Late Pleistocene glaciation of the Mt Giluwe volcano, Papua New Guinea

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1. Introduction

The tropics receive the greatest energy per unit area from the sun anywhere on Earth. Heat received in the tropics is distributed by wind, surface ocean currents and by latent heat to the rest of the planet, controlling the distribution of climate. The mountains of New Guinea lie in the western Pacific Ocean tropics near the centre of the Indo-Pacific warm pool. This region is the largest body of warm ocean on the planet and spans almost half way around the globe. Changes in the temperature of this water body not only strongly influence the climate of New Guinea but have hemispheric climatic consequences.

Knowledge of the evolution of the warm pool through the Pleistocene has grown rapidly over the last 30 years. The highly influential CLIMAP project (CLIMAP, 1976, CLIMAP Project Members, 1981; Thompson, 1981) documented remarkably little change in tropical sea surface temperature (SST) during the last glacial maximum (LGM), an event when the lowest temperatures in the late Pleistocene would be expected. Since this time, several new techniques have been developed to estimate SST in deep sea sediments, and a large number of new cores with good chronologies covering the LGM have become available. Despite this, recent revisions of SST during the LGM indicate that the CLIMAP conclusion of little change remains robust (MARGO Project Members, 2009). Estimates of cooling in SST from three independent methods around the perimeter of New Guinea in the Western Pacific Ocean range from 0 to 3 °C (Rosell-Melé et al., 2004; Barrows and Juggins, 2005; Barker et al., 2005). This minimal cooling strongly contrasts with the observation that the snowline high in the mountains of New Guinea was as much as 1100 m lower at the same time, requiring a temperature difference of 5–6 °C (Löffler, 1972). This difference in surface cooling has been perceived as an apparent 'paradox' and has fuelled a long standing debate attempting to reconcile the two sets of estimates (Webster and Streten, 1978; Rind and Peteet, 1985; Peterson et al., 2002). No consensus has emerged as to an explanation that satisfactorily explains this phenomenon. Some steepening of tropical lapse rates probably occurred but this does not fully resolve the issue (e.g., Blard et al., 2007).

Despite significant progress in refining SST estimates, less progress has been made in quantifying temperature changes in...
tropical mountains. Glaciers that existed in the mountains of New Guinea are ideally placed to record temperature changes in the western Pacific Ocean through time. However, the maximum lowering of the snowline is not directly dated, and hence it is possible that cooling estimates may have been incorrectly attributed to the LGM. In recent reviews, the lack of chronological control was highlighted (Mark et al., 2005). The island of New Guinea represents an ideal area to reassess the tropical 'paradox', both because it records extensive glaciation and because the surrounding Indo-Pacific warm pool is central to the debate on the magnitude of maximum cooling in the tropics. Most moraines on New Guinea are assigned to the LGM on the basis of limiting radiocarbon dates in only a few locations (Prentice et al., 2005), and because of a lack of evidence for more than one glaciation (Löfler, 1972). Therefore the outermost are frequently assigned an LGM age. Clearly, the correct moraine sequences need to be identified before temperature estimates from snowline changes can be reliably attributed to the LGM.

In this paper we present a new study of the moraine systems on the southwest side of Mt Giluwe on the island of New Guinea. We present maps of the glacial geology and review previous glacial stratigraphies. To constrain the age of glaciation we present exposure ages on boulders from representative moraine sequences. Together with radiocarbon dates from basal sediments in associated basins, we construct the first glacial chronostratigraphy for a mountain in New Guinea. These ages are used as the basis for a regional review of the timing of glaciation in Southern Hemisphere sector of the western Pacific Ocean. Lastly, based on the heights of dated moraines, we review previous estimates of temperature change attributed to changes in the equilibrium line-altitude (ELA).

2. Glacial geology

2.1. Background

The mountains on the island of New Guinea were the most extensively glaciated area in the Asian tropics during the late Pleistocene (Fig. 1). On the western side of the island over 3400 km² was glaciated, mostly in the form of an almost continuous ice field from Mt Jaya and Mt Idenburg to the Star Mountains Prentice et al., (2011). On the eastern side of New Guinea, there is evidence for glacial activity on at least 20 mountains. The ice cap on Mt Giluwe was the most extensive area of glaciation (170–188 km²) followed by Mt Wilhelm (~ 107 km²) (Löfler, 1972; Bik, 1972). Löfler (1972) placed the LGM snowline at an altitude of 3500–3550 m in central Papua New Guinea, rising to between 3600 and 3700 m in the Saruwaged and Owen Stanley Ranges. Today, there is no active glaciation on the eastern side of the island and only three of the highest peaks were recently glaciated on the western side of the island (Mt Jaya, Mt Idenburg, Mt Trikora, and Mt Mandala). In the early 1970's, the ELA lay near ~4600 m (Löfler, 1972).

Mt Giluwe is an extinct shield volcano and the second highest peak (4368 m) in Papua New Guinea, rising above the western highlands surface at 1500–1700 m (Fig. 2). The mountain is deeply incised by radial drainage, and only the lower half still bears original volcanic geomorphology (Blake and Löfler, 1971). The upper half is extensively modified by glaciation (Fig. 2a). Löfler (1972) and Blake and Löfler (1971) concluded that Mt Giluwe had been glaciated at least twice, with an intervening interglacial period. The first event was associated with volcanic activity on the east of the mountain. The second event was responsible for the prominent moraines around the mountain. This most recent event (the 'main' glaciation) probably consisted of two stages separated by an unknown amount of time, and deposited five groups of recessional moraines (Blake and Löfler, 1971; Löfler, 1972). Bik (1972) also observed two stages of glaciation based on the positions of the terminal moraines, but concluded that there must be at least a stadial between the glaciations. Bik (1972) also suggested that there were 5 main groupings of moraines on the southwest side. In contrast, earlier work by Verstappen (1952) and Reiner (1960) concluded that there was evidence for only one glaciation in New Guinea.

2.2. Previous dating

There are few limiting dates from New Guinea for the timing of maximum glaciation. On Mt Giluwe, the basal radiocarbon dates from peat behind terminal moraines provide minimum ages

![Fig. 1. Regional map of New Guinea showing the location of Mt Giluwe and deep sea cores ODP Hole 806B and MD97-2140.](image-url)
Radiocarbon dates are calibrated using Calib 6.0 (Stuiver and Reimer, 1993) utilising the INTCAL09 calibration curve (Reimer et al., 2009). Uncalibrated radiocarbon dates are expressed in radiocarbon years (14C yr BP) and calibrated radiocarbon dates are expressed as calibrated years (cal yr BP). Exposure ages are given in thousands of years as 'ka'. Where referred to, oxygen isotope chronozones (OIC = the informal 'marine isotope stage') are based on the LR4 chronostratigraphy of Lisiecki and Raymo (2005).

Löffler (1972) limited the ages of the second youngest and youngest of his five recessional groupings to at least 7890 ± 120/−170 cal yr BP and 3840 ± 130/−140 cal yr BP years based on dates on peat samples from pits in gullies incised through deposits behind moraines (Fig. 1, Table 1). Radiocarbon dates limit the outermost moraines to greater than 4170 ± 240/−180 cal yr BP years old (Blake and Löffler, 1971). On the basis of these radiocarbon dates, Löffler (1972) concluded that Mt Giluwe only became ice free.

Fig. 2. a) Shaded relief map of Mt Giluwe, b) Topographic map of Mt Giluwe showing sites of radiocarbon dating (sites 1–6 in Table 1) and dated palagonite (‘V’ symbols). The digital terrain model used is the SRTM 90 m digital elevation data. Solid triangles indicate the east and west peaks of Mt Giluwe. Inset box indicates the location of Fig. 3.
in the late Holocene. However, this dated sequence most probably reflects erosion following human disturbance of the forest cover.

On the northeast side of the mountain in the Kaugel Plain where the glaciated catchments of the Gorgon and Tamal rivers drain the mountain (Fig. 2), a 40 m section through the glaciated catchments of the Gorgon and Tamal rivers drain the mountain (Fig. 2), a 40 m section through the floor plain terrace reveals stratified sediments intercalated with thick beds of peat (Löf, 1972). The uppermost layer is underlain by peat dated at 28,270 ± 1240/−1300 cal yr BP and deeper peat layers are beyond the radiocarbon limit. Consequently, Löf (1972) placed the final stage of the last glaciation at 28,000 cal yr BP.

Subsequent radiocarbon dating casts doubt on how closely the above dates limit the glaciation (Table 1; Figs. 2–3). A date of 15,620 ± 1170/−980 cal yr BP was obtained on basal clayey gyttja in a small Carpha alpina quaking fen nestled between moraines in the Tongo valley labelled ‘group 3’ by Löf (1972), confirming the pre-Holocene age of these moraines (Hope and Peterson, 1975). Further up the valley, a further date of 11,340 ± 420/−540 cal yr BP was obtained on basal gyttja from a small basin in ‘group 4’ moraines (Hope and Peterson, 1975). This is a similar stratigraphic position to the 7890 cal yr BP age of Löf (1972). Finally, an age of 11,570 ± 490/−400 cal yr BP (N 1317) cal yr BP on basal pond muds overlying sand was obtained within a small cirque basin at 4170 m about 300 m southeast of the highest peak (site 5 in Fig. 2). The above ages strongly contrast with the interbedded till. We refer here to this early evidence of ice cover as the Gogon Glaciation, and infer it occurred between 239 and 306 ka. The Gogon glaciation was more extensive than the last glaciation because the interbedded till site lies outside its limits.

Palagonite is also found on the east peak of Mt Giluwe (Blake and Löf, 1971). Here the volcanic activity is much older and the eruptions took place under only thin ice. Lava near the palagonite dates to 701 ± 41 ka (G42; 652 ± 60 ka, 743 ± 55 ka). Palagonitized lava from the main peak dates to 810 ± 40 ka (G23; 753 ± 70 ka, 885 ± 80 ka). These are some of the oldest ages for lava from Mt Giluwe. Only one other site near the main peak is older, dating to 121–128 Ma (Löf et al., 1980) which signifies the first time the mountain reached its present altitude. The younger ages at 810 ka may therefore represent the first recorded time the regional snowline intersected the mountain and provide circumstantial evidence for ice cover at this time.

2.3. Glacial stratigraphy

In Fig. 3 we present a new map of the glacial features of the southwest quadrant of the mountain. Preservation of glacial features is generally excellent, with virtually no modification of the depositional form and even minor moraine ridges are preserved. Only the crests of the outermost moraines show rounding (Bik, 1972). Bik (1972) attributed the preservation to the presence of thick grass cover that protects the slopes from erosion. We observed that glacial till under boulders was compact and plastic due to the high clay content, and therefore resistant to erosion from surface runoff.

The style of glaciation is strongly controlled by topography (Fig. 3). The low angle upper slopes of the mountain bore ice cap glaciation, which terminated as either outlet glaciers in the valleys or as a broad shield on the flanks of the mountain. On the southwest side of the mountain, ice originated from two major centres on either side of an arete with headwalls up to ~ 3800 m. The northern cirque (N) supplied ice mainly into the Tongo River valley. The southern cirque (S) formed an ice cap over most of the southwest flank of the mountain. The cirques are not over-deepened indicating that the ice cover was not strongly erosive. The number and spacing of moraines preserved is a function of the thickness of the ice and the rate of retreat. Where the ice cap spread out radially on the southwest and the ice was thinnest, small changes in thickness necessitated considerable retreat of the ice front (Bik, 1972). Consequently, this
area has the largest number of moraines. Conversely, the fewest moraines are located where the ice was channelled into valleys, such as the Tongo Valley in the north, and ice reached its lowest elevation at these sites. Post-glacial fluvial incision on the valley floor is minimal and in most cases has only dissected moraines and has not deeply incised into the bedrock.

There are at least three distinct sequences of moraines marking individual glaciations. The moraines are mainly separated on the basis of stratigraphic cross-cutting relationships. The best stratigraphy on this part of the mountain occurs on the west side of the Tongo River and in the Mengane River catchment (Fig. 4a). Successive glaciations and fluvial incision have lowered the base level of the Tongo River valley floor, channelling more ice into the valley and preventing it from flowing over the col into the Mengane catchment (Fig. 3). Consequently, the ice from the northern cirque was progressively more confined to the Tongo valley through time, and older moraines, particularly the western lateral moraines, were preserved instead of being over-ridden by subsequent ice cover.

The oldest moraines recognised to date occur in the Mengane River valley and we assign them to the Mengane Glaciation

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**Fig. 3.** Glacial geology of the southwest flank of Mt Giluwe. The map is based on unrectified aerial photography due to poor local height control. Contours are estimated from Royal Australian Survey Corps, Sheet 7685, Series T683, 1974. Inset boxes indicate locations of Fig. 4a and b. Numbers refer to sites of radiocarbon dating (sites 7–11 in Table 1).
This glaciation was the most extensive and terminal moraines occur as low as 3200 m. During this phase, glaciers from the northern and southern cirques coalesced through the Mengane River col to form an ice cap that reached almost as far down slope as the confluence of the Mengane and Menken Rivers. The presence of rillen karren on the sides of basalt boulders and subaerial weathering rinds up to 1 cm thick indicate a considerable age for this glaciation. Moraine crests are rounded, and rivers have cut into bedrock in the valley floor upstream of the terminal moraines, particularly in the lower Menken River. Between the Yamboro and the Mengane Rivers lies the oldest stratigraphic evidence for an ice cap. These moraines seem to be at least in part conformable with the terminal moraines in the Mengane valley, but the outermost moraines may pre-date them.

The second youngest moraines lie between the Mengane moraines and the youngest glaciation in the upper Komia River
valley, as a western lateral moraine of the Tongo River (Fig. 4a). We assign these moraines to the Komia Glaciation. Stratigraphically, we also place the moraines between the Mengane moraines and the youngest glaciation in the Menken river valley into this glaciation, but these are not contiguous with the moraines adjacent to the Tongo River valley, making this assumption difficult to assess. It is also difficult to separate moraines from the younger glaciation between the Mengane River and the Yamboro River on the basis of preservation alone.

The most recent glaciation, the Tongo glaciation, has very detailed moraine stratigraphy preserved over much of the mountain, particularly in the Tongo River valley and on the southwest flank of the mountain where there are at least 52 moraines. The Tongo Glaciation was less extensive than the preceding glaciations. Moraines occur down to 3400—3300 m in the southwest, extending about 2.5 km from the back wall of the cirque.

The lateral moraines of the Tongo Valley (Fig. 4a) are the largest depositional landforms on this side of the mountain. From the height difference between the valley floor and the top of the lateral moraine, ice was ~ 100 m thick in the Tongo Valley (=Komia Valley of Löfler (1972)) depositing a series of large lateral moraines that extend down to ~3100 m and 4.1 km from the headwall. Löfler (1972) attributed the lateral moraines of the Tongo River to the last glaciation, but thought that it consisted of two stages. The western lateral lies at the junction of the Tongo Valley glacier and a lobe of ice from the northern cirque (Fig. 3).

Five moraines about the large western lateral moraine of the Tongo valley (Fig. 4a). Traced to the south these moraines progressively split into as many as 20 narrower, lower moraines where the glacier was thinnest. The number of retreat moraines is mostly a function of topography. Both Löfler (1972) and Bik (1972) placed the moraines in this glaciation into 5 groups or ‘bundles’ of moraines. This grouping may be climatic, or a product of the topography of the southwest side of the mountain because it breaks down laterally to the north and south of this area. At maximum ice extent, the gradient of the ice cap was lowest in the south. In the region of Bren Tarn (Fig. 4b), the mean gradient to the top of the backbone was of the order of 9°. Bik (1972) noted that the distal slope of the terminal moraines was 5—6°, probably indicating the slope near the ice front.

Ice was at least 100 m thick in each cirque, and split into two individual ice fields when it thinned to less than ~100 m (Fig. 3). As ice retreated through the col at the top of the Mengane valley (Fig. 3), the Tongo glacier continued to occupy its lateral moraines. The Tongo glacier deposited a series of inset lateral moraines as it retreated and became an ice field, similar to that in the southern cirque. As ice retreated in the cirques it left multiple low moraines, numerous kettle holes and well-preserved subglacial topography such as flutted till. The spacing between the moraines is greatest in the south where the bedrock slope was shallowest. From the moraine orientation the last ice persisted in a southwest-facing cleft in the northern cirque and under the northern wall of the southern cirque (N and 5 on Fig. 3). There does not appear to have been any strong aspect control on ice retreat.

As noted by Löfler (1972), the most recent ice cap draped till over the older moraines without modifying them. An example of this is the northern side of the Menken River valley (marked by a bold line in (Fig. 4b)). This ridge is probably mostly bedrock, but is overlain by medial moraine deposited during the Mengane Glaciation when glaciers occupied both the Mengane and the Menken valleys. The ice during the Tongo Glaciation deposited low terminal moraines over the ridge at right angles to the medial moraine. The till sheet is probably no more than 0.5—5 m thick (Löfler, 1972). Down valley from Bren Tarn, older terminal moraines have been over-ridden and are still visible through the thin till sheet (Fig. 3).

3. Exposure dating

The exceptionally preserved moraines of Mt Giluwe provide good targets for exposure dating. We chose sampling sites to cover the full range of glacial features from the outermost icecap moraines to the retreat sequences. MP and GH collected the GIL samples in 2001 and TTB, MP and GH collected the GLW samples in 2003. In addition, GH and MP collected peat cores in 2001 and 2003 for radiocarbon dating. Because of poor regional elevation control, we used the Shuttle Radar Topography Mission (SRTM) 90 m digital elevation data for sample site heights in conjunction with GPS field measurements. Geographic coordinates for the 2001 samples were estimated. We sampled only the largest boulders, but in some cases for the GIL samples only small boulders (<1 m in height) were available. Post-depositional weathering on the Tongo Glaciation boulders appears negligible. However, on some boulders on the Mengane Glaciation moraines we observed significant surface weathering in the form of rillen karren indicating potential removal of centimetres of rock and avoided those surfaces. Surface spalling from the boulders was not observed because of the massive nature of the basaltic rock. We focussed sampling on moraine crests to reduce the risk of sampling boulders exhumed by erosion. Snow cover in the potential source of shielding on Mt Giluwe is negligible, except during glacial periods but this is difficult to quantify. The entire sample area is virtually forest free today, and would have lain above the treeline until ~11 000 cal yr BP. A low subalpine forest probably occupied the well drained sites from 11,000- ca 3000 cal yr BP, possibly building up a layer of root mat and litter. The date of widespread clearance is currently unknown but the charcoal dates of Löfler (1972) indicate that clearing occurred soon after 4000 cal yr BP. We assume any shielding had similar but negligible effects on all samples. Site data for the sampled boulders is presented in Table 2.

The exposure ages on the thin moraine belts are likely to provide an accurate account of the retreat of ice. However, the larger lateral moraines are thick and are sequentially constructed. Therefore our exposure ages are inherently biased towards the most recent time ice stood at the moraine and provide a minimum age for the ice advance. Because of the minimal erosive power of the ice on low angle slopes, there is potential for survival of old surfaces. This, together with reworking of boulders from older moraines, acts to skew the age distribution towards older ages. Conversely on older moraines, there is a likely bias towards younger ages, because through time erosion flattens the profile of moraines, exhuming previously buried boulders, and weathering lowers the surface of boulders.

Because of the availability of basalt with high potassium content, we chose 36Cl for exposure dating. Chlorine was extracted from the whole rock because the fine-grained nature of the basalt made mineral separation impossible. The isotopic ratios of 36Cl/Cl were measured by accelerator mass spectrometry on the 14UD accelerator at the Australian National University. Because of the very high production rates at high altitude, blank corrections were negligible. The abundance of major target elements in the dolerite samples was determined using X-ray fluorescence. The concentrations of trace elements with large neutron capture cross sections (B, Gd, and Sm) and neutron-producing elements (U and Th) were measured by inductively-coupled plasma mass spectrometry. Chlorine content was determined by isotope dilution. Chemical data are summarized in Table 3. Other major element and trace element geochemical data are listed in the Appendix which is also available from the WDC-A for Paleoclimatology, Boulder, USA or the Australian Quaternary Data Archive.
Exposure ages are calculated as detailed in Barrows et al. (2002), with the modifications made by Barrows et al. (2004). We used the production rates of Stone et al. (1996a, 1996b, 1998), Evans (2001) and Masarik and Reedy (1995). We scaled the production rates using the procedures of Stone (2000) for consistency with regional data sets. We found excellent agreement between exposure ages and radiocarbon ages indicating that this scaling approach is satisfactory for high altitude at low latitude. For $^{36}$Cl production by neutron capture we followed the procedures of Liu et al. (1994), Phillips et al. (2001) and Stone et al. (1998).

Age calculations are presented as conventional exposure ages, calculated using geographic latitude and including $^{36}$Cl production from titanium and iron. No corrections are made for geomagnetic changes through time due to uncertainties in scaling low latitude sites. Based on the exceptional hardness of the basalt and the lack of evidence of weathering on the boulders we sampled, all ages are calculated assuming that there has been no weathering since initial exposure. We cannot rule out at least some erosion on the boulders on the outermost moraines on the scale of centimetres. To bear in mind the sensitivity of exposure ages to weathering, 1 cm of weathering would lower the age of a boulder that has been exposed for 150,000 years by less than 1%. All measurement errors, including production rate errors, are fully propagated and moraine ages are weighted means. All discussion in the text refers to the corrected ages and all ages and statistics are reported at one standard deviation.

4. Results

Exposure ages calculated using the $^{36}$Cl data are presented in Table 4. The exposure ages are discussed below in order of decreasing age.

4.1. Mengeglaciation

All exposure ages on moraines from the Mengeglaciation are older than 100,000 years, and range from 109 to 192 ka. Of the seven ages on the outermost moraines, four ranging from 136 to
Due to their great age, it is likely that the boulders on the Mengane of 64.2 ka have a mean age of 146 ± 10 ka but with scatter beyond what would be expected due to random error alone (χ²/ν = 2.1). Due to their great age, it is likely that the boulders on the Mengane moraines have not all experienced a simple exposure history. The three younger ages younger than the group of four are likely to have been partially shielded by till early in their exposure history, and the remaining age (192.3 ka; GLW-31) probably represents a case of previous exposure. Finally, an age of 208 ± 12.1 ka (GLW-17) came from a boulder on the outer western lateral moraine of the Tongo River. Although only a single age and older than expected, it probably indicates that this moraine belongs to the Mengane Glaciation.

### 4.2. Komia Glaciation

The only locations where we dated the Komia Glaciation were on a terminal moraine mostly overridden by the Tongo Glaciation and on an adjacent western lateral moraine (Fig. 4a). Two ages on a single boulder (GLW-15) on the terminal moraine give an average of 64.2 ± 11.8 ka. The boulder on the lateral moraine has an exposure age of 61.8 ± 3.1 ka (GLW-16). The two boulders give a weighted mean age of 61.9 ± 3.0 ka (χ²/ν = 0.2) for the Komia Glaciation. An additional boulder from the leading edge of the terminal moraine, where it was deeply embedded, was dated at 38.4 ± 1.9 ka (GLW-14), and the top of the boulder was probably shielded by till during part of its history.

#### 4.3. Tongo Glaciation

Most of our dating was concentrated on the youngest and best-preserved glaciation to define the age of the maximum ice extent and timing of retreat during the last glacial maximum. Our exposure ages agree well with the general timing of ice retreat from the radiocarbon dating of Hope and Peterson (1975). The outermost terminal moraine abutting the large western lateral moraine of the Tongo valley is dated at 19.4 ± 1.0 ka (GLW-11). This is the older of a series of six moraines adjoining the Tongo left lateral moraine. The age agrees well with a new radiocarbon date 20.350 ± 1300 ± 140 cal yr BP we obtained from the base of a bog formed behind a branch of a lateral moraine on the Mengane River (Table 1). The third oldest moraine, which eventually merges with the first two into a single crest, has an average age of 18.3 ± 1.4 ka (GLW-01, 02),
The second youngest moraine has a single age of 16.5 ± 0.9 ka (GLW-13). Together, these ages indicate that the ice front occupied the 5 moraines near its maximum extent for 3000 years. The volume of the moraines is similar, and assuming debris supply was constant, indicates ice occupied each limit for only ~600 years.

Shortly after deposition of the last of the 5 moraines, ice retreated through the col in the upper Mengane Valley (Fig. 3). The Tongo left lateral appears to have been occupied continuously from 19.0 ± 0.9 (GLW-18) until 15.5 ± 0.5 ka (GLW-03, 04, 06) (χ² / v = 0.38). A fifth age on the moraine is an outlier (41.6 ± 1.6 ka; GLW-05). Further south, 15 ages on 13 boulders constrain the subsequent retreat of the ice cap into the southern cirque. A new radiocarbon date (Fig. 3; site 13) on this flank indicates that ice stayed near its maximum limits until shortly before 15,760 ± 320/500 cal yr BP. The 13 boulders range in age from 15.4 to 9.9 ka and overlap in age to some extent between adjacent moraines. Two boulders (GIL-01/GLW-23; GIL-02/GLW-20) were sampled on the separate field trips and the exposure ages agree within error (13.1 ± 0.6/13.8 ± 0.7 ka; 11.5 ± 0.5/11.2 ± 0.4 ka respectively). Some boulders appear to record cases of small amounts of prior exposure (e.g., GIL-04, GIL-07, GIL-10) and two boulders (GIL-06 and GIL-08) probably represent reworked boulders from the underlying older till. The oldest moraine has an average exposure age of 13.4 ± 0.5 ka (GIL-01/GLW-23).

The ice cap split into the northern and southern lobes shortly after 11.3 ± 0.3 ka (GIL-02/GLW-20). The youngest moraine has an average age of 11.5 ± 2.3 ka (GIL-04, GL-12). No further boulders were sampled up the valley. However, this final age of retreat is very similar to the minimum limiting age of the glaciation from the summit bog (11,570 ± 490/400), and the rate of retreat suggests that the last few moraines were deposited in less than 1000 years.

5. Discussion

Exposure dating of moraines has identified at least four distinct stages of glaciation on Mt Giluwe within the last 150,000 years. These moraines are difficult to subdivide on the basis of weathering and preservation alone and this has led to them being previously classified as one (Blake and Löffler, 1971; Löffler, 1972) or two glaciations (Bik, 1972).
To interpret the glacial history of Mt Giluwe in a regional context, we have assembled two key regional SST records from the equatorial western Pacific Ocean (Fig. 5). Core ODP Hole 806B (Lea et al., 2000) is located on the Ontong Java Plateau and MD97-2140 (de Garidel-Thoron et al., 2005) is located north of New Guinea (Fig. 1). Both cores come from within the bounds of the Indo-Pacific warm pool of SST >28 °C and are derived using Mg/Ca in planktonic foraminifera tests. Comparison of the timing of glaciation with these SST records allows us to interpret the extent to which glaciation over New Guinea is regionally representative and driven by temperature changes.

The oldest evidence for glaciation on Mt Giluwe is dated by K/Ar at 293–306 ka early in OIC 8, associated with the last dated volcanic activity on the mountain (Blake and Löffler, 1971; Löffler, 1972; Löffler et al., 1980). No evidence of this glaciation was found on the southwest flank of the mountain. However, because of its age, it may have little surface expression remaining. Core ODP Hole 806B records warm SST at 300 ka, but rapid cooling into OIC 8 at 289 ka whereas MD97-2140 (de Garidel-Thoron et al., 2005) records SST during OIC 8 at 293 ka as the coldest in the last 500 ka (Fig. 5). Considering the likely errors on the conventional K/Ar ages, the Gogon Glaciation was probably associated with cooling early in OIC 8.

The oldest directly dated glaciation in our study, the Mengane Glaciation, reached its maximum extent during OIC 6. Minimum SST in ODP Hole 806B occurs at 143 ka and is recorded earlier in MD97-2140 at 169 ka. The OIC 6 interval is one of strong cooling in the temperate latitudes of the western Pacific Ocean (Barrows et al., 2007a), but in contrast to New Guinea, no ice advance has yet been directly dated to this time in Australia or New Zealand. In the central North Pacific Ocean, the oldest evidence for glaciation on Mauna Kea in Hawaii, the

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**Fig. 5.** Regional sea surface temperature records in comparison to dated glacial maxima on Mt Giluwe. a) Distance ice reached from the headwall for each of the dated glacial maxima. Error bar is for the weighted mean age for the Tongo, Komia and Mengane Glaciations, and is an age range for the Gogon Glaciation. b) SST record of ODP Hole 806B (Lea et al., 2000). c) SST record of MD97-2140 (de Garidel-Thoron et al., 2005). d) LR4 benthonic oxygen isotope record stack of Lisiecki and Raymo (2005). This stack is derived from δ18O records from around the world and primarily shows eustatic sea level changes. The age models of ODP Hole 806B and MD97-2140 are tuned to the chronostratigraphy of the LR4 stack based on tie points listed in the Appendix.
Pohakula Glaciation, is poorly dated from about 100–150 ka to 150–200 ka (Wolfe et al., 1997) and possibly records glaciation at the same time as on Mt Giluwe. Elsewhere, well preserved moraines of OIC 6 age conducive to exposure dating are rare (Blard et al., 2009).

The Komia Glaciation reached its maximum extent on Mt Giluwe at 62 ± 3 ka. This timing is slightly later than the OIC 4 SST minima in ODP Hole 806B at 66 ka and 65 ka in MD97-2140. This glaciation is only well preserved in some locations because it had a similar extent to the Tongo Glaciation, and so it is poorly preserved and we know little about the timing of ice retreat. Glaciation of this age is not known from elsewhere in New Guinea, but Hope (2009) inferred cooling from pollen records around this time in the Albert Edward Range. On Hawaii, the penultimate glaciation or the Waihu Glaciation, has a single exposure age of 150 ka at 100 and 150 ka, but it could be as young as 70 ka (Wolfe et al., 1997). There is no direct evidence for an ice advance at 60 ka in New Zealand or on the island of Tasmania, but the greatest extent of ice in the Snowy Mountains of Australia is dated at 59 ka during the Snowy River advance (Barrows et al., 2001). Mid latitude SST records indicate that Snowy River advance occurred at the end of the cold phase of OIC 4 which ended at 60 ka and began, with variable timing between cores, at 75–67 ka (Barrows et al., 2007a). It is important to note that minimum SST in the western Pacific Ocean is similar during both OIC 4 and 6, despite large differences in sea level and therefore global terrestrial ice volume.

Cooling of a magnitude similar to OIC 2 and 4 occurs intermittently during OIC 3 in both ODP Hole 806B and MD97-2140 cores, but no glacial record is known on Mt Giluwe from this time, presumably because ice did not advance to the same extent as the ensuing LGM. However, Highland vegetation histories suggest low treelines from 35,000-28,000 cal yr BP (Prentice et al. 2005). In contrast, the temperate glaciations of Australia and New Zealand are characterised by large advances of ice late in OIC 3 (Barrows et al., 2001; Suggate and Almond, 2005; Mackintosh et al., 2006) when temperate latitude SST cools to levels similar to the LGM (Barrows et al., 2007a).

The youngest glaciation, the Tongo Glaciation, reached its maximum extent at 19.4 ka during OIC 2. The oldest limiting radiocarbon age of 20,300 + 130/-140 cal yr BP (Table 1) places maximum ice advance shortly before this. Exposure ages are likely to be minimum estimates for the age of a moraine because of the lag time between deposition and geomorphic stabilisation/colonisation by vegetation. Bearing this in mind, the volume of the terminal moraine at the site dated (GLW-11) is similar to that of the other four terminal moraines in that set. This indicates that the ice was only briefly at the maximum position, probably less than 1000 years.

The maximum extent of the Tongo Glaciation occurs during the maximum phase of cooling in the equatorial Pacific Ocean between 19.8 ka (ODP Hole 806B) and 16.8 ka (MD97-2140). In the central Pacific Ocean on Mauna Kea in Hawaii, maximum glaciation also occurs between 19–16 ka (Blard et al., 2007). This timing coincides with the last glacial maximum, unlike some other locations in the tropics (e.g., Smith et al., 2005b). Differences in ages between Blard et al. (2007) and Pigati et al. (2008) from similar moraines may be largely due to production rate calibration and scaling schemes. The age of maximum ice extent during the LGM in the Pacific Ocean tropics is very similar in timing to the age of maximum ice extent in the temperate mountains of the western Pacific Ocean. Barrows et al. (2001, 2002) found that maximum ice extent occurred in Australia between 17–20 ka, after a period of periglacial activity centred at 22 ka (Barrows et al., 2004). The last major West Coast glacial advance culminated at 21.5 ka in New Zealand (Suggate and Almond, 2005) with retreat well underway by 17 ka (Schafer et al., 2006). Throughout the Indo-Pacific region, sea-surface temperature (SST) was at a minimum at 21 ka (Barrows et al., 2000; Barrows and Juggins, 2005).

Ice on Mt Giluwe retreated only slowly from its maximum extent for 4000 years, before a more rapid retreat from 15.4 ka over the subsequent 4000 years. On the southwest flank, there is on average one moraine per 150 years. Our radiocarbon ages indicate rapid establishment of bogs behind the moraines as the ice retreated (Table 1). Ice was absent from the highest elevations after 11.5 ka with no late Holocene advances. Similarly on Hawaii, retreat of ice after 16 ka was marked by a brief pause at 15.5 ka, then rapid retreat leaving the mountain ice free by 15 ka (Blard et al., 2007). The pattern of tropical ice retreat varies from that seen in the temperate western Pacific Ocean. In both Australia and New Zealand, the deglaciation is marked by one major readvance of ice at 16–19 ka (Barrows et al., 2001, 2002; Suggate and Almond, 2005). However, across 45° of latitude there is very rapid retreat of ice after 15 ka. In the mid latitude SST records, warming is rapid after 18–19 ka, and reaches near Holocene temperatures by 15 ka (Barrows et al., 2007a). Similarly in the tropical deep-sea cores, warming was very rapid after 16.8–19.8 ka increasing to Holocene level temperatures at 14.9 ka (ODP Hole 806B) and 12.3 ka (MD97-2140).

The strong relationship between glacial maxima and SST minima across the central and western Pacific Ocean indicates a common regional climatic forcing of glaciation through atmospheric cooling. The above comparison between the glacial record of Mt Giluwe with records in the temperate latitudes reveals a distinctly Southern Hemisphere pattern (Barrows et al., 2007a) in the western Pacific Ocean tropics. Blard et al. (2007) suggested that the warming pattern in the central Pacific followed the North Atlantic pattern, as recorded in the Greenland ice cores, but there are few differences between Hawaii and the western Pacific Ocean glacial records. In both locations there is no evidence for a significant readvance of ice or North Atlantic Ocean style cooling during the Younger Dryas Chronozone (YD), which is actually a time of warming in the western Pacific Ocean (Barrows et al., 2007b; Williams et al., 2009). Suggested cooling during the YD nearby at New Caledonia (Correge et al., 2004), has no support from independent estimates of SST, nor from the glacial records. There are also no glacier advances associated with Dansgaard-Oeschger events which are classically North Atlantic Ocean phenomenon. At present we are not aware of advances of ice during OIC 3 in the Tropical glacier record but these are prominent in the temperate glaciated areas. These latter advances appear to be events associated with large changes in sensible heat content in the Southern Ocean (Barrows et al., 2007a).

Despite the Southern Hemisphere character of the timing of deglaciation in the western Pacific Ocean, the first order timing of glacial maxima is of a global nature. There is a strong rhythm matching glacier maximum timing on Mt Giluwe (20 ka, 62 ka, 146 ka) and the periodicity of obliquity in the Earth’s orbit (40 ka). These times match minima in solar radiation during summer at high latitudes in the Northern Hemisphere. These times also match troughs in carbon dioxide concentrations as recorded in ice cores (Barnola et al., 1987; Petit et al., 1999). This provides support that synchronisation of global cooling during glacial maxima was aided by low greenhouse gas concentrations.

5.1. Climate change during the last glacial maximum from glacial evidence

Direct dating on the moraines of Mt Giluwe allows us to reassess the Tropical LGM temperature paradox in the western Pacific Ocean.
Ocean. We found the maximum extent of ice during the LGM less than previously mapped (Löfler, 1972). An estimate of the temperature difference relative to present can be approximated from the elevation of the ice cap on the southwest flank. Löfler (1972) used a THAR (toe to headwall altitude ratio) of 0.5 to estimate the equilibrium line altitude (ELA) on Mt Giluwe, and we use the same approach here to provide a comparison. A more sophisticated glaciated reconstruction will follow. Ice extended from a maximum of about 3800 m at the highest point of the headwall down to 3300 m. A THAR of 0.5 gives an ELA of 3550 ± 100 m. This is consistent with the highest elevation of lateral moraine in the Tongo River Valley of about 3600 m. This ELA estimate is at the upper limit of the original estimate of 3500–3550 m made by Löfler (1972).

There are presently no glaciers in Papua New Guinea (PNG). Mt Giluwe reaches a maximum elevation of 4367 m and is currently ice and permanent snow patch free. Löfler (1972) estimated that the 1970’s equilibrium line at about 4600 m. The climate of montane PNG within the humid inner tropics is monsoonal, so the ELA will most closely correspond to the summer 0 °C isotherm (Benn et al., 2005). Using 4600 m as a first estimation of the modern ELA, the difference between the LGM and present ELA is about 1050 m. Combining this estimate with a lapse rate of 6 °C/1000 m indicates that the LGM was approximately 6 °C colder than present. Even given possible errors in the position of the ELA (±100 m) and the likelihood that the lapse rate would be a minimum for the drier mountain air during the LGM, the temperature difference is unlikely to be less than 5 °C. This estimate of cooling is similar to the estimate of 5–7 °C for Hawaii (Blad et al., 2007). Because previous ice caps that formed on Mt Giluwe were larger in magnitude, this temperature difference is likely to be a minimum for these glaciations.

A 5–6 °C temperature difference contrasts with the most recent estimates of an SST difference of a maximum of 3 °C for the tropical western Pacific Ocean (MARGO Project Members, 2009). This supports previous observations that greater cooling occurred at high altitudes in New Guinea compared to the sea-surface. This differential is not unique to the tropics and is also observed in the temperate latitudes of eastern Australia (Barrows et al., 2000). Given the improved dating and the refinement in SST methodologies, a climatological explanation must be sought to explain the difference. Although the magnitude of cooling in the terrestrial environment is magnified, the timing of cooling between the terrestrial and marine realms is very similar, implying a common climatic control, probably through atmospheric cooling.

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Appendix. Supplementary data

Supplementary data associated with this article can be found in the online version, at doi:10.1016/j.quascirev.2011.05.022.

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