



Short Paper

Late Quaternary glaciation and equilibrium-line altitudes of the Mayan Ice Cap, Guatemala, Central America

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ABSTRACT

The Sierra los Cuchumatanes (3837 m), Guatemala, supported a plateau ice cap and valley glaciers around Montaña San Juan (3784 m) that totaled ~43 km² in area during the last local glacial maximum. Former ice limits are defined by sharp-crested lateral and terminal moraines that extend to elevations of ~3450 m along the ice cap margin, and to ca. 3000–3300 m for the valley glaciers. Equilibrium-line altitudes (ELAs) estimated using the area–altitude balance ratio method for the maximum late Quaternary glaciation reached as low as 3470 m for the valley glaciers and 3670 m for the Mayan Ice Cap. Relative to the modern altitude of the 0°C isotherm of ~4840 m, we determined ELA depressions of 1110–1436 m. If interpreted in terms of a depression of the freezing level during maximal glaciation along the modern lapse rate of $-5.3^{\circ}\text{C km}^{-1}$, this ΔELA indicates tropical highland cooling of ~ 5.9 to $7.6 \pm 1.2^{\circ}\text{C}$. Our data support greater glacial highland cooling than at sea level, implying a high tropical sensitivity to global climate changes. The large magnitude of ELA depression in Guatemala may have been partially forced by enhanced wetness associated with southward excursions of the boreal winter polar air mass.

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Introduction

Background and context

Former tropical glacier extents and equilibrium-line altitudes (ELAs) provide important evidence about paleoclimates during the last glacial period (Porter, 2001; Kaser and Osmaston, 2002; Benn et al., 2005; Mark et al., 2005; Osmaston, 2005), yet few data are available to constrain past temperature reductions at tropical high altitudes. Glaciation in Central America is limited to those peaks older than the late glacial and having altitudes exceeding ca. 3500 m in Guatemala and Costa Rica (Anderson, 1969a; Anderson et al., 1973; Hastenrath, 1974; Horn, 1990; Orvis and Horn, 2000; Lachniet, 2007; Marshall, 2007). Glacial evidence suggests a 5–8°C highland cooling that is considerably more than at sea level (CLIMAP, 1976; Crowley, 2000; Lea, 2004), an observation dubbed the “tropical paleotemperature paradox” (Rind and Peteet, 1985; Ballantyne et al., 2005). Global last glacial maximum (LGM) sea-surface temperature (SST) and land-surface temperature depressions were $2.7 \pm 0.5^{\circ}\text{C}$ and $5.4 \pm 0.3^{\circ}\text{C}$, respectively, with more cooling in the tropical Atlantic than the tropical Pacific (Ballantyne et al., 2005). However, further paleotem-

perature data are needed from locations sensitive to sea-surface temperature forcing in order to evaluate last glacial highland temperature reductions. Our study area is a key location for such a test because it is closely bordered by the Caribbean Sea, the Gulf of Mexico, and the Pacific Ocean. To this end, we have completed glacial geologic field mapping in the Sierra los Cuchumatanes in the northern Guatemalan highlands, reconstructed past equilibrium-line altitudes (ELAs) for the last local glacial maximum (LLGM), and provide estimates of tropical highland cooling.

Study area

The Sierra los Cuchumatanes (15°30.0'N, 91°30.0'W) is a high limestone plateau, located within the Mayan Highlands at the southernmost limit of North America (Fig. 1) (Anderson et al., 1973; Bundschuh et al., 2007a). The plateau contains the highest non-volcanic point in Central America at 3837 m (Bundschuh et al., 2007a), and traces of former glaciation were documented by previous studies (Enjalbert, 1967; Anderson, 1969b; Anderson et al., 1973; Hastenrath, 1974). Although no radiometric ages for the glaciation exist, the slightly weathered till suggested a late Wisconsin age (Anderson, 1969b). The late Quaternary ELA was estimated at ca. 3650 m (Hastenrath, 1974). The plateau also shows karst landforms with largely internal, subterranean drainage (Blake, 1934; Enjalbert, 1967; Anderson et al., 1973; Hastenrath, 1974). The glacial geology of nearby Montaña San Juan, located south of the main plateau region, has not previously been described.

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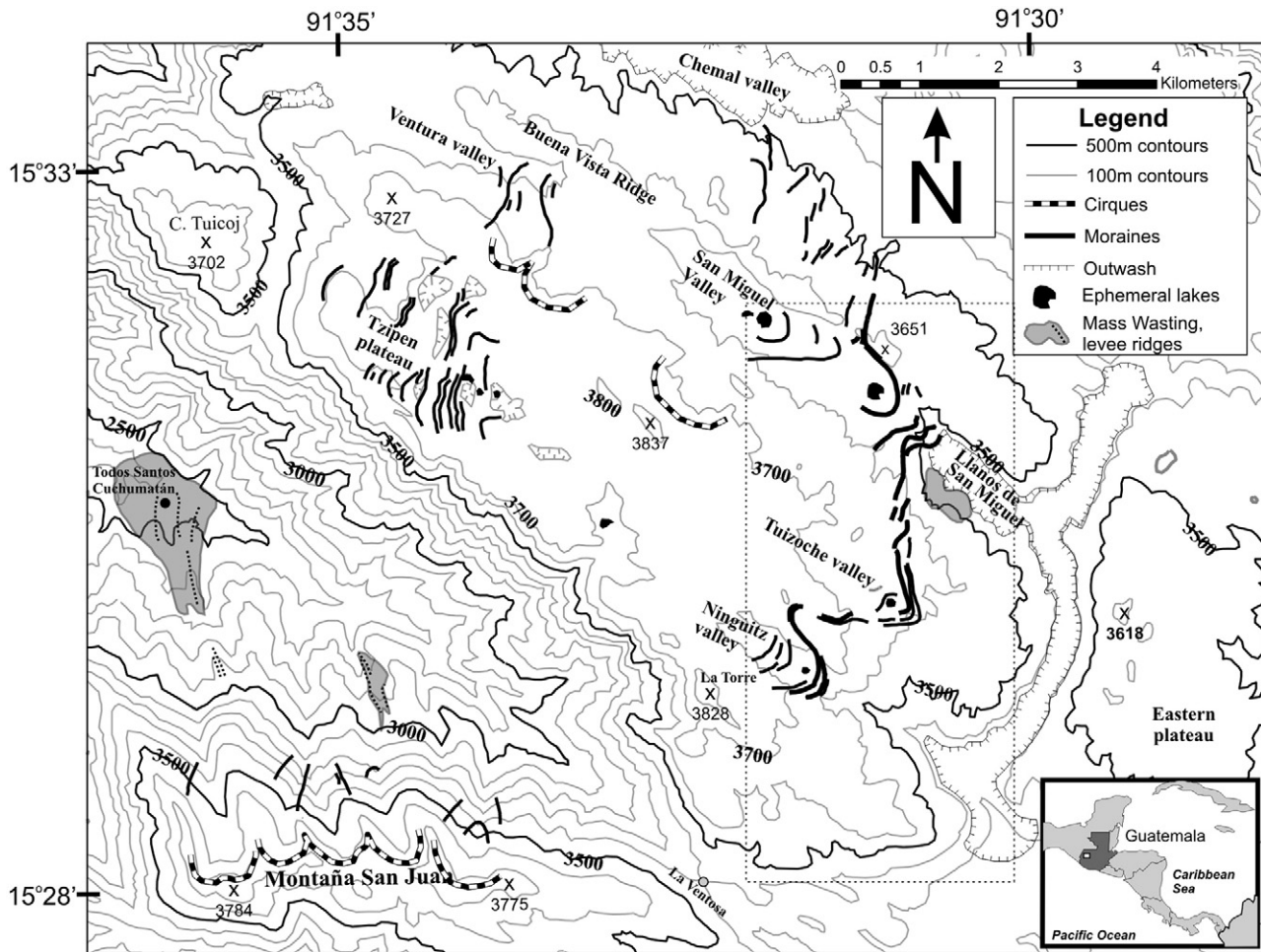


Figure 1. Topographic map of the field area and glacial geology in the Sierra los Cuchumatanes region, based on field and aerial photographic mapping. The rectangle delineated by dashed lines is the approximate area covered in the stereopair in Fig. 2. Contour lines are at 100-m intervals. Selected peak altitudes are indicated, as well as locations mentioned in the text. Inset is Central America, with Guatemala shaded and study area denoted by the small white box.

Modern climate of Guatemala

The climate of Guatemala is humid tropical, with a boreal summer wet season (May–October) and a winter dry season (November–April) when temperatures are influenced by cold surges (*nortes*) of polar air (Portig, 1965; Bundschuh et al., 2007b). Mean annual temperature varies from $\sim 27^{\circ}\text{C}$ near sea level to 14.2°C in Todos Santos Cuchumatán ($15^{\circ}30.5'\text{N}$, $91^{\circ}36.3'\text{W}$, 2480 m, ~ 5 km south of field area), and 18.1°C at Huehuetenango ($15^{\circ}19.3'\text{N}$, $91^{\circ}28.1'\text{W}$, 1870 m), the two stations nearest the glaciated plateau (INSIVUMEH, 2008). Todos Santos receives ~ 1155 mm annual precipitation, which falls as rain primarily during the wet season. Based on analysis of climate data (INSIVUMEH, 2008), the temperature/altitude lapse rate is $-5.3^{\circ}\text{C km}^{-1}$ for all stations and $-6.3^{\circ}\text{C km}^{-1}$ between Huehuetenango and Todos Santos. The altitude of the 0°C isotherm is 4840 ± 230 m, which is typical of humid tropical stations, and our observations of ice formation in March 2006 suggest the night time freezing level descends to 3500 m during the winter.

Tropical glaciation and Late Quaternary paleoclimate

Humid-region inner tropical glaciers respond most strongly to variations in mean annual temperature and, to a lesser degree, to precipitation (Seltzer, 1994; Kaser and Osmaston, 2002; Seltzer et al., 2002; Benn et al., 2005). Small inner tropical glaciers—such as those formerly present in Guatemala—may be particularly sensitive to climate changes (Nesje and Dahl, 2000), and commonly have

response times of 10 to 50 yr (Oerlemans, 1994). Climate variations are recorded in the glacial equilibrium-line altitude, which is lowered during positive mass balance years and *vice versa*. ELAs may be estimated from the position of glacial moraines and reconstructed ice topography (Furbish and Andrews, 1984; Benn et al., 2005), and on inner tropical glaciers are found in close proximity to the 0°C isotherm (Kaser and Osmaston, 2002). Inferences of past temperature changes at high altitudes relative to today may be calculated by either 1) differencing the modern temperature at the altitude of the LGM ELA from 0°C , the inferred temperature at the former ELA; or 2) by multiplying the change in ELA (ΔELA) by the environmental lapse rate.

Equilibrium lines of LGM tropical glaciers were depressed by 400 to 1400 m, averaging 900 ± 135 m (Porter, 2001), and were associated with temperature depressions of 4.2° to 7.5°C (Mark et al., 2005). ΔELA in the circum-Caribbean ranged from 500 to 1625 m (Lachniet and Vazquez-Selem, 2005). In Costa Rica the LGM cooling inferred from glacial evidence was estimated to be ~ 7 to 9°C (Orvis and Horn, 2000; Lachniet and Seltzer, 2002) and is constrained by a minimum age of $10,140 \pm 140$ ^{14}C yr BP (Orvis and Horn, 2000). The LGM temperature depression inferred from glacial moraines in the Colombian Andes (Mark and Helmens, 2005) is $\sim 8^{\circ}\text{C}$ and is $\sim 8.8 \pm 2^{\circ}\text{C}$ for the Cordillera de Mérida, Venezuela (Stansell et al., 2007).

Pollen and isotopic data from the Guatemalan lowland Lake Quexil were used to infer an arid climate and an “Ice Age” temperature depression of 6.5 to 8.0°C (Leyden, 1993), or alternatively temperature depression of 4.7 to 6.5°C (Leyden et al., 1994). However, a new

chronology for high-resolution proxy data from a Lake Petén Itzá sediment core confirms that the LGM (23 to 18 ka) was moist in the Guatemalan lowlands (Hodell et al., 2008; Bush et al., 2009). Following drying associated with Heinrich event 1, the Petén remained dry between 18.0 and 14.7 ka, just prior to a return to wet conditions during the Bølling–Allerød interstade, which was in turn followed by a relatively dry Younger Dryas stade (Hodell et al., 2008). The wet conditions between 23 and 18 ka may have been the result of increased *nortes*, which bring orographic precipitation to leeward mountains, and is consistent with a pollen-based temperature reduction of 4 to 5°C (Bush et al., 2009). The LGM-to-Holocene transition in Costa Rica was marked by an ~8°C temperature rise at 2300 m elevation (Hooghiemstra et al., 1992), and the Younger Dryas stade was marked by a 2.5 to 3.0°C temperature reduction (Islebe and Hooghiemstra, 1997).

Methods

Glacial geology and ELA estimation

Glacial landforms were mapped in the field and with topographic maps and aerial stereophotographs (INSIVUMEH, 2007). The maximum ice extent was delineated by the lowest limit of terminal moraines and extrapolated lateral moraines on Montaña San Juan. To constrain realistic basal shear stress for the former glaciers, ice thicknesses inferred from the moraine limits were adjusted to match the glacial geomorphology (Paterson, 1981; Benn and Evans, 1998) via the equation

$$\tau = \rho_i g h \sin \alpha$$

where τ is the shear stress, ρ_i the density of ice (917 kg m⁻³), g the gravitational acceleration (9.81 m/s⁻²), h is the ice thickness (m) and

$\sin \alpha$ is the surface slope of the ice given in degrees. For typical glaciers the range of basal shear stress values range from 50 to 150 kPa (Paterson, 1981).

We calculated ELAs using the area–altitude balance ratio (AABR) and accumulation–area ratio (AAR) methods (Benn and Evans, 1998; Kaser and Osmaston, 2002; Benn et al., 2005; Rea, 2009). The AABR method takes into account the vertical mass-balance gradients (Osmaston, 2005; Rea, 2009) and was calculated for balance ratios (BRs, the ratio of the mass-balance gradients in the ablation and accumulation zones) of 1.0 to 5.0 (Benn and Evans, 1998; Osmaston, 2005). The tropical balance ratio is affected by year-round ablation and an almost constant elevation of the 0°C isotherm, resulting in higher ratios in the tropics (>2.0 in Kaser and Osmaston, 2002) than the mid-latitudes values of 1.8 (Furbish and Andrews, 1984). For the AAR method we used glacier hypsometries and ratios ranging from 0.6 to 0.7 (Kaser and Osmaston, 2002). Temperature reductions and uncertainties were calculated by multiplying the modern lapse rate of $-5.3^\circ\text{C km}^{-1}$ by the ΔELA and the ± 230 m uncertainty in the altitude of the 0°C isotherm, respectively.

Results

Glacial geomorphology

The plateau has a subdued relief of ~400 m and presents well-defined glacial erosional features. Most valleys show a characteristic broad U-shape and lee-side plucked bedrock surfaces. Striations appear to have been mostly removed by chemical erosion. The central portion of the plateau contains three cirques that supported ice that flowed northward into the Ventura and San Miguel valleys after ice recession from maximum limits (Fig. 1). We identified two prominent moraine groups, based on relative size and moraine-crest morphology. The first group consists of

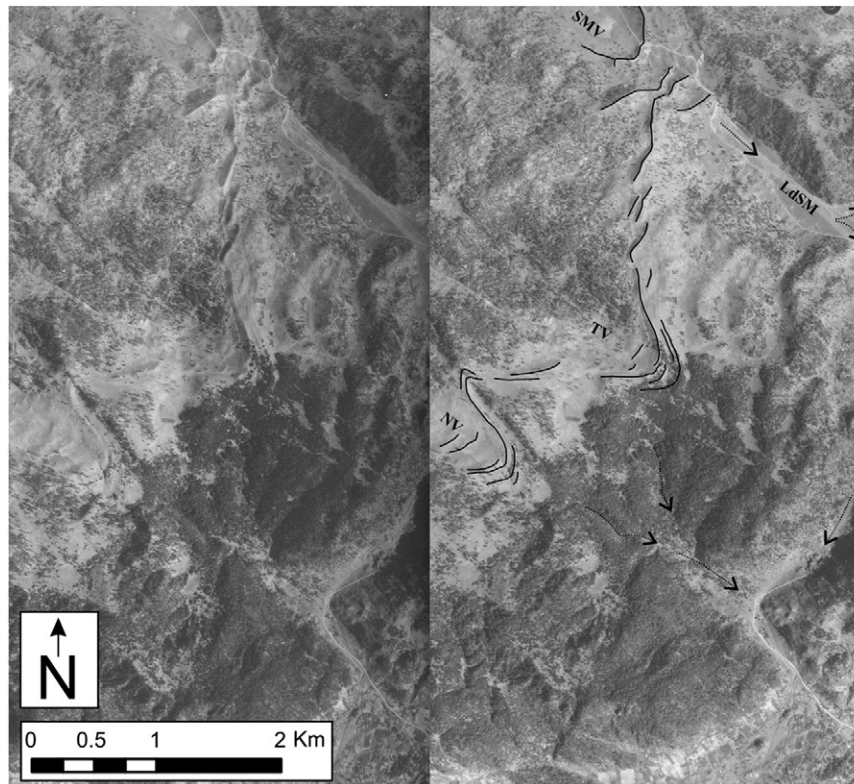


Figure 2. Stereopair of the eastern plateau region glaciated by the former Mayan Ice Cap. Some of the prominent moraine ridges are delineated with black lines. The dashed arrows show former glaciofluvial drainage directions. Locations are San Miguel Valley (SMV), the Llanos de San Miguel (LdSM) Valley outwash train, Tuizoche Valley (TV), and Ninguitz Valley (NV). The karstic character of the high limestone plateau is also evident in the numerous closed depressions.

lateral moraines and closely spaced and overlapping terminal moraines along the plateau's eastern valleys (Fig. 1), which are clearly visible in the stereopair image (Fig. 2). The largest (LLGM) moraines range up to 15 m high and have distinctly peaked crests, while others, interpreted as a possibly older sequence, are observed with more rounded broad crests that are located immediately downvalley but in contact with the sharper-crested moraines. Photographs of several of the glacial geomorphic features are shown in Figure 3. Lake basins have formed in several locations where they were impounded by terminal moraines and in karst dolines (Fig. 3).

The second moraine group consists of smaller, muted, broad-crested overlapping moraines which are found up-valley from the larger LLGM moraines, and nested in groups across a large portion of the northwestern Sierra Los Cuchumatanes (the Tzipen plateau). A few poorly developed moraines indicate that ice overtopped Buena Vista ridge and flowed north to the Chermal Valley as narrow outlet ice tongues. Steep topography (45° slopes and higher) on the southern plateau margin would have limited the extent of outlet tongues, which may have rapidly lost mass due to crevasse-induced calving along the steep fault scarp (Fig. 1). The glacial evidence observed on the plateau corresponds to a history of multiple glacial advances and a

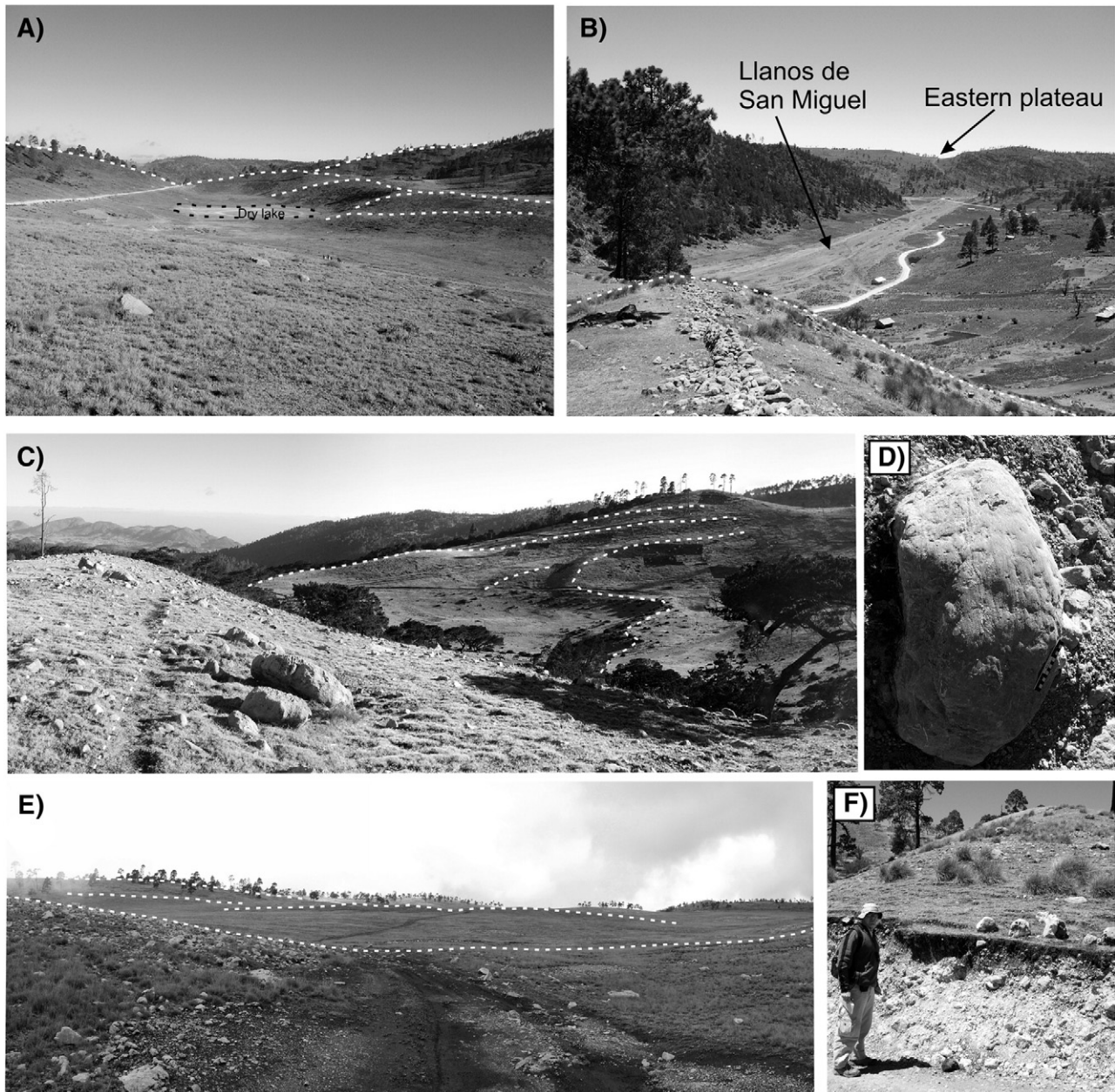


Figure 3. Selected photographs of glacial geologic features in the Cuchumatanes region. A) is a view southeast down the San Miguel Valley, which shows lateral and end moraines which dam an ephemeral dry lake basin. B) is a view looking east downvalley of the Llanos de San Miguel glacial outwash plain. Where the plain encounters the limestone ridge of the Eastern plateau, the drainage channels split into northerly and southerly flowing branches. Photo taken from the crest of a prominent lateral/end moraine ridge crest (outlined with dashed white line) in the San Miguel Valley. C) View of right-lateral moraines (white dashed lines) to the south, including a few generations of recessional moraines, in the Tuizoché Valley. Photo was taken from the crest of the most prominent left-lateral moraine in the valley. D) is a photograph of a striated and faceted cobble from a quarry in glacial till, San Miguel Valley; ruler for scale. E) is a view to the northwest of the muted recessional moraines (crests denoted by dashed white lines) on the Tzipen plateau. F) Organic-rich soil development (~5–10 cm thick) on the crest of a terminal moraine in San Miguel Valley, formed on a till composed of chemically unweathered limestone clasts. Trees for scale in A, B, C, and E.

retreat of the ice cap, with maximal advances not significantly exceeding the LLGM limits.

Below the San Miguel valley terminal moraines, a now-dry narrow outwash valley train extends for ~2 km (Fig. 1). Diamict deposits, interpreted as a subglacial till, were observed at the western edge of the Ventura Valley to an altitude of ~3500 m, which is overlain by outwash in places. Between ~3560 and 3600 m a set of four small muted moraines are evident in Ventura Valley, one of which dams a dry lake. Most exposed morainal boulders have moderately developed rillenkarren surfaces and relief that indicate post-depositional chemical weathering. The average rillenkarren depth measured on three boulders was 15.6 mm ($n=29$; max 32 mm), and chert nodules on one boulder stand an average height of 9.8 mm ($n=13$) above the weathered limestone. These measurements provide minimum erosion amounts since moraine stabilization.

The Montaña San Juan (3784 m) is flanked by several north-facing cirques that contain rounded and scoured bedrock and small closed basins. Cirque headwalls at ~3650 m are joined by well-defined arêtes and small horns. Below the cirques, small lateral and end moraine segments descend downvalley. Below the glacial limit on Montaña San Juan, several linear ridges lead to a large hummocky lobe of sediment with a central depression that forms the town site of Todos Santos Cuchumatán. We interpret these linear ridges as debris-flow levees, and the hummocky lobe as a debris flow that partially dammed the Rio Limón valley (Fig. 1). We examined numerous glacial sedimentary exposures in several quarries, home sites, and in the walls of hand-dug wells in moraines in the San Miguel, Tuizoche, and Ninguitz valleys. Bouldery tills consist of a poorly rounded clast-

supported diamict, typically lacking facets or striations. Despite an extensive search, no organic material suitable for ^{14}C dating was found within or associated with glacial deposits, from which we infer scarce vegetation prior to and during glaciation.

Ice cap and valley glacier reconstruction

The LLGM extent of the here-named Mayan Ice Cap is well defined by moraines on the eastern plateau, reaching lowest altitudes of ~3450 m. Fewer moraines along the western plateau are present, so ice limits there were estimated by the transition from U-shaped to V-shaped valleys. The Mayan Ice Cap covered an area of ~39.2 km² with ice thicknesses reaching >200 m, to attain altitudes of ca. 3800 m. At least four well-defined valley glaciers on the north-facing slopes of Montaña San Juan total ~3.8 km² in area and reached lower limits of 3000–3300 m (numbered I through IV on Fig. 4).

Equilibrium-line altitudes and temperature depression

Using a balance ratio of 2.0, the LLGM ELA was 3700 m for the Mayan Ice Cap and 3520 m for the valley glaciers (Table 1). The ELAs are 3670 m for the ice cap and 3470 m for the valley glaciers using a balance ratio of 5.0, considered more typical of inner tropical glaciers (Kaser and Osmaston, 2002). The LLGM ELA estimates are closely bounded by the small altitudinal range of glacial evidence of the plateau (3470 to 3840 m) and valley glaciers (3100 to 3650 m). An ELA of ~3500 m would place several areas in the Sierra los Cuchumatanes

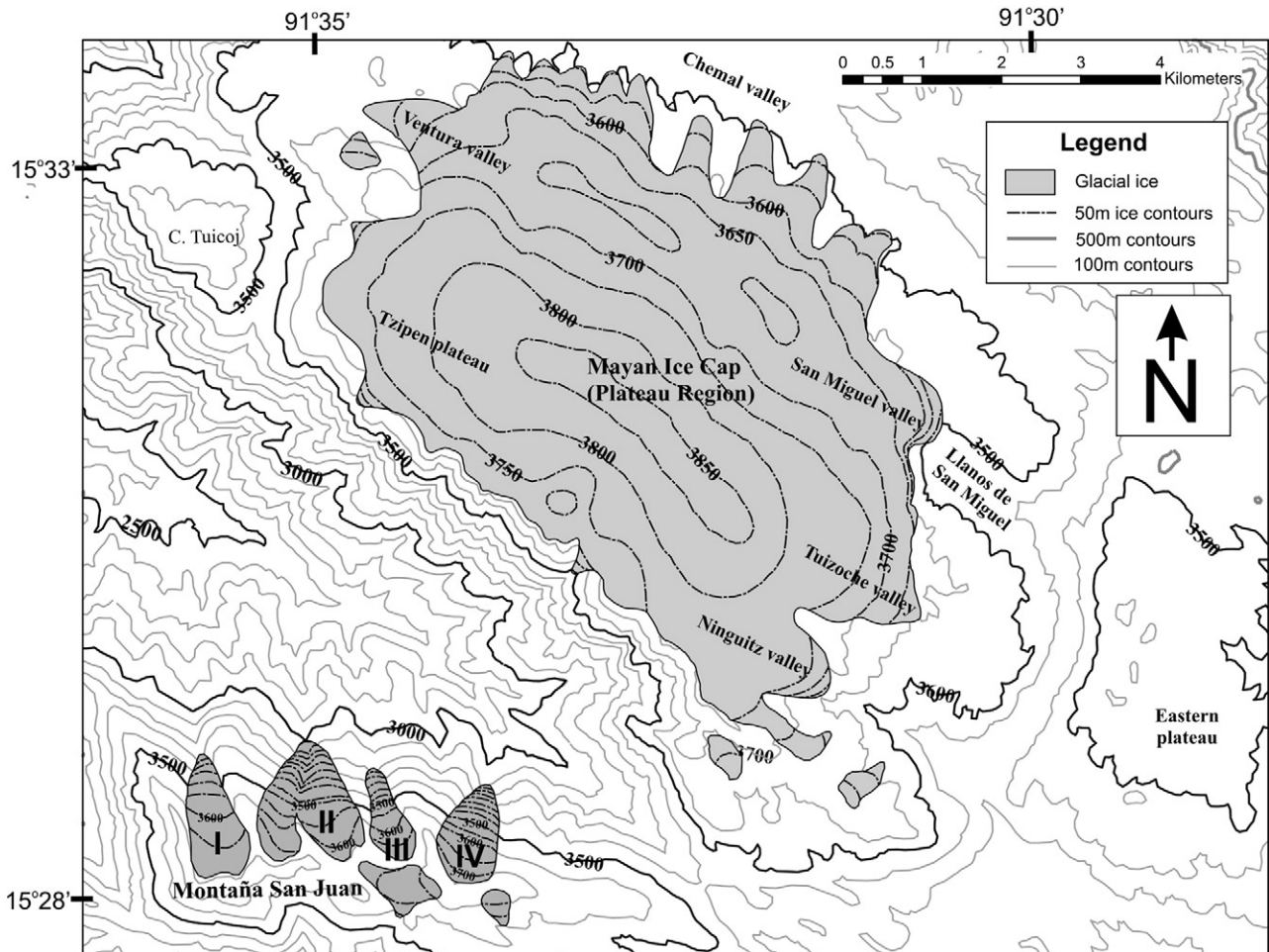


Figure 4. Map of the reconstructed Mayan Ice Cap and valley glaciers during maximal glacial limits. Ice limits are based on mapped maximum moraine limits and glacial geomorphology. The ice contours (black dashed lines) have 50-m interval for an enhanced representation of ice thickness.

Table 1
Equilibrium-line altitudes for the Mayan Ice Cap and four valley glaciers, rounded to nearest 10 m.

Paleoglacier	Headwall altitude (m)	Terminus altitude (m)	AAR ratio				Balance ratio			
			0.6	0.65	0.7	1.0	2.0	3.0	4.0	5.0
Ice Cap	3800	3450	3660	3650	3640	3730	3700	3690	3680	3670
Valley glacier I	3650	3300	3550	3540	3520	3600	3580	3560	3550	3540
Valley glacier II	3650	3000	3460	3440	3390	3490	3450	3420	3400	3420
Valley glacier III	3650	3200	3510	3500	3480	3550	3520	3490	3470	3460
Valley glacier IV	3700	3100	3510	3490	3470	3560	3520	3500	3480	3470
Valley mean	3663	3150	3508	3493	3465	3550	3518	3493	3475	3473
Valley std. dev.	22	112	32	36	47	39	46	50	53	43

above the LLGM snowline: for example, Cerro Tuicoy (3702 m, ~ 3 km²), a broad upland plateau ~ 7 km² just east of the Llanos de San Miguel (reaching 3618 m); and an ~ 1.5 km² plateau north of Valle Chermal. Aerial photographic analysis of these areas does not reveal obvious glacial landforms, but they may have supported perennial ice at maximum cooling. These areas, along with snow/ice fields on the southern flank of Montaña San Juan, may have exceeded 15 km².

The ELA depressions (Δ ELA) exceed 1100 m, with a maximum of 1420 m for valley glacier II (Table 2). The Δ ELAs are larger than the average “tropical” value of 900 ± 135 m (Broecker, 1997; Porter, 2001). The Δ ELAs have not been corrected for a fall in sea level (up to -120 m at maximum ice volume) because 1) the timing of glaciation is unknown, and 2) the accuracy of making such a correction is uncertain (Benn et al., 2005; Osmaston, 2006). Paleotemperature depressions determined by multiplying the lapse rate by the Δ ELA during the LLGM were 5.9 to $7.6 \pm 1.2^\circ\text{C}$ relative to modern (Table 2). These depressions compare well to those estimated by differencing the modern temperature at the altitude of the putative LLGM ELA from 0°C , and are 6.4°C and 7.5°C for past ELAs of 3670 m (Ice Cap), and 3470 m (valley glacier average).

Discussion

The Guatemalan ELAs at the LLGM of 3470–3700 m are statistically identical to those from Costa Rica (Orvis and Horn, 2000; Lachniet and Seltzer, 2002), which suggests that the freezing level was spatially consistent over Central America. The LLGM temperature depressions of 5.9 to $7.6 \pm 1.2^\circ\text{C}$ are larger than the average LGM tropical sea-surface temperature depressions of $2.7 \pm 0.5^\circ\text{C}$ (Ballantyne et al., 2005), regardless of the precise timing of glaciation in Guatemala. We interpret the Δ ELA as forced by neotropical SST, which is the dominant control on tropical freezing levels (Kageyama et al., 2005; Bradley et al., 2009). In contrast, only very large changes in snow accumulation result in significant shifts of glacial ELAs (Seltzer, 1994). Our data show greater highland cooling than at sea level in the

Caribbean Sea and subtropical North Atlantic Ocean (~ 4 to 6°C ; Bard et al., 2000; Lea et al., 2003), and the 4°C cooling at the LGM in the Gulf of Mexico (Flower et al., 2004). Such cooling has been demonstrated during periods of decreased thermohaline circulation, in some cases associated with Heinrich events (Bard et al., 2000; Lea et al., 2003). During these cold periods, a higher frequency of polar outbreaks (*nortes*) (Schultz et al., 1998) due to a southward displaced polar jet stream may have resulted in both cooler and wetter conditions (Hodell et al., 2008; Bush et al., 2009). The record from Lake Peten-Itzá (Hodell et al., 2008) clearly shows that the LGM was wet in Central America, and also coincided with cold temperatures in the Cariaco basin that were depressed ca. 4°C from early Holocene values (Lea et al., 2003).

If the maximum ice limits in Guatemala coincided with a wet LGM, then some portion of the Δ ELA may be attributed to increased mass balance, so that our high-altitude land surface temperature depressions of 5.9 to $7.6 \pm 1.2^\circ\text{C}$ may be overestimated. Because a large accumulation increase (ca. 1000 mm/yr) can depress tropical glacier ELAs by ~ 300 m for a constant temperature (Seltzer, 1994), our temperature depression estimates may be up to $\sim 1.6^\circ\text{C}$ too high at the lapse rate of $-5.3^\circ\text{C km}^{-1}$. On the other hand, if the maximum glaciation occurred during drier-than-average conditions, such as during Heinrich event one, an environmental lapse rate of $-6.3^\circ\text{C km}^{-1}$ (as between the two highland stations of Huehuetenango and Todos Santos) would yield temperature depressions of 7.0 to $8.6 \pm 1.5^\circ\text{C}$. In either a dry vs. a wet LGM, highland surface temperature depression was larger than at sea level, suggesting an alpine amplification of climate change, possibly due to changes in lapse rate (Bush et al., 2009) or anomalous atmospheric circulation.

A lack of suitable material for radiometric dating prohibited establishment of an absolute glacial chronology. In Mexico, ³⁶Cl dates show glacial advances between 20.0 and 17.5 ka (Hueyatlaco-1), 17.0 to 14.0 ka (Hueyatlaco-2), a recessional moraine phase between 14.0 and 13.0 ka, and a later advance at 12.0 to 10.0 ka (Milpulco-1). The

Table 2
Depression of equilibrium-line altitudes (Δ ELA) and temperature (Δ T) for the Mayan Ice Cap and four valley glaciers for the various AAR- and AABR-determined ELAs.

	AAR ratio				Balance ratio			
	0.6	0.65	0.7	1.0	2.0	3.0	4.0	5.0
Δ ELA (m)								
Ice Cap	1180	1190	1200	1110	1140	1150	1160	1170
Valley glacier I	1290	1300	1320	1240	1260	1280	1290	1300
Valley glacier II	1380	1400	1450	1350	1390	1420	1440	1420
Valley glacier III	1330	1340	1360	1290	1320	1350	1370	1380
Valley glacier IV	1330	1350	1370	1280	1320	1340	1360	1370
Valley mean	1333	1348	1375	1290	1323	1348	1365	1368
Valley std. dev.	32	36	47	39	46	50	53	43
Δ T ($^\circ\text{C}$)								
Ice Cap	6.3	6.3	6.4	5.9	6.0	6.1	6.1	6.2
Valley glacier I	6.8	6.9	7.0	6.6	6.7	6.8	6.8	6.9
Valley glacier II	7.3	7.4	7.7	7.2	7.4	7.5	7.6	7.5
Valley glacier III	7.0	7.1	7.2	6.8	7.0	7.2	7.3	7.3
Valley glacier IV	7.0	7.2	7.3	6.8	7.0	7.1	7.2	7.3
Valley mean	7.1	7.1	7.3	6.8	7.0	7.1	7.2	7.2
Valley std. dev.	0.2	0.2	0.3	0.2	0.2	0.3	0.3	0.2

Hueyatlaco-1 advance occurred during wet and cold conditions (Bard et al., 2000; Hodell et al., 2008), to which we tentatively correlate our LGM moraines. Similarly, we correlate the recessional moraine group with either the Hueyatlaco-2 or later recessional moraines of the Mexican glacial sequence. Regional deglaciation may have been associated with either the Bølling–Allerød interstade or Holocene epoch warming. A last glacial age of the moraines is suggested by the <20 mm average depth of boulder weathering from the rillenkarren and chert nodules. Our glacial moraine correlation is tentative, and future work on coring of moraine-dammed lakes in the Cuchumatanes is planned.

Acknowledgments

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