occupy an area of roughly 7060 ± 137 km² with a total ice volume of 344 ± 37 km³ that has the potential to raise sea level by 0.86 ± 0.09 mm sea level equivalent. Although this contribution is small, glaciers in tropical regions are particularly sensitive to changes in climate and play an important role as water reservoirs (Vuille et al., 2008). Andean cities such as La Paz (Bolivia), Lima (Perú), and Quito (Ecuador) are close to glaciers and depend heavily on glacier-derived water flows for their water supply and for a large part of their energy needs (Chevallier et al., 2011).

Climatic characteristics in the tropics are different from those at higher latitudes. On tropical glaciers, ablation and accumulation periods occur simultaneously and ablation is not confined to a summer season. Moreover, temperature does not change much from summer to winter as opposed to the pronounced seasonal changes of humidity and precipitation (Mölg et al., 2008; Giesen and Oerlemans, 2012).

A recent review of the current state of glaciers in the tropical Andes by Rabatel et al. (2013) showed that in terms of changes in surface area and length, glacial retreat over the last three decades is unprecedented since the maximum extent of the Little Ice Age. In terms of changes in mass balance, they concluded that, although several glaciers have shown sporadic gains, the overall trend has been negative over the last 50 years. In addition, even if glaciers are currently retreating everywhere in the tropical Andes, the retreat is much more pronounced for small glaciers with a maximum altitude lower than 5400 m a.s.l.; these do not have permanent accumulation zones and could disappear within years or decades. In fact, Morizawa et al. (2013) found that the Condoriri glacier (5 km², 4800–5600 m a.s.l.) lost 41% of its area from 1988 to 2010, and, at the current rate of retreat, it could virtually disappear by 2035.

Surface melt rates can be calculated by means of two approaches that differ on their data requirements, complexity, and whether they use physically-based calculations or conceptual equations, in other words, energy balance or temperature-index models. Previous studies in the tropical Andes applied energy balance models to simulate melt and runoff (e.g., Caballero et al., 2007; Juen et al., 2007; Sicart et al., 2011). Although these methods have a strong physical basis and compute relevant energy fluxes at the glacier surface, they are often not practical because of their large data requirements. On the other hand, applications of temperature-index models have been few, owing to their decreasing accuracy with increasing temporal resolution and shortcomings in modelling the spatial variability of melt rates induced by the effects of topography and surface type. In tropical glaciers, simple models similar to the standard degree-day approach have been used to simulate the mass balance without detailed examination of the physical basis supporting
the model (Rabatel et al., 2013). There have been many attempts to overcome these limitations of temperature-index models by including additional variables, such as radiation components, humidity, and wind speed (e.g., Hock, 1999; Pellicciotti et al., 2005). However, none of them have been tested in tropical glaciers.

Hydrological and glaciological observations have already been carried out on Zongo glacier to estimate glacier mass balance (Sicart et al., 2007). Moreover, a distributed energy balance model was applied to investigate atmospheric forcing that controls meltwater discharge and seasonal variations in mass balance of this glacier (Sicart et al., 2011). The main aim of this study is to evaluate the applicability of an enhanced temperature-index glacier melt model (ETI) in tropical environments. ETI was developed on a glacier in the Swiss Alps and has been tested extensively in the Alps and tropical environments. ETI was developed on a glacier in the Andes of Chile (Pellicciotti et al., 2011). The program involves the assessment of surface mass balance (Francou et al., 1995), meltwater discharge (Ribstein et al., 1995), and energy balance (Wagnon et al., 1999). In addition, Zongo has been cataloged as one of 37 reference glaciers by the World Glacier Monitoring Service. These reference glaciers have well-documented long-term mass balance observations based on the direct glaciological method and are especially valuable for interpreting climate/glacier relationships (WGMS, 2013).

We used two sources of meteorological, hydrological and glacial data: the free-access GLACIOCLIM database (http://www-lgge.ujf-grenoble.fr/ServiceObs/) and previously published reports (e.g., Berger et al., 2006; Perroy et al., 2007). More details about these data can be found in Supplement Text S1.

From these data we created a daily database of ventilated air temperature (°C), global radiation (W m–2), albedo, precipitation (mm), relative humidity (%), wind speed (m s–1) and discharge (m3 s–1) for the period 1 September 2004 to 31 August 2006 (Supplement Figure S2). Gaps in the data (less than 9% of measured data) were filled by using linear regression. Daily albedo was calculated as the ratio of daily mean reflected shortwave radiation to daily mean downward shortwave radiation. Whenever albedo was determined to be larger than one, which happened occasionally because snowflakes adhered to the upward-looking sensor during snowfall, the incoming shortwave radiation was adjusted to yield a maximum value of 0.90.

Comparisons between discharge and mass balance have been performed by Sicart et al. (2007) using the hydrological and glaciological methods. They concluded that, although the two methods reproduce similar interannual variations, the hydrological budget gives values that are approximately 600 mm w.e. (water equivalent) per year lower than those of the glaciological budget. These authors attributed the large difference to the catch deficiency of rain gauges and the absence of precipitation measurements at high altitudes.

METHODS

Glacier mass balance

For most glaciers, the surface mass balance controls the mass exchange between the glacier and the atmosphere. Contributions from several processes (e.g., snowfall, avalanche deposition, melt, refreezing of water, sublimation and wind deposition) determine the surface mass balance rate at a point (Cuffey and Paterson, 2010). If it is assumed that snowfall, melt and sublimation dominate the mass budget, then the glacier mass balance change \((b, \text{ in mm w.e.})\) can be approximated by:

\[
b = a - M - s \tag{1}
\]

Here, \(a\) (mm w.e.) represents snowfall, \(M\) (mm w.e.) is melt, and \(s\) (mm w.e.) stands for sublimation. A preselected critical temperature, \(T_{\text{cr}}\) (°C) was used in the model to decide whether precipitation was rainfall or snow. In the former case, precipitation immediately contributes to runoff but not to mass balance while, in the latter case, new snow is retained as part of the snowpack until melting conditions occur. Lejeune et al. (2007) found that \(T_{\text{cr}}\) ranged from –1 to 3°C on Zongo glacier. Here, we used a constant value of 1.5°C.
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which is similar to the value applied by Sicart et al. (2011).

Mass balance varies substantially from place to place on a glacier; there is generally a broad upper region of mass surplus and a broad lower region of mass deficit. The equilibrium line is the boundary between these zones, where accumulation equals ablation; its elevation defines the equilibrium line altitude (ELA). ELA is typically estimated from the mass balance profile and corresponds to the elevation where annual mass balance equals zero. Another parameter of interest is the accumulation area ratio (AAR), which is defined as the ratio of the area of the accumulation zone to that of the entire glacier.

Integrating the annual balance over the total area of the glacier, $A_g$, gives the balance of the whole glacier for one year, $B_g$:

$$ B_g = \frac{1}{A_g} \left( \sum b_{ni} - A_{gi} \right) $$

(2)

where $b_{ni}$ (mm w.e.) is the annual balance at a given elevation band $i$, and $A_{gi}$ ($\text{km}^2$) is the glacier surface area of the $i$th band.

Melt model

A general description of the standard degree-day method can be found in Supplement Text S2. In this research, the model proposed by Pellicciotti et al. (2005) was selected for estimating the melt rates on Zongo glacier. This enhanced temperature-index model (ETI) calculates melt as the sum of two components:

$$ M = \begin{cases} \text{TF} \cdot T + \text{SRF} \cdot (1-\alpha) \cdot G & \text{if } T > T_f \\ 0 & \text{if } T \leq T_f \end{cases} $$

(3)

where $T$ (°C) is the daily mean temperature, $\alpha$ is albedo, $G$ (W m$^{-2}$) is incoming shortwave radiation, and $\text{TF}$ (mm d$^{-1}$ °C$^{-1}$) and $\text{SRF}$ (mm d$^{-1}$ km$^{-2}$ W$^{-1}$) are two empirical coefficients, namely the temperature factor and the shortwave radiation factor. Their values for the Swiss Alps are $\text{TF} = 1.2$ and $\text{SRF} = 0.226$ (Pellicciotti et al., 2005). The threshold temperature ($T_f$) for melt to occur is $1.0$ °C.

Melt, mass balance, precipitation and sublimation are not constant in space. Their spatial variations were accounted for by dividing the Zongo watershed into 13 elevation zones, each having an elevation range of about 100 meters (Supplement Figure S1). A detailed description of the routines for extrapolating meteorological data and estimating sublimation, runoff, and model performance can be found in Supplement Texts S1 and S3–S5.

Albedo parameterization

Daily values for albedo at a given elevation were simulated using the approach presented by Oerlemans and Knap (1998) and compared with observations. This parameterization was considered because of its simplicity in terms of required data and control parameters. Albedo of the snow-covered glacier site on day $n$ was decreased by an exponential function of the days since the last snowfall $n_s$ as follows:

$$ \alpha_{\text{snow}} = \alpha_{\text{fim}} + (\alpha_{\text{fim}} - \alpha_{\text{fim}}) \exp \left( \frac{n_s - n}{t^*} \right) $$

(4)

where $\alpha_{\text{fim}}$ represents the characteristic albedo of firm, $\alpha_{\text{fim}}$ is the characteristic albedo of fresh snow, and $t^*$ (d) is a time-scale parameter that determines how fast the snow albedo approaches the firm albedo after a snowfall.

To account for the transition to the characteristic ice albedo ($\alpha_{\text{ic}}$) when snow depth is small, albedo of thin snow layers was calculated as:

$$ \alpha = \alpha_{\text{snow}} + (\alpha_{\text{ic}} - \alpha_{\text{snow}}) \exp \left( \frac{-\text{SWE}}{d^*} \right) $$

(5)

where SWE (mm w.e.) is the snow water equivalent (w.e.) of the snowpack over the glacier, and $d^*$ (mm w.e.) is a depth-scale parameter. Characteristic albedos of firm, snow and ice were initially obtained from a literature review by Cuffey and Paterson (2010) and then were manually adjusted to albedo observations. The time-scale and depth-scale parameters were taken from the research of Sicart et al. (2011). Their values are summarized in Table I.

Finally, the daily change in SWE was simulated as the difference between snowfall, melt and sublimation:

$$ \text{SWE}_{n+1} = \begin{cases} \text{SWE}_{n} + a_{\text{snow}} - (M_s + s_n) & \text{if } \text{SWE}_{n} + a_{\text{snow}} - (M_s + s_n) > 0 \\ 0 & \text{Otherwise} \end{cases} $$

where SWE, equals zero, ice-melt took place in the glacierized areas of the basin.

This scheme posed an interesting problem for thin snow-packs: at every time step, precipitation was calculated, snow depth was updated, albedo and melt were computed, and then snow depth was adjusted again. Therefore, albedo was solved in an iterative way.

RESULTS

First of all, model parameters for ETI were either determined from observations, taken from previous studies, or tuned by calibration. Estimated daily mass balance was converted to monthly mass balance and compared with the readings at ablation stake 9K. The site was snow-covered at the beginning of hydrological year 2004–2005 with an initial SWE of 0.8 mm w.e. This value was derived from monthly observations of snow depth and converted into units of water equivalent using a fresh snow density of 250 kg m$^{-3}$ (Sicart et al., 2011). Ablation stake 9K was selected because of its proximity to the AWS (Automatic Weather Station) on the glacier, thus reducing the errors due to data extrapolation.

Table I summarizes the optimum parameter set for Zongo glacier. The melt model parameters were obtained using an optimization procedure. The generalized reduced gradient solving method (Lasdon and Waren, 1981), which is included with the standard version of Excel under the data analysis toolbox, was used for nonlinear optimization of $\text{TF}$ and $\text{SRF}$. The mean absolute error (MAE) was selected as the objective function for which the optimal parameters produced the minimum value. The optimization problem was defined by the constraint $T > T_f$, where $T_f$ was unknown. Its optimal value was manually adjusted until the best agreement between observed and simulated mass balance was found.

A distinct set of parameters was derived for the wet and the dry seasons to account for the different climatic conditions typical of each season. $\text{SRF}$ reached a value close to its
Physically-based value of 0.259 mm d^{-1} m^{-2} W^{-1} (in a physical melt model, this value corresponds to the energy to water depth conversion) in the wet season, showing a higher contribution of solar radiation to melt because of the high elevation and low latitude of the studied glacier, but it was considerably lower in the dry season. Similarly, the calibrated value of TF was equal to that of Pellicciotti et al. (2005) in the wet season, indicating a positive contribution of the temperature-dependent terms of the energy balance (longwave radiation and turbulent fluxes) to melt, while it was close to zero in the dry season.

Pellicciotti et al. (2008) and Carenzo et al. (2009) linked SRF and TF to the prevailing meteorological conditions of the glacier site and found that “high SRFs are obtained on clear-sky days, whereas higher TFs are typical of locations where glacier winds prevail and turbulent fluxes are high”. On Zongo glacier, solar radiation increases to its annual maximum value and net shortwave radiation provides most of the melt energy in the wet season (Supplement Figure S2e and Sicart et al., 2011). Therefore, SRF is higher for the period September–February, although cloudy conditions are frequent, as shown by the relative humidity data (Supplement Figure S2d). Moreover, turbulent fluxes of sensible and latent heat almost offset each other in the wet season (Sicart et al., 2011), and, although they have a higher contribution in the dry season when energy losses due to sublimation are expected (Wagnon et al., 1999), they remain low compared to radiative fluxes. For this reason, TF took a low constant value of 0.2 mm d^{-1} °C^{-1}.

The threshold temperature remained unexplained, although a first assessment reveals that it might be related to the glacier surface temperatures: in the wet season, surface temperatures on Zongo remain near the melting point, so that any contribution of energy to the glacier surface is consumed by melting (low TF). On the other hand, during the dry season, surface temperatures remain well below the melting point for most of the time and the energy available for melt is used first to heat up the snow or ice pack (high TF).

The resulting model parameters were then applied to the other stake sites. The results are shown in Figure 1 for six selected sites and summarized in Table II for all stakes. Also shown in the table are mean MAE and bias for the ensemble of the 16 stakes. ETI with the original parameters of Pellicciotti et al. (2005) underestimated melt at most sites, while ETI with recalibrated parameters performed better at 10 out of the 16 stakes and fitted measured ablation data quite well (Figure 1). MAE for the ensemble reached 3.5 mm d^{-1}, for measured daily melt rates in the range of 0.0 to 26.0 mm d^{-1}. The best fits were produced at locations near the central axis of the glacier (e.g., stakes 2G, 9K, 10K and 12K) with moderate slopes (3–15°) facing east to southeast directions, where MAE ranged between 0.9 and 3.5 mm d^{-1}.

We found the best agreement at ablation stake 9K, which was installed on a regular and flat area of the glacier and hence received the maximum amount of solar radiation. Other locations (e.g., stakes X1, 11K, 9N, 10N, 11N and 12N, aspect angles from 67 to 90°) showed an improvement in mass balance estimations when using recalibrated parameters, but the differences between observed and simulated mass balance were larger. Ablation stake X1 was located near a steep slope of seracs, formed in the transition between the accumulation and ablation zones between 5100–5300 m

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Table I. Optimum parameter set used for running ETI. Details of how their values were determined are described in the text. The ranges of the various parameters were taken from the literature.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Unit</th>
<th>Value</th>
<th>Range</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>General parameters</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Temperature lapse rate (2004–2005)</td>
<td>γ</td>
<td>°C per 100 m</td>
<td>0.48</td>
<td>0.32–0.58</td>
<td>Berger et al. (2006)</td>
</tr>
<tr>
<td>Temperature lapse rate (2005–2006)</td>
<td>γ</td>
<td>°C per 100 m</td>
<td>0.52</td>
<td>0.48–0.60</td>
<td>Perroy et al. (2007)</td>
</tr>
<tr>
<td>Precipitation altitude gradient</td>
<td>τₚ</td>
<td>% per 100 m</td>
<td>0.00</td>
<td>0.00</td>
<td>Sicart et al. (2007)</td>
</tr>
<tr>
<td>Critical temperature</td>
<td>Tₑ</td>
<td>°C</td>
<td>1.5</td>
<td>–1.0–3.0</td>
<td>Lejeune et al. (2007)</td>
</tr>
<tr>
<td>Threshold temperature for the wet season</td>
<td>Tₑw</td>
<td>°C</td>
<td>0.4</td>
<td>1.0</td>
<td>Pellicciotti et al. (2005)</td>
</tr>
<tr>
<td>Threshold temperature for the dry season</td>
<td>Tₑd</td>
<td>°C</td>
<td>1.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>ETI model</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Temperature factor for the wet season</td>
<td>TF</td>
<td>mm d⁻¹ °C⁻¹</td>
<td>1.19</td>
<td>–0.24–1.92</td>
<td>Carenzo et al. (2009)</td>
</tr>
<tr>
<td>Temperature factor for the dry season</td>
<td>TF</td>
<td>mm d⁻¹ °C⁻¹</td>
<td>0.20</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Shortwave radiation factor for the wet season</td>
<td>SRF</td>
<td>mm d⁻¹ m² W⁻¹</td>
<td>0.247</td>
<td>0.211–0.254</td>
<td>Carenzo et al. (2009)</td>
</tr>
<tr>
<td>Shortwave radiation factor for the dry season</td>
<td>SRF</td>
<td>mm d⁻¹ m² W⁻¹</td>
<td>0.171</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Characteristic albedo of fresh snow</td>
<td>α₁snow</td>
<td></td>
<td>0.80</td>
<td>0.75–0.98</td>
<td>Cuffey and Paterson (2010)</td>
</tr>
<tr>
<td>Characteristic albedo of firm</td>
<td>α₁firn</td>
<td></td>
<td>0.50</td>
<td>0.50–0.65</td>
<td>Cuffey and Paterson (2010)</td>
</tr>
<tr>
<td>Characteristic albedo of ice</td>
<td>α₂ice</td>
<td></td>
<td>0.30</td>
<td>0.30–0.46</td>
<td>Cuffey and Paterson (2010)</td>
</tr>
<tr>
<td>Time scale parameter</td>
<td>t*</td>
<td>d</td>
<td>10</td>
<td>10–21</td>
<td>Sicart et al. (2011)</td>
</tr>
<tr>
<td>Depth scale parameter</td>
<td>d*</td>
<td>mm w.e.</td>
<td>6.0</td>
<td>6.0</td>
<td>Sicart et al. (2011)</td>
</tr>
<tr>
<td>Runoff model</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Runoff coefficient for rainfall</td>
<td>cₑ</td>
<td>—</td>
<td>0.8</td>
<td>0.1–1.0</td>
<td>Ribstein et al. (1995)</td>
</tr>
<tr>
<td>Runoff coefficient for melt</td>
<td>cₛₘ</td>
<td>—</td>
<td>0.8</td>
<td>0.1–1.0</td>
<td>Ribstein et al. (1995)</td>
</tr>
<tr>
<td>Reccession coefficient</td>
<td>x</td>
<td>—</td>
<td>0.83</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Reccession coefficient</td>
<td>y</td>
<td>—</td>
<td>0.05</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lag time</td>
<td>L</td>
<td>h</td>
<td>6</td>
<td>0–24</td>
<td></td>
</tr>
</tbody>
</table>
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affected mass balance through mechanisms not well understood, given the simplicity of the model. At lower elevations, crevasses and well-developed penitents are not rare and led to overestimations of melt because of the large surface exposed to sunbeam and multiple reflections of radiation between the walls (see Corripio, 2002; Sicart et al., 2007 for further information). Finally, we found a poor agreement at six ablation stakes (6K, 4F, 7N, 8N, 13N and 14N) where MAE for the model with recalibrated parameters was higher. Most of the differences occurred in the hydrological year 2005–2006. This overestimation might be attributed, in addition to the aforementioned reasons, to the assumption of temporal stationarity of the melt parameters by unrealistically assuming a constancy of the average weather during both years.

These results suggest that the local characteristics at each ablation stake (slope, aspect, surface irregularities and topographic shading) play an important role in the melt processes that were not considered in this study. In addition, temporal and spatial transferability of the melt parameters need to be investigated. We will address these issues in a following paper.

Table II. MAE and bias in the calculated melt rates for ETI with original and recalibrated parameters at the selected sites and for the ensemble of 16 stakes

<table>
<thead>
<tr>
<th>Stake</th>
<th>Hydrological year</th>
<th>Elevation [m a.s.l.]</th>
<th>Slope and aspect</th>
<th>ETI original MAE [mm d⁻¹]</th>
<th>ETI original bias [mm d⁻¹]</th>
<th>ETI recalibrated MAE [mm d⁻¹]</th>
<th>ETI recalibrated bias [mm d⁻¹]</th>
</tr>
</thead>
<tbody>
<tr>
<td>X1</td>
<td>2004–05</td>
<td>5236</td>
<td>31° SE</td>
<td>2.9</td>
<td>2.7</td>
<td>1.8</td>
<td>0.0</td>
</tr>
<tr>
<td>2G</td>
<td>2004–05</td>
<td>5145</td>
<td>15° SE</td>
<td>2.8</td>
<td>2.8</td>
<td>1.4</td>
<td>–1.1</td>
</tr>
<tr>
<td>6K</td>
<td>2004–05</td>
<td>5116</td>
<td>10° SE</td>
<td>2.1</td>
<td>–1.2</td>
<td>4.4</td>
<td>–4.4</td>
</tr>
<tr>
<td>4F</td>
<td>2004–05</td>
<td>5103</td>
<td>6° ESE</td>
<td>1.5</td>
<td>–0.4</td>
<td>4.3</td>
<td>–4.3</td>
</tr>
<tr>
<td>9K</td>
<td>2004–05</td>
<td>5089</td>
<td>3° E</td>
<td>3.8</td>
<td>3.8</td>
<td>0.9</td>
<td>–0.2</td>
</tr>
<tr>
<td>7N</td>
<td>2005–06</td>
<td>5072</td>
<td>11° ESE</td>
<td>2.1</td>
<td>2.1</td>
<td>4.5</td>
<td>–4.5</td>
</tr>
<tr>
<td>10K</td>
<td>2004–05</td>
<td>5072</td>
<td>11° ESE</td>
<td>7.0</td>
<td>7.0</td>
<td>2.5</td>
<td>2.4</td>
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<tr>
<td>8N</td>
<td>2005–06</td>
<td>5064</td>
<td>10° ENE</td>
<td>3.7</td>
<td>3.7</td>
<td>3.6</td>
<td>–3.5</td>
</tr>
<tr>
<td>11K</td>
<td>2004–05</td>
<td>5064</td>
<td>10° ENE</td>
<td>6.3</td>
<td>6.3</td>
<td>2.5</td>
<td>1.6</td>
</tr>
<tr>
<td>9N</td>
<td>2005–06</td>
<td>5046</td>
<td>4° E</td>
<td>4.7</td>
<td>4.7</td>
<td>3.7</td>
<td>–2.6</td>
</tr>
<tr>
<td>10N</td>
<td>2005–06</td>
<td>5036</td>
<td>19° E</td>
<td>7.7</td>
<td>7.7</td>
<td>3.5</td>
<td>0.5</td>
</tr>
<tr>
<td>12K</td>
<td>2004–05</td>
<td>5024</td>
<td>15° E</td>
<td>9.3</td>
<td>9.3</td>
<td>3.5</td>
<td>3.5</td>
</tr>
<tr>
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<td>2005–06</td>
<td>5018</td>
<td>13° ENE</td>
<td>4.5</td>
<td>4.5</td>
<td>3.9</td>
<td>–2.8</td>
</tr>
<tr>
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<td>2005–06</td>
<td>5004</td>
<td>22° ENE</td>
<td>5.4</td>
<td>4.9</td>
<td>4.8</td>
<td>–1.0</td>
</tr>
<tr>
<td>13N</td>
<td>2005–06</td>
<td>4971</td>
<td>10° ENE</td>
<td>4.2</td>
<td>1.1</td>
<td>5.4</td>
<td>–5.1</td>
</tr>
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<td>2005–06</td>
<td>4957</td>
<td>22° ESE</td>
<td>4.1</td>
<td>2.3</td>
<td>4.9</td>
<td>–4.0</td>
</tr>
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</table>

Ensemble mean 4.5 3.8 3.5 –1.5

Figure 1. Observed and simulated cumulative mass balance at six selected ablation stakes for the enhanced temperature-index model (ETI) with original and recalibrated parameters. The left and right panels show results for hydrological years 2004–2005 and 2005–2006, respectively.
Figure 2 illustrates the annual variations of mass balance with altitude. Observed and simulated annual mass balance show similar patterns in the ablation zone for the two hydrological years, but differ in the accumulation zone for year 2005–2006. The vertical variability of mass balance is mainly controlled by precipitation, the temperature lapse rate, and the critical temperature. Sensitivity analysis (see below) shows that both the critical temperature and a poor estimation of the actual precipitation depths in the accumulation zone are responsible for this difference.

Table III summarizes the observed and simulated mass balance parameters for the two hydrological years of our study. Simulated glacier-wide mass balance for years 2004–2005 and 2005–2006 differed from the observed values by –62 and –104 mm w.e., respectively, less than the overall error on the glaciological mass balance of ±400 mm w.e estimated by Sicart et al. (2007).

ELA was predicted with an error of approximately ±70 m a.s.l. which caused an over (under) estimation of AAR in 2004–2005 (2005–2006). This error resulted in an increase (decrease) in net mass balance of +71 (–67) mm w.e.

To ensure that the models were able to predict glacier mass balance as well as streamflow, we calculated runoff with the optimum parameter set. Additionally, we tested the performance of ETI with original parameters and DDM, which uses a single DDF to compute melt. The value of the latter was estimated based on the findings of Ribstein et al. (1995).

Figure 3a illustrates the resulting monthly hydrographs for DDM and ETI. Here, $R^2$ of DDM reached 0.330 while ETI with original parameters underestimated discharge with $R^2 = 0.085$. The best simulation was obtained for ETI with recalibrated parameters for which $R^2$ reached 0.772.

Figure 3b shows observed and simulated daily discharge for ETI with recalibrated parameters. These data reveal another limitation of ETI: during the core of the wet season, when clouds cover the glacier for most part of the time, thus reducing the contribution of solar radiation, ETI underestimates discharge. An example of this outcome was observed in December–March of 2005–2006. A similar observation was reported by Carenzo et al. (2009) and they concluded that the lower model performance cannot be overcome by recalibration, but it is an intrinsic characteristic of the model, which works less well for overcast conditions. This conclusion may hold true for other situations when solar radiation

<table>
<thead>
<tr>
<th>Year</th>
<th>$B_n$ (mm w.e.)</th>
<th>ELA (m a.s.l.)</th>
<th>AAR (%)</th>
<th>$Q$ (mm w.e.)</th>
<th>$P$ (mm w.e.)</th>
<th>$s$ (mm w.e.)</th>
<th>$M$ (mm w.e.)</th>
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<tbody>
<tr>
<td></td>
<td>Obs</td>
<td>Sim</td>
<td>Obs</td>
<td>Sim</td>
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</tbody>
</table>
Table IV. Sensitivity of modelled net mass balance to changes in model parameters. ΔBn represents the variation in modelled annual glacier mass balance

<table>
<thead>
<tr>
<th>Parameter change</th>
<th>ΔBn (mm w.e.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>γ = 0.48°C per 100 m</td>
<td>−11</td>
</tr>
<tr>
<td>γ = 0.60°C per 100 m</td>
<td>49</td>
</tr>
<tr>
<td>γp = 3% per 100 m</td>
<td>271</td>
</tr>
<tr>
<td>Tr = −1.0°C</td>
<td>−681</td>
</tr>
<tr>
<td>Tr = 3.0°C</td>
<td>15</td>
</tr>
<tr>
<td>TF = −0.24 mm d⁻¹ °C⁻¹</td>
<td>60</td>
</tr>
<tr>
<td>TF = −1.92 mm d⁻¹ °C⁻¹</td>
<td>−72</td>
</tr>
<tr>
<td>SRF = 0.21 mm d⁻¹ m⁻² W⁻¹</td>
<td>58</td>
</tr>
<tr>
<td>SRF = 0.25 mm d⁻¹ m⁻² W⁻¹</td>
<td>−107</td>
</tr>
<tr>
<td>αfrsnow = 0.92</td>
<td>132</td>
</tr>
<tr>
<td>αfr = 0.65</td>
<td>52</td>
</tr>
<tr>
<td>αcry = 0.20</td>
<td>−63</td>
</tr>
</tbody>
</table>

is reduced. In fact, Table III shows that ETI underestimated discharge by 93 and 232 mm w.e. for years 2004–2005 and 2005–2006. These differences occurred mainly in the wet season.

Albedo was also used as a validation variable: Supplement Figure S3 compares the simulated albedo with observations made at the AWS at 5050 m a.s.l. The simulations agreed well with observations with a correlation coefficient of 0.64 (n = 730 days). Discrepancies between estimated and measured albedo were due to the incorrect estimation of precipitation volume, phase and timing caused by the need to extrapolate precipitation events recorded at the base station to the glacier based station. Using observed albedo slightly improved the performance of ETI model (MAE = 3.2) but overestimated net mass balance by 256 mm w.e. (2004–2005) and 96 mm w.e. (2005–2006), and reduced R² to 0.516.

The sensitivity of the model to parameters and input data is presented in Supplement Text S6. Table IV shows the results of the sensitivity analysis for Zongo glacier. The results show that the model is not very sensitive to the temperature lapse rate (γ) when the range in γ is small and the mean value is used. On the other hand, considering a precipitation vertical gradient of 3% per 100 m significantly increases net mass balance and γp is therefore an important factor. A decrease in the critical temperature decreases the proportion of precipitation falling as snowfall, which, in turn, causes the ice surface to be frequently exposed, enhances melt, and underestimates net mass balance. In contrast, an increase in the critical temperature does not have an important effect on the results. TF and SRF have important effects on the estimated net balance and are therefore the primary candidates for adjustment. Finally, using the maximum recommended values for the characteristic albedos for snow, firn and ice (Cuffey and Paterson, 2010) showed that the model is very sensitive to these parameters and so a careful consideration of their values is needed before the model is applied.

The climate sensitivity of the glacier net balance was quantified by the seasonal sensitivity characteristic (SSC, Oerlemans and Reichert, 2000). The model was run for temperature perturbations of +1°C and precipitation perturbations of +10%. The results are illustrated in Figure 4 for hydrological year 2005–2006. SSCs for Zongo glacier were different from those determined for New Zealand, Norway, Austria, Canada, Kirghizstan (all shown in Oerlemans and Reichert, 2000) and Argentina (Stuefer et al., 2007). The peculiarity of this site is that both temperature and precipitation effects dominate in the wet season, with significant SSC values of up to 0.25 m month⁻¹ for temperature perturbations of 1°C and a maximum precipitation SSC of roughly 0.05 m month⁻¹ in January. Temperature was the most important factor throughout the period of analysis, even for winter when temperature SSC reached 0.05 m month⁻¹. Precipitation did not have any effect in the dry season.

**SUMMARY AND CONCLUSIONS**

We evaluated the feasibility of modelling melt, discharge, and glacier mass balance for the tropical Zongo glacier by using an enhanced temperature-index model (ETI). The model explicitly included the shortwave radiation balance on one term and represented net longwave radiation and the turbulent fluxes by a temperature-dependent term (see Equation 3), thus having a stronger physical meaning than the standard degree-day method (DDM). This study was the first to explore the applicability of ETI in a tropical glacier over two complete hydrological years that included a wet and a dry season with clearly differentiated climatic characteristics.

Surface mass balance was estimated using the original parameter values of ETI and a recalibrated parameter set on Zongo, and the results were compared with monthly observations. Then, runoff was estimated using the two versions of ETI and DDM. The optimum parameter set for the wet season was similar to those determined for Haut Glacier d’Arolla and Juncal Norte. However, during the dry season, both TF and SRF were significantly reduced because of the combined effect of two processes: the absence of clouds (low humidity, Supplement Figure S2) and the large energy consumption by sublimation (Supplement Text S3). The threshold temperature (Tr) remained an empirical parameter...
of the model, which needs to be further investigated in relation with the surface energy balance.

ETI with the recalibrated parameter set showed a better performance for monthly predictions of discharge and mass balance. The results showed that mass balance was highly dependent on the local characteristics at each site. Additional comparisons of ETI outputs with observations of albedo, net mass balance over the whole glacier, the variation of annual mass balance with elevation, the equilibrium line altitude, and the accumulation area ratio, demonstrated the robustness of the model. However, a major drawback of ETI was that it worked less well for overcast conditions. The model could be further improved using the findings of Giesen and Oerlemans (2012). In that study, they concluded that in tropical glaciers an energy flux relation dependent on both air temperature and relative humidity would be more appropriate. Using this kind of function and an algorithm to discriminate between overcast and clear-sky conditions might improve the limitation of ETI.

We can conclude that the ETI model is applicable to the conditions found in the Zongo glacier and that it may be applicable to other tropical glaciers as well if careful consideration of model parameters is undertaken. We give an optimal parameter set that can be used as a starting point for other studies.

ACKNOWLEDGEMENTS

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SUPPLEMENTS

Text S1. Data collection and extrapolation methods
Text S2. Degree-day approach
Text S3. Estimation of sublimation
Text S4. Runoff routing
Text S5. Evaluation of model performance
Text S6. Sensitivity analysis
Figure S1. Map of Zongo drainage basin
Figure S2. Climatic conditions during hydrological years from 2004 to 2006
Figure S3. Observed and simulated average albedo at 5050 m a.s.l.
Figure S4. Comparison of annual precipitation, sublimation and melt

REFERENCES

MELT MODELLING ON TROPICAL GLACIERS


