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Is the decline of ice on Kilimanjaro unprecedented in the Holocene?

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Abstract

Glaciers on Kilimanjaro's highest peak, Kibo, are currently regarded as a persistent feature of the Holocene. Here we synthesize all available measurements, observations, and our understanding of current processes on Kibo – gained from intensive research over the past decade – to formulate an alternative hypothesis about the age of these ice fields. This suggests a shorter, discontinuous history of the tabular-shaped glaciers on Kibo's plateau, where typical 'life cycles' of the ice may last only a few hundred years. If life cycles overlap, they are likely the cause of the observed steps in the plateau glaciers. Thus, it is likely that ice has come and gone repeatedly on Kibo's summit plateau, throughout the Holocene. Such a cyclicity is supported by lake-derived proxy records.

Keywords

climate, glaciers, Holocene, hypothesis, Kilimanjaro, Tropics

Introduction

Knowing the history of a glacier or an ice cap allows insight into past climate conditions at a particular site, as well as into regional and larger-scale palaeoclimate dynamics. In the case of glaciers on high mountains at low latitudes careful examination reveals the history of tropical or subtropical mid-troposphere climate (e.g. Hardy *et al*., 1998; Kull and Grosjean, 2000; Mölg *et al*., 2009a; Wagnon *et al*., 2003), where climatic data and proxy records are very rare (e.g. Trenberth *et al*., 2007).

Besides, or in addition to, mapping and dating of moraines and other pro-glacial features (e.g. Osmaston, 1989; Patzelt *et al*., 1984; Rabatel *et al*., 2006; Smith *et al*., 2005; Solomina *et al*., 2007), analyses of ice cores taken from appropriate sites are increasingly used to reconstruct ice and climate history at low-latitude high mountain sites (e.g. Thompson *et al*., 2005). Glacier–climate interactions in the low latitudes are primarily related to varying hygric conditions (Favier *et al*., 2004; Francou *et al*., 2003, 2004; Hastenrath, 2010; Hastenrath and Greischar, 1997; Kaser, 2001; Kaser and Georges, 1997; Kruss and Hastenrath, 1987; Mölg and Hardy, 2004; Mölg *et al*., 2003a, b, 2006a, b, 2008a, 2009a; Vuille *et al*., 2008a, 2008b; Wagnon *et al*., 2001) and, thus, particular care needs to be taken when analyzing and interpreting ice cores (Ginot *et al*., 2001; Hardy *et al*., 2003; Schotterer *et al*., 2003; Seimon, 2003; Stichler *et al*., 2001). Direct observations and measurements of present-day conditions and governing processes, for instance, allow for a more rigorous interpretation of reconstructed glacier extents and ice core records.

Following the spatial delimitations proposed by Kaser *et al*. (1996), tropical glaciers still exist in Irian Jaya (New Guinea), in the South American Andes, and in East Africa. Among them, the glaciers on Kilimanjaro's highest peak, Kibo, have attained particular attention, not least since Irion (2001) attributed the modern time changes to increased air temperature in the context of global warming and an ice-core analysis by Thompson *et al*. (2002) proposed the near extinction of the ice on Kibo as being unprecedented over the last 11 700 years.

In 2001 and 2002, we compiled a comprehensive understanding of Kibo's glaciers, and on the processes governing their persistent shrinkage since the late nineteenth century (Kaser *et al*., 2004). This effort resulted in the formulation of a hypothesis about the glacier–climate interaction on Kibo, and subsequent joint field investigations by Innsbruck and Massachusetts Universities resulted in the installation and ongoing maintenance of three automatic weather stations. Results to date corroborate the 2004 hypothesis of governing processes, indicating that atmospheric moisture primarily controls the modern-time glacier changes on Kibo by dynamics across different spatio-temporal scales in the climate system (Cullen *et al*., 2006, 2007; Mölg and Hardy, 2004; Mölg *et al*., 2003b, 2006a, 2009a, c). This evidence not only rules out rising local air temperature (i.e. on the peak of Kibo) as the main driver of observed changes during the last 120 years, but also puts the currently accepted 11 700 years age in question.

Here we present an alternative hypothesis about the age of the Kilimanjaro glaciers and their Holocene history. We first briefly

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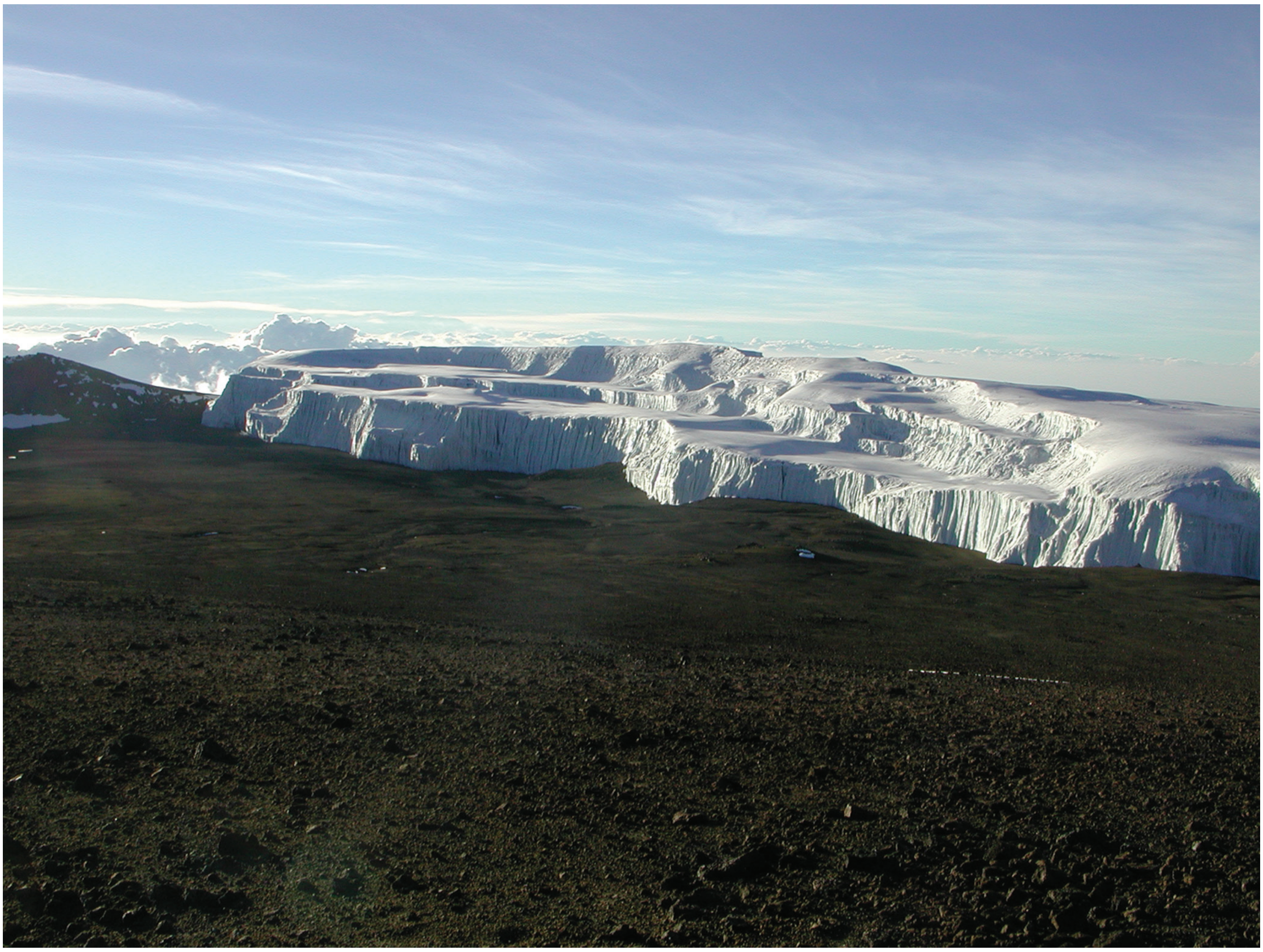

Figure 1. Tabular ice on Kibo's summit plateau. The Northern Icefield (NIF) as seen from the rim of the Reusch crater. Note the characteristic steps (photo: NJ Cullen, July 2005)

review the ice core dating (Thompson et al., 2002) and uncertainties that arise (following section); Gasse (2002) has previously discussed chronology issues in the ice core record. Then we compile evidence to construct our hypothesis in sections 'An extrapolation from observed areal glacier changes – deriving a characteristic decay time period'; 'How can plateau glaciers grow?'; and 'A millennium-scale plateau glacier history and beyond'. The 'Summary and conclusions' section summarizes the presented discussion and identifies the open questions.

The present view on past and future glaciers on Kilimanjaro

The ice history from ice-core analyses

In 2000, several ice cores were taken from the flat parts of Kilimanjaro's Northern Icefield (NIF), Southern Icefield (SIF) and Furtwängler Glacier (FWG) (Thompson *et al*., 2002), three distinct and particular ice entities situated on the summit of Kibo. Thompson *et al*. (2002) reach the following major conclusions:

(1) The Northern Icefield on top of Kibo, Kilimanjaro, began to grow ~ 11.7 ka ago.
(2) The ice core mirrors cycles of abundant precipitation and droughts in East Africa.
(3) 'The disappearance of Kilimanjaro's ice fields is expected between 2015 and 2020'.
(4) This 'will be unprecedented for the Holocene'.

We underscore here that, despite not clearly stated by Thompson *et al*. (2002), the discussed age issue exclusively concerns the tabular-shaped plateau glaciers (Figure 1), but not the slope glaciers on Kibo's steep flanks. The differences between the two glacier systems are important (e.g. Cullen *et al*., 2006; Kaser *et al*., 2004) and are discussed in more detail in sections 'The future of the glaciers' and 'An extrapolation from observed areal glacier changes – deriving a characteristic decay time period'.

To establish the past to present chronology in the ice core, Thompson *et al*. (2002) assign three time horizons to specific depths in the reference ice core (NIF3). In the uppermost section, elevated ^{36}Cl concentration from a thermonuclear bomb test serves to precisely identify the 1952 layer (Gasse, 2002), which is used as a starting point to develop a depth–age relation.

At the bottom of the core, the basal age of 11.7 ka is assigned by 'comparing the NIF3 $\delta^{18}O$ record with the precisely dated … speleothem record from Soreq Cave … in the eastern Mediterranean region' (Thompson *et al*., 2002). However, small samples of organic material from the bottom region (47.9–48.97 m) yield ages of 9.36, 6.7, and 4.09 ka before AD 1950 (Thompson *et al*., 2002: table S1). Gasse (2002) attribute this scatter to low mass samples, but it is also unclear (1) whether the organic material was deposited at the ice surface directly after death of the plant/animal or preserved elsewhere before deposition on the ice (the scatter of ages points to the latter), and (2) why the organic samples are considerably younger than the assigned basal age.

Between the top and bottom time horizons two large $\delta^{18}O$ depletions are identified in 'the upper part of the Kilimanjaro $\delta^{18}O$ record' (Thompson *et al*., 2002). The date AD 1325 is assigned to the earliest and largest depletion, which coincides with the Wolf solar minimum and a highstand of Lake Naivasha. Interestingly, two solar minima (Spörer and Maunder) can be observed to have taken place after the Wolf solar minimum, which are both of a higher magnitude than the Wolf and feature higher lake levels as well (Thompson *et al*., 2002: figure 3), yet neither of the two large $\delta^{18}O$ depletions was assigned to the Spörer solar minimum. This, and the absence of a solar minimum for the most recent highstand of Lake Naivasha in the late nineteenth century, suggest that the 'close association' (Thompson *et al*., 2002) between lake levels, solar minima, and $\delta^{18}O$ minima is somewhat tenuous (Gasse, 2002).

These three time horizons (as described above) are used to determine a mean annual accumulation rate (c_r), which is held constant to allow a Nye age model for a steady-state glacier to be applied (Thompson *et al*., 2002). Apart from the fact that our measurements on Kibo clearly show that c_r is highly variable and therefore very unlikely to be constant through time (e.g. Mölg and Hardy, 2004; Mölg *et al*., 2009a, c), a different assignment of time horizons would also change c_r and thus profoundly impact the age and history of the ice core.

The end result from the ice core analysis is an attractive paleoclimate history that agrees with proxies of environmental and social dynamics (Thompson *et al*., 2002), an agreement which is not unexpected – as the NIF3 record was assimilated to the well-dated Soreq Cave record. On the other hand, there is uncertainty in the dating as indicated above, which needs no further mention here since the point has already been made by Gasse (2002). This means, however, that such uncertainties leave room for an alternative view, which we exploit in the present study to stimulate more discussion about the age of plateau glaciers on Kibo.

The future of the glaciers

Thompson *et al*. (2002) project the disappearance of Kibo ice between 2015 and 2020 as determined from a linear extrapolation of not fully linear shrinkage rates derived from former glacier area mappings. Cullen *et al*. (2006) scrutinized this interpretation more carefully and reiterated that glaciers on Kibo are basically of two different kinds, as first recognized by Geilinger (1936): the tabular-shaped plateau glaciers and the slope glaciers (further discussion in section 'An extrapolation from observed areal glacier changes – deriving a characteristic decay time period'). Whereas the latter are subject to gravitational movement along the slopes, the former are on mostly flat ground and too shallow for any considerable ice movements. Thus, ice dynamics are negligible and the plateau ice masses are an undistorted reflection of climate conditions. Their areal shrinkage rate has been almost linear since 1912 (Cullen *et al*., 2006, discussed in section 'An extrapolation from observed areal glacier changes – deriving a characteristic decay time period') and has most likely been this way since about 1880 (Lemke *et al*., 2007; Mölg *et al*., 2003b), which implies a constant driving mechanism has been in place that has changed little over the past 120 years. Both conceptually (Geilinger, 1936) and from modelling, the energy from solar radiation absorbed at the ice cliffs appears to be such a driver (Kaser *et al*., 2004; Mölg *et al*., 2003b), leading to a constant lateral recession and, thus, constant areal shrinkage of the plateau glaciers. Mölg *et al*. (2003b) simulated the areal shrinkage of a plateau ice cap that covers the summit plateau in 1880, using solar radiation as the only forcing. The modelled pattern of 2000 clearly resembles the observed one, and the last remnants of the plateau ice disappear in model year 2046 (in their basic scenario A). Slope glaciers, in turn, show a smaller tendency to vanish even in the dry conditions of the present regional climate (Geilinger, 1936; Cullen *et al*., 2006; Mölg *et al*., 2008b), so a complete deglaciation of Kilimanjaro in the next decade is not certain at all.

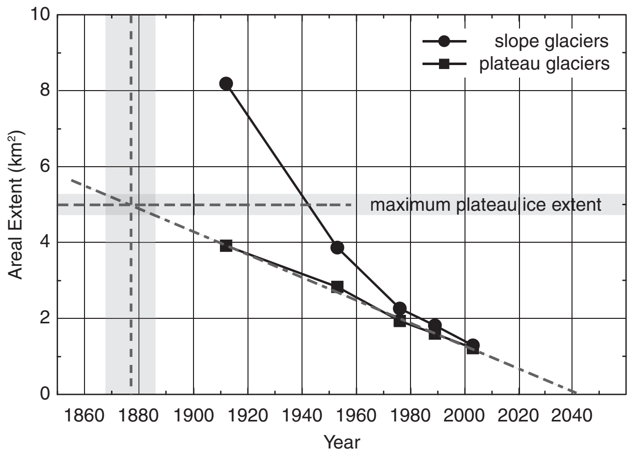

Figure 2. The change in horizontally projected surface areas of slope glaciers and plateau glaciers. Solid lines are from measured values as compiled by Cullen *et al.* (2006). Extensions from the observed plateau glacier extents are made (dashed line) to the approximate area of the summit plateau and to the extinction of the plateau ice, respectively

An extrapolation from observed areal glacier changes – deriving a characteristic *decay time period*

Whereas the slope glaciers show a constantly decelerating surface area loss between 1912 and 2003, the plateau glaciers shrink at a constant rate (Cullen *et al*., 2006, and observed values in Figure 2). Despite the limited number of observation dates one would expect them to reflect major climate fluctuations but neither one nor the other resolve twentieth-century variations, such as a near equilibrium state around 1970, as most glaciers worldwide do, including those in the tropics (Cogley, 2005; Dyurgerov and Meier, 2005; Greene, 2005; Kaser, 1999; Kaser *et al*., 2006; Lemke *et al*., 2007; Ohmura, 2004). Cullen *et al*. (2006) interpret the observed area shrinkage rates as resulting from a strong offset of glaciers from equilibrium some time prior to 1912, exposing them to a considerably and persistently drier condition than before. The change from wetter to drier conditions after the late nineteenth century was most probably related to a rather rapid change of sea surface temperature gradients in the Indian Ocean, which decreased the frequency of easterly winds in the central Indian Ocean and moisture transport to East Africa considerably (Hastenrath, 2001; Mölg

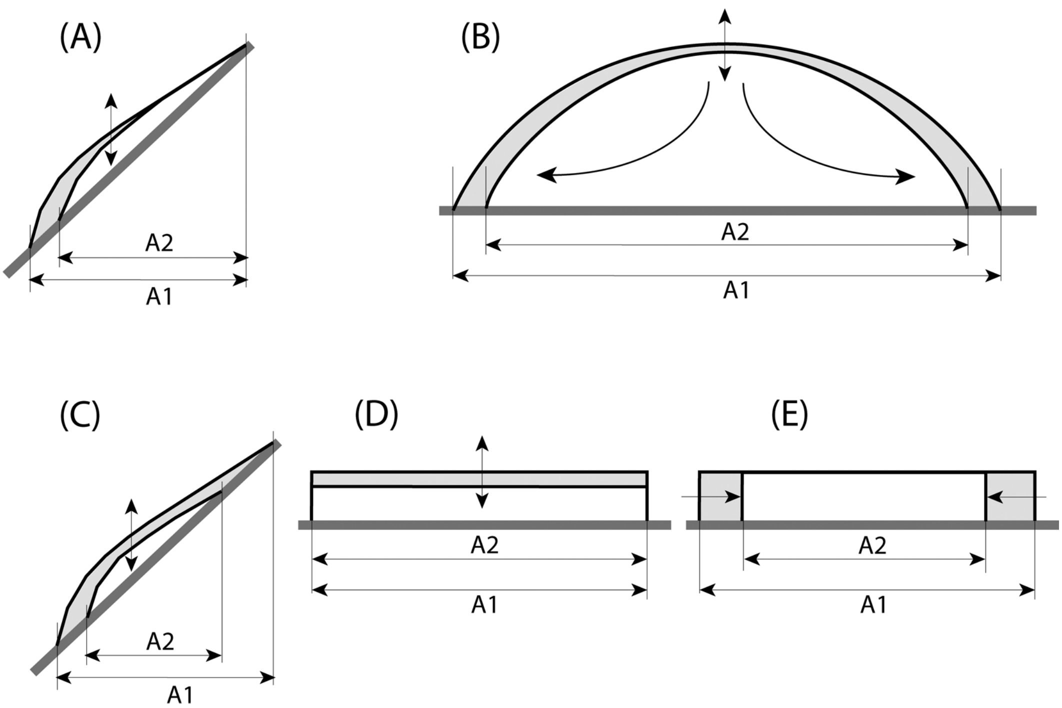

Figure 3. Changes of horizontally projected surface areas from A_1 to A_2 (as e.g. displayed on maps) from different changes in mass balance on a mountain glacier (A), an ice cap (B), the slope glaciers of Kibo (C) and the ice tables on Kibo summit (D) and (E). The bent arrows in (B) indicate ice flux by deformation that keeps the ice cap dome shaped. Ice tables on Kibo are too thin to be deformed this way. Slope glaciers on Kibo (C) show shrinkage from both the tongue and the top, possibly toward a smaller size around an equilibrium line. See Kaser *et al.* (2004) for more details

et al., 2006a). During the most recent decades, Indian Ocean warming appears to have been responsible for the continuation of disrupted moisture supply from the ocean (Funk *et al*., 2008). In fact, the combination of measurements and backward energy and mass balance modelling has indicated that the late nineteenth-century maximum extent of Kersten glacier on Kibo's southwestern slope resulted mainly from 160–240 mm higher annual precipitation amounts than at present (Mölg *et al*., 2009a). The geometry of the observed area changes on Kibo also indicates drying as a driver on Kibo. Slope glaciers typically adjust to drier conditions by shrinking from both tongue and top (see Figure 3C). This is the case with the slope glaciers on Kibo (Kaser *et al*., 2004; Mölg *et al*., 2009a), yet if warmer air conditions were the driver, glaciers would mainly shrink from the tongue (Figure 3A).

For reconstructing glacier history, the history and age of the plateau glaciers is the focus, as with Thompson *et al*. (2002) (see section 'The present view on past and future glaciers on Kilimanjaro' above). The rather steep slope glaciers even if a more persistent feature on Kibo, cannot contain old ice because even under prevailing dry conditions their mass turnover is large relative to a small storage volume. As a first reconstruction step, without questioning its validity for the time being, we extrapolate from the constant rate of mapped changes to both a maximum possible extent of the plateau glaciers in the past (5 ± 0.25 km^2, see section 'The potential maximum extent of plateau glaciers' below) and the extinction of the plateau glaciers in the future (Figure 2). From this exercise we reach a date of onset of the present shrinkage at around 1875 ± 10 years. If the shrinking rate continues, the plateau glaciers will vanish around 2040. This results in a 165 ± 10 years *plateau glacier decay time period*, which is very similar to the 166 years obtained independently from the ice cap model by Mölg *et al*. (2003b) (see section 'The future of the glaciers' above). Note that the simple geometric extrapolation in Figure 2 is covered by measurements for the middle 50% time span and extends to 25% extrapolation in both directions only. Nevertheless, we need to examine the validity of the assumptions implicit here.

- Is the plateau area correctly estimated and has a maximum plateau glacier really covered the entire plateau?
- How can the shrinkage rates be constant and can we assume the shrinking rate to be constant over the entire time period of 165 ± 10 years?
- If so, are the drivers of shrinkage also of a constant nature?
- Can constant drivers be extrapolated to potential previous decay cycles?

The potential maximum extent of plateau glaciers

As a first approximation, a maximum plateau glacier extent has been determined using the 5700 m contour line as a maximum outer boundary and the rim of the Reusch Crater as an inner boundary (Cullen *et al*., 2006; Mölg *et al*., 2003b) (Figure 4), leading to a surface area of approximately 5.0 km^2. One can assume that inside the Reusch Crater a continuous geothermal heat flux (e.g. Spink, 1944) prevents ice from accumulating and provides one possible onset mechanism of the vertical ice walls (Kaser *et al*., 2004), as will be later discussed as a prerequisite to understand the plateau glaciers. However, it is difficult to give a precise number for the plateau surface area and the maximum plateau ice extent. First of all, the plateau is not perfectly horizontal. This means that its edge, or boundary with the steep slopes, is not at exactly the same elevation everywhere. We also have no precise knowledge of the bed topography underneath the present-day ice. The deviation from a perfect horizontal also means that, in detail, small-scale

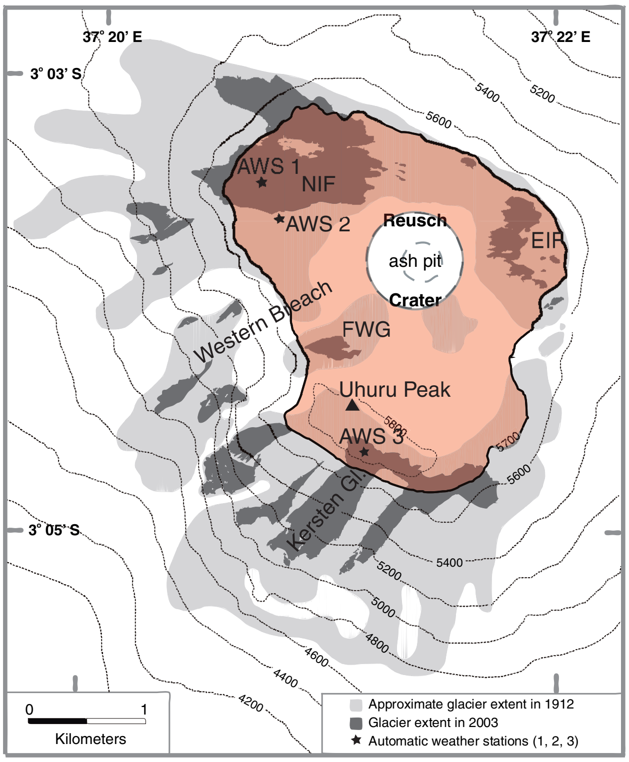

Figure 4. A simplified delineation of the near horizontal summit plateau following the 5700 m contour line and the rim of the Reusch Crater (transparent red). Glacier extents on Kibo are those in 1912 and 2003, respectively. NIF is the Northern Ice Field, EIF the Eastern Ice Field, and FWG Furtwängler Glacier (after Cullen *et al.*, 2006)

relief features may cause shading and exposure from solar radiation and wind, respectively, resulting in a scatter in maximum ice cover. For example, the steep rock cliffs beneath the Uhuru summit crest and from the crest between the western breach and the NIF onto the summit plateau experience much morning sun and have perhaps never supported glacier growth. In contrast, the Reusch crater cone provides shading in the depression southwest of it where the FWG is situated, which has resulted in a much longer survival period for FWG compared with other ice entities around the crater.

These uncertainties about a maximum plateau ice extent cannot be remedied, but are considered to be of a minor effect on the maximum plateau ice cover in view of our hypothesis. A size of 5.0 ± 0.25 km^2, as defined by the 5700 m contour line and the rim of the Reusch crater, seems to be a reasonably good estimate for the maximum horizontal extent the plateau glacier can reach. Minor deviations as discussed above would have little impact on the decay time length.

Linear shrinkage rates

While mountain glaciers or ice caps typically change their horizontally projected surface areas because of both ice movement and thinning or thickening of their ice bodies (see Figure 3A and B, respectively), Kibo's approximately 20–40 m thick tabular plateau glaciers do not (Kaser *et al.*, 2004). Gliding can be excluded for the near horizontal bed. Deformation cannot be totally disregarded and must play a role, but appears to be minor as it does not result in the usual shape of an ice cap being formed. Instead, the ice surface is largely parallel to the base, with some exceptions where geothermal hot spots cause basal melt (Kaser *et al.*, 2004), which result in caves. Deformation most likely closes caves over time, while depressions formed on the top of the ice as a result of this localized deformation are most likely levelled out by snow drift. As a consequence, distorted and non-continuous layering occurs on these sites. This, by the way, must also affect dating of any ice core that passes through such a location but this will not be further discussed here. The edges of the tabular ice bodies are bound by 70° to 90° steep walls and cliffs (Figure 3D, E). A change in thickness of the tabular ice would not cause a change in horizontally projected surface area until thickness reaches zero; only lateral changes result in decreases in areal extent (Figure 3D and E, respectively). In fact, despite a most probable thinning since 1962 (Hastenrath, 2010; Thompson *et al.*, 2002, 2009), shrinkage of the plateau glaciers over the past 120 years (from Hans Meyer's first visit on the summit plateau in 1889 up to present) are almost exclusively controlled by the retreat of ice along its walled margins (Kaser *et al.*, 2004). These ice walls are a particularly stable feature, at least within the climatic variations after the late nineteenth century. They have been described by the first observer (Meyer, 1900) and by subsequent visitors on the mountain, and remain one of the most striking features of this unique environment.

It is not trivial to understand why area shrinkage rates of Kibo's plateau glaciers are constant. In principle, objects of any shape change their surface area following a quadratic relationship if the retreat of all its edges is uniform (e.g. a cirque decreases its surface area quadratically if the radius decreases linearly). Only if the retreat takes place exclusively from either one or two mutually opposite sides, the relationship would be linear. This is not entirely true for the ice walls on Kibo that essentially exist (and retreat) on all sides of the glaciers. Yet, north- and south-facing walls dominate (Kaser *et al.*, 2004) and the retreat is dominated in normal direction to them. If they, in addition, retreat at near constant rates, areal shrinkage can be nearly linear. The constant retreat and respective drivers will be discussed next but another feature deserves attention first. If such an ice body disintegrates into several parts, more walls are created and areal shrinkage rates from edge retreat will increase (Downie and Wilkinson, 1972). When looking into the near future of the plateau glaciers, this accelerating component may become dominant toward the final decay of the ice, with minor implications on the decay time (less than 10 years).

Processes, drivers, sensitivities

Though we have a detailed understanding of the energy and mass balance of the horizontal surfaces of the NIF (Cullen *et al.*, 2007; Mölg and Hardy, 2004) and on the slope glaciers (Mölg *et al.*, 2008a, 2009a) we still do not completely understand the processes governing the existence of and ablation from the ice walls. However, data collected from AWS 2 (Figure 5, equipped in the same way as AWS 3 on the upper slopes of Kersten glacier (Mölg *et al.*, 2008a) but with radiation sensors oriented to the wall) have provided important insights into the atmospheric processes controlling the retreat.

Air temperatures and ice wall surface temperatures as derived from emitted long-wave radiation, as measured at AWS 2, indicate the importance of solar radiation on the wall itself (being consistent with former exploratory modelling: Mölg *et al.*, 2003b). Distinct differences between the season the wall is in its shadow and

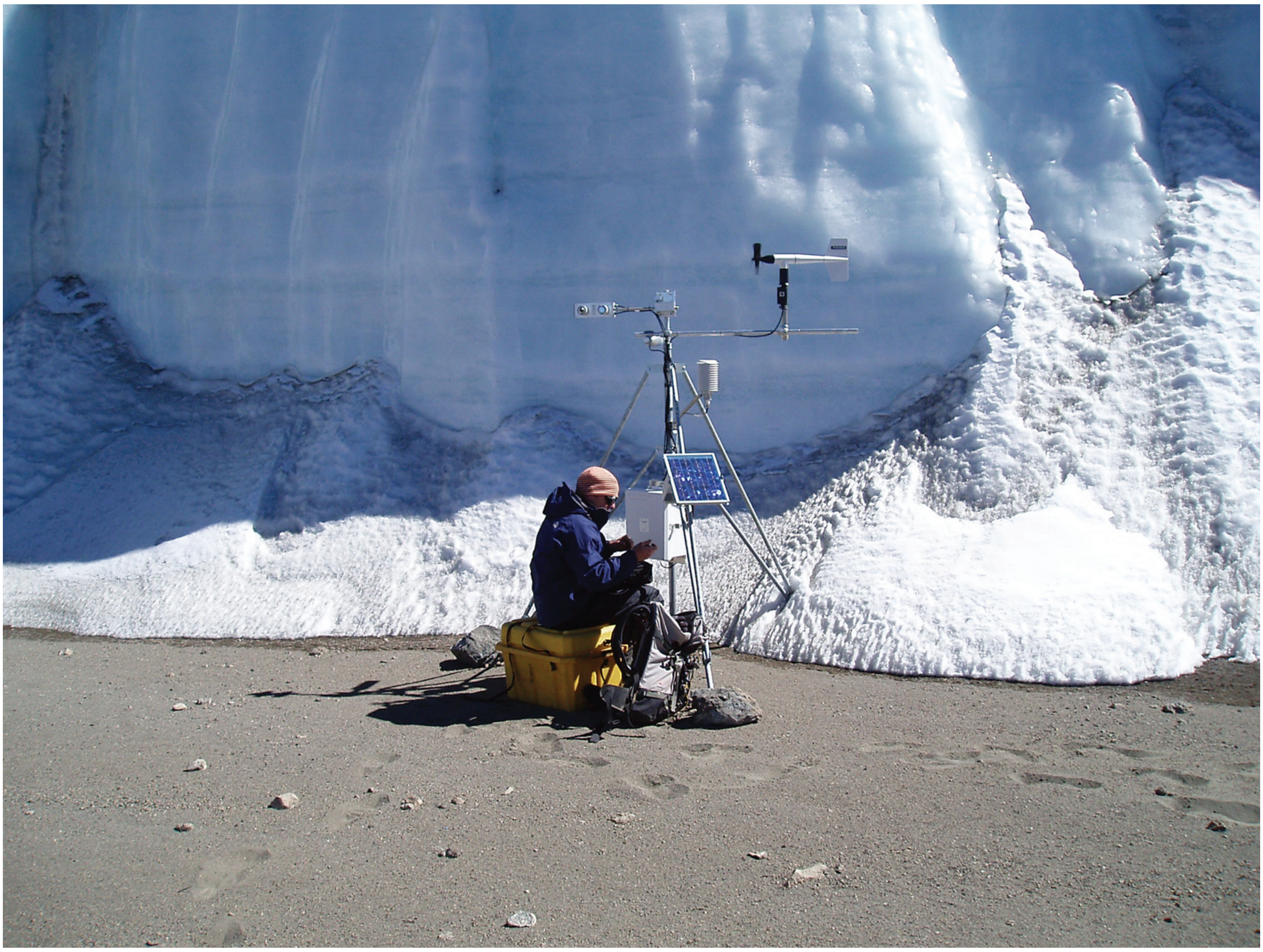

Figure 5. The automatic weather station AWS 2 has radiation sensors and a sonic distance ranger orientated normal to the south facing NIF ice wall (the position of AWS 2 is shown in Figure 4). It has been in operation since February 2005 (photo: G Kaser, July 2005 and NJ Cullen pictured working on AWS)

when the wall is irradiated by the sun are shown in Figure 6. The mean daily variations of air temperature do not vary much between the two seasons – reaching about −3 to −2°C around noon – and in both cases a minimum of about −9°C occurs shortly before sunrise. In contrast, the wall temperatures change considerably, with a crucial effect on ablation. Whereas during the shaded season the wall remains as cold as −7 to −15°C, it reaches melting conditions throughout several hours a day when direct solar radiation provides the required energy to the wall surface.

The ablation rates as measured from a sonic ranging sensor initially mounted 2.1 m from the ice wall (Figure 7 top) clearly reflect the energy conditions as portrayed by the ice wall surface temperature regimes. These rates are more than seven times higher during the direct sunlight season (−3.3 mm ice/day) compared with the shady season (− 0.47 mm ice/day) (Figure 7 top). Whereas it is exclusively sublimation during the period dominated by shade, melting and even enhanced sublimation (because the wall-air vapour pressure difference is larger when the ice wall is at melting point) account for higher ablation rates during direct sunlight. It is striking how constant the ablation rates are within each season. One reason for this can be inferred from the observed temperatures. In order for melting to occur during conditions characterized

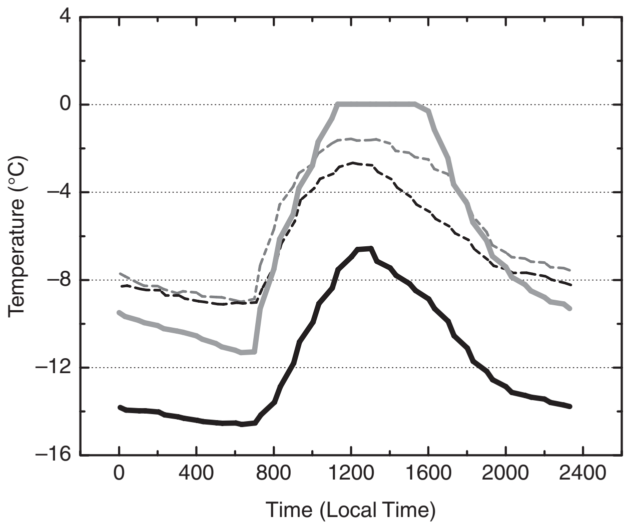

Figure 6. Mean diurnal cycles of air temperature (dashed lines) and ice surface temperatures (solid lines) measured at AWS 2 (ice wall station), contrasting representative intervals with the wall in shadow (black lines; 14 April to 26 July 2005), and when illuminated (grey lines; 8 February to 14 March 2005)

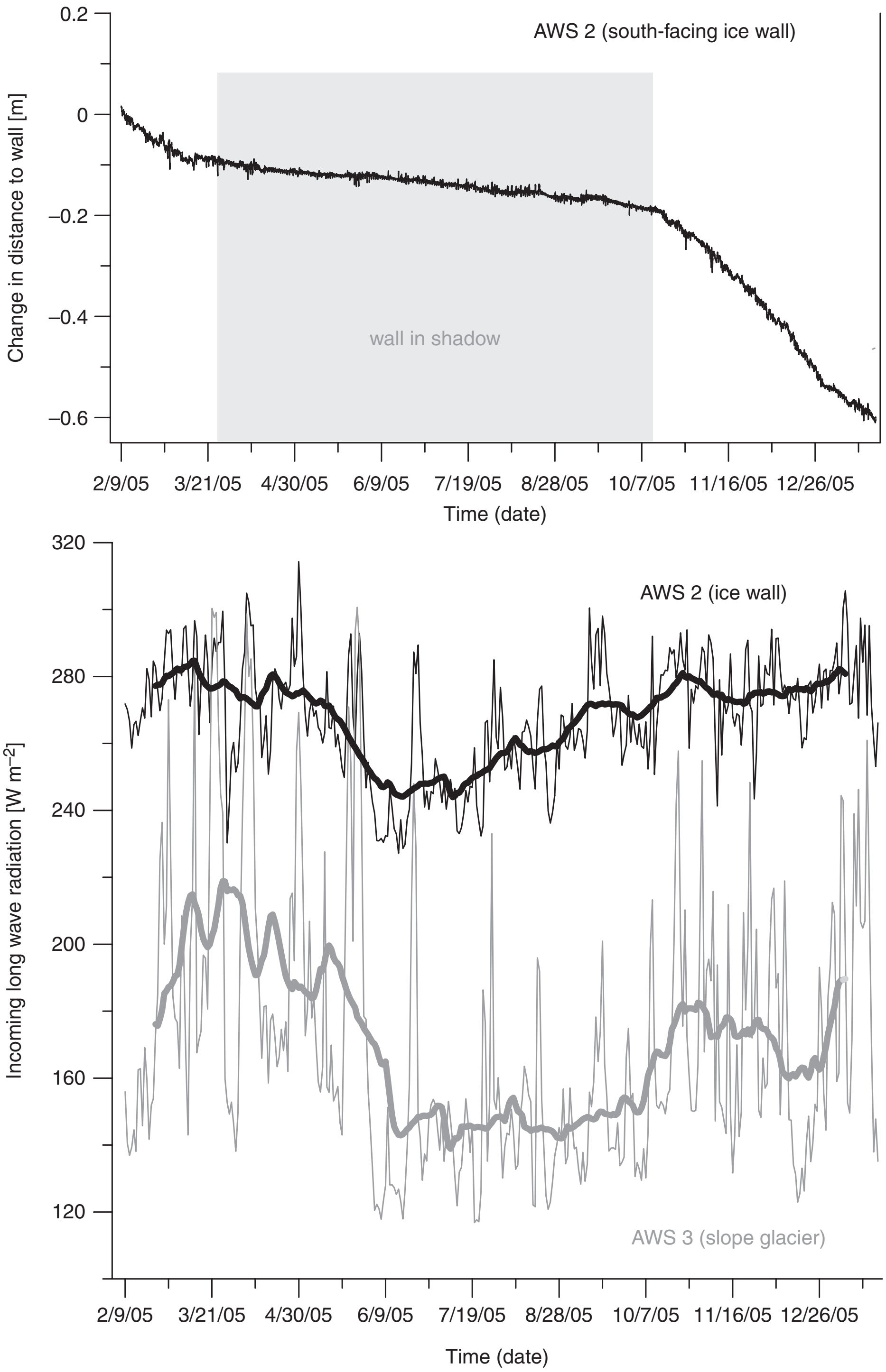

Figure 7. NIF ice wall. Top panel: accumulated changes in distance from AWS2 to the ice wall through 2005. The grey rectangle indicates the astronomical period when a precisely south-facing wall on Kibo (a condition nearly met for the wall in front of AWS2) is in shadow. Bottom panel: daily means of incoming longwave radiation at AWS2 (black) and AWS3 (grey). The smoothed bold lines represent the 30-day running means

by subfreezing air temperatures, turbulent heat exchange must be negligible or very small. If so, melting can even occur at much lower air temperatures than those observed near the ice wall on Kibo (Kuhn, 1987). Consequently, if turbulence and radiation conditions do not change considerably, the alternation between small amounts of sublimation (shadow) and melting plus considerably higher amounts of sublimation (sunlight) can be maintained over a wide range of air temperatures. Though melting hours may be less under lower air temperatures and sublimation somewhat smaller, it is the concept of alternating the seasonal cycle of direct sunlight and shadow that provides rather constant amounts of total ablation on the wall margins of the plateau glaciers.

Additionally, the dark ash surfaces adjacent to the ice wall base provide a strong, and, if not prevented by a lasting snow cover, rather constant energy supply to the walls. Figure 7 (bottom) compares long-wave incoming radiation at AWS 2 (ice wall) and AWS

3 (slope glacier; see Figure 4 for the position of the two AWSs), the latter originating from atmospheric emission only and at the ice wall combining emission from both the atmosphere and the snow-free ash surfaces. Values at the wall are almost 1.5 times as high as those on the slope glacier. Both series vary in concert with atmospheric moisture but the AWS 2 record is only marginally affected by whether the wall itself is in its shadow or not. The near constant strong input from the ash surfaces' long-wave radiation (that cannot reach the horizontal or sloping glacier surfaces) contributes to both the maintenance of the walls as well as their constant retreat rates. Only lasting snow cover and, usually combined with, changing intensity and duration of wet seasons can change this stabilizing effect considerably (see section 'How can plateau glaciers grow?').

Another stabilizing effect on the walls can be detected by examining the composition of incoming solar radiation. Owing to both little cloud coverage and a relatively short solar path through the tropical atmosphere, the shortwave incoming radiation on Kibo is dominated by direct solar radiation. Mölg *et al*. (2009b) show that the two-year average of global radiation measured at AWS3 (333 W/m^2) from February 2005 to January 2007 consists of 286 W/m^2 direct and only 47 W/m^2 diffuse solar radiation. Since the direct component of solar radiation appears to be the primary forcing of retreat along the ice wall margins, as discussed above and in Mölg *et al*. (2003b), the dominance of direct solar radiation on Kibo makes the maintenance of ice walls rather insensitive to changes in other climatic variables (e.g. air temperature and/or humidity). Even in the East African wet seasons, when atmospheric transmissivity of solar radiation is lowest over Kibo, it remains higher than the maxima found on extra-tropical mountain glaciers (Mölg *et al*., 2009b).

How constant is the decay time period?

Concluding from the discussion provided so far we obtain a decay time period for the plateau ice, which is governed by near constant surface area shrinkage rates due to near constant ablation rates of the ice walls. Since these rates are mainly driven by radiative forcing from solar and infrared radiation, variations in air temperature, as long as they do not reach values above the melting point, have little effect on the ablation rates of the ice walls (assuming turbulence conditions remain largely constant). There is currently no evidence that daytime maximum air temperatures have more frequently reached values > 0°C throughout the Holocene. This leads to the conclusion that, starting from a maximum extent of the plateau glaciers on Kibo summit, an intrinsic decay time exists with little potential variation. Disturbance of any major consequence could only come from either air temperatures regularly exceeding 0°C or from a substantial change in the amount of precipitation received on the top of the mountain. In the first case, the tabular glaciers would change their form and behaviour entirely and would probably be gone within a relatively short period of time (Kaser *et al*., 2004). The second case implies that the crater surface became snow covered for a longer duration. This, in turn, would initiate the formation of a new plateau glacier.

How can plateau glaciers grow?

The next step toward an ice-history hypothesis after having determined the decay time, is establishing how a plateau glacier may grow. Toward the end of the 2006 October to December (OND) wet season, very high snowfall events added another 50 cm to the already high seasonal amount of 40 cm and left 90 cm of snow measured on top of the NIF in early January 2007. The entire Kibo cone was covered with snow down to an altitude of approximately 4500 m. On the fully covered summit our detailed measurements showed an average snow depth of about 60 cm in January 2007 (Figure 8). Usually, snow cover is much less and individual events typically do not accumulate more than a few centimetres of snow. In such a case, sun penetrates the snow (Brandt and Warren, 1993) and heats the dark ashes underneath, causing the snow to disappear within hours to days. This was different after OND 2006 and the additional moderate accumulation in March to May 2007 resulted in localized fields of penitentes still being observed in October 2007 (Figure 9).

These snowy conditions stimulated us to carry out the following quantitative exercise. Using the verified mass balance model of Mölg *et al*. (2008a, 2009a), we simulated the annual snow cover evolution over ice for a typical flat location on the summit plateau (5750 m a.s.l., no slope, sky view factor = 0.98), with the results of this shown in Figure 10. The simulation is initiated assuming a 60 cm deep snowpack that has a bulk density of 400 kg/m^3 (Mölg *et al*., 2008a), and is then driven with meteorological data from AWS3 which are representative of a dry (2005), wet (2006) and average year such as 2007 (see Mölg *et al*., 2009a). Note that the initial condition of 60 cm deep snow is true for 2007 but hypothetical for the other years. Figure 10 suggests a critical snow depth of 10 cm, at or below which penetrating solar radiation would accelerate the degradation of the snow pack over volcanic ashes (see discussion above) relative to over ice. The simulations indicate that dry and normal (for the present climate) conditions probably can not maintain snow cover over the course of the year (where the snow cover in 2007 disappears indeed toward the end of the JJAS dry season as observed, see above). However, once a persistent snow cover is provided one can imagine that a series of strong precipitation seasons such as OND 2006 could easily build up a snow and firn cover within a few years (Figure 10). A few decades of more persistent wet conditions could therefore allow a plateau glacier to grow several tens of meters thick. Also the transition from snow into ice is rather quick on tropical mountains where melting occurs on many days around noon and refreezing takes place soon after (e.g. Mölg *et al*., 2008a), which results in the snow pack rapidly becoming dense (Sicart *et al*., 2002). In conclusion, a maximum plateau glacier extent could start with one single strong snowfall season that is followed by others in a way that allow a continuous snow cover and then annual accumulation. With this and with the *decay time period*, a single *life cycle* of plateau glaciation on Kibo is determined.

A millennium-scale plateau glacier history and beyond

In order to resolve a history of plateau ice on Kibo, one needs to identify humid periods that may have allowed the glaciers to grow. Lake level records as reconstructed from paleolimnological evidence in lake sediments give a hint at least of these periods over the last couple of centuries. In Figure 11 we use Lake Naivasha level variations reconstructed by Verschuren *et al*. (2000) as a first step to do this (Figure 11A; grey shading). We illustrate two options of growing periods and respective decay onsets, one shifting into decay after lake level rise (Figure 11B), a second allowing the decay to take place only after lake levels start to lower (Figure 11C). Each combination of plateau glacier growth followed by

Figure 8. Deep snow cover (approximately 60 cm) from strong snowfall events towards the end of the climatological OND wet season (November 2006 to early January 2007) at Crater Camp near Furtwängler Glacier, Kibo summit plateau (photo courtesy of Bernd Noggler, January 2007)

decay provides a complex picture of how the plateau glaciers could form and disappear. In each of the two cases cycles overlap resulting in a longer temporal period of glaciation when several ice layers sit on top of each other causing the observed steps on the NIF (Figure 1). This takes into account that glaciers have also thinned while shrinking in areal extent (e.g. Thompson *et al*., 2002). Otherwise the highest parts of the NIF would have grown very thick and deformation would have started to control the shape of the ice. Under both of our growth and retreat scenarios (Figure 11B and C) an absence of plateau glaciers before 1200 is shown and it is very probable that such extended dry periods have occurred several times throughout the Holocene (e.g. Bonnefille and Chalie, 2000). Besides the Naivasha record (Figure 11A), the reconstructed shallow water diatom (SWD, low percentages during high water levels) record in Lake Victoria (Figure 11D) supports the mentioned hiatus in plateau glaciation, showing above-average values ($\geq$ ~40%) after AD 1000 and a pronounced increase towards AD 1200 (Stager *et al*., 2005). Other probable periods without plateau glaciers around 1600 and in the early nineteenth century also coincide with marked peaks in the SWD and are preceded by the most prominent increases in SWD on the record (Figure 11D). The strongly decreasing SWD in the early twentieth century is stimulated by cultural eutrophication (Stager *et al*., 2005), so that part of the record cannot be used to infer a significant lake high-stand on the millennium scale and a related wet period. In the twentieth century, only the early 1960s experienced a sharp rise of lake levels because of a few years of strong precipitation (e.g. Nicholson and Yin, 2001). These heavy rains coincide with an abrupt, high-amplitude $\delta^{18}O$ diatom minimum from alpine lakes on Mt Kenya (Barker *et al*., 2001).

Summary and conclusions

From a compilation of all available information on modern time and present-day phenomena and processes controlling the glaciers on Kilimanjaro, and a careful glaciological evaluation, we propose a new hypothesis about the evolution of ice on Kilimanjaro's central peak Kibo. This suggests a climate history differing considerably from the data based on the ice core interpretation provided by Thompson *et al*. (2002).

From observations and measurements we conclude that:

(1) The Kibo summit plateau glaciers need to be analysed differently from the slope glaciers; here we primarily consider the former.
(2) Possible minor changes in thickness have no impact on the changing surface area of the tabular plateau glaciers.
(3) Plateau glacier area decrease has been strikingly constant over the twentieth century as the result of both constant ice-edge retreat rate and a progressive disintegration of ice bodies into yet smaller entities.

Figure 9. In some areas of the plateau, 2006/2007 wet season snow survived until at least October 2007. Here, a snowfield is shown on the southwest facing slope descending from the summit crater rim (photo: DR Hardy, 7 October 2007)

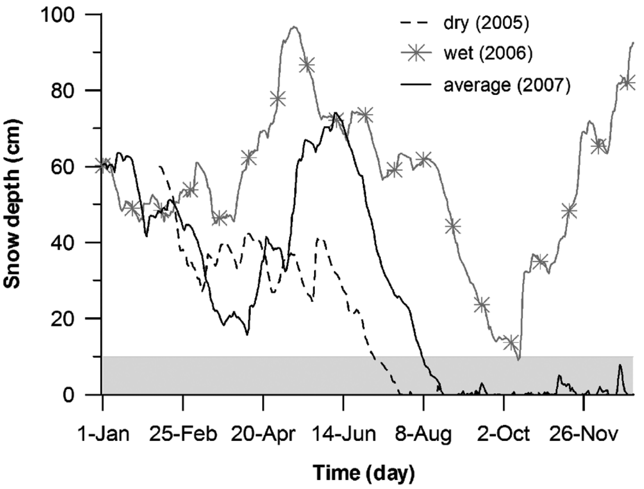

Figure 10. Simulation of the annual snow cover for summit plateau conditions on Kibo under dry, wet and average atmospheric forcing, starting with a 60 cm deep snow pack of bulk density = 400 kg/m^3. Note slightly delayed start of the 'dry' simulation, since measurements at AWS3 began on 9 February 2005. The grey-shaded area indicates a 'critical' snow depth (see text)

(4) Ablation rates of the ice walls are persistently constant for geometric reasons (alternating between shadow and sun conditions), which can only be markedly disturbed by either air temperatures rising regularly above 0°C over a longer period, or by considerably wetter conditions. The latter would hardly become effective since they, in turn, initiate glacier growth.

(5) From a maximum possible extent and constant shrinkage rates a characteristic *decay time period* of the plateau glaciers is estimated to be 165 ± 10 years.

(6) A maximum glacier extent can start with one major snowfall season followed by a series of exceptional wet seasons that would allow a tabular ice body to grow within a few decades.

(7) We interpret wet periods from lake level stands in East Africa as glacier growth periods on Kibo and postulate that after each wet period the shrinkage of plateau ice starts again. As a consequence, the application of the *decay time period* to the time series of lake level stands leads to the ice history of the Kibo summit plateau.

(8) The hypothesized ice history indicates that the plateau ice may have come and gone repeatedly throughout the Holocene. The near extinction of the plateau ice in modern times is controlled by the absence of sustained regional wet periods rather than changes in local air temperature on the peak of Kilimanjaro (consistent with Mölg *et al*., 2009a). Wet season prevalence is likely to be affected in future by global warming (e.g. Held and Soden, 2006).

We are confident that our concept of the age of Kibo's glaciers provides a reasonable alternative to the view of persistent glacierization of Kibo throughout the Holocene (Thompson *et al*., 2002). The concept also implies that plateau glaciers on Kilimanjaro are inappropriate indicators of global warming or cooling, unlike the

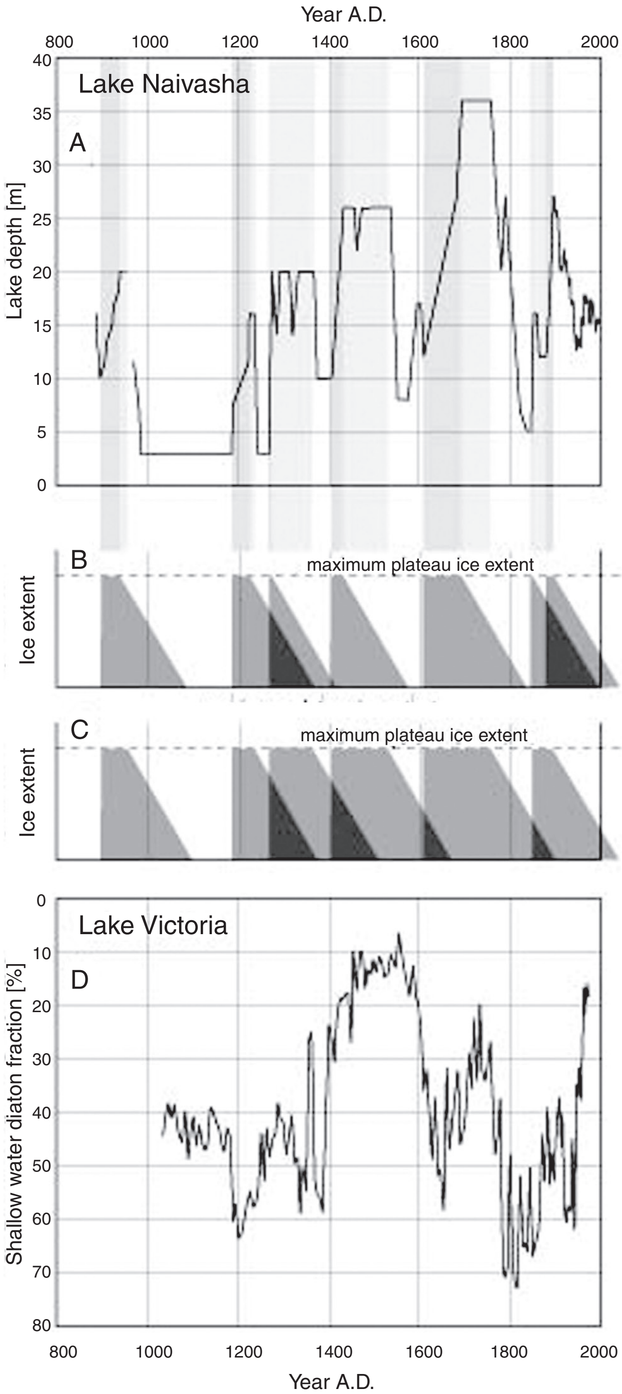

Figure 11. (A) Lake levels of East African Lake Naivasha from 900 to present (Verschuren *et al.*, 2000), (B) and (C) respective history of Kibo summit plateau glaciation (discussion provided in the text), and (D) shallow water diatom fractions from Lake Victoria (Stager *et al.*, 2005; note that the y-axis is reversed)

majority of glaciers on Earth. In near future we intend to simulate multicentury time series of glacier mass balance from a combination of measurements and a physically based model approach, which will help evaluate the short-term cyclicity of plateau glaciers on Kibo.

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