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The montane circulation on Kilimanjaro, Tanzania and its relevance for the summit ice fields: Comparison of surface mountain climate with equivalent reanalysis parameters

N.C. Pepin [a,*], W.J. Duane [b], D.R. Hardy [c]

[a] University of Portsmouth, UK
[b] Universiti Brunei Darussalam, Brunei
[c] University of Massachusetts, Amherst, United States

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ABSTRACT

We compare surface climate (temperature and moisture) measured on an hourly basis at ten elevations on Kilimanjaro with equivalent observations in the free atmosphere from NCEP/NCAR reanalysis data, for September 2004–July 2008. On the lower forested slopes the mountain surface is consistently cooler and moister than the atmospheric boundary layer. In contrast, temperatures and moisture on the higher slopes above treeline (~3000 m) are decoupled from the free atmosphere, showing substantial heating/cooling by day/night and import of moisture up from lower elevations during daylight hours. The mountain is universally warmer than the background atmosphere at 1500 EAT, the sparsely vegetated upper slopes acting as the focus for the most intense heating. The persistent vapour pressure excesses (>5 mb) in the forest zone move upslope during daylight and subside downslope at night. Strong seasonal contrasts are shown in the vigour of this process, the resultant mountain thermal circulation and its consequences. The synoptic forcing of this process (as represented by flow indices developed from reanalysis wind components), although evident, is relatively weak. This means that upslope flow from the forest zone is an important supplementary source of moisture for the upper slopes of the mountain and that free-air variability, although important, alone cannot account for all the variability in the summit moisture regime. Long-term ice retreat at the summit of Kilimanjaro therefore is most likely to be influenced by changes in local land-use as well as more regional free-air changes.

1. Introduction

The summit ice fields on Kilimanjaro are undergoing substantial retreat (Hastenrath and Greischar, 1997; Thompson et al., 2002; Kaser et al., 2004; Cullen et al., 2006; Thompson et al., 2009). A long history of studies on the mountain (Hastenrath, 1984, 2001; Thompson et al., 2002; Mölg et al., 2003; Mölg and Hardy, 2004; Cullen et al., 2006; Mölg et al., 2008, 2009a,b) has shown that an increased drying of the summit climate is largely responsible for current net ablation of ice. A drier climate will decrease cloud cover, increase the saturation deficit, increase direct solar radiation, and decrease precipitation. All these changes will decrease glacier mass balance through a combination of reduced accumulation, reduced albedo (Mölg and Hardy, 2004) and enhanced sublimation driven by intense radiation (Kaser, 2001; Kaser and Osmaston, 2002). Measurements and modelling of glacier–climate interactions on Kilimanjaro (e.g., Mölg et al., 2009b) therefore show limited sensitivity to air temperature in comparison with changes in hydrological indicators.

Long term drying throughout the 20th century in East Africa is supported by changes in vegetation (Hemp, 2005), proxy data (Hastenrath, 2001) and paleoclimatic modelling (Mölg et al., 2006). Many factors are thought to play a part, including both regional and global climate forcings (Kaser, 1999; Mölg et al., 2009a) and more local change induced by land-use change (Altmann et al., 2002; Hemp, 2005).

Recent work has demonstrated the increased inter-annual variability in atmospheric moisture at high elevations in the tropics in comparison with lower elevations. Mölg et al. (2009a) show that precipitation on the higher slopes of Kilimanjaro is extremely sensitive to variations in free-atmospheric structure and forcing by SSTs over the Indian Ocean to the east. Variability within the OND wet season (short rains) is particularly pronounced, dependent on El-Nino related forcing (Indeje et al., 2000; Chiang and Lintner, 2005; Hastenrath et al., 2007). There is also evidence that global forcing may influence the position of the convective zone in the tropics (Lintner and Neelin 2007) which leads to regional drying.

* Corresponding author. Department of Geography, Buckingham Building, Lion Terrace, University of Portsmouth, UK, PO1 3HE. Tel.: +44 2392 842507; fax: +44 2392 842512.
E-mail address: Nicholas.pepin@port.ac.uk (N.C. Pepin).

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Although there has been some debate (Shugart et al., 2001; Hemp, 2005), it is also widely suggested that there has been considerable deforestation on the lower slopes of Kilimanjaro in the last 40 years (Agrawala et al., 2003; FAO, 2006) due to expansion of agricultural cultivation (Soini, 2005), use of forest for firewood, and increase in local population. These pressures have increased bushfire frequency which has depressed the treeline on the southern slopes (Hemp, 2005) in contrast to the expected movement of vegetational zones upslope which would be expected with a warming trend (Grabherr and Pauli, 1994). Local land-use change could therefore contribute to the local drying of the climate which is increasingly held responsible for the current patterns of ice retreat (Kaser et al., 2004; Mölg and Hardy, 2004; Mölg et al., 2008, 2009b). However, a definitive link between local land-use change and the drying of the mountain climate as a whole remains unproven (Agrawala et al., 2003).

Although many observations of summit climate (including snow surface change as a proxy of summit mass balance) have been taken, and modelling attempts undertaken (Mölg et al., 2009a) to examine the distribution of hydrological parameters across the mountain, there have been no detailed observations of surface temperature/moisture or airflow across the mountain as a whole. Thus the summit climate is typically observed in isolation. In particular there has been limited study of the linkages between atmospheric moisture on the lower slopes and the climate at the summit of the mountain. This is important because the climatological rainfall maximum is on the lower south-western slopes (Coutts, 1969; Hemp, 2006) despite prevailing free-air flow from the east. Traditional models of orographic enhancement (reviewed by Roe, 2005) based on mechanical interaction need to be modified in the tropics, and particularly on Kilimanjaro, not least because of inherent stability in the large scale synoptic flow. A moisture budget of the forest zone was estimated by Hemp (2005) and rainfall has been measured at various elevations on the southern slopes (Hemp, 2002; Rohr and Killingtveit, 2003; Hemp, 2006) but there are limited hydrological observations (moisture, evaporation, precipitation) on the mountain slopes, particularly above treeline.

This paper fills a gap in current knowledge by comparing recent air temperature and humidity records from ten sensors installed on a transect up the south-western slope of the mountain (Duane et al., 2008) over a four year period (2004–2008) with background flow data interpolated from NCEP/NCAR reanalysis to equivalent elevations and locations. The transect covers an elevational range of nearly 4000 m. In this way the modification of this flow by the mountain surface can be quantified (see Pepin and Seidel, 2005), which allows examination of the mechanisms of the montane circulation (as first postulated by Troll and Wien, 1949 with reference to Mt Kenya) in so far as they apply to Kilimanjaro. This comparison allows separation of surface effects from the larger scale free atmospheric forcings. In particular we examine the role of the mountain surface in controlling atmospheric moisture and how excess moisture develops and moves upslope to the summit crater. We show that the workings of the diurnal thermal circulation, in tandem with mechanical forcing as suggested by Mölg et al. (2009a), are critical to this process.

After data sources are outlined in Section 2, Section 3.1 examines background temperature and moisture profiles for northern Tanzania. These determine the background climate. Low temperatures and lack of atmospheric moisture at summit level are common features of the current free atmosphere in this region. Section 3.2 then examines how the surface climate of Kilimanjaro on the broad scale of the mountain differs from this background flow. We derive temperature and moisture differences at each of the ten surface stations and examine how these differences change diurnally and seasonally. Section 4.1 develops a descriptive model of the current mountain circulation in the two trade-wind seasons (January and July), based on the above observations. Some additional field wind measurements collected in July and August 2008 are used to substantiate the July model. In Section 4.2 we compare our results with the dynamical modelling approach of Mölg et al. (2009a). Finally Section 5 discusses the ramifications of these findings in terms of the summit moisture regime and ice field mass balance.

2. Data and study area

Three main sources of data are used in this study, including

a) NCEP/NCAR reanalysis free-air temperatures and humidities, and
b) autonomous surface temperature and humidity measurements, and
c) surface wind speed and direction from field observations.

a) Background flow temperature and moisture were derived from the NCEP/NCAR reanalysis R1 project (see Kalnay et al., 1996; Kistler et al., 2001). Pressure level grids for four times a day (0, 6, 12 and 18 UTC) were extracted from the Climate Diagnostics Center ftp.cdc.noaa.gov. CDC, (2008). We interpolated the temperatures and relative humidities to the exact equivalent location (in x, y and z) of each of our ten surface stations described in Section 2b) (Hartmann, pers. comm. 2005). The grids correspond to 3, 9, 15 and 21 EAT (East African Time) respectively. All interpolation was linear. Available reanalysis grids ran up until 6 July 2008.

Reanalysis u (zonal) and v (meridional) free-air flow components were also interpolated from each six hourly grid in a similar manner. Positive/negative u is westerly/easterly flow and positive/negative v southerly/northerly flow respectively. Because the reanalysis does not contain topographic forcing and/or surface data (being an amalgam of radiosonde and satellite data), it assumes that Kilimanjaro does not exist. Note we are not using modelled reanalysis "surface" data but free-air pressure level data (i.e. there is no direct "surface" effect in this data). The mean surface elevation of the four reanalysis grid points surrounding the mountain is only 1231 m (Mölg et al., 2009a). Thus the interpolation is well above the surface of the reanalysis grid, and representative of background flow conditions. Although there is a modelling component in the reanalysis in the tropics, the product does include radiosonde data from Nairobi and Dar es Salaam (Kistler et al., 2001; Durre et al., 2006) and is used frequently to represent background flow conditions (Mölg et al. 2009a). The reanalysis profile does include atmospheric boundary layer effects below approximately 3000 m. In this article we use the term background flow to include both the free atmosphere (above 3000 m) and the atmospheric boundary layer (below this elevation), i.e. the whole atmospheric profile independent from the mountain surface itself.

b) Hobo Pro Series dataloggers were employed at 10 locations listed in Table 1 and outlined on Fig. 1 on the south-western slope of the mountain. The Machame route provides the basis for the transect, all logger locations having a similar south-westerly aspect and avoiding topographic hollows (for further discussion see Duane et al. (2008)). The loggers, which measure both air temperature and relative humidity, have a quoted accuracy of +/−0.2 °C at 20 °C and +/−3% (4%) in a non-condensing (condensing) environment. In the cloud forest where humidities are often near 100% condensation is common. This type of logger was evaluated in field trials by Whiteman et al. (2000) who concluded that they were "suitable for a variety of meteorological applications". We calculate partial pressures of water vapour (above a water surface) using equations given by Kuemmel (1997). Loggers were mounted 1.5 m above ground level using a 6 mm thick white cylindrical uPVC radiation shield positioned nominally horizontally with the open ends orientated north and south. This prevented direct radiation from entering the shield (the sensor is mid-way along the tube). The sampling interval was 1 hour and loggers were installed in September 2004. The most recent data collection was

Table 1
Site and temperature/moisture data characteristics.

Logger no.	Elevation	Site description	Missing data
1	1890 m	Dense montane rainforest	Sep 2004–Jan 2006
2	2340 m	Dense montane rainforest	None
3	2760 m	Sparse montane rainforest	None
4	3170 m	Transitional zone between rainforest and sub-alpine heathland	Jan 2006–Sep 2007
5	3630 m	Sub-alpine heathland	Jan 2006–July 2008
6	4050 m	Alpine with limited vegetation	Sep 2004–Jan 2006
7	4570 m	Alpine with limited vegetation	Sep 2004–Jan 2006[a]
8	4970 m	Bare rock/desert	None
9	5470 m	Bare rock/desert	None
10	5800 m	Ice field	None

[a] No humidity data is available for this station.

in July 2008 so there are nearly four years of data at most sites. There are some gaps due to logger malfunction and/or loss (see Table 1) but at all sites except site 5 there is more than two years of complete data. Our sensor network covers around 4000 m in elevation range and represents all the major vegetation zones present on Kilimanjaro, discounting the foothills (Fig. 2). Sites 1–3 are broadly in the rainforest, sites 4 and 5 in the giant heather zone, 6 and 7 in the alpine zone, 8 and 9 in the high elevation desert, and site 10 on the summit ice sheet.

c) At the six higher sites (5–10), logging anemometers were installed during an expedition in late July/early August 2008 (Fig. 3). The lower sites were not used because they were in dense forest. Around 5–6 days of wind data (mean wind speed, gust and wind direction) was obtained with a sampling interval of 1 minute. Although this is only a short period it gives insight into the development and decay of slope winds on a diurnal basis at this time of year. Table 2 lists the wind data available at each site, along with site details.

The quality of the surface data was assessed through calibration with AWS1 on the summit ice-field (see Mölg and Hardy, 2004; Duane et al., 2008) where the radiative environment is the most extreme. Mean differences in temperature at this location between the two instruments were small (<1 °C at all hours of the day). Radiative heating within the radiation shield at site 10 is minimal and is expected to be smaller at the other sites where radiative loading is reduced. Any radiative heating would also have a much reduced effect (if any at all) on vapour pressure measurements which would tend to be conserved in a hypothetical sealed environment (as opposed to relative humidity). All wind sensors were checked daily in the field for free rotation and there is no evidence of data contamination (e.g. sticky sensors, tendency to a fixed wind direction during periods of little or no wind).

3. Results

3.1. Background flow climate in northern Tanzania

We examine mean air temperature, relative humidity and vapour pressure profiles interpolated to the Kilimanjaro sites using reanalysis (Fig. 4). Temperature shows a broadly linear temperature profile with

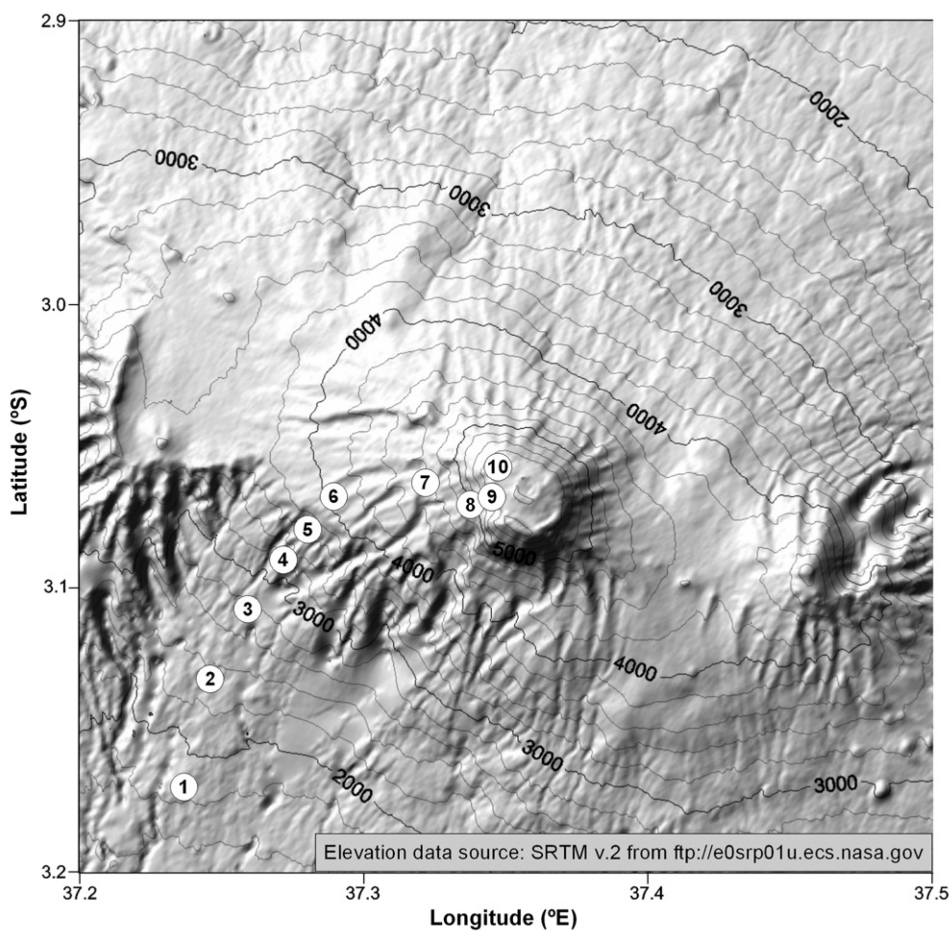

Fig. 1. Map of the 10 logger sites on the south-west slope of Kilimanjaro.

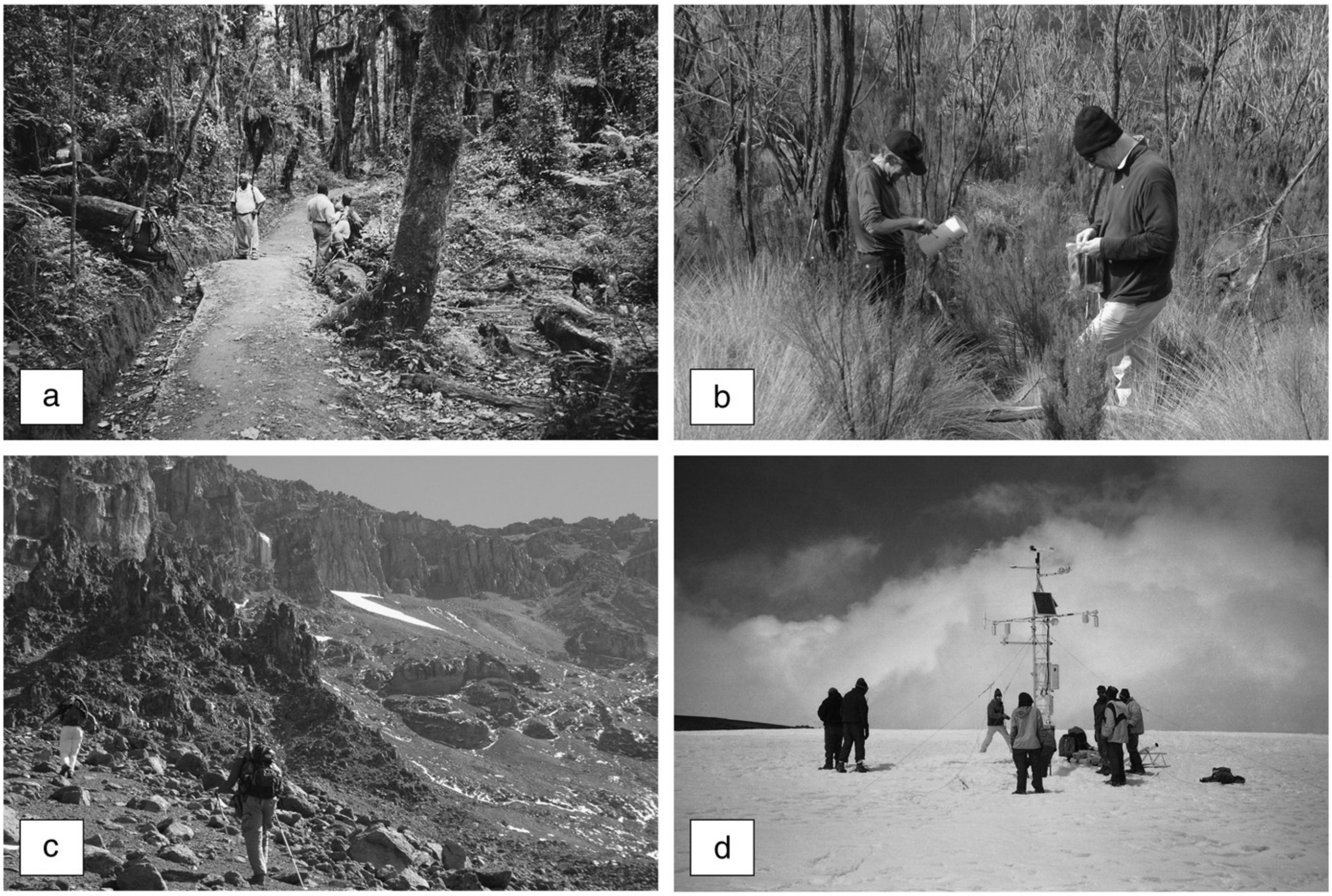

Fig. 2. Pictures showing some of the vegetational zones on Kilimanjaro, a) forest zone (loggers 1–3), b) giant heather zone (loggers 4–5), c) desert zone (logger 8), d) ice field (logger 10).

a very slight weakening of lapse rate above 5000 m. The mean lapse rate over the whole elevational range is −5.5 °C/km (annual) which is fairly shallow (Fig. 4a), with a slightly steeper rate in January and a weaker one in July. Background relative humidities (Fig. 4b) decrease less consistently (at slightly less than 1%/100 m) from a mean of 70.8% at 1890 m (equivalent to our lowest surface logger) to 38.0% at 5803 m (equivalent to the highest logger) on an annual basis. The gradient steepens between 3000 and 4500 m, but especially during

Fig. 3. Temperature/humidity and wind loggers set up in the field at site 8.

Table 2
Wind data characteristics.

Logger no.	Elevation	Site description	Available wind data
5	3630 m	Sub-alpine heathland	July 29–August 2 2008
6	4050 m	Alpine with limited vegetation	July 30–August 4 2008
7	4570 m	Alpine with limited vegetation	July 30–August 4 2008
8	4970 m	Bare rock	July 31–August 5 2008
9	5470 m	Bare rock	July 31–August 5 2008
10	5800 m	Summit crater	August 3–6 2008

July when it is particularly dry at higher elevations. The rate of decrease for the annual mean profile is fastest at medium elevations. The humidity decrease, combined with the rapid decrease in temperature, means a sharp decrease in mean vapour pressure from 12.9 mb at 1890 m to 1.6 mb at 5803 m (i.e. nearly an order of magnitude), and the profiles for all seasons (Fig. 4c) are curved (meaning a more rapid decrease at lower elevations). It is therefore normal for the upper air to be extremely dry. This decrease occurs in all seasons but particularly during the south-east trade winds (July) when the mean free-air vapour pressure at summit elevation falls below 1 mb. Although it is difficult to compare directly due to differences in the types of readings reported, these figures agree broadly with those reported by Mölg et al. (2009a) in their Table 1.

Reanalysis diurnal changes (not shown), although systematic, remain small in absolute terms and the mean diurnal temperature range at ~5800 m is only 1.0 °C as opposed to 3.9 °C at 1890 m. Diurnal changes are therefore concentrated below 3000 m where regional atmospheric boundary layer effects are evident. Diurnal changes in vapour pressure are surprisingly small (around 1 mb at the lowest elevations, but <0.1 mb above 3000 m). Since mean temperatures at ~5800 m are well below freezing (around −5 °C) the temperature on its own is likely to minimize any melting of the summit ice sheets (Mölg et al., 2009b). However cloud is infrequent and incoming radiation intense. The relative humidity only tops 95% on 0.15% of occasions (i.e. very infrequently).

Table 3 lists mean u and v components at each elevation for January and July, along with the most frequent flow direction and the mean wind speed F (both derived using Pythagoras from the raw reanalysis wind components). Scalar mean flow directions would be misleading. We choose these two months as representative of the two distinct trade wind regimes. During the north-east trade winds (January), a weak strengthening of mean flow with height is evident and the flow veers from north-east to east at the higher elevations. There is less distinct wind shear during July (south-east trade winds). Site 1 elevation has the strongest flow, representative of the East African Low Level Jet (EALLJ) which penetrates the furthest inland at this time (Findlater, 1977). Above this the flow weakens and turns more indistinct. At moderate elevations representative of the Shira Plateau a northerly flow becomes the most frequent regime, before reverting to south-easterly at the highest elevations. At low levels (especially the elevations representative of sites 1 and 2) there is a simple switch from north-easterly flow during January to south-easterly flow in July but at higher levels the difference between the two seasons becomes more indistinct and upper level flow usually becomes more easterly all year round. This agrees broadly with the findings of Mölg et al. (2009a,b) who using the same reanalysis data

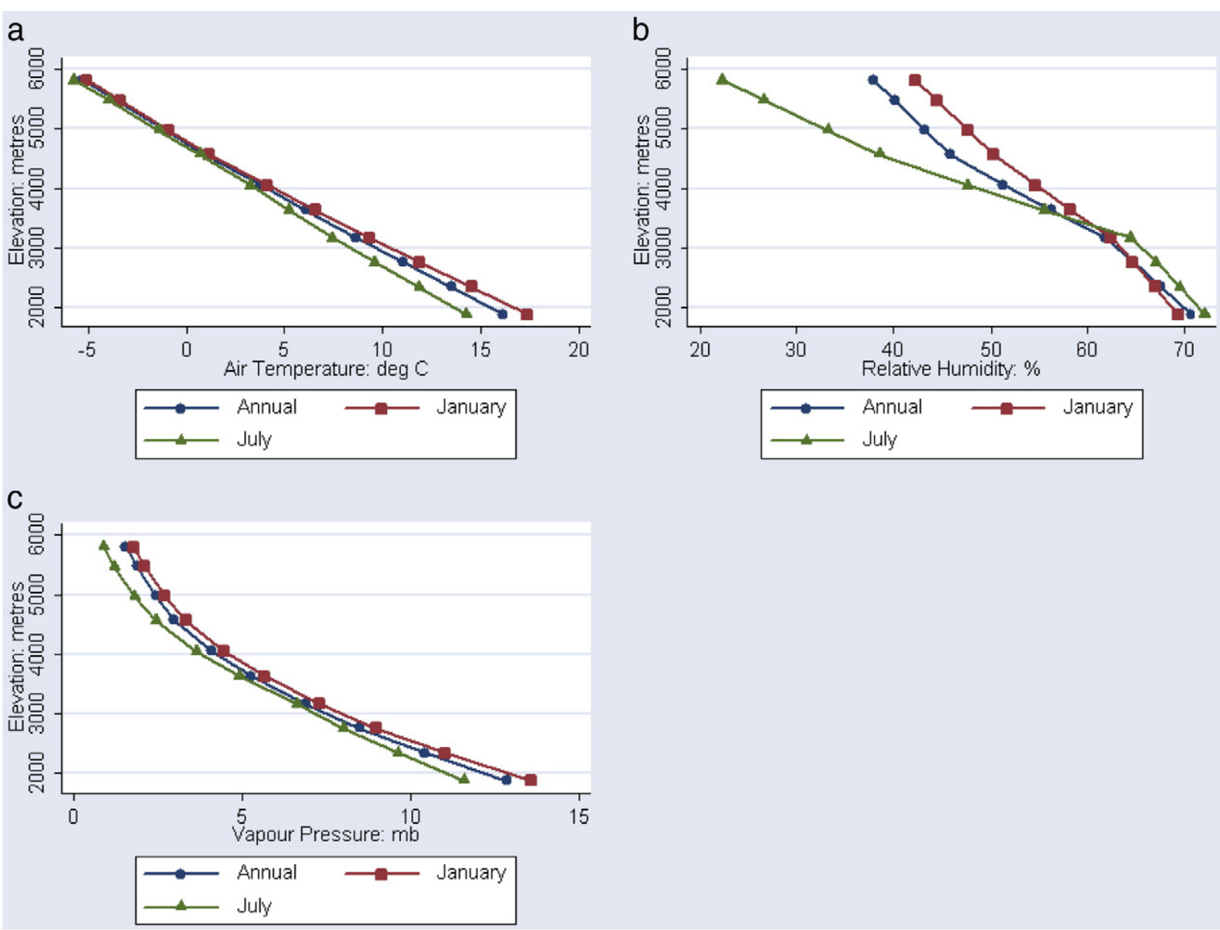

Fig. 4. Vertical profiles of a) mean free-air temperature, b) mean free-air relative humidity, and c) mean free-air vapour pressure from reanalysis January 2004–July 2008. Curves are shown for the annual mean, January and July.

Table 3
Mean free-air components at each logger elevation (September 2004–July 2008).

Logger no.	Elev. (m)	U (m/s)	V (m/s)	F (m/s)	Flow dir (mode)	U (m/s)	V (m/s)	F (m/s)	Flow dir (mode)
		Jan	Jan	Jan	Jan	July	July	July	July
1	1890	−2.88	−2.64	4.57	NE	−3.16	3.58	5.05	SE
2	2340	−2.83	−2.76	4.72	NE	−1.85	2.65	3.79	SE
3	2760	−2.78	−2.88	4.94	NE	−0.63	1.76	3.34	SE
4	3170	−2.75	−2.97	5.22	NE	0.51	0.90	3.76	N
5	3630	−3.27	−2.52	5.18	NE	0.49	0.60	3.53	N
6	4050	−3.74	−2.12	5.32	E	0.47	0.33	3.70	N
7	4570	−4.35	−1.67	5.65	E	0.26	0.15	3.97	N
8	4970	−4.88	−1.45	5.88	E	−0.25	0.28	3.81	N & SE
9	5470	−5.55	−1.17	6.40	E	−0.90	0.45	4.03	N & SE
10	5800	−6.00	−0.99	6.85	E	−1.31	0.56	4.42	N & SE

Table 4
Surface/free-air temperature/vapour pressure correlations. Figures in brackets represent the correlations for 1500 EAT (diurnal cycle removed).

Logger no.	Elevation	Annual temp	Annual VP	January temp	January VP	July temp	July VP
1	1890 m	0.74 (0.65)	0.14 (0.38)	0.69	0.06	0.72	−0.38
2	2340 m	0.70 (0.72)	0.21 (0.30)	0.62	−0.13	0.61	0.22
3	2760 m	0.45 (0.68)	0.15 (0.21)	0.19	−0.02	0.40	0.21
4	3170 m	0.18 (0.27)	0.16 (0.09)	0.13	0.00	−0.08	0.17
5	3630 m	0.14 (0.10)	0.30 (0.09)	0.06	0.03	0.06	0.46
6	4050 m	0.04 (0.08)	0.33 (0.15)	0.12	0.22	−0.10	0.36
7	4570 m	0.03 (0.17)	NA	0.08	NA	−0.09	NA
8	4970 m	0.14 (0.22)	0.41 (0.30)	0.36	0.24	−0.03	0.31
9	5470 m	0.13 (0.26)	0.36 (0.28)	0.29	0.19	−0.06	0.23
10	5800 m	0.05 (0.27)	0.49 (0.39)	0.07	0.44	−0.07	0.30

source show that the most frequent wind direction between the surface and 500 hPa lies between 45 and 90°.

3.2. Surface climate on Kilimanjaro and its comparison with background conditions

The climate represented by the reanalysis is equable. Conditions we measure on the mountain surface however are far more extreme. This is because the mountain surface responds to the local energy balance, heating/cooling with respect to the free atmosphere as the balance becomes positive/negative respectively. This will be especially important in the tropics where the diurnal radiative cycle is strong and convection/precipitation occurs mainly in the afternoon (Hedberg, 1964; Hastenrath, 1991). Free-air/surface temperature differences have been intensely studied (Peppler, 1931; McCutchan, 1983; Richner and Phillips, 1984; Pepin and Losleben, 2002; Pepin and Seidel, 2005) and the processes are fairly well understood. Cloud cover usually minimizes temperature differences, as does strong free-air advection (Pepin and Norris, 2005). Snow cover usually makes the surface universally colder than the free atmosphere at the same elevation. The same broad patterns can be seen in this data.

On Kilimanjaro correlations between raw surface and background temperatures are greatest at lower elevations, particularly at sites 1–3 (Table 4). This is not surprising since the reanalysis and mountain surface contain common boundary layer effects at the lowest elevations (but not higher up). However, correlations often remain high at the lowest sites even when sub-times of day are analysed (e.g. 0.884 at site 1 at 900 EAT) meaning that the common diurnal effect is not wholely responsible for these higher correlations. Thus figures for 1500 EAT alone (figures in brackets in Table 4) show a broadly similar pattern to those for the whole day. As we move above the forest and heather zones, the correlation between surface and background temperatures weakens, dramatically above site 5. This pattern also remains when any seasonal influence is removed (not shown). The temperatures on the higher slopes are thus decoupled from the free atmosphere suggesting intense heating/cooling. Note that this does not necessarily imply that the differences above the forest zone are larger, just that they are more variable in time. Conversely, the forest zone, although it has very large mean differences, has more inertia and responds to diurnal and synoptic changes more in tandem with the atmospheric boundary layer.

Correlations are also shown for surface and free-air vapour pressures based on all data (Table 4). In this case correlations are universally low, perhaps increasing slightly at higher elevations, but remaining below 0.5 in all cases. Removal of the diurnal signal (figures for 1500 EAT in brackets) has an inconsistent influence on results, strengthening correlations at low elevations, but weakening them at high elevations. This means that the diurnal component becomes more important higher up the mountain, and the synoptic control stronger lower down. In broad terms background and surface moisture regimes are decoupled from one another.

3.2.1. Temperature differences (ΔT)

Hourly temperature differences (Δ*T*) (surface minus reanalysis) were calculated at each location. Since the reanalysis is produced only four times daily the resultant differences are derived for 300, 900, 1500 and 2100 EAT. The differences are much larger than could be explained by any errors in the surface data (for example due to the minimal heating in the shield — see Section 2) and in general decrease with elevation.

The mean daily difference (Fig. 5) is strongly negative at sites 1–4 in all months, showing that the forest is a heat sink. This is unsurprising, since the area beneath the canopy where temperatures are measured is cut-off from the intense tropical sun, and most energy is put into the latent flux rather than sensible heating. Differences are more strongly negative in June–July. The mean Δ*T* becomes positive at site 5 which emerges above the forest zone on the edge of the Shira plateau. This is a sharp discontinuity. The intense sunlight and resultant surface heating makes mean differences positive, especially in the less cloudy periods at this elevation (Aug/Sep and Dec/Jan). Sites 6 and 7 also become much warmer than the free-air in the dry relatively cloud-free period of June–September, but less so in January/February. However, mean daily differences are much smaller (<1 °C) in comparison with site 5. Sites 8 and 9 have very small mean differences. At this elevation, high above the Shira plateau, free-air advection is likely to be strong and thus there will be a reduced mass heating/cooling effect (Tabony, 1985). However, as the sites are on the lee side of the crater, they are frequently trapped in a lee eddy, and have reduced exposure to upwind free-air flow. Nevertheless, mean differences are less than 1 °C, compared with more than 2 °C lower down the mountain. There are slight variations with season but these are not systematic. Finally site 10 is a heat sink (surface colder than the free atmosphere) because of the ice-field, but particularly in the dry seasons (Jan/Feb and July/August). The mean surface temperature is −6.5 °C, more than a degree lower than the mean free-air temperature. In summary most surface heating occurs just above the tree-line on mid level slopes and on the Shira Plateau.

The mean daily differences in Fig. 5 hide the strong diurnal cycle in Δ*T* (Fig. 6) which agrees with previous studies (Richner and Phillips, 1984; Dreiseitl, 1988; Pepin and Losleben, 2002). At 0300 EAT all sites are colder than the background flow. The biggest deficits are at sites 3 and 4, and 10. The negative temperature difference at site 10 is unsurprising, the ice-field acting as a strong heat sink due to thermal emission. One would expect deficits to be minimal below the cloud base in the rainforest. Thus the large deficits at sites 3 and 4 are unexpected and difficult to explain. It may be that the cloud base falls below site 3 (2760 m) on many nights, leading to rapid cooling in the clearer air. However this does not explain why deficits at the less cloudy higher elevations are smaller than at these sites.

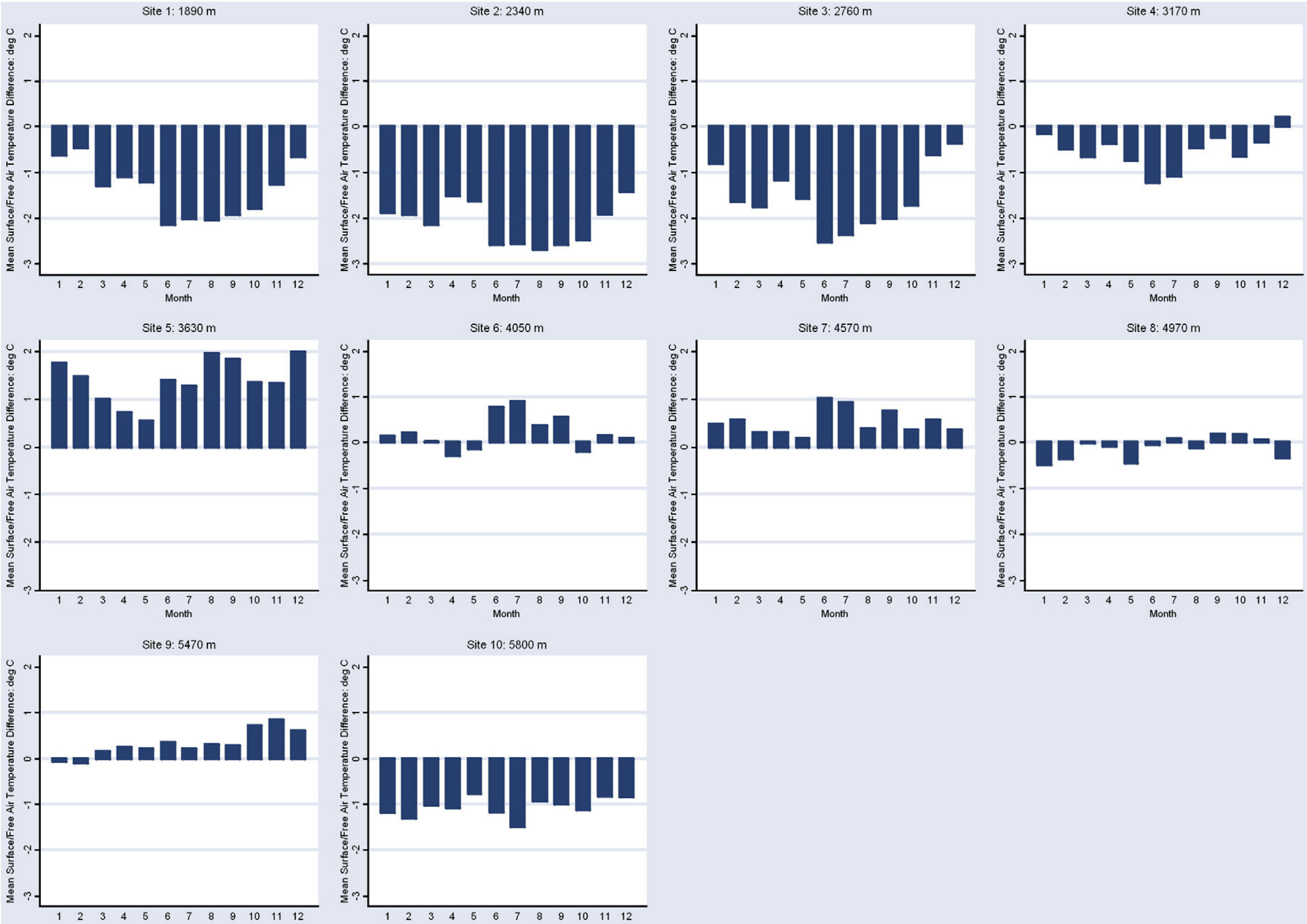

Fig. 5. Mean surface/background flow temperature differences for each month of the year at the ten logger locations. A positive value means that the surface is warmer than the background flow.

By 0900 EAT the sun's heating has turned Δ*T* positive. This is focused between sites 4 and 7, but especially at site 5 towards the top of the heather zone. This is probably because the Shira Plateau and surroundings respond rapidly to the increase in solar radiation, showing a mass-elevation effect (De Quervain, 1904) as well as more limited interception of radiation load by vegetation. Δ*T* remains negative at Sites 1 and 2 below the dense forest canopy. The upper levels of the mountain (sites 8–10) show relatively little heating at this time of day. This is probably a result of solar geometry as well as increased elevation and exposure since the Western Breach, where sites 8 and 9 are situated, is not in direct sunlight at 0900 EAT.

By 1500 EAT surface heating has spread upslope (and downslope) to all sites apart from 1 and 2. The largest excesses are reported at sites 4, 5 and 9 but all of the sites above the treeline record Δ*T* in the range 2–3 °C. The exception is site 10 where the heat sink effect of the ice partially compensates for the strong positive radiation balance and differences remain small.

By 2100 EAT (Fig. 6d) there is rapid cooling, with surface temperatures falling below equivalent free-air values at all sites, especially below treeline (sites 1–4) and on the ice field (site 10). It may be thought that the forest canopy should prevent temperatures dropping well below background levels but this is not the case. Cooling at sites 5–9 above tree-line is less severe, where it is also less cloudy. The concept of nocturnal cooling of the surface being strongest above treeline, starting downslope katabatic flow from above the cloud base (Troll and Wien, 1949) is therefore probably misleading. The explanation of this relative lack of surface cooling above treeline is challenging, but it may result from compression and adiabatic warming in down slope winds which can be strong at this time of night, especially at sites 8 and 9 (nocturnal wind speeds measured in July/August 2008 at these sites exceeded 10 m/s). The topography of the upper slopes of Kilimanjaro, and especially along the logger transect, is not conducive to cold air ponding, so this is unlikely to be a factor.

There are differences in the diurnal cycle of Δ*T* between seasons. The most marked contrasts are illustrated by considering the core of the north-east (January) and south-east (July) trade wind periods (Fig. 7). In July the morning (0900 EAT) slope heating is subdued in comparison with January when Δ*T* is both larger and extends to lower elevations. There is much greater stratiform cloud cover at low/mid levels in the north-east monsoon (July), particularly in the vicinity of sites 4 and 5 which limits heating during the first part of the day. However, by 1500 EAT the situation is reversed and in July temperature excesses of 4–5 °C extend from site 4 to site 9. In January the peak heating appears to be reached over a more extensive area much earlier and by afternoon Δ*T* has decreased at most medium level sites (the exception being sites 8–10). These contrasts are explained

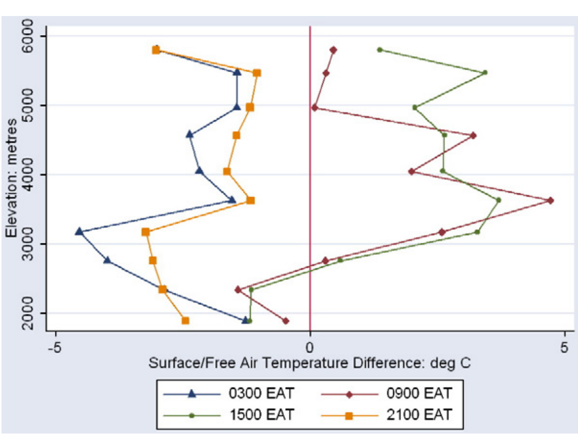

Fig. 6. Vertical profiles of the mean surface/background temperature difference for four times of day: 300, 900, 1500 and 2100 EAT.

through construction of a conceptual circulation model of Kilimanjaro (Section 4.1).

3.2.2. Moisture differences (ΔVP)

Moisture excesses/deficits in comparison with the reanalysis are also helpful in understanding how the mountain slope creates its own climate. Mean annual vapour pressure differences (ΔVP) are positive at all sites (Fig. 8a), showing an excess of moisture at the mountain surface in comparison with the background flow. The reanalysis does not include direct surface effects such as slope effects and vegetation. The moisture excess is evidence of substantial latent heat flux. The excess is greatest at the lower sites, averaging 3–4 mb and can reach more than 5 mb during the day. We believe that the role of the forest in creating these excesses is clearly important since there is a consistent drop in absolute excess with increased elevation and on our transect, which is representative of the broad vegetational transition on the south-western slope of the mountain, vegetation changes are extreme. At the ice field site (site 10) the mean excess is only 0.6 mb, more than five times lower than in the forest zone. Thus on average the moisture "created" by the mountain stays on the lower slopes.

Under certain conditions however this excess moisture moves upslope. Fig. 8b shows the diurnal change in the vertical profile of vapour pressure excess. At 0300 EAT the mean excess is just over 1 mb, and is evenly distributed from site 1 to 6. By 0900 EAT this has trebled to between 3 and 4 mb at all vegetated sites (1–5). This rapid moistening of the atmosphere is most likely a result of rapid evapotranspiration. At this time however there is very little change above site 6, suggesting minimal *in situ* evapotranspiration in the alpine and desert zones. By 1500 EAT the excesses increase further, reaching nearly 6 mb at site 1. However there is a progression/movement of this excess moisture upslope and the most dramatic increases in comparison with 0900 EAT are now at higher sites (e.g. sites 6 to 9) while the rate of increase has slowed in the forest zone. For example site 6 (in the alpine zone) now has over 4 mb excess, while site 3 has shown relatively little increase compared with 0900 EAT, even though the rainforest is where daytime evapotranspiration should be at a peak. The systematic delay in response moving upslope, along with increased excesses well above the vegetation limit, is strong evidence that this excess afternoon moisture on the upper slopes of Kilimanjaro has been imported from lower elevations rather than created solely through *in situ* evapotranspiration. By 2100 EAT the excesses have decreased everywhere and fallen to around 2–3 mb between sites 1 and 6, and more or less stabilized at night time values (~1 mb) at higher locations.

Differences in ΔVP also occur on a climatological basis between January and July. Fig. 9 shows that moisture moves upslope far more effectively in January in comparison with July. By 1500 EAT excesses exceed 5 mb at 5 sites in January compared with at no sites in July. Explanations concerning the relative strength and vertical extent of this transfer in these two seasons will be shown in Section 4.

In both seasons, not all of the available moisture is transported upslope and the excess still increases in the afternoon, albeit less rapidly, at the lower sites, even when the slope circulation is acting to transport moisture up and out of the cloud forest. The steepest altitudinal gradient in absolute excess occurs in the afternoon at all times of year which shows that the atmospheric reservoir in the forest zone at present is not being depleted by the convective process and that trees are very effective at continuously supplying water even when upslope winds act to transfer this moisture away from the forest.

Another important point is that the afternoon excess at site 10 is not out of line with sites 8 and 9 immediately below (despite being over the ice field). Thus afternoon sublimation over the ice field itself appears relatively unimportant in humidification of the air at crater elevation, in comparison with larger scale processes. During an expedition in July/August 2008 another temperature/humidity

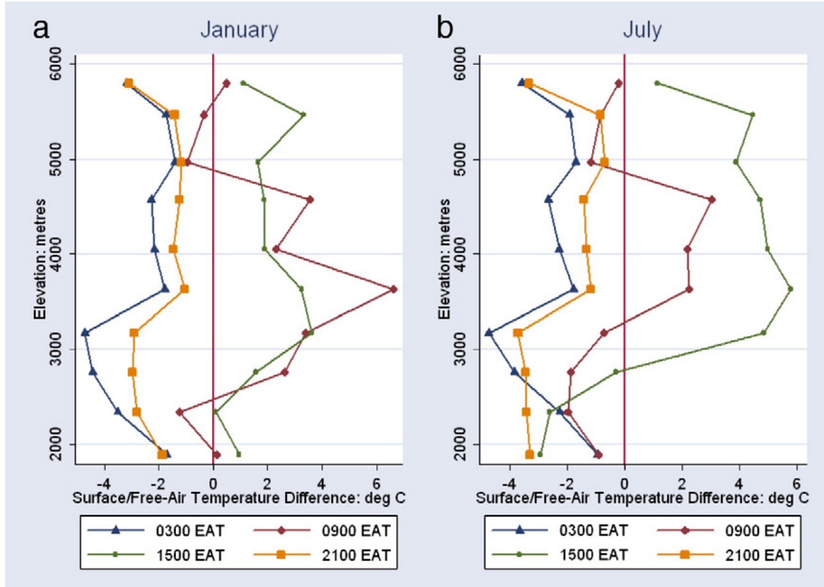

Fig. 7. Same as Fig. 6, except for a) January and b) July.

measurement point was installed temporarily on the summit crater around 500 m away from the ice-field (at the same elevation as point 10). Differences between vapour pressures at this site and those over the ice-field were minimal (averaging less than 0.1 mb) which provides further short-term evidence that large-scale moisture import is the dominant process at this elevation. However longer term comparisons at crater elevation between sites on the ice-field and further away are needed to confirm this.

Examination of the patterns of absolute vapour excess (Figs. 8 and 9) may give the impression that most of the diurnal variability occurs at the lower elevations, and it is clear that most of the atmospheric moisture typically resides in the forest zone. However, even though absolute excesses are smaller at summit level, the importance of these excesses in relative terms cannot be overstressed. As a percentage of total surface vapour pressure (ΔVP/VP), the median relative surface excess increases with elevation from 19% at site 1 to 36% at site 6 then falling to 27% at summit level (site 10), although these figures hide huge variability. At 1500 EAT the equivalent figures are 33%, 50% and 47%. Thus significant proportions of the moisture at summit level are imported, and at the low temperatures common at this elevation even a small increase/decrease in vapour pressure can result in condensation/dissolution of cloud respectively.

3.2.3. Relationships between ΔT, ΔVP and background flow

Although we have shown moisture to be imported to the upper slopes through a diurnal process (the diurnal signal is strong), this process may not be independent of free-atmospheric forcing. Much research has shown that strong upper level gradient winds suppress thermal circulation in mountain regions (Whiteman, 1990; Bossert and Cotton, 1994; Garreaud et al., 2003). Personal observation

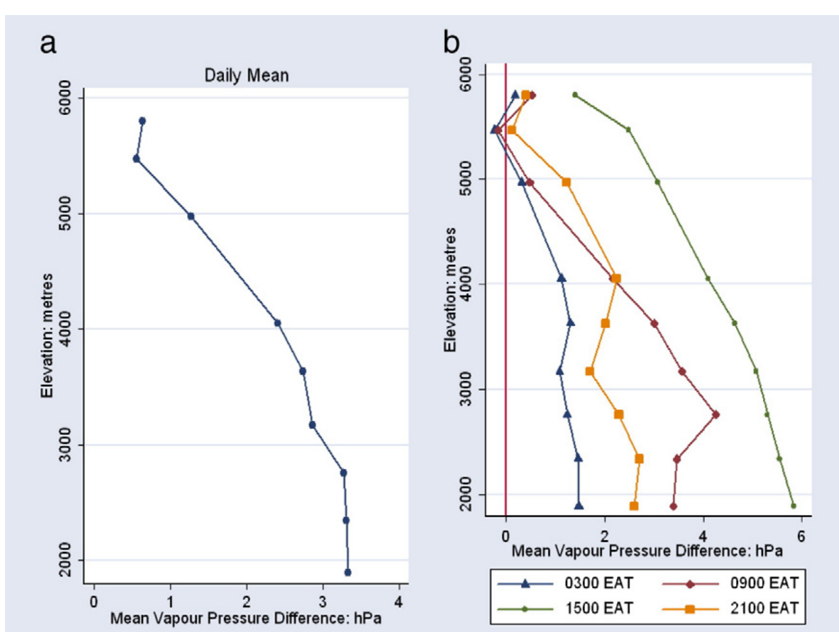

Fig. 8. Vertical profiles of the mean surface/background vapour pressure difference for a) diurnal mean and b) four times of day: 300, 900, 1500 and 2100 EAT.

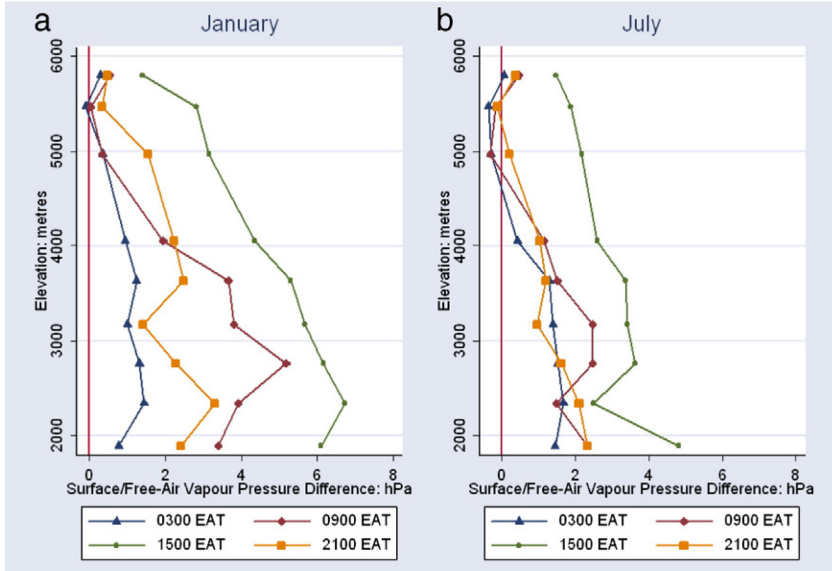

Fig. 9. Same as Fig. 8b, except for a) January and b) July.

suggests that on windy days it is difficult for cloud to attain the crater (Losleben pers. comm. 2010). It is reasonable therefore to expect negative relationships between $\Delta T/\Delta VP$ and upper level flow strength, especially at the higher sites.

Perhaps surprisingly, however, there are no simple relationships between background flow direction or strength and ΔT at any elevation when all the data is combined together. There is a slight weakening of the absolute magnitude of ΔT at site 10 when wind speed at summit level is very strong and all the major heating episodes (strongly +ve ΔT) occur when wind speed is weak (Fig. 10a). Similar statements can be said as regards ΔVP which also is not strongly correlated with flow strength (Fig. 10b) or flow direction.

Consideration of time-specific relationships eradicates any common diurnal forcing. At site 10 (the ice field) at 1500 EAT for example,

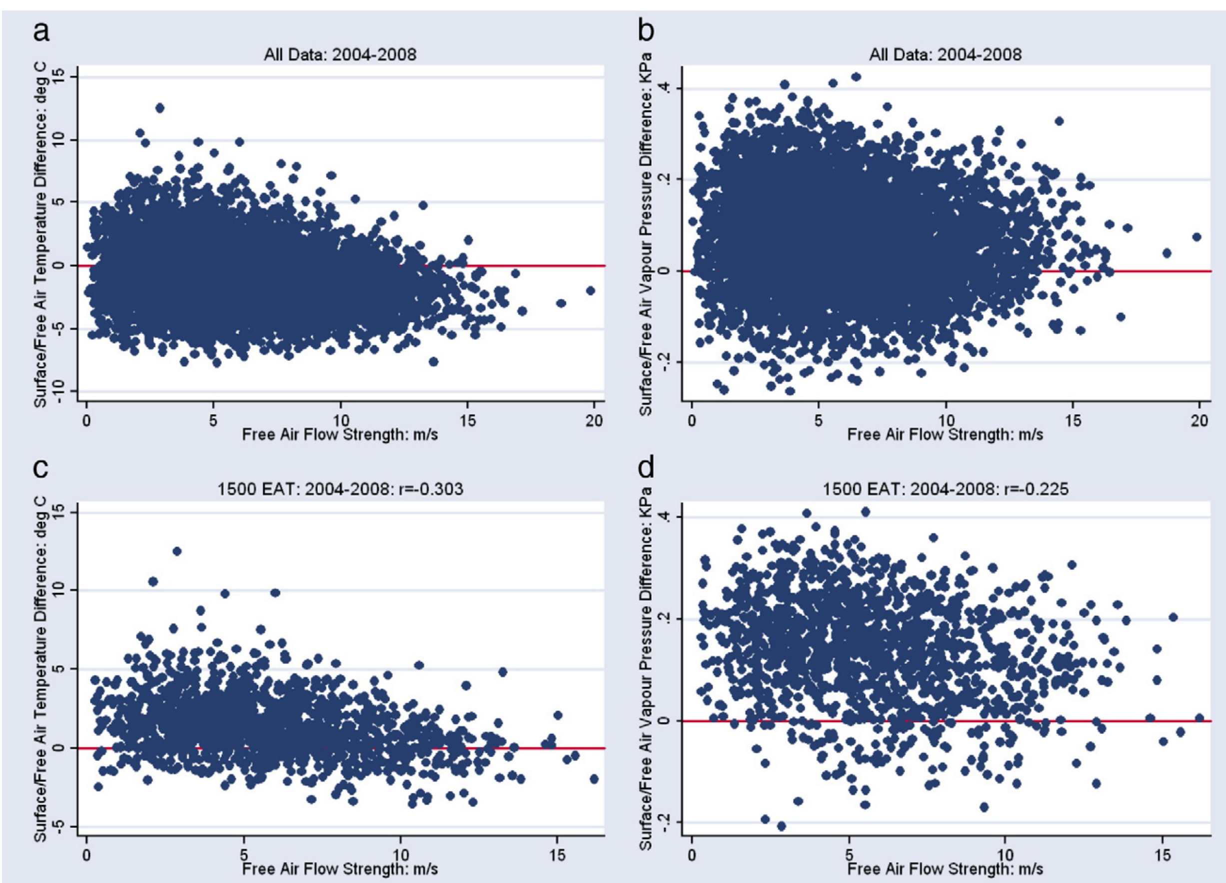

Fig. 10. Relationship between temperature and moisture excesses and background flow strength for a) ΔT versus flow strength (all data), b) ΔVP versus flow strength (all data), c) ΔT versus flow strength (1500 EAT), d) ΔVP versus flow strength (1500 EAT).

both the temperature excess and moisture excess are weakly controlled by upper level wind speed($r = -0.224$ for moisture, $r = -0.303$ for temperature, $p < 0.001$) (Fig. 10c and d). The relationship is weaker at other times of day. Although these patterns fit with Mölg et al.'s (2009a) observation that heavy precipitation episodes on the summit occur when summit wind speeds are relatively weak, it is clearly not a simple case of upslope import and free-air advection in direct competition. Weak upper airflow appears to be a necessary but not sufficient condition for large moisture excesses to develop at summit level (Fig. 10b) and it may well be that some of the upslope import is a direct result of the mechanical interaction of free-air flow and the mountain (see Section 4.2).

4. An observationally-derived model of the mountain circulation

Recent research has examined the moisture regime on Kilimanjaro using a dynamical model (Mölg et al., 2009a). To complement this, we develop a conceptual model based on our observations (Section 4.1) and then compare our results with the dynamical findings of that paper (Section 4.2). Although dynamical modelling attempts already exist, it is important to develop independent ideas based on observations which open the opportunity for cross-comparison and validation.

4.1. Conceptual model

The above patterns of temperature difference and moisture excess on the south-west slope of Kilimanjaro, along with supplementary field measurements of wind speed and direction from sites 5–10 during July/August 2008, are used to construct models of the mountain circulation for the two main trade wind periods (January and July) (Fig. 11). Figures listed below each site indicate the percentage frequency of relative humidities above 95%, a conservative estimate of presence of cloud based on analysis of summit webcam images (Duane et al., 2008). The positions of clouds are in part based on these figures, but the diagram is not strictly scientific in construction. The models are based on climatological mean conditions.

During January (Fig. 11a) the morning heating is particularly strong under clear skies. The typical top of the low level cloud deck is around 3000 m at 0900 EAT. Direct solar radiation received on the mountain slope is increased by the time of year (austral summer), orbital forcing (perihelion) and the southerly component to the slope aspect. As a result strong upslope winds develop by mid-morning focussed on where large surface temperature excesses develop (sites 4 and above), in turn transferring vapour transpired from the upper forest and giant heather zones rapidly upslope. This flow leads to rapid cloud development at mid and high elevations which in turn acts as a negative feedback loop on the system. Thus ΔT is reduced by 1500 EAT (Fig. 11b). The consequence however of this process is transfer of moisture from the upper forest and heather zones up to the higher slopes and cloud formation in the ice-field vicinity.

During July the morning heating (Fig. 11c) is much more subdued. This is partly a result of a slightly lower solar angle (austral winter); a solar path less perpendicular to the slope (which faces south-west); and orbital forcing (aphelion) combined, despite a small contrast in

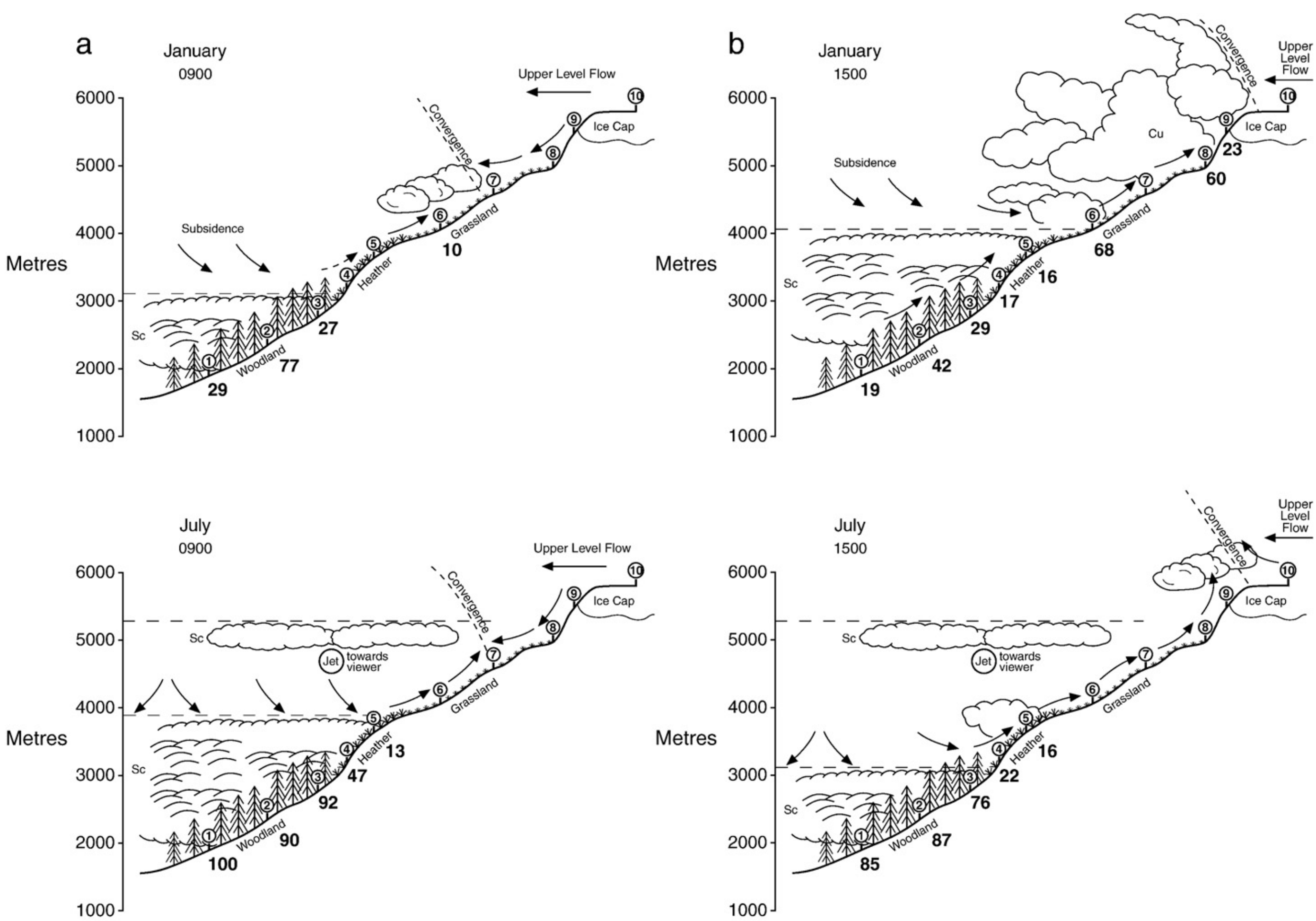

Fig. 11. Schematic diagram of the thermal circulation in a) January and b) July on the south-western slope of Kilimanjaro. Stations are represented by numbered circles. Figures beneath the stations represent the % of observations with relative humidity >95% for that time and month (broadly representative of the probability of cloud cover). Values less than 5% are omitted.

noon solar elevation. An additional synoptic contribution is the cool and moist EALLJ which brings in additional moisture at the lowest elevations. Importantly the upper forest and giant heather zones (sites 4 and 5) are often covered by early morning cloud at this season, the stratocumulus cloud deck typically extending from 2000 m to around 4000 m. This cloud limits the surface area of the mountain exposed to the morning sun, and thus any thermal upslope component would be confined to the upper slopes, at least in the first part of the day. Although this can be effectively developed (see the later discussion of observed surface winds in Fig. 12), the upslope flow is not tapping the main moisture source which is the forest lower down the mountain under the stratiform cloud layer. Vegetation is sparse above site 5 and *in situ* evapotranspiration small. Even though large temperature excesses are eventually reported in the afternoon (ΔT exceeding 4 °C at most upper sites by 1500 EAT unlike in January–Fig. 11d), much less moisture is brought up the mountain at this time of year and cloud formation at the higher sites is less frequent. Much of the solar flux in the vicinity of sites 4 and 5 is probably used in evaporating the stratiform cloud layer, leaving less available to heat the surface. By the time the top of the stratiform cloud layer is lowered to be below the forest limit (afternoon) it is too late for extensive upslope moisture transport.

Fig. 12 shows field measurements of wind direction measured at sites 5 to 10 during a five day period in late July/early August 2008 which substantiates the model. Different days are overlaid on the same axis since the focus is the mean diurnal signal in wind direction. 360° has been added to winds with an easterly component to make any daily progression clearer. Although there is some inter-diurnal variability, at all sites (with the possible exception of site 10) there is a consistent backing of the surface wind from a more northerly direction to a more westerly direction during daylight hours, showing consistent development of upslope winds. Site 5 shows a more gradual and limited shift than other sites, with upslope winds not reaching their full development until much later (usually around 1200 EAT) in comparison with higher sites. At sites 6 and 7 there is a more pronounced shift between 900 and 1000 EAT. So for much of the morning at this season the forest and heather zones (sites 5 and below) remain disconnected from the developing thermal circulation on the upper slopes. Site 9 has the most rapid and distinct upslope/downslope transition, probably because it is on the steepest part of the crater wall. Sites 7 and 8 are slightly more complex, with some inter-diurnal variation in the daytime development of upslope winds.

Wind speeds at all sites (not shown) increase on average at all sites during the afternoon, usually peaking around 1600 EAT but there is much inter-diurnal variability. However, occasional strong winds can occur at night, and the downslope nocturnal component can be particularly strong at sites 8 and 9 (>10 m/s) where the slope gradient is steep.

4.2. Comparison of field observations with model results

Mölg et al. (2009a) simulate precipitation patterns on the mountain using a dynamical model, the input parameters for which

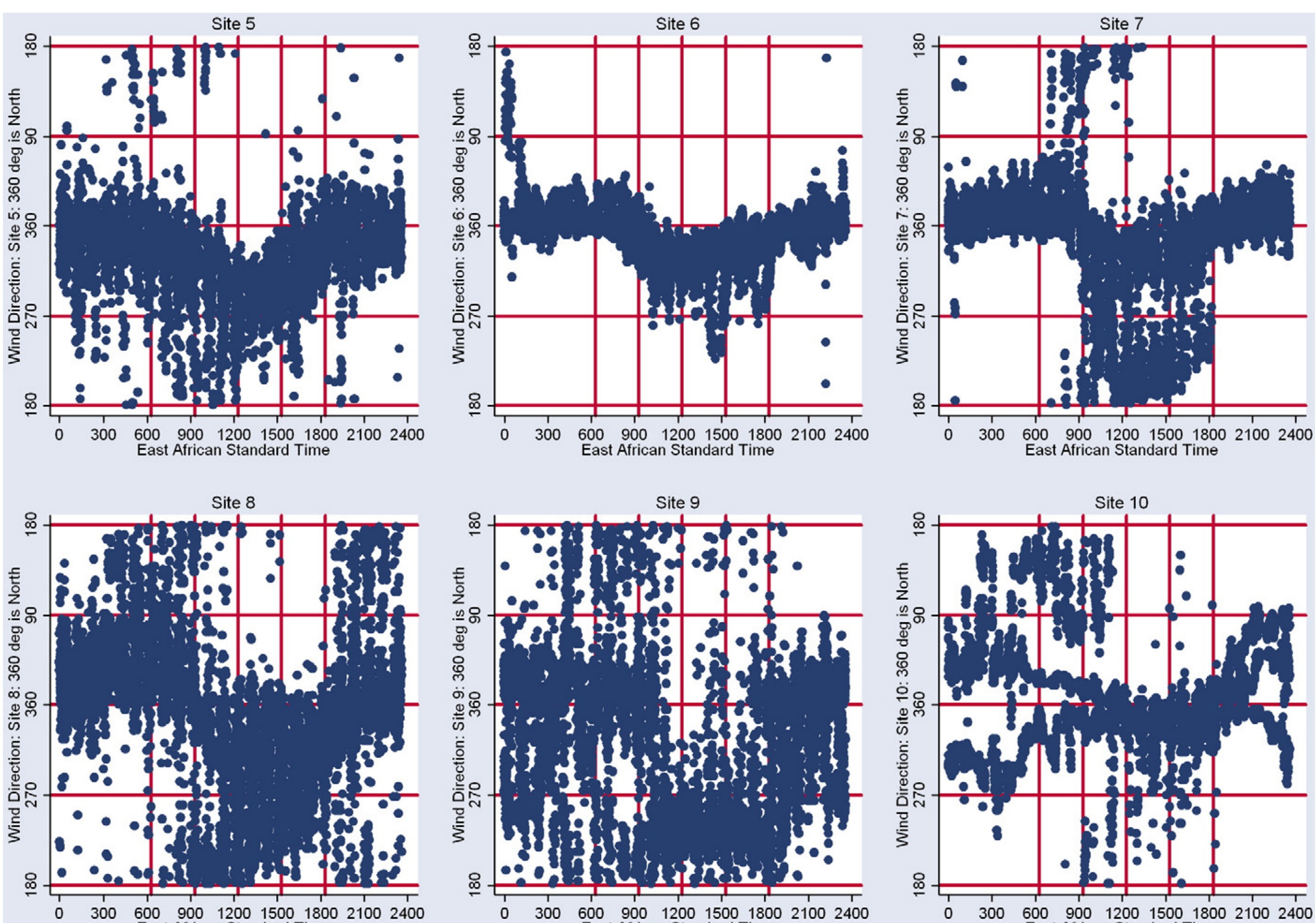

Fig. 12. Mean diurnal signal in wind direction at each of the six upper sites (sites 5–10) for 29 July–6 August 2008 (see Table 2). The red vertical lines represent 3 h periods between sunrise (0638 EAT) and sunset (1837 EAT).

are derived from characteristic background flow temperature and moisture profiles observed during heavy precipitation events at the summit. Although the conceptual models outlined above are based on mean monthly conditions and Molg et al. separate heavy, moderate precipitation events and dry conditions, it is helpful to assess how our findings interrelate.

The above authors find a correlation between heavy precipitation events on the summit of Kilimanjaro and free-air moisture content between 500 and 700 mb upstream (north-east) of the mountain. Our work does not contradict this finding. In fact we find similar correlations between summit vapour pressure and free-air moisture content on a climatological basis in our data. However, the relationship is weakest at 1500 EAT ($r = 0.456$ between free-air vapour pressure integrated from 3000–5800 m and summit vapour pressure), suggesting that any supplementary slope source is most relevant at this time. Conversely, higher correlations at other times of day suggest a reduced surface contribution at that time.

Mölg et al. suggest that it is the dynamics of the circulation which encourages a counter-flow to develop on the south-western forested slope. This counter-flow transports moisture upslope and results in precipitation at higher elevations than average when conditions are conducive for heavy precipitation events. At the same time however, it is acknowledged that surface heating is an important aid to convective development in the afternoon simulations. Our data in general support this analysis but there are some clarifications which arise from our observations.

Upslope moisture transport in our data typically only occurs during the afternoon, meaning that the thermal forcing of this process is important. Dynamical forcing alone does not tend to produce heavy precipitation on the higher slopes, particularly since Kilimanjaro is an isolated peak and the flow can go around, rather than over the crater, in nearly all circumstances (Schär, 2002). Conversely though, the development of a thermal upslope circulation alone (as in our July field data) is not sufficient to transport moisture upslope automatically, particularly if it is only developed at higher elevations above tree-line. Therefore, a combination of dynamical and thermal forcing is probably responsible for the moistening of the summit area, and the downward extent of any counter-flow is probably critical, as suggested by Mölg et al. (2009a).

We also examine observed vapour pressure conditions at all our locations for the same occasions of heavy precipitation at summit level as defined by Mölg et al. (2009a). For the 14 days listed in Molg's paper as significant summit precipitation events (in their Fig. 2), we calculated background and surface vapour pressures at all our sites. We use the whole day's records to increase sample size to 56. A significant increase above all other days in both surface and reanalysis values is shown at all locations. However the surface increase is larger than the simultaneous free-air increase at all sites, and a distinct excess “vapour trail” independent of the free-air contribution is present leading back to the forest zone (Table 5). At all sites the surface vapour excess (ΔVP) above free-air values (column 3) therefore increases during these heavy precipitation events, even more so at low elevations. Free-air flow strength on the other hand shows strengthening at low elevations, but weakening at the highest elevations (site 7 and above). Thus relative moistening of the mountain surface under such conditions occurs at all elevations.

Table 5
Mean surface and free-air vapour pressures (and their difference ΔVP), and free-air flow strength (*F* – m/s) for days of significant summit precipitation (defined by Mölg et al., 2009a,b – Fig. 2) in comparison with all other days. Figures in brackets represent the long-term mean based on all other days.

Logger no.	Surface VP (mb)	Free-air VP (mb)	ΔVP (mb)	*F* (m/s)
1	–	14.4 (12.8)	–	4.84 (4.91)
2	15.0 (13.4)	10.7 (10.0)	4.3 (3.3)	4.82 (4.39)
3	12.9 (11.4)	8.8 (8.1)	4.1 (3.2)	4.96 (4.26)
4	11.3 (9.6)	7.3 (6.7)	4.0 (2.9)	5.21 (4.52)
5	9.7 (7.7)	6.0 (5.0)	3.7 (2.8)	5.03 (4.59)
6	8.2 (6.6)	5.7 (4.4)	2.5 (2.2)	4.99 (4.89)
7	–	4.5 (3.0)	–	5.02 (5.27)
8	5.6 (3.7)	3.8 (2.4)	1.8 (1.2)	4.94 (5.33)
9	4.3 (2.6)	3.1 (1.8)	1.2 (0.8)	5.29 (5.71)
10	3.9 (2.2)	2.7 (1.5)	1.2 (0.7)	5.70 (6.10)

5. Summary and discussion

Our study has examined temperature and humidity records from ten locations up the south-western slope of Kilimanjaro and has compared them with simultaneous free-air measurements derived from NCEP/NCAR reanalysis.

1) Background flow conditions show an equable climate with little temperature variation and dry air at high levels. Diurnal changes are small.
2) The mountain slope and summit climate are substantially different. In particular there is strong surface heating during the daytime (particularly above the treeline which lies around 3000 m) and strong surface moisture excesses develop at all sites in the morning, reaching their maximum during the afternoon.
3) The forest zone is the source region for this moisture, excesses declining systematically with increased distance upslope. This moisture is sometimes taken up the mountain by upslope flow. This process is most successful during January but weaker during July because any upslope circulation appears restricted to the upper slopes for much of the morning.
4) Synoptic conditions such as upper level wind speed, which could in theory provide opposition to this process, appear to have limited influence on the inter-diurnal variability of moisture excess at the higher sites on the mountain. The lack of a simple inverse relationship between dynamical and thermal forcings, suggests a more complex interaction between the two than commonly envisaged.

There have been many studies of free-air properties in northern Tanzania and over the Indian Ocean upstream to the east (Latif et al., 1999; Mölg et al., 2006), in particular examining the roles of changes in moisture advection and the influence on rainfall (Kabanda and Jury, 1999). It has also been shown that ENSO influences free-air properties in the region through changes in sea-surface temperatures (Ropelewski and Halpert, 1987; Plisnier et al., 2000). Most recently, heavy precipitation events on the summit are shown to be correlated with upstream free-air moisture content (Mölg et al., 2009a). It is becoming clearer therefore that the highest elevations of Kilimanjaro are extremely sensitive to small changes in free-atmospheric moisture.

Although background free-air changes are important, our study also shows the strong influence of the mountain surface on summit climate. Free-air properties alone cannot account for all of the summit variability. This makes long term changes in surface properties (as could be influenced by local land-use change such as deforestation) a possible contributor to the long-term summit ice-field decline, as well as regional-scale drying which would be caused by a change in synoptic conditions. It is clear therefore that the role of the rainforest in supplying moisture to the upper reaches of Kilimanjaro needs much further examination.

At the summit crater the afternoon vapour pressures reach over twice the level of the ambient free atmosphere. This means that locally sourced vapour (either from the ice as sublimation or brought upslope) is at least as important as any direct free-air source. That this is also the case at sites 8 and 9 (away from, below and upwind of the ice-field) suggests that the lower slopes (and not the ice field itself) are the dominant source of this excess moisture. A depression of the upper treeline on Kilimanjaro which has been quantified through aerial survey (Lambrechts et al., 2002) and satellite analysis (Hemp, 2005) could therefore in theory have important consequences as it would move the moisture reservoir lower down the mountain and further away from the summit area, thus reducing the supply of moisture at crater level.

The current decline in vapour pressure excess (ΔVP) with increasing distance from the forest zone is approximately linear. The current mean annual rate of decline is 1.48 mb/km, based on a comparison of sites 5 and 10. Although the figures are approximate, this translates to a loss of all the remaining afternoon excess at summit level (~1.4 mb) if this rate of decline were to remain fixed in the future and the treeline were to be depressed by 1000 m. Current estimates of treeline depression are reported as "several hundred metres" since 1976 (Hemp, 2005) and this could in theory therefore cause a significant reduction in moisture availability at summit level. Even though these figures are speculative, they do illustrate the potential importance of a continuous forest belt at moderate elevations on Kilimanjaro in sustaining upper level moisture.

To avoid misinterpretation of our findings we clarify what we can and cannot deduce from our current data. This primarily concerns the search for definitive moisture "sources", and attribution of moisture fluxes from our observations of atmospheric stores.

There are three main scales of moisture variability on the mountain

a) The large scale contrast between the mountain and the background flow (examined in this paper)
b) Vertical gradients at a point, which result from differential vegetation types
c) Horizontal gradients within an elevational zone, which result from differences in land-use.

While we have detailed data on temperature and moisture conditions in a regional transect moving away from the forest up to the crater (representative of the large scale mountain environment), we do not have the data to comment on b) and c). We deliberately kept our measurements at a common height (screen level) to facilitate large-scale comparison between mountain and background flow. Thus our lower site measurements are below the vegetation canopy (forest) and are influenced by it, but our alpine site measurements (sites 6 and 7) are well above the low-lying cushion vegetation, and less influenced by it. It is difficult therefore to define the exact source of the moisture at screen level at sites 6 and 7 for example. We suspect the lower slopes are important (reasons are outlined in Section 3.2.2.) but more work is required to substantiate the importance of various sources.

We also do not have direct measurement of the influence of local land-use change, such as deforestation, on a micro-scale (at a given elevation). Therefore we cannot claim that deforestation has occurred or that this has decreased moisture availability and is the overriding cause of summit ice-field decline. There is no time series component to this paper. We can only claim that, because of the importance of the mountain surface in supplying moisture, it has the potential to do so.

There is a clear need to expand the network of surface measurements and to:

a) Examine specific areas which have been deforested and compare the moisture regime in and around these areas to those areas which still have a healthy forest cover (preferably at the same elevation). This should include locations within the Chagga agroforest belt and at lower elevations outside the Kilimanjaro National Park.
b) Examine the influence of vegetation on the vertical moisture profile within each of the elevation zones.
c) Obtain longer term observations of surface wind speed and direction, including in seasons other than the south-east trade winds, to quantify the diurnal wind cycle and validate further our models of the circulation as presented here.
d) Expand our surface data to include other transects, especially on the northern and eastern sides of the mountain. All evidence suggests that the thermal circulation and its attendant moisture transport are probably not symmetrical, the northern and eastern slopes being much drier and more sparsely vegetated. Additionally any background flow is from the east and would act to supplement any daytime upslope circulation on that slope (instead of opposing it). Thus for the mountain as a whole the thermal/mechanical interaction is not fully understood.

The more we can understand how the land-use on the mountain influences the summit regime, and how thermal and mechanical forcings interact in moisture transport, the more impetus this will provide for local conservation efforts.

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References

Agrawala, S., Moehner, A., Hemp, A., van Aalst, M., Hitz, S., Smith, J., Meena, H., Mwakifwamba, S.M., Hyera, T., Mwaipopo, O.U., 2003. Development and climate change in Tanzania: focus on Mount Kilimanjaro. OECD Report, Paris, France. 72 pp.
Altmann, J., Alberts, S.C., Altmann, S.A., 2002. Dramatic change in local climate patterns in the Amboseli basin, Kenya. African Journal of Ecology 40, 248–251.
Bossert, J.E., Cotton, W.R., 1994. Regional-scale flows in mountainous terrain. Part I: A numerical and observational comparison. Monthly Weather Review 122, 1449–1471.
CDC, 2008. Climate Diagnostic Center NCAR/NCEP Reanalysis page, http://www.cdc.noaa.gov/cdc/data.ncep.reanalysis.html Accessed August 2008.
Chiang, J.C.H., Lintner, B.R., 2005. Mechanisms of remote tropical surface warming during El Nino. Journal of Climate 18, 4130–4149.
Coutts, H.H., 1969. Rainfall of the Kilimanjaro area. Weather 24, 66–69.
Cullen, N.J., Mölg, T., Kaser, G., Hussein, K., Steffen, K., Hardy, D.R., 2006. Kilimanjaro Glaciers: Recent areal extent from satellite data and new interpretation of observed 20th century retreat rates. Geophysical Research Letters 33, L16502. doi:10.1029/2006GL027084.
De Quervain, A., 1904. Die Hebung der atmosphärischen Isothermen in der Schweizer Alpen und ihre Beziehung zu deren Höhengrenzen. Gerlands Beitrage Zur Geophysik 6, 481–533.
Dreiseitl, E., 1988. Slope and free air temperature in the Inn Valley. Meteorology and Atmospheric Physics 39, 25–41.
Duane, W.J., Pepin, N.C., Losleben, M.L., Hardy, D.R., 2008. General characteristics of temperature and humidity variability on Kilimanjaro, Tanzania. Arctic, Antarctic and Alpine Research 40, 323–334.
Durre, I., Vose, R.S., Wuertz, D.B., 2006. Overview of the integrated global radiosonde archive. Journal of Climate 19, 53–68.
FAO, 2006. Global Forest Resources Assessment 2005: Progress towards sustainable forest management, Food and Agriculture Organisation Forestry Paper 147. Viale delle Terme di Caracalla, Rome, Italy. 348 pp.
Findlater, J., 1977. Observational aspects of the low-level cross-equatorial jet stream of the western Indian Ocean. Pure and Applied Geophysics 115, 1251–1262.
Garreaud, R., Vuille, M., Clement, A.C., 2003. The climate of the Altiplano: observed current conditions and mechanisms of the past. Palaeogeography, Palaeoclimatology, Palaeoecology 3054, 1–18.
Grabherr, G., Pauli, H., 1994. Climate effects on mountain plants. Nature 369, 448.
Hastenrath, S., 1984. The glaciers of Equatorial East Africa. Solid Earth Sciences Library, Dordrecht, Boston, Lancaster.
Hastenrath, S., 1991. Climate dynamics of the tropics. Kluwer, Dordrecht, Boston, London.
Hastenrath, S., 2001. Variations of East African climate during the past two centuries. Climatic Change 50, 209–217.
Hastenrath, S., Greischar, L., 1997. Glacier recession on Kilimanjaro, East Africa, 1912–89. Journal of Glaciology 43, 455–459.
Hastenrath, S., Polzin, D., Mutai, C., 2007. Diagnosing the 2005 drought in equatorial East Africa. Journal of Climate 20, 4628-4637.
Hedberg, O., 1964. Features of afroalpine plant ecology. Acta Phytogeographica Suecica 49, 1–144.
Hemp, A., 2002. Ecology of the pteridophytes on the southern slopes of Mt. Kilimanjaro – I. Altitudinal distribution. Plant Ecology 159 (2), 211–239.
Hemp, A., 2005. Climate change-driven forest fires marginalize the impact of ice cap wasting on Kilimanjaro. Global Change Biology 11, 1013–1023.

Hemp, A., 2006. Vegetation of Kilimanjaro: hidden endemics and missing bamboo. African Journal of Ecology 44, 305–328.
Indeje, M., Semazzi, F.H.M., Ogallo, L.J., 2000. ENSO signals in East African rainfall seasons. International Journal of Climatology 20, 19–46.
Kabanda, T.A., Jury, M.R., 1999. Inter-annual variability of short rains over northern Tanzania. Climate Research 13, 231–241.
Kalnay, E., et al., 1996. The NCEP/NCAR 40-year reanalysis project. Bulletin of the American Meteorological Society 77, 437–471.
Kaser, G., 1999. A review of the modern fluctuations of tropical glaciers. Global and Planetary Change 22, 93–103.
Kaser, G., 2001. Glacier–climate interaction at low latitudes. Journal of Glaciology 47, 195–204.
Kaser, G., Osmaston, H., 2002. Tropical glaciers. (International Hydrology Series). Cambridge University Press, Cambridge. 207 pp.
Kaser, G., Hardy, D.R., Mölg, T., Bradley, R.S., Hyera, T.M., 2004. Modern glacier retreat on Kilimanjaro as evidence of climate change: observations and facts. International Journal of Climatology 24, 329–339.
Kistler, R., et al., 2001. The NCEP/NCAR 50-year reanalysis. Bulletin of the American Meteorological Society 82, 247–268.
Kuemmel, B., 1997. Temp, humidity and dew point ONA. http://www.faqs.org/faqs/meteorology/temp-dewpoint/ accessed 7 May 2007.
Lambrechts, C., Woodley, B., Hemp, A., Hemp, C., Nnyiti, P., 2002. Aerial survey of the threats to Mt. Kilimanjaro forests. UNDP, Dar es Salaam.
Latif, M., Dommenget, D., Dima, M., Grötzner, A., 1999. The role of Indian Ocean SST in forcing East African rainfall anomalies during December/January 1997/98. Journal of Climate 12, 3497–3504.
Lintner, B.R., Neelin, J.D., 2007. A prototype for convective margin shifts. Geophysical Research Letters 34, L05812. doi:10.1029/2006GL027305.
Mölg, T., Hardy, D.R., 2004. Ablation and associated energy balance of a horizontal glacier surface on Kilimanjaro. Journal of Geophysical Research 109, D16104. doi:10.1029/2003JD004338.
Mölg, T., Hardy, D.R., Kaser, G., 2003. Solar radiation maintained glacier recession on Kilimanjaro drawn from combined ice-radiation geometry modelling. Journal of Geophysical Research 108 (D23), 4731. doi:10.1029/2003JD003546.
Mölg, T., Renold, M., Vuille, M., Cullen, N.J., Stocker, T.F., Kaser, G., 2006. Indian Ocean zonal mode activity in a multicentury integration of a coupled AOGCM consistent with climate proxy data. Geophysical Research Letters 33, L18710. doi:10.1029/2006GL026384.
Mölg, T., Cullen, N.J., Hardy, D.R., Kaser, G., Klok, L., 2008. Mass balance of a slope glacier on Kilimanjaro and its sensitivity to climate. International Journal of Climatology 28, 881–892.
Mölg, T., Chiang, J.C.H., Gohm, A., Cullen, N.J., 2009a. Temporal precipitation variability versus altitude on a tropical high mountain: observations and mesoscale atmospheric modeling. Quarterly Journal of the Royal Meteorological Society 135. doi:10.1002/qj.461.
Mölg, T., Cullen, N.J., Hardy, D.R., Winkler, M., Kaser, G., 2009b. Quantifying climate change in the tropical mid troposphere over East Africa from glacier shrinkage on Kilimanjaro. Journal of Climate 22, 4162–4181.
Pepin, N.C., Losleben, M., 2002. Climate change in the Colorado Rocky Mountains: free air versus surface temperature trends. International Journal of Climatology 22, 311–329.
Pepin, N.C., Norris, J.R., 2005. An examination of the differences between surface and free-air temperature trend at high-elevation sites: relationships with cloud cover, snow cover, and wind. Journal of Geophysical Research 110, D24112. doi:10.1029/2005JD006150, 200.
Pepin, N.C., Seidel, D.J., 2005. A global comparison of surface and free-air temperatures at high elevations. Journal of Geophysical Research 110, D03104. doi:10.1029/2004JD005047.
Peppler, W., 1931. Zur Frage des Temperaturunterschiedes zwischen den Berggipfeln und der freien Atmosphäre. Beiträge Zur Physik der Atmosphäre 17, 247–263.
Plisnier, P.D., Serneels, S., Lambin, E.F., 2000. Impact of ENSO on East African ecosystems: a multivariate analysis based on climate and remote sensing data. Global Ecology and Biogeography 9 (6), 481–497.
Richner, H., Phillips, P.D., 1984. A comparison of temperatures from mountaintops and the free atmosphere: Their diurnal variation and mean difference. Monthly Weather Review 112, 1328–1340.
Roe, G.H., 2005. Orographic precipitation. Annual Review of Earth and Planetary Sciences 33, 645–671.
Rohr, P.C., Killingtveit, T., 2003. Rainfall distribution on the slopes of Mt. Kilimanjaro. Journal of Hydrological Sciences 48 (1), 65–77.
Ropelewski, C.F., Halpert, M.S., 1987. Precipitation patterns associated with El Niño/Southern Oscillation. Monthly Weather Review 115, 1606–1626.
Schär, C., 2002. Mesoscale mountains and the larger scale atmospheric dynamics: a review. In: Pearce, R.P. (Ed.), Meteorology at the Millenium. Academic Press, New York, pp. 29–42.
Shugart, H.H., French, N.H.F., Kasischke, E.S., Slawski, J.J., Dull, C.W., Shuchman, R.A., Mwangi, J., 2001. Detection of vegetation change using reconnaissance imagery. Global Change Biology 7, 247–252.
Soini, E., 2005. Land use change patterns and livelihood dynamics on the slopes of Mt. Kilimanjaro, Tanzania. Agricultural Systems 85, 306–323.
Tabony, R.C., 1985. The variation of surface temperature with altitude. Meteorological Magazine 114, 37–48.
Thompson, L.G., Mosley-Thompson, E., Davis, M.E., Henderson, K.A., Brecher, H.H., Zagorodnov, V.S., Mashiotta, T.A., Lin, P.-N., Mikhalenko, V.N., Hardy, D.R., Beer, J., 2002. Kilimanjaro ice core records. Evidence of holocene climate change in tropical Africa. Science 298, 589–593.
Thompson, L.G., Brecher, H.H., Mosley-Thompson, E., Hardy, D.R., Mark, B.G., 2009. Glacier loss on Kilimanjaro continues unabated. Proceedings of the National Academy of Sciences 106 (47), 19770–19775.
Troll, C., Wien, K., 1949. Der Lewisgletscher am Mount Kenya. Geografiska Annaler 31, 257–274.
Whiteman, C.D., 1990. Observations of thermally developed wind systems in mountainous terrain. Atmospheric Processes over Complex Terrain, Meteor. Monogr., No. 45. Amer. Meteor. Soc. 5–42.
Whiteman, C.D., Hubbe, J.M., Shaw, W.J., 2000. Evaluation of an inexpensive temperature datalogger for meteorological applications. Journal of Atmospheric and Oceanic Technology 17 (1), 77–81.