Wintertime high-altitude surface energy balance of a Bolivian glacier, Illimani, 6340 m above sea level

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[1] The objective of this study is to evaluate the surface energy balance (SEB) of a cold, high-altitude tropical glacier, Illimani (16°39′S; 67°47′W, 6340 m above sea level (asl)), where a 137 m ice core was drilled down to the bedrock in June 1999. During the dry austral winter, tropical glaciers are known to experience strong sublimation, which may be responsible for snow composition changes through postdepositional processes. In order to help toward the interpretation of this climatic archive, SEB experiments were carried out in 1999, 2001, and 2002, during the dry season (mostly clear and cold atmosphere, strong westerly winds). The daily net all-wave radiation is usually negative during this austral dry winter because of the highly reflective snow surface and because of reduced incoming long-wave radiation due to a low cloudiness compared to outgoing long-wave radiation. The turbulent heat fluxes were evaluated using the bulk aerodynamic approach, including stability correction. The roughness parameters are deduced from direct sublimation measurements and serve as calibration parameters. The sensible heat flux strongly heats the surface at night but changes to negative values during daytime unstable conditions (between 1000 and 1600 LT). The latent heat flux is always negative, which means that the surface loses mass through sublimation, particularly in the daytime (sublimation rates are -1.2 mm w.e. d^{-1}, -0.7 mm w.e. d^{-1}, and -0.8 mm w.e. d^{-1} during the 2001, 2002, and 1999 measuring periods, respectively, where w.e. is water equivalent). The winter SEB of this high-altitude cold tropical glacier is comparable to the summer SEB over snow surfaces of the intermediate slopes of Antarctica. *INDEX TERMS:* 0342 Atmospheric Composition and Structure: Middle atmosphere—energy deposition; 0358 Atmospheric Composition and Structure: Thermosphere—energy deposition; 3307 Meteorology and Atmospheric Dynamics: Boundary layer processes; *KEYWORDS:* tropical glaciers, high altitude, sublimation, turbulent fluxes, net all-wave radiation, energy balance

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1. Introduction

[2] Little is known about the surface energy balance (SEB) of high-altitude glaciers especially in the tropics, although it is of prime importance for the interpretation of the ice cores from the tropical Andes, which contain important climatic information (1983: Quelccaya Ice Cap (5670 m asl, Peru) [*Hastenrath*, 1978; *Thompson et al.*, 1984, 1985, 1986]; 1993: Huascaran (6048 m asl, Peru) [*Thompson et al.*, 1995]; 1997: Sajama (6540 m asl, Bolivia) [*Thompson et al.*, 1998]; 1999: Illimani (6340 m asl, Bolivia) [*Hoffmann et al.*, 2003]; 2000: Chimborazo (6250 m asl, Ecuador)). Analysis of these ice cores suffer from a lack of knowledge of the SEB of these high-altitude sites especially because sublimation is known to be very high on the lower part of tropical glaciers [e.g., *Wagnon et al.*, 1999] and may strongly affect the chemical and isotopic compositions of the snow and ice (postdepositional processes). The climatic signal contained in these cores may therefore be altered by local sublimation, which makes the climatic interpretation of these archives somewhat hazardous as shown by *Ginot et al.* [2001] and *Stichler et al.* [2001] on Cerro Tapado Ice Cap (5536 m asl, Chilean Andes). Thus, two SEB experiments were conducted on the drilling site of Illimani (6340 m asl, 16°39′S; 67°47′W) during 6 days of May 2001 and during 29 days of May–June 2002 together with bi-daily snow surface samplings in order to better understand how much sublimation influences the chemical and isotopic compositions of surface snow. In this paper, we mainly focus on the results of these SEB experiments conducted between May 21 and 26, 2001 and between May 5 and June 4, 2002. During the 1999 drilling expedition on Illimani, a 2 week SEB experiment was also

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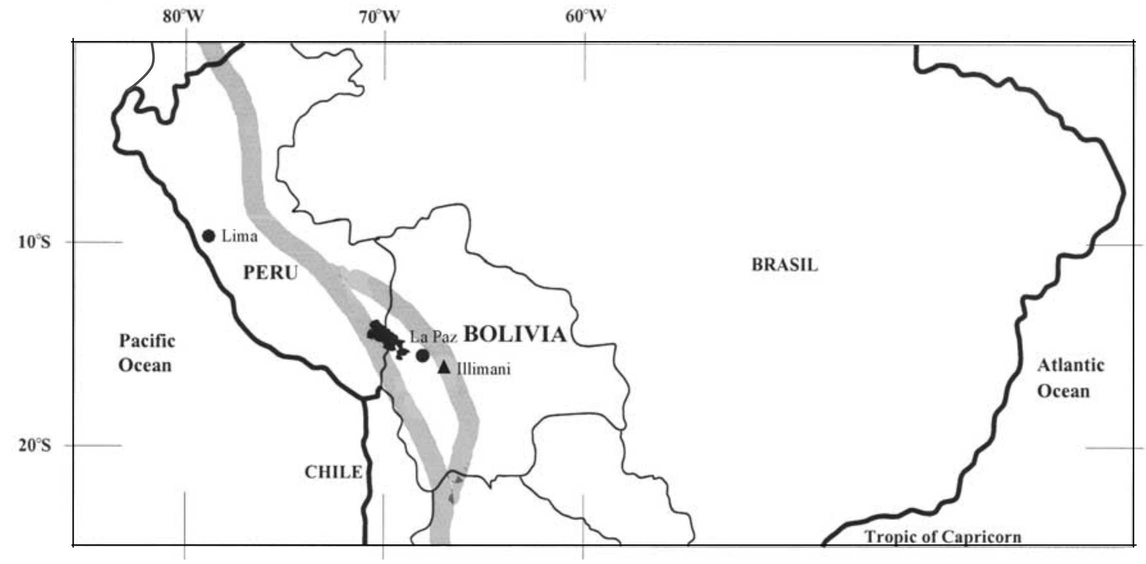

Figure 1. Simplified map of South America showing the location of Illimani, Cordillera Real, Bolivian Andes.

carried out between May 24 and June 7, the results of which are presented as a comparison.

2. Location and Measurement Program

[3] Illimani is a massif with a 10 km long ridge above 6000 m asl, located at the extreme southern end of the Cordillera Real, approximately 30 km south-east of La Paz, the capital of Bolivia (Figure 1). In May–June 1999, a drilling expedition was carried out at 6340 m asl on a flat and large pass between the main summit to the south (6450 m asl) and the central summit to the northeast (6380 m asl). A 137 m ice core was collected and is still under analysis [*Hoffmann et al.*, 2003]. At this high-altitude site, based on the measurements on the ice core and on a field trip conducted during the warm wet season on February 18–20, 2002 during which air temperature never exceeded −3°C, it seems that melting never occurs throughout the year. Between May 21, 2001, 1710 LT and May 26, 2001, 1345 LT, and also between May 5, 2002, 1230 LT and June 4, 2002, 0130 LT, an automatic weather station (AWS) collected data at the site of the 1999 drilling expedition on Illimani (16°39′S; 67°47′W, 6340 m asl). A picture of this AWS is shown in Figure 2. This AWS was installed on a relatively flat surface and was in sunlight as soon as the Sun rose at 0645 LT and was in shade at 1745 LT around half an hour before the night.

[4] The measurements were made within the surface boundary layer using sensors sold by Campbell Scientific (United Kingdom). Wind speed, air temperature and humidity were recorded as half-hourly means over 15-s time intervals. Instantaneous values of wind direction were collected every half hour. Incident and reflected short-wave radiation, incoming and outgoing long-wave radiation were recorded as half-hourly means over 15-s time intervals. In May 2001, 25 m east of the AWS, 5 Cu-Cst thermocouples were installed respectively at 5 cm, 10 cm, 15 cm, 20 cm and 25 cm into the snow, below the surface using a sixth thermocouple in an isothermal liquid water-ice bath as a 0°C reference thermometer. Table 1 gives a list of the sensors with their characteristics and their height above the surface. In 2001, the Vaisala hygro-thermometer was not artificially ventilated but the wind was always strong enough to maintain an efficient natural ventilation and to prevent measurement errors due to radiation. In 2002, the Vaisala hygro-thermometer was artificially ventilated in the daytime using a solar panel. Due to power supply problems, night values (from 0100 to 0700 LT) are missing between May 5 and 13, 2002.

[5] In addition to the data collected by the AWS, direct qualitative observations of the meteorological conditions (wind regime, weather type, cloudiness, cloud type) and of the surface state were made on the field site during the

Figure 2. Picture of the AWS installed on the field site on Illimani at 6340 m asl (May 25, 2001). Photo: P. Wagnon.

Table 1. List of Different Sensors With Their Specifications, Installed on the Weather Station at 6340 m Above Sea Level in May 2001, May–June 2002, and May–June 1999

Quantity	Sensor Type (Height) in 2001 and 2002	Sensor Type (Height) in 1999	Accuracy According to the Manufacturer
Air temperature, °C	Vaisala (0.9 m)	Vaisala (1.9 m)	±0.2°C
Relative humidity, %	Vaisala (0.9 m)	Vaisala (1.9 m)	±2 %
Wind speed, m s^{-1}	Young (2.4/2.5 m)[a]	Young (2.5 m)	±0.3 m s^{-1}
Wind direction, deg	Young (2.4/2.5 m)[a]	Young (2.5 m)	±3 deg
Incident short-wave radiation, W m^{-2}	Kipp&Zonen CM3 (0.85 m) $0.305 < \lambda < 2.8\mu m$	Campbell sp1110 (1 m) $0.35 < \lambda < 1.1\mu m$	±3 %
Reflected short-wave radiation, W m^{-2}	Kipp&Zonen CM3 (0.85 m) $0.305 < \lambda < 2.8\mu m$	Campbell sp1110 (1 m) $0.35 < \lambda < 1.1\mu m$	±3 %
Incoming long-wave radiation, W m^{-2}	Kipp&Zonen CG3 (0.85 m) $5 < \lambda < 50\ \mu m$	–	±3 %
Outgoing long-wave radiation, W m^{-2}	Kipp&Zonen CG3 (0.85 m) $5 < \lambda < 50\mu m$	–	±3 %
Net all-wave radiation, W m^{-2}	Calculated from the 4 precedent quantities	Campbell Q7 (1 m) $0.25 < \lambda < 60\mu m$	from ±3 % to ±10 %[b]
Snow/ice temperature, °C	Cu-Cst thermocouples (−5, −10, −15, −20, −25 cm)	–	±0.3°C

[a]Height was 2.4 m in 2001, 2.5 m in 2002.
[b]Depending on the horizontality of the sensor.

entire 2001 measuring period and between May 5 and 16, 2002. Sublimation was measured using 7 "poor-man lysimeters" (translucent round plastic pails of 360 cm^2 surface and 10 cm deep, buried into the snow and filled with snow in order to reconstruct natural surfaces as well as possible) weighed with an accuracy of ±1 g (corresponding to ± 0.03 mm water equivalent (w.e.)) at around 0800 and 1700 LT every day. An ablation stake was regularly measured with an accuracy of ±1 mm.

[6] During the 1999 drilling expedition on Illimani, a previous AWS was run at the same site between May 24, 1400 LT and June 7, 1000 LT. This AWS was slightly different from the one installed in May 2001 especially because the net radiometer did not provide each term of the surface radiative balance. Table 1 gives a list of the sensors with their specifications. Qualitative observations of the surface state were made during this drilling expedition. Daily sublimation was measured directly between May 26 and June 3 using a network of 4 "poor-man lysimeters."

3. Climatic Conditions

[7] Illimani belongs to the outer tropics, as defined by *Kaser* [2001] characterized by one dry and one wet season throughout the year. On one hand, the winter dry season (May–August) is produced by the northward displacement of the mid- and upper-tropospheric westerlies, which in turn prevent moisture influx from the east. On the other hand, the rainy season (October–March) is a result of the seasonal expansion of the equatorial easterlies in the upper troposphere, allowing near-surface moisture influx from the east due to turbulent entrainment of easterly momentum over the Andean ridge [*Garreaud*, 1999, 2000; *Vuille et al.*, 1998, 2000].

[8] Figures 3 and 4 show some meteorological quantities of the 2001 and 2002 measuring periods respectively and Table 2 gives the mean values of these quantities. The weather during the measuring period of May 2001 consisted of steady, dry and cold conditions with strong winds from the west. The 2002 measuring period was also characterized by strong westerly winds (except on May 8, 23, 24 and 25) but experienced regular weak snowfalls and a higher relative humidity. The scatter in the wind direction is all the larger as the wind speed is low and because wind direction values reported on Figures 3 and 4 are instantaneous records and not means. Nevertheless, field observations report a wind regime always from the west even during calm weather.

[9] Actually, due to the relatively high frequency of precipitation at the beginning of the period, May 2002 can be considered as a transition month between the wet season and the dry one, with strong westerly winds usual of the dry season but still with a high cloudiness and precipitation, whereas the days of May 2001 characterized by low temperature, dry air, clear sky most of the time and strong winds from the west are typical of the dry season at high elevation sites [*Hardy et al.*, 1998; *Vuille*, 1999]. Thus, although short, both measuring periods can give a good insight into the climatic conditions that prevail at the beginning of the dry season on Illimani, which is confirmed by comparison with the 2 week measurement period of May–June 1999 (Figure 5). Indeed, during this 1999 period, the weather was also most of the time cold, dry, cloudless with strong winds from the west except for 5 days (May 29 and 30 experienced light winds and an extremely low humidity and May 25, June 4 and 5 were overcast with precipitation during the night between June 4 and 5). Due to the reduced wind speed, the wind directions recorded on May 29 and 30 are scattered over 360 degrees but the wind regime was still from the west. The extremely low values of relative humidity of May 29 and 30 are probably measuring artifacts related to the absence of artificial or natural ventilation (light winds) of the Vaisala sensor, leading to an overestimation of air temperature and an underestimation of relative humidity. During this 1999 period, we can distinguish thus two weather regimes according to the cloudiness: 1. a typical weather regime of the dry season with westerly strong winds at altitude and a clear and dry atmosphere (like during the 2001 measuring period) with sporadic situations with light winds (May 29 and 30) and 2.

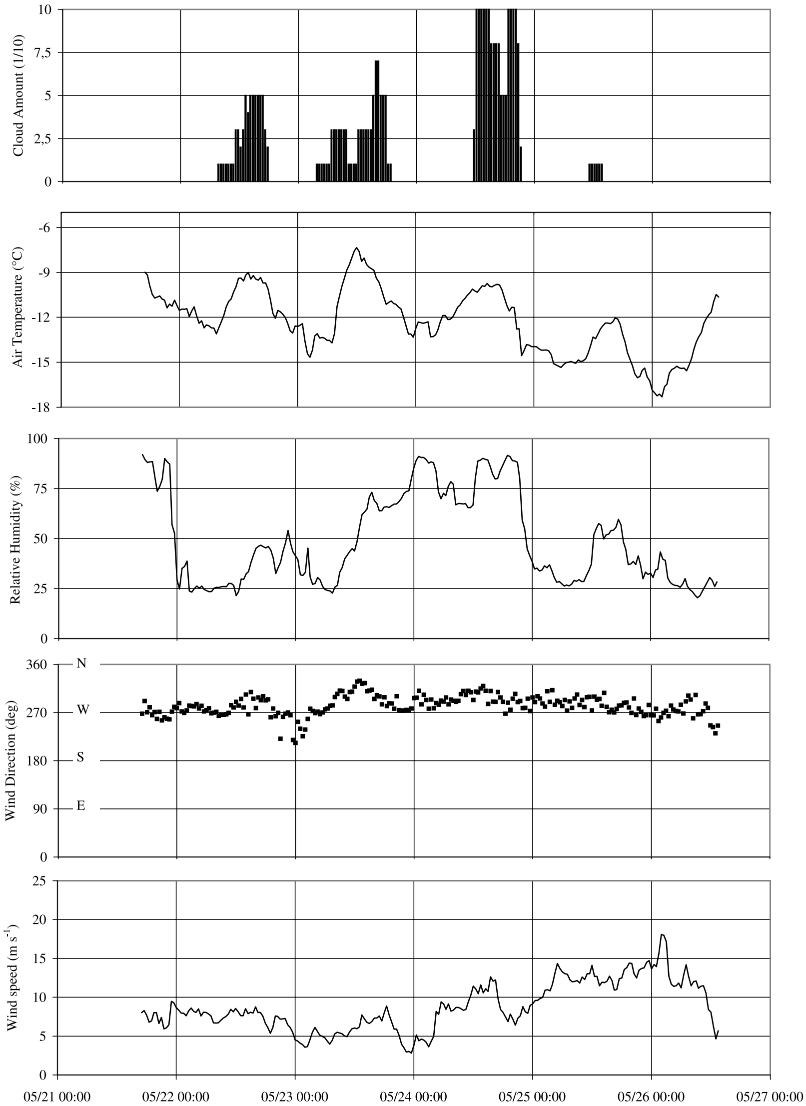

Figure 3. Half-hourly mean values of total cloud amount, air temperature (0.9 m), relative humidity (0.9 m), wind direction and speed (2.4 m) on Illimani, 6340 m asl, between May 21 and May 26, 2001. No precipitation has been observed during this period.

cold fronts coming with strong westerly winds bringing thick clouds and sometimes precipitation related to outbursts of cold polar air as shown by *Vuille* [1999]. Table 2 shows the mean values of the meteorological quantities for the entire 1999 measuring period and for both weather regimes.

4. Energy Balance Study

[10] For a non-melting cold surface, the SEB equation can generally be written as (fluxes toward the surface positive) [e.g., *Oke*, 1987]:

$$R + H + LE + G = 0 \qquad \left(\text{in W m}^{-2}\right) \tag{1}$$

where R is net all-wave radiation, H is the turbulent sensible heat flux, LE is the turbulent latent heat flux and G is the upward energy flux inside the snow/ice. Melting was not observed at the field site and is therefore not dealt with. Since precipitation was nil in May 2001 or weak in May–June 2002 or on June 4–5, 1999, the heat supplied by precipitation remains insignificant.

4.1. Net All-Wave Radiation

[11] The net radiation is the balance of the incident and reflected short-wave radiations and the incoming and outgoing long-wave radiations (fluxes toward the surface positive).

$$R = S\downarrow - S\uparrow + L\downarrow - L\uparrow = S\downarrow (1 - \alpha) + L\downarrow - L\uparrow \quad \left(\text{in W m}^{-2}\right) \tag{2}$$

where $S\downarrow$ is the incident short-wave radiation, $S\uparrow$ is the reflected short-wave radiation ($S = S\downarrow - S\uparrow$ is the net short-wave radiation), α is the short-wave albedo of the snow surface, $L\downarrow$ is the incoming long-wave radiation, and $L\uparrow$ is

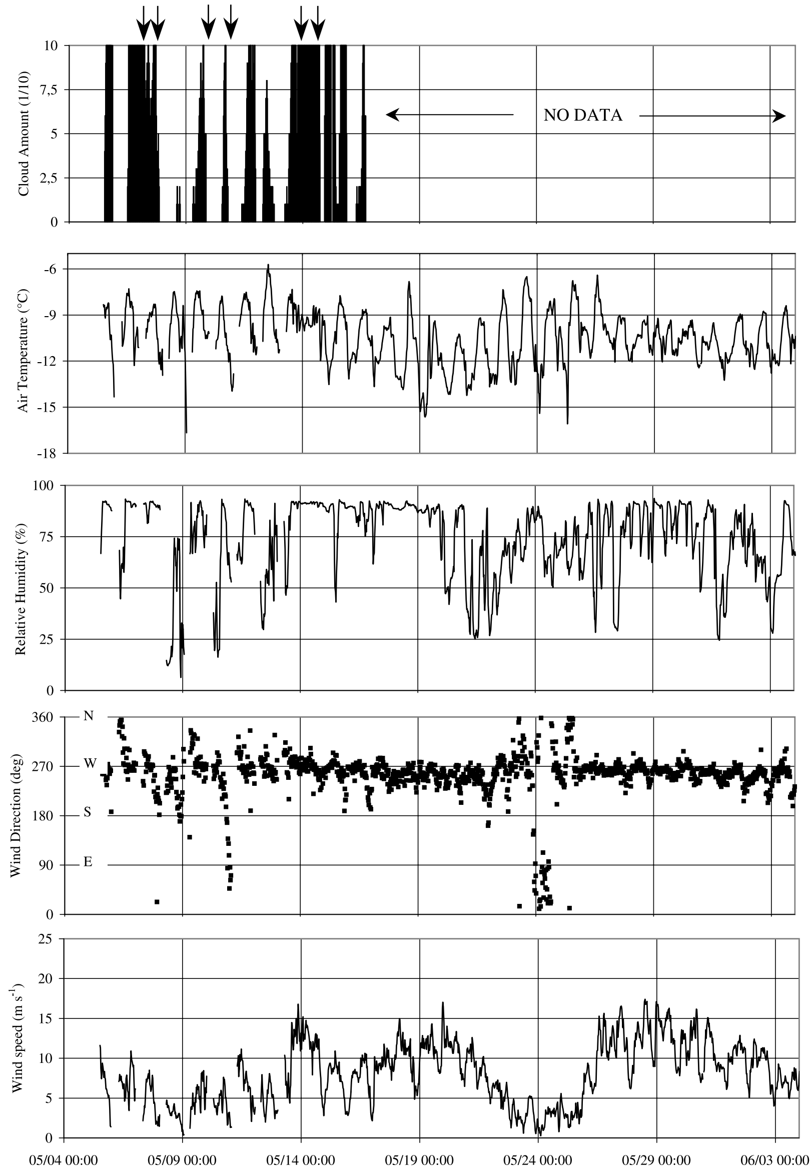

Figure 4. Half-hourly mean values of total cloud amount (until May 16, 1600 LT only), air temperature (0.9 m), relative humidity (0.9 m), wind direction and speed (2.5 m) on Illimani, 6340 m asl, between May 5 and June 4, 2002. Arrows stand for snowfalls.

the outgoing long-wave radiation ($L = L\downarrow - L\uparrow$ is the net long-wave radiation). In May 2001 and May–June 2002, the four terms of the surface radiative balance were measured directly on the field site by a Kipp & Zonen net radiometer which is composed of 2 pyranometers (short waves: $0.305 < \lambda < 2.8$ μm) and 2 pyrgeometers (long waves: $5 < \lambda < 50$ μm). Because of the strong wind, which maintained a natural ventilation of the sensor, the radiative heating of the sensor was sharply reduced and therefore, we have not applied any correction to the long-wave radiation data as suggested by *Obleitner and de Wolde* [1999]. In May–June 1999, incoming and outgoing longwave radiations were not available from the AWS; only $S\downarrow$, $S\uparrow$ and R were measured directly in the field by 2 pyranometers ($0.35 < \lambda < 1.1$ μm) and a net radiometer ($0.25 < \lambda < 60$ μm) respectively. Usually, in the early morning between 0645 and 0900 LT, sun beams were very

Table 2. Mean Meteorological Data Evaluated From Half-Hourly Values

	T, °C	RH, %	u, m s^{-1}	Direction	Cloudiness (1/10)
2001[a]					
Mean (May 22–25)	−12.1	50	8.4	W	1.6
2002[b]					
Mean (May 5 to June 3)[c]	−10.4	74	8.0	W	4.1[d]
1999[e]					
Regime 1[f]	−10.3	44	6.7	W	–
Regime 2[g]	−10.6	84	11.2	W	–
Mean	−10.4	53	7.8	W	–

[a]For 2001: T (0.9 m), RH (0.9 m), u (2.4 m), and direction (2.4 m).
[b]For 2002: T (0.9 m), RH (0.9 m), u (2.5 m), and direction (2.5 m).
[c]Night values (0100–0700 LT) are missing from May 5 to 13.
[d]Mean value from May 5 to 16.
[e]For 1999: T (1.9 m), RH (1.9 m), u (2.5 m), and direction (2.5 m).
[f]Regime 1: May 26–June 3 and June 6, 1999 (clear windy weather with occasional calm periods).
[g]Regime 2: May 25 and June 4–5, 1999 (cloudy sometimes snowy windy weather).

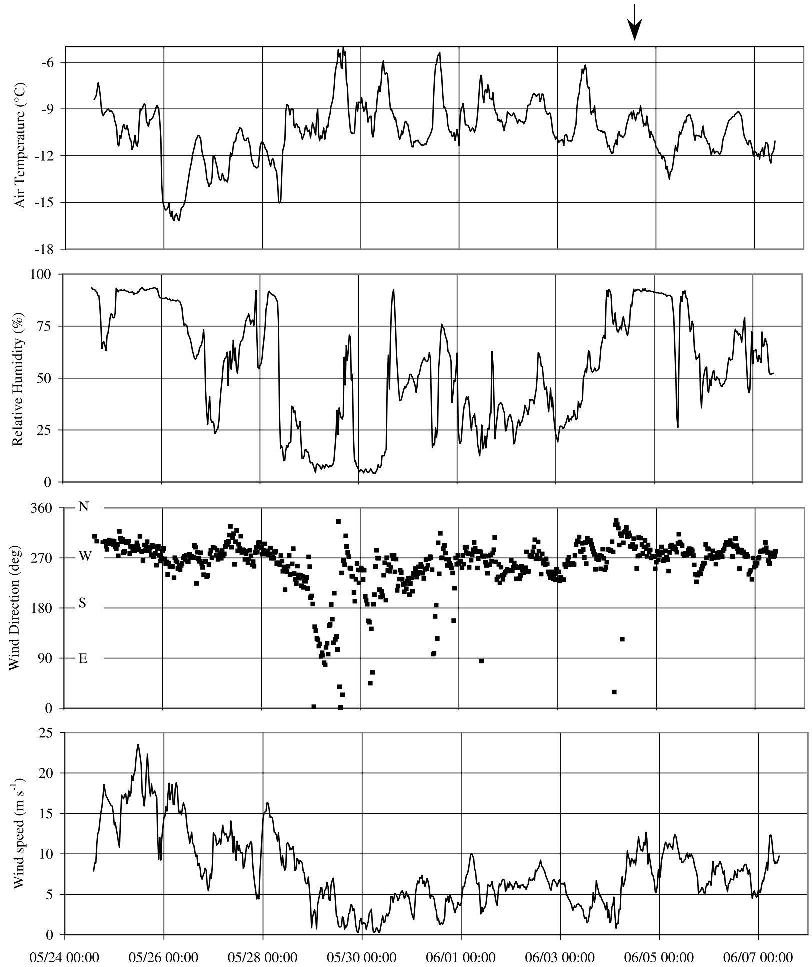

Figure 5. Half-hourly mean values of air temperature (1.9 m), relative humidity (1.9 m), wind direction and speed (2.5 m) on Illimani, 6340 m asl, between May 24 and June 7, 1999. The arrow stands for the snowfall of the night June 4–5.

inclined over the surface, values of $S\uparrow$ were small and consequently errors on pyranometers were large leading to $S\uparrow$ greater than $S\downarrow$. In those cases where the albedo measurements were higher than 0.9, we applied a correction to $S\downarrow$ considering a maximum snow albedo of 0.9: $S\downarrow = S\uparrow/0.9$. Anyway, this correction always concerned early morning values of incident short-wave radiation, which are very small compared to daytime values.

4.2. Turbulent Fluxes

4.2.1. Bulk Method

[12] The turbulent heat fluxes are calculated using the bulk aerodynamic approach, including stability correction. This method is usually used for practical purposes because it allows us to estimate the turbulent heat fluxes from one level of measurement [*Arck and Scherer*, 2002]. Moreover, *Denby and Greuell* [2000] showed that this method is usually the most appropriate over snow surfaces with a maximum of katabatic wind. *Arck and Scherer* [2002] also showed that this method yields the best correlation to the eddy-covariance measurements, particularly as long as air temperature is negative. In this approach, a constant gradient is assumed between the level of measurement and surface and thus, surface values have to be evaluated. The stability of the surface layer is described by the bulk Richardson number Ri_b, which relates the relative effects of buoyancy to mechanical forces [e.g., *Brutsaert*, 1982; *Moore*, 1983; *Oke*, 1987]:

$$Ri_b = \frac{g\frac{(T-T_s)}{(z-z_{0m})}}{T\left(\frac{u}{z-z_{0m}}\right)^2} = \frac{g(T-T_s)(z-z_{0m})}{Tu^2} \quad (3)$$

where T and u are respectively the mean values of air temperature (in K) and horizontal wind speed (in m s^{-1}) at

Table 3. Error on the Results of the Turbulent Fluxes as a Function of the Surface Temperature T_s for the 1999 Measuring Period

	$T_s - 2$°C	$T_s - 1$°C	T_s	$T_s + 1$°C	$T_s + 2$°C
ΔH	+45%[a]	+23%	0%	−23%	−47%
ΔLE	−97%	−51%	0%	+57%	+120%

[a] $\Delta H = [H(T_s - 2) - H(T_s)]/H(T_s)$, where $H(T_s)$ and $H(T_s - 2)$ are mean values for the 1999 entire measuring period. Idem for ΔLE.

Table 4. Values of z_0 During the Different Measuring Periods

Period	z_0, mm	Surface State
May 21–24, 01	0.5	Smooth surface of fresh crusty clean snow
May 25, 01	From 0.5 to 1[a]	Development of zastrugis due to strong wind. The surface becomes a little bit rougher
May 26, 01	1	Crusty smooth snow with small zastrugis (10–20 cm high) spread out irregularly around the AWS
May 5 to June 4, 02	0.3	Smooth surface of fresh and clean snow
May 24 to June 7, 99	1.5	Irregular zastrugis (20–30 cm) around the AWS

[a]Linear evolution of z_0 from 0.5 to 1 mm.

the level of measurement z. g is the acceleration of gravity (g = 9.81 m s^{-2}), T_s is the surface temperature (in K) and z_{0m} is the surface roughness parameter for momentum (in m). By definition, z_{0m} is the height where the horizontal component of the wind speed is zero, $u(z_{0m}) = 0$. Ri_b is positive in a stable atmosphere. Assuming that local gradients of mean horizontal wind speed u, mean temperature T and mean specific humidity q are equal to the finite differences between the measurement level and the surface, it is possible to give analytical expressions for the turbulent fluxes [e.g., *Oke*, 1987]:

$$H = \rho \frac{C_p k^2 u(T - T_s)}{\left(\ln \frac{z}{z_{0m}}\right)\left(\ln \frac{z}{z_{0T}}\right)} (\Phi_m \Phi_h)^{-1} \quad (4)$$

$$LE = \rho \frac{L_s k^2 u(q - q_s)}{\left(\ln \frac{z}{z_{0m}}\right)\left(\ln \frac{z}{z_{0q}}\right)} (\Phi_m \Phi_v)^{-1} \quad (5)$$

where q_s is the mean specific humidity at the surface (in g kg^{-1}), ρ = 0.55 kg m^{-3} is the air density at 6340 m asl (440 hPa), C_P is the specific heat capacity for air at constant pressure ($C_p = C_{pd}$ (1 + 0.84q) with C_{pd} = 1005 J kg^{-1} K^{-1}, the specific heat capacity for dry air at constant pressure), L_s is the latent heat of sublimation of snow or ice (L_s = 2.834 10^6 J kg^{-1}), and k is the von Karman constant (k = 0.4). z_{0m}, z_{0T} and z_{0q} are the surface roughness parameters for momentum, temperature and humidity respectively. The nondimensional stability functions for momentum (Φ_m), for heat (Φ_h) and moisture (Φ_v) may be expressed in terms of Ri_b:

$$\text{For } Ri_b \text{ positive (stable)}: (\Phi_m, \Phi_h)^{-1} = (\Phi_m, \Phi_v)^{-1} = (1 - 5Ri_b)^2 \quad (6)$$

$$\text{For } Ri_b \text{ negative (stable)}: (\Phi_m, \Phi_h)^{-1} = (\Phi_m, \Phi_v)^{-1} = (1 - 16Ri_b)^{0.75} \quad (7)$$

[13] To apply the bulk aerodynamic approach for the measurements on Illimani of May 2001 and May–June 2002, the surface temperature T_s was derived from the outgoing long-wave radiation $L\uparrow$ using the Stefan-Boltzmann equation assuming that the emissivity of the snow is unity, $L\uparrow = \sigma T_s^4$ with σ = 5.67 10^{-8} W m^{-2} K^{-4} [e.g., *Brugman*, 1991; *Bintanja and Van den Broeke*, 1995]. Saturation was assumed for the surface temperature to calculate the surface specific humidity. During the 1999 measuring period, since the outgoing long-wave radiation was not available from the AWS, it was not possible to calculate the surface temperature of the field site. Therefore, since most of the time the weather regime was similar in May–June 1999 and in May 2001, we arbitrary applied the mean diurnal cycle of T_s calculated from May 22 to May 25, 2001 (4 entire days) to the 1999 measuring period. This assumption is very coarse but it was the only way to calculate the turbulent fluxes from the measurements performed during the 1999 drilling expedition. An error analysis is given in Table 3. H and LE are very sensitive to T_s but since the turbulent fluxes are calibrated using the roughness parameters (see the following section), the error induced by this assumption on T_s might be compensated by the calibration. Anyway, the turbulent fluxes calculated during the 1999 measuring period are likely to be less accurate and are presented just for the purpose of comparison.

[14] The bulk method was applied between the surface and the level of measurement of T and q. Since u was measured at

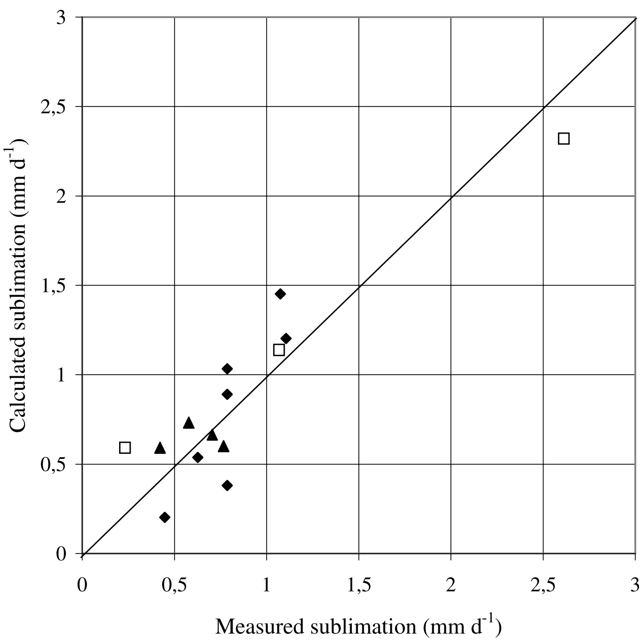

Figure 6. Comparison of calculated and measured daily sublimation values for 14 days: 7 days in 1999 (May 26, 1600–June 3, 1600 LT except May 31–June 1; black diamonds), 3 days in 2001 (May 23, 0730–May 26, 0730 LT; white squares) and 4 days in 2002 (May 8, 0830–May 9, 0830 LT; May 11, 0830–May 13, 0830 LT and May 14, 1700–May 15, 1700 LT; black triangles). The good correlation (r^2 = 0.83 or 0.60 without the highest value) yields confidence in the turbulent flux results. Also shown is the 1:1 line.

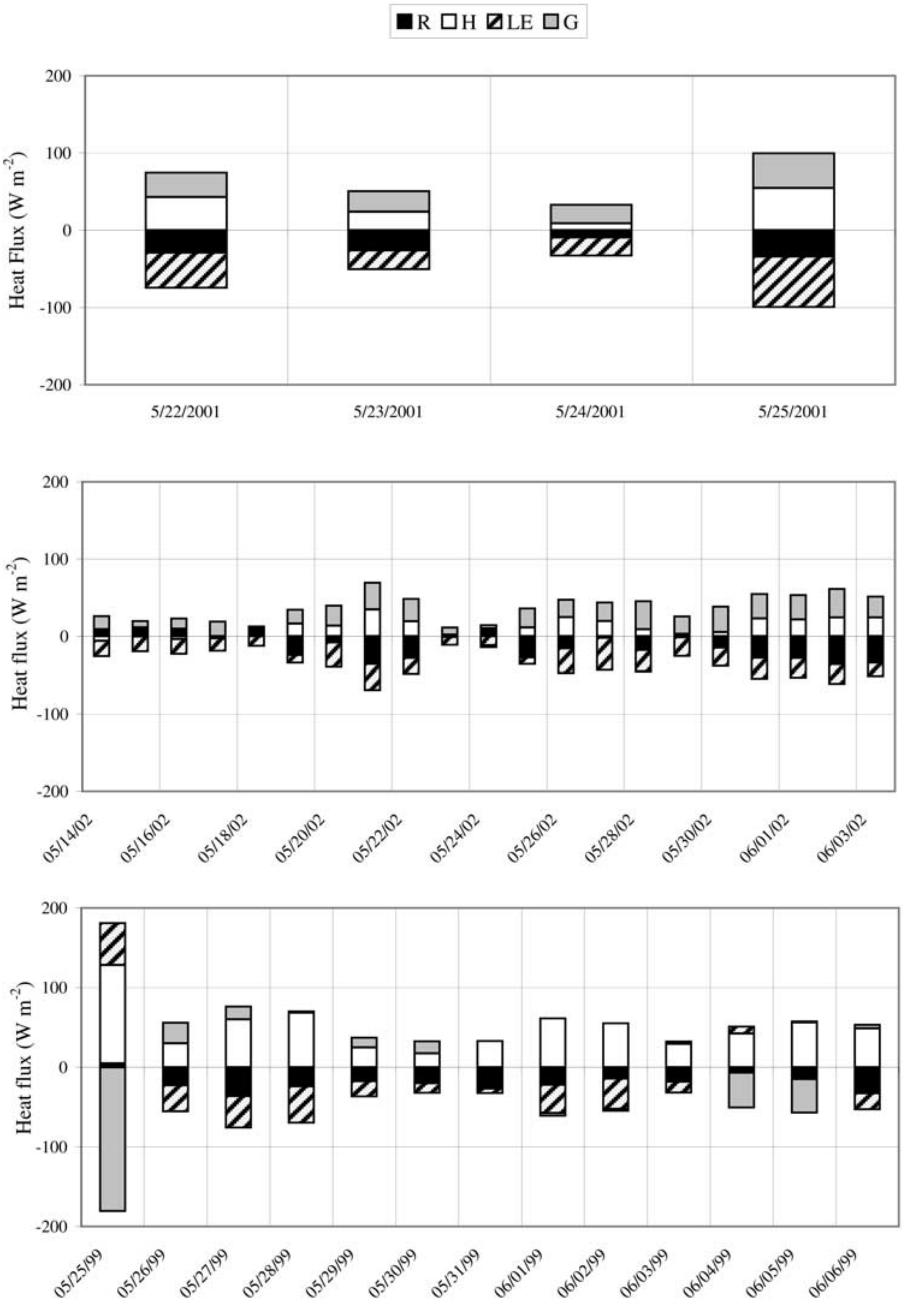

Figure 7. Daily mean values of the energy balance terms during the 2001 measuring period (upper panel), the 2002 measuring period (middle panel) and the 1999 measuring period (lower panel). The net radiation is indicated by black areas, sensible heat flux by white areas, latent heat flux by hatched areas, and the total subsurface energy flux by gray areas.

a higher level to prevent perturbations from the AWS mast, the wind speed was recalculated at the level of T and q assuming that the vertical wind speed profile is logarithmic (neutral conditions): $u = (u^*/k)\ln(z/z_{0m})$ with u^*, the characteristic scale of velocity, also called the friction velocity.

4.2.2. Roughness Parameters

[15] Since the turbulent fluxes are very sensitive to the choice of the roughness lengths [e.g., *Hock and Holmgren*, 1996; *Wagnon et al.*, 1999], allocating an entire sub-section to these parameters is justified. As given by *Wagnon et al.* [1999], the surface roughness parameters are all chosen equal $z_{0m} = z_{0T} = z_{0q} = z_0$ and are used as calibration parameters. Indeed, z_0 is calibrated in order to fit the mean calculated sublimation (obtained from *LE* derived from the bulk method) over the longest period with direct measurements to the mean measured sublimation on the same period obtained by averaging the values of the lysimeters. In May 2001, the calibration was done on the mean value of 3 entire

Table 5. Mean Values of Each Term of the SEB

	R, W m^{-2}	S↓, W m^{-2}	S↑, W m^{-2}	α	L↓, W m^{-2}	L↑, W m^{-2}	H, W m^{-2}	LE, W m^{-2}	G, W m^{-2}	Sublim. mm w.e. d^{-1}
2001										
Mean 5/22–25	−24.6	277.2	218.8	0.79	167.8	250.8	32.7	−39.9	31.8	−1.2
2002										
Mean 5/14–6/3	−12.0	253.4	200.7	0.82	198.4	263.1	11.7	−21.8	22.1	−0.7
1999										
Regime 1[a]	−23.5	257.4	225.1	0.87			42.8	−26.4	7.1	−0.8
Regime 2[b]	−5.8	203.3	179.0	0.88			73.9	20.9	−89.0	+0.6
Mean	−19.4	244.9	214.5	0.87			50.0	−15.5	−15.1	−0.5

[a]Regime 1: May 26–June 3 and June 6. 1999 (clear windy weather).
[b]Regime 2: May 25 and June 4–5. 1999 (cloudy sometimes snowy windy weather).

days where direct measurements were available: −1.3 mm w.e. d^{-1} (average value of 7 lysimeters from May 23, 0730 to May 26, 0730 LT). In May–June 2002, due to regular snowfalls which have perturbated the field trip, direct measurements of sublimation were available only for four days: May 8–9, 0830 LT; May 11–12 and 12–13, 0830 LT and May 14–15, 1700 LT. The mean value of these four days (−0.6 mm w.e. d^{-1}) was used to calibrate z_0 assuming that sublimation was nil during the night (from 0100 to 0700 LT) when meteorological data were missing (before May 13). In May–June 1999, the calibration was done on the mean value of 7 entire days: −0.8 mm w.e. d^{-1} (average value of 4 lysimeters from May 26, 1600 to June 3, 1600 LT except May 31–June 1). This approach is may be questionable but with only one level of measurement, it is better than choosing some values of the roughness parameters from the literature which are very scattered and highly dependent on the field site and on local meteorology (see *Braithwaite* [1995] for a review). A geometrical determination of z_{0m} [*Lettau*, 1969] was impossible due to the irregular spatial repartition of the roughness elements. Except *King and Anderson* [1994] who found z_{0T}, $z_{0q} \gg z_{0m}$ over an Antarctic ice shelf in winter, most of the authors and theoretical predictions suggest that z_{0T} and z_{0q} are 1 or 2 orders of magnitude lower than z_{0m} [e.g., *Ambach*, 1986; *Andreas*, 1987; *Morris*, 1989; *Hock and Holmgren*, 1996; *Meesters et al.*, 1997]. In front of this scattering, it was more justified to keep a unique value. Since these parameters are calibrated on direct measurements, taking different values for z_{0m}, z_{0T} and z_{0q} would have changed the values of these parameters but not the final results of the turbulent fluxes. Anyway, this assumption of equality of the roughness parameters is probably correct for smooth surfaces, which was usually the case for the surface immediately below the AWS [*Bintanja and Van den Broeke*, 1995]. Therefore, z_0, although calibration parameter, is likely to be not far from real roughness lengths for momentum, heat or moisture. Table 4 gives the values of z_0 for the different measuring periods together with a description of the surface state. On May 25, 2001, in the field site, we observed a rapid evolution of the surface state due to strong winds, which have mechanically eroded the surface letting zastrugis emerge around the AWS (erosive forms due to the wind which look like snow waves [*Lliboutry*, 1964]). That is why we have linearly increased z_0 from 0.5 to 1 mm as a function of the qualitative observations of the surface roughness below and around the AWS. In May–June 2002, due to regular snowfalls, the surface was covered by smooth fresh

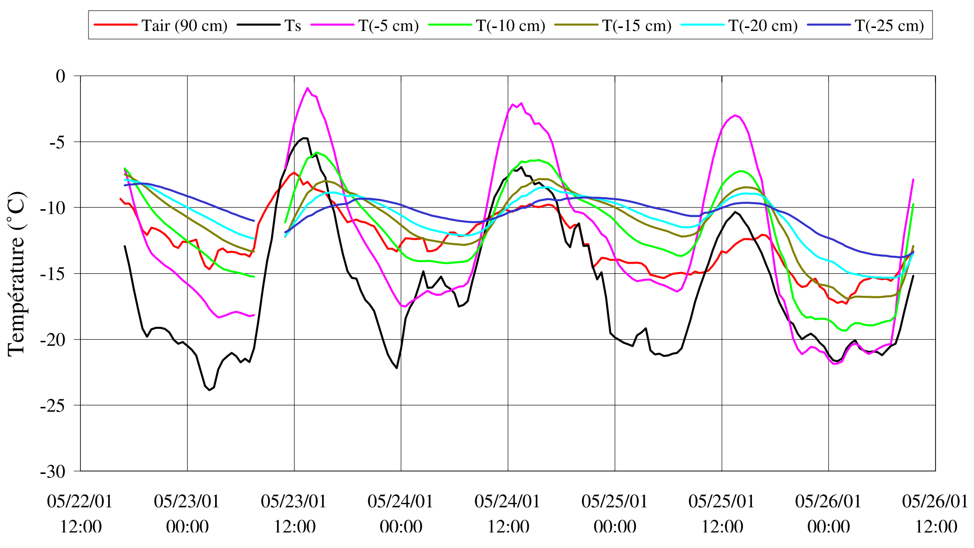

Figure 8. Half-hourly mean values of air temperature (0.9 m), surface temperature and snow temperatures (−5, −10, −15, −20 and −25 cm) on Illimani, 6340 m asl, between May 21 and May 26, 2001.

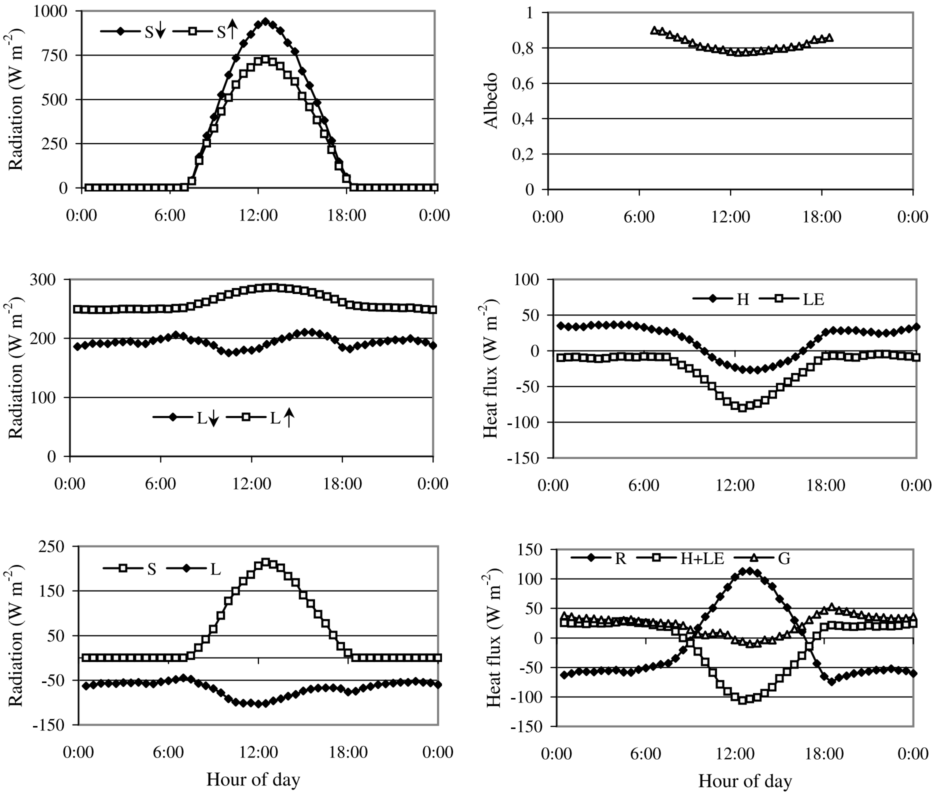

Figure 9. Mean diurnal cycle of the various terms of the SEB corresponding to both 2001 and 2002 measuring periods (25 days in total: May 22–25, 2001 and May 14–June 3, 2002): incident short-wave radiation *S*↓, reflected short-wave radiation *S*↑, incoming long-wave radiation *L*↓, outgoing longwave radiation *L*↑, net short-wave radiation *S*, net long-wave radiation *L*, albedo α, sensible heat flux *H*, latent heat flux *LE*, net all-wave radiation *R* and total subsurface energy flux *G*.

snow with small spread out zastrugis and remained unchanged during the entire measuring period (constant value of z_0). In May–June 1999, the surface experienced no significant change during the measuring period and was relatively similar to the surface of the end of the 2001 measuring period: local smooth surface of clean and crusty snow below the AWS but surrounded by irregular zastrugis a little higher than in 2001. Nevertheless, during this 1999 measuring period, due to the coarse assumption on T_s, z_0 must partly compensate the errors on T_s. Indeed, if the 1999 surface temperatures had been available on Illimani, the calculations of the turbulent fluxes would have been modified and the calibration of z_0 would have certainly been different. Therefore, the calibration of z_0 is an interesting way to reduce the errors on *LE* and *H* but for this reason, it is not very representative of the surface conditions.

[16] Once the calibration completed over the periods of several days, it is possible to compare daily values of sublimation measured with the lysimeters and calculated with the bulk method (Figure 6). Even if the two sets of data are not independent since mean values over 3, 4 and 7 days for 2001, 2002 and 1999 respectively were calibrated, we observe a good agreement between measured and calculated daily sublimation (r^2 = 0.83, n = 14 days; r^2 = 0.60 if we do not take into account the highest value). Therefore, the bulk method lets calculate correctly the daily values of the turbulent fluxes even if an interval of accuracy cannot be evaluated.

5. Results

5.1. Daily Mean Values of the Energy Balance Terms

[17] The daily mean SEB values for the 2001, 2002 and 1999 measuring periods are presented in Figure 7 and are reported in Table 5. In this usually clear and dry austral Bolivian winter, *R* is most of the time negative but varies mainly due to changes in cloudiness: *R* increases with increasing cloudiness because the increase of net long-wave radiation is higher than the decrease in net short-wave radiation. This feature of net radiation over high reflective surfaces is known as the ''radiation paradox'' [e.g., *Ambach*, 1974; *Wendler*, 1986; *Bintanja and Van den Broeke*, 1995]. When the cloudiness is very high like on May 14–16, 18, 23, 24 and 29, 2002, or on May 25, 1999, daily values of *R* can even reach positive values more typical of the wet season than of the beginning of the Bolivian winter. *LE* is usually negative which means that the surface loses mass through sublimation. *H* heats the surface and is an important source of energy at this high-altitude site especially when the atmosphere is clear. These two turbulent fluxes are very sensitive to the wind speed and remain very small during calm weather like on May 23–24, 2002. The other positive term is the subsur-

face energy flux G obtained as a residual of equation (1). This means that at the beginning of the austral winter, the snow below the surface loses heat to the benefit of the surface, which receives energy. This is confirmed while looking at the half-hourly snow temperatures recorded in May 2001 (Figure 8). During this period, a cold front penetrates into the glacier and snow temperatures slightly decrease toward the end of the measuring period. The variability of the various energy fluxes is large and connected with different weather regimes. This will be explored in more detail in section 5.3.

5.2. Diurnal Cycle of the SEB

[18] The mean diurnal cycle of the SEB terms for both 2001 and 2002 measuring periods (4 days in 2001: May 22–25 and 21 complete days in 2002: May 14–June 3 averaged all together) is shown in Figure 9. The diurnal cycle of all components is mainly determined by the net all-wave radiation R. At night, R is negative due to a strongly imbalanced net long-wave radiation in relation with the clear and dry atmosphere at very high altitude. R is all the more negative as the cloud amount is low. That is the reason why R is lower during the 2001 measuring period than in May–June 2002, which was a more cloudy and wetter period with therefore a higher incoming long-wave radiation $L\downarrow$ than in May 2001 (Table 5). This loss of energy at night is almost totally compensated by the positive sensible heat flux H. LE is reduced at night but still negative (mass loss through sublimation) and therefore, G remains positive which means that there is an upward heat conduction inside the top layers of the glacier. In the daytime, R follows the diurnal variations of the net short-wave radiation, S. Day to day variations of S are directly connected to changes in the cloud amount and S is also greatly reduced compared to $S\downarrow$ due to a very high albedo of this high-altitude site (mean value of $\alpha = 0.79$ between May 22 and 25, 2001 and $\alpha = 0.82$ between May 5 and June 3, 2002). The energy absorbed by the surface and the first layers of the glacier as net short-wave radiation compensates the energy loss due to the negative net long-wave radiation L. The remaining energy, which is the net all-wave radiation R, is then mainly used to increase the turbulence of the surface boundary layer: H switches to negative values (unstable conditions) and sublimation is strongly enhanced. The upward energy flux inside the snow is positive during the night in relation with low surface temperatures due to the negative net long-wave radiation L, and in the daytime G slowly decreases to reach negative values in the middle of the day, which means that the snowpack gains heat. There is a strong anticorrelation between half-hourly values of R and half-hourly values of the sum of the turbulent fluxes $H + LE$ ($r^2 = 0.84$, slope = -1.0, n = 234 values for the 2001 measuring period; $r^2 = 0.83$, slope = -1.3, n = 1314 values for the 2002 measuring period) which strengthens the idea that net all-wave radiation is responsible for the diurnal cycle of the turbulent fluxes and therefore drives the SEB of this high-altitude cold site.

5.3. Stability of the Surface Boundary Layer

[19] The bulk Richardson number Ri_b is a measure of stability and is connected to the diurnal cycle of u^* and θ^* together plotted on Figure 10 where u^* and θ^* are the characteristic scales of velocity and potential temperature respectively: $\theta^* = -H/(\rho C_p u^*)$. Ri_b is most of the time positive and small corresponding to moderately stable conditions of the surface boundary layer which means that at night, in the morning and in the late afternoon, the lower atmosphere is cooled down by contact with the cold snow surface. Between 1000 and 1600 LT, daytime heating of the surface causes unstable conditions (negative Ri_b): the lower atmosphere is heated in contact with the surface. The night variability of Ri_b must be related to the peculiar conditions that have prevailed on May 23 and 24, 2002 where the wind speed was very low leading to high values of Ri_b. For instance, night peaks of Ri_b (at 0230 and 2100 LT) are due to very occasional high values of Ri_b (>0.5) corresponding to unusual very stable conditions of May 24, 2002 with small wind speeds (<1 m s^{-1}) and strong vertical gradients

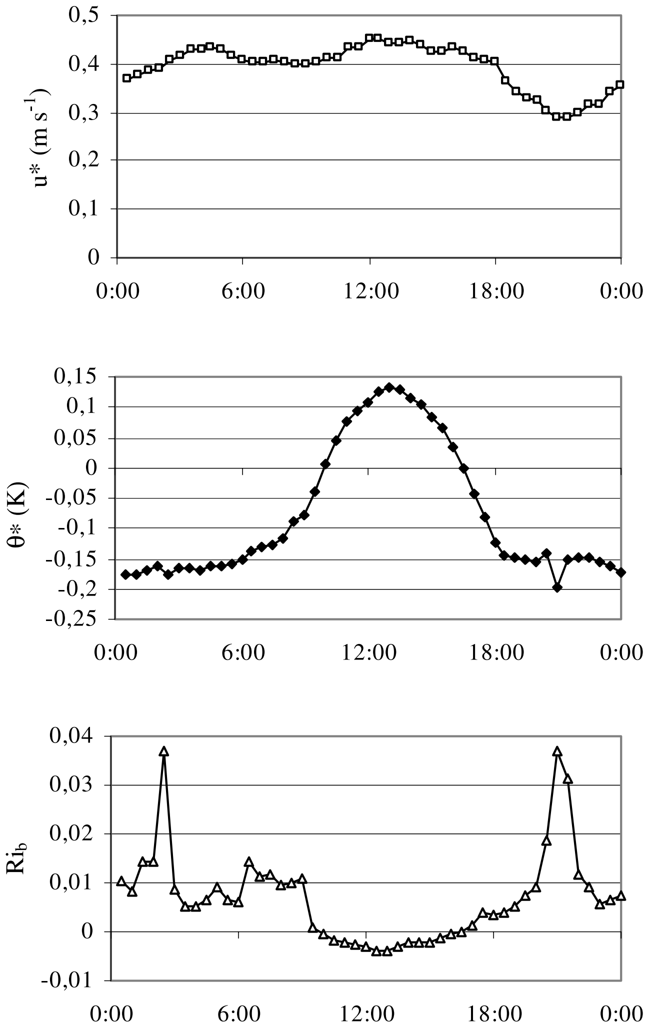

Figure 10. Mean diurnal cycle of friction velocity u^*, temperature scale θ^*, and bulk Richardson number Ri_b corresponding to both 2001 and 2002 measuring periods (25 days in total). Most of the time, moderately stable conditions prevail on the glacier surface except between 1000 and 1600 LT where Ri_b is negative (unstable conditions).

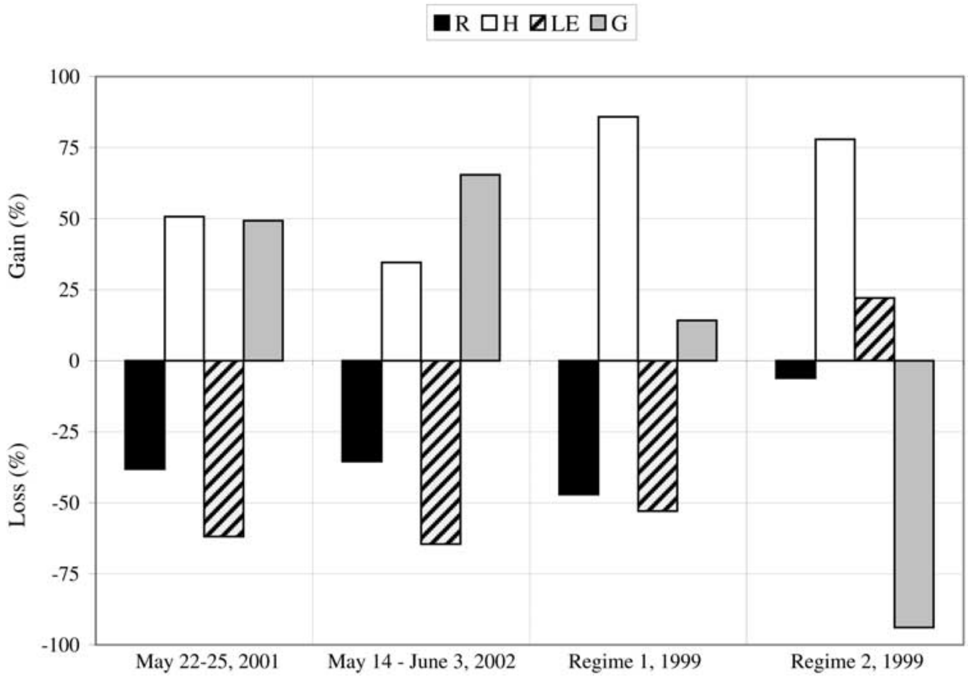

Figure 11. Relative importance of the mean surface heat fluxes for the 2001 and 2002 measuring periods and for the 2 weather regimes of the 1999 measuring period. All positive fluxes are summed and scaled to 100%.

of air temperature (up to 10 K m^{-1}). The negative peak of θ^* at 2100 LT is also related to these peculiar conditions of May 24, 2002 ($\theta^* = -1.5$ K).

5.4. Daily Variability of the SEB

[20] The daily variability of the energy fluxes is large during the various measuring periods mainly in relation with the cloudiness, which enhances *R*, and the wind speed, which directly influences the turbulent fluxes. For instance, looking at the 2002 measuring period, the energy fluxes of overcast days (May 14–18; 23–24 and 29) remain all small because the radiative budget is balanced (*R* is close to 0 or slightly positive) and therefore, there are few energy exchanges at the glacier surface as turbulent fluxes (except during very windy days like May 27) or subsurface conductive flux. On the other hand, without any clouds, *R* is strongly negative and this energy sink is compensated by the other fluxes (*H*, *LE* and *G*) which are all the higher as the wind blows fast. Such clear days with important energy exchanges at the glacier surface are more frequent at the end of the 2002 measuring period, which again suggests that this period is a transition toward the dry season. Consequently, the weather regime typical of the Bolivian dry season (entire 2001 measuring period, end of the 2002 measuring period and regime 1 of 1999) is characterized by sharply negative net radiation values due to a high albedo and to a reduced incoming long-wave radiation of the dry, cold and clear atmosphere of this very high-altitude site. It is also characterized by high turbulent fluxes with positive values of *H* and negative values of *LE* due to strong westerly winds and a positive upward energy flux inside the snow (a cold front penetrates into the glacier). During this typical weather of the dry season, sublimation is high (the average values of the 2001 and 2002 entire measuring periods and the 1999 regime 1 period are −1.2 mm w.e. d^{-1}, −0.7 mm w.e. d^{-1} and −0.8 mm w.e. d^{-1} respectively) and is comparable to the sublimation already observed lower in altitude, at 5150 m asl, on Zongo Glacier, 40 km north of Illimani (the average values for the dry seasons (May–August) 1997 and 1998 are −0.95 mm w.e. d^{-1} and −0.66 mm w.e. d^{-1} respectively) [*Wagnon et al.*, 2001]. Nevertheless, these sublimation rates on Illimani remain lower than the values reported by *Ginot et al.* [2001] in the drier Chilean subtropical mountains (−1.9 mm w.e. d^{-1} on Cerro Tapado, 5536 m asl).

[21] The 1999 regime 2 (May 25 and June 4–5) corresponds to outbursts of cold polar air bringing clouds and precipitation [*Vuille*, 1999]. Therefore, the net all-wave radiation increases (“radiation paradox”). Moreover, relative humidity is very high leading to positive vertical moisture gradients above the glacier surface and condensation can occur on the field site. The energy exchanges are higher as the wind is strong (May 25, 1999). During these days, *R* is negative but small, *H* and *LE* are positive and sometimes high depending on the wind speed and therefore, the snow gains energy mainly transferred by the turbulent fluxes (*G* is high and negative).

6. Discussion and Conclusion

[22] The purpose of this study was to characterize the SEB of a cold, high-altitude tropical glacier during the dry season. The relative importance of the energy fluxes for the different measuring periods and for various weather regimes can be inferred from Figure 11. The sum of the positive fluxes is scaled to 100%. During the dry season (austral winter), mean daily values of *R* are always negative but this heat sink is reduced during occasional overcast days due to

a more balanced long-wave radiative balance (beginning of the 2002 measuring period and regime 2, 1999). Except during cloudy and snowy days, *LE* is a heat sink at the glacier surface and all the larger as the wind speed is high. *H* brings energy to the surface and, at a daily scale, counter-balances *LE* during clear days. During the dry season, the energy flux inside the snow is directed upward which means that a cold front penetrates into the glacier except during occasional overcast days where the turbulent fluxes bring heat to the surface. The main feature of the SEB of this high-altitude tropical site is the strongly negative latent heat flux during typical days of the dry season (low temperature and humidity, strong westerly winds) which means that the surface loses mass through sublimation at a rate sometimes higher than 1 mm d^{-1} w.e. Figure 12 shows the cumulative ablation through sublimation on Illimani for the three measuring periods. Since accumulation is very reduced during the 6 months of the dry season, the snow surface remains several months in contact with the atmosphere and therefore, it is likely that the chemical and isotopic contents of the surface layers of the glacier are altered by this high sublimation. Therefore, the climatic interpretation of the tropical ice cores must be done with caution.

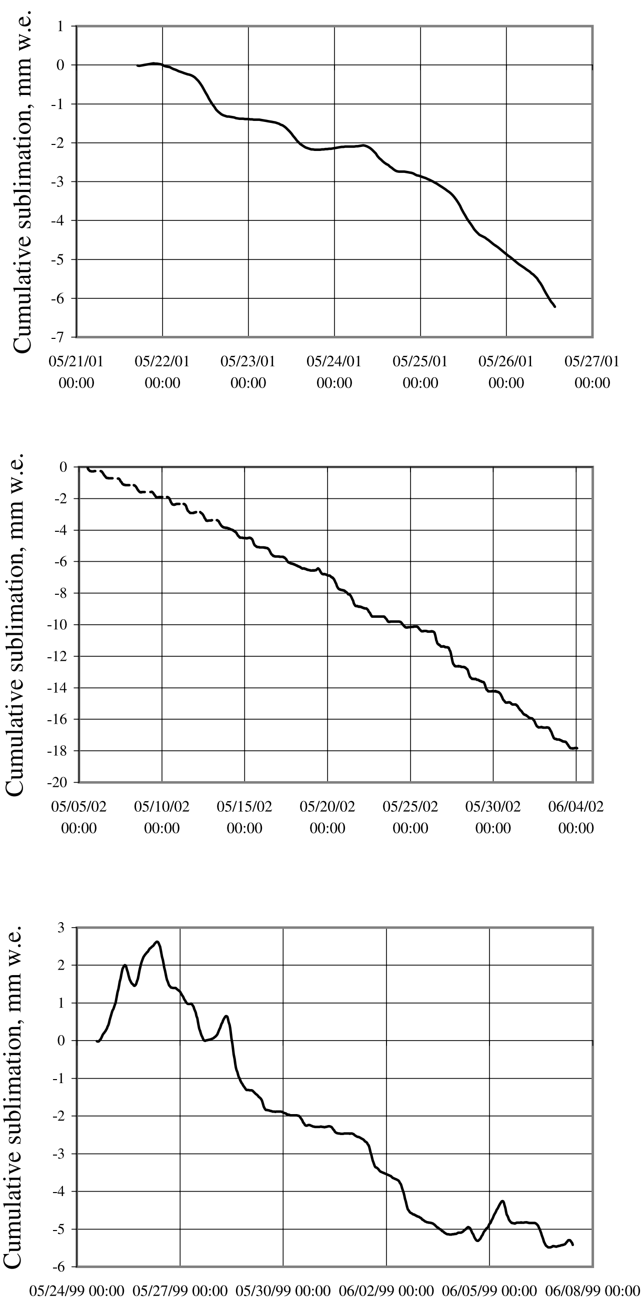

Figure 12. The cumulative sublimation calculated on the three measuring periods: 2001 (upper panel), 2002 (middle panel) and 1999 (lower panel).

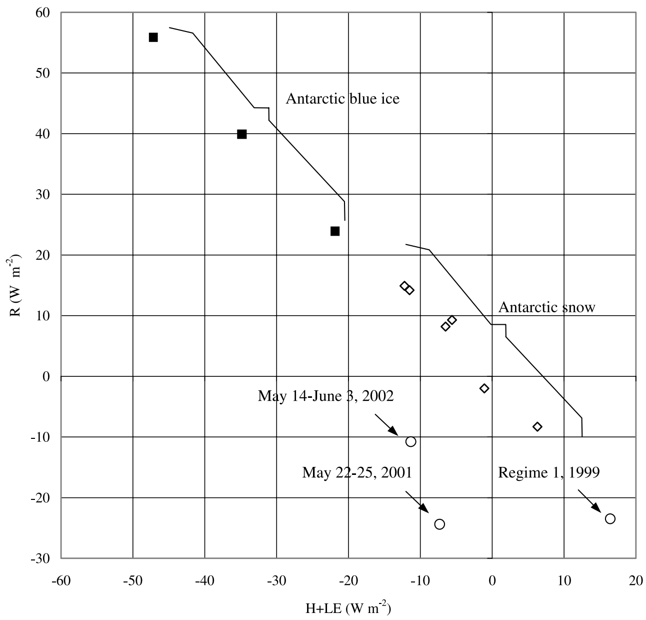

Figure 13. Mean summertime values of net radiation *R* as a function of the sum of the turbulent fluxes ($H + LE$) for Antarctic sites: blue ice areas (black squares) in the Dronning Maud Land (1200 m asl) and snow areas in the same area (white diamonds) (from Figure 19 of *Bintanja and Van den Broeke* [1995]) in comparison with this study on Illimani, 6340 m asl (mean values of the days typical of the dry season: Regime 1, 1999, 2001 and 2002 entire measuring periods; white circles).

[23] It is noteworthy that the characteristics of the SEB of this cold, high-altitude tropical site are comparable to the summer SEB of snow surfaces on the intermediate slopes of Antarctica (around 1200 m asl) [*Bintanja and Van den Broeke*, 1995]. Turbulent fluxes have the same order of magnitude with a positive daily sensible heat flux, which counter-balances the daily latent heat flux. For an identical albedo, the daily values of *R* are positive in Antarctica and negative on Illimani mainly because measurements in Antarctica were conducted in summer (24 hours a day of sunshine leading to higher daily values of $S\downarrow$ and identical values of $L\downarrow$ and $L\uparrow$) although they were done in winter on Illimani. The reverse is observed for *G*: in summer, a warm front penetrates into the glacier in Antarctica although the glacier on Illimani experiences a cold front in winter (see Figure 13 of *Bintanja and Van den Broeke* [1995] as a comparison to Figure 11). Figure 13 displays the mean summertime net radiation as a function of the sum of the turbulent fluxes for Antarctic sites together with the mean

values of R and $H + LE$ for the wintertime conditions on Illimani (2001 and 2002 entire measuring periods and regime 1 of 1999). The data point of the 1999 regime 1 together with the data points of Antarctic intermediate slopes roughly stand in a straight line crossing the zero, which suggests that the SEB is characterized by the net radiation and turbulent fluxes both in Antarctica and on high-altitude cold tropical glaciers. On both locations, the energy available as net radiation is used to increase the turbulence of the surface boundary layer and is converted into turbulent fluxes. The main conditions for this similarity are the high reflective snow surface, the absence of melting and the comparable local meteorology on both locations. Nevertheless, the data points of the 2001 and 2002 measuring periods stand below the line, which means that the sum $H + LE$ is smaller than expected and is then dominated by the negative latent heat flux. Therefore, although sublimation rates are comparable between both locations for the represented measuring periods, for the same amount of net radiation sublimation on Illimani is likely to be stronger than on the intermediate slopes of Antarctica. In conclusion, postdepositional processes due to sublimation might be more vigorous on high-altitude tropical sites than in Antarctica. This point will be examined in detail in a future paper.

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