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Reconstruction of glacier equilibrium-line altitudes for the Last Glacial Maximum on the High Plain of Bogotá, Eastern Cordillera, Colombia: climatic and topographic implications

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ABSTRACT: The High Plain of Bogotá in the Andes of Colombia provides an exceptionally detailed record of glaciation. A two-stage Last Glacial Maximum (LGM) is noted in Bogotá; the older stage (max) presents an opportunity to reconstruct individual valley glaciers and explore spatial patterns. Well-mapped geomorphic features on topographic base maps permit the reconstruction of 23 palaeoglacier surfaces. Glacier extent varies across the region, with lower altitudes reached farther to the east. Equilibrium line altitudes (ELAs) are reconstructed using the area–altitude balance ratio (AABR) method, with BRs in three groups reflecting the W–E gradient in glacier extent and selected by minimising variation from group means. Average LGM ELA for all palaeoglaciers is 3488 m with a standard deviation of 182 m. The average lowering in ELA from LGM to modern of ca. 1300 m is best explained by a considerable drop in temperature. Significant intra-regional variance in LGM ELA can be ascribed to topography and its influence on precipitation and/or glacier form, with lower headwall elevations being correlated to larger accumulation areas. Copyright © 2005 John Wiley & Sons, Ltd.

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KEYWORDS: Last Glacial Maximum; equilibrium line altitudes; hypsometry; Colombia; glacial geology.

Introduction

The mountainous highlands in the region around Bogotá, Colombia, comprise an important region for evaluating the extent, timing and distribution of glacial advances that occurred during the Last Glacial Maximum (LGM) in South America. Located at 4–5 °N on the South American continent, within the northern reaches of the Andes, this tropical region lies at a climatic juncture between hemispheres and features clear evidence of previous glaciations. It is within the influence of the northeastern trade winds, and likewise has been grouped with circum-Caribbean sites in previous glacial histories (e.g. Lachniet and Vazquez-Selem, 2005). Located in the easternmost branch of the Andes, the region receives moisture from both the Amazon basin to the east, as well as the Pacific to the west (Helmens, 1988). ENSO dominates the interannual climate variability (Poveda *et al.*, 2001). Moreover, well-preserved glacial geomorphic evidence has yielded constraints on the chronology of tropical Andean glaciation and allows to test important hypotheses related to the interaction of glacier climate and topography.

At an altitude just below 4000 m, the mountains around the Sabana de Bogotá are currently not glacierised, but are high enough to have experienced repeated glaciations that pre-date the Holocene. There is an excellent record of moraines and geomorphologic features that have been mapped (Helmens, 1988, 1990). Moreover, the high plain on which Bogotá is located, the Sabana de Bogotá, is a closed sedimentary basin that supported large lakes during the Pleistocene. This basin was situated at an elevation between the extreme tree levels of glacial and interglacial periods, and contains well-dated lake sediment sequences and multiple pollen records (van der Hammen and Hooghiemstra, 1995, 2003). These independent chronostratigraphic records provide ecological proxy records to constrain contemporaneous climate conditions at the time the moraines were formed.

Analysis of the extent and timing of glacial events in this region provides an important point of comparison to link glacial activity between hemispheres during the last glacial cycle. It is a critical location between well-dated sites farther to the south along the Andes and to the north into the North American Cordillera not only to test the inter-hemispheric synchroneity of glaciation (i.e. Clapperton, 2000), but also to compare the

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relative spatial extent and climatic forcing of earlier glacier advances. The bounding dates of the last glacial cycle in tropical South America are not well constrained, but the longest sedimentary records from Lake Titicaca would imply that the initiation (the last interglacial–glacial transition) occurred after ca. 64 000 yr BP (Seltzer *et al.*, 2003). Furthermore, the same records indicate that the onset of deglaciation pre-dated that of the Northern Hemisphere by a couple of millennia (Seltzer *et al.*, 2002). Other sites in the Central Andes have yielded mapped moraines with cosmogenic dates recording multiple glacial advances during this time, with a much diminished glacier extent during the LGM (Smith *et al.*, 2005). Radiocarbon-dated geomorphic evidence from the mountains around Bogotá have indicated five successively less extensive glaciations during the last 50 000 yr BP (Helmens, 1988, 1990; Helmens and Kuhry, 1995; Helmens *et al.*, 1997). While the glacial chronology (discussed below) suggests much closer timing to the global maximum in ice volume at the LGM as defined by the EPILOG project at 21 000 $\pm$ 2000 yr BP (Mix *et al.*, 2001), the evidence indicates a two-stage LGM separated by a distinct warming episode and potentially abrupt deglaciation (Helmens *et al.*, 1996).

The existence of good moraine maps and valley topographic data for this region allows individual palaeoglacier surfaces to be reconstructed. A preliminary reconstruction of individual palaeoglaciers was incorporated as a regional case study into a global database of individual tropical palaeoglacier equilibrium line altitudes (ELAs) at the LGM (Mark *et al.*, 2005), but remained undocumented. Other previously published reconstructions of palaeoglacier ELAs have utilised only the average moraine altitudes to derive a regional average using simple altitude ratios (Lachniet and Vazquez-Selem, 2005). However, regional averages based on such simplified altitude-based methods are problematic since local topography can alter regional climate and affect glacier morphology (i.e. Kerr, 1993). Analyses of the global database concluded that local topographic influence can predominate over regional climate, causing intra-regional variation in ELA to complicate any regional climatic interpretation (Mark *et al.*, 2005). Therefore, meaningful regional climatic interpretations from glacier ELAs must be based on individual palaeoglaciers over the spatial range of glaciation.

In this paper, we review the glacial geology and present detailed methodology and mapped models of individual glacier forms used to estimate palaeo-ELAs in the Bogotá region. We discuss the results in the context of local precipitation gradients and also explore other topographical factors influencing the intra-regional variance in ELA.

Geographical setting

The high plain of Bogotá, or Sabana de Bogotá, is a large tectonic basin flanked by mountains reaching close to 4000 m a.s.l. located in the eastern-most chain (Cordillera Oriental) of the tropical Andes between 4°30′–5°15′N latitude. Presently, the mountains are glacier-free, but extensive glacial landforms (Helmens, 1988, 1990) give evidence of former glaciation. The nearest presently glacierised mountains are the Sierra Nevada del Cocuy, with a modern snowline above 4700 m a.s.l (Helmens *et al.*, 1997). This study focuses on three sub-regions surrounding Bogotá where the Late Pleistocene glacial sequence was mapped in detail by Helmens (1988): the Páramo de Palacio, the Páramo de Sumapaz, and the Páramo de Peña Negra (Fig. 1). Present-day climate conditions in the Northern High Andes are reflected in the name; *páramo* refers to the altitudinal belt between the tree line and permanent snow, as well as to the vegetation that occurs in it (Rull, 1998). Climate in the *páramos* is generally cold and humid, with a mean annual temperature in the nearby Venezuelan highlands ranging from zero to 10 °C (Salgado-Labouriau, 1979). Over the Sabana de Bogotá, situated at an altitude of about 2600 m a.s.l. in the upper part of the Andean forest belt, mean annual temperatures are about 13–15 °C, with an estimated lapse rate of 0.66 °C per 100 m (van der Hammen and Gonzalez, 1963). As is typical of tropical highlands, the temperature shows little seasonal but large diurnal variation, and is mainly controlled by altitude and cloudiness (Schubert and Clapperton, 1990).

The primary feature of regional climatic variability in this isothermal tropical highland setting is the strong seasonality in precipitation, which is locally modified by topography. The tropical region features two annual precipitation maxima, March–June and September–November, associated with the solar zenith and meridionally migrating equatorial trough. Between these high mountains, the Sabana de Bogotá forms the upper drainage of the Río Bogotá that drains to the SW. The general NE–SW trend of the Andes in this region directs the outward drainage from the mountains to the SE, down the steep 'Llanos slope' on the side of the Llanos Orientales, and to the NW, down the 'Magdalena slope' towards the inter-Andean Río Magdalena (Fig. 1). The principal source of atmospheric moisture to the region is the Pacific Ocean to the west, except for the eastern slopes. The easterly trade winds enhance the moisture flux from the east, resulting in only one annual dry season, and feature a higher peak in strength in July–August (Helmens, 1988). Total annual precipitation is strongly controlled by local topography, with maxima of over 2000 mm to ca. 1400 mm along the Llanos and Magdalena slopes, respectively. A rainshadow effect causes a decrease in precipitation totals elsewhere, down to annual totals of 600–900 mm on the Sabana proper and in its adjacent valleys. Interannual variability in the hydroclimatology of the region is most significantly impacted by ENSO (Poveda *et al.*, 2001).

Glacial geology

Glacial landforms in the high mountain ranges of the Páramo de Guerrero, the Páramo de Peña Negra, the westernmost part of the Páramo de Palacio and the northern part of the Páramo de Sumapaz have been mapped in detail by Helmens (1988) and reviewed in the context of the Quaternary glacial record of the Colombian Andes (Helmens, 2004). The glacial geological mapping focused on the mountain slopes that are drained towards the high plain of Bogotá, because it formed part of a project on the Pliocene–Quaternary stratigraphy, palaeoenvironments and landscape evolution of the Bogotá basin and surrounding mountains (Helmens, 1990). For comparison studies, Helmens (1988) also mapped glacial landforms in one of the valleys of the Páramo de Palacio along the outer, eastern slope of the Cordillera, and in the northern valleys of the Páramo de Sumapaz that are drained directly towards the inter-Andean Magdalena River. Here we summarise the principal moraine formations and ages. To simplify notation, all radiocarbon dates in this paper are in uncalibrated years before present, where ka = ^{14}C yr BP.

Based on general morphology and degree of modification by erosion and denudation, the maps by Helmens (1988) differentiate four morainic complexes, defined from oldest to

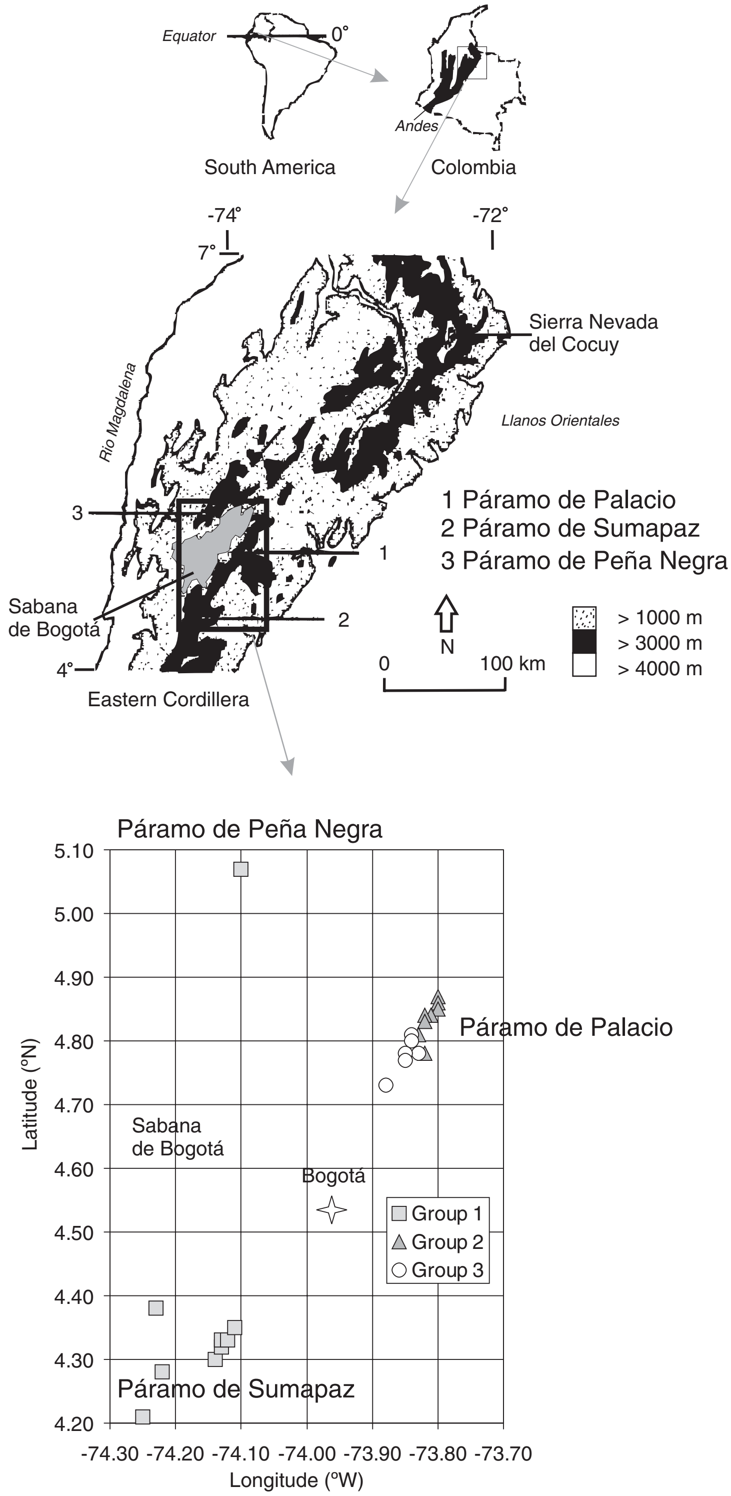

Figure 1 Map showing the Sabana de Bogotá in the Eastern Cordillera of Colombia with locations of the three Páramo regions indicated, with the lower panel showing the groupings of the reconstructed palaeoglaciers

youngest as morainic complexes 1 to 4. Yet sites with 'older glacial deposits', which no longer display a morainic topography, have been additionally encountered beyond the sequence of moraines (van der Hammen *et al.*, 1980; Helmens, 1990). Therefore, radiocarbon dates of organic-rich sediments and palaeosols associated with the glacial moraine complexes, in addition to older evidence provided by the radiocarbon-dated palaeosol sequence in the Bogotá area, place glacial events between probably 43 and 38 ka, 36 to 31 ka (complex 1), 23.5 to 19.5 ka (complex 2), 18.0 to 15.5 ka (complex 3), and 13.5 to 12.5 ka (complex 4) (Helmens, 1988, 1990; Helmens and Kuhry, 1995; Helmens *et al.*, 1997). The study of the Bogotá morainic sequence is characterised by its great detail, and provides one of the best-dated Late Pleistocene glacial records for the Andes (Clapperton, 2000).

Maximum glaciation in the Bogotá mountains pre-dated the Marine Isotope Stage 2 (MIS 2) glacial maximum (LGM). One (possibly two) major glacier advance(s) during MIS 3 are recognised, the moraines of which are only very locally preserved (complex 1), occurring at elevations some 200 m below the LGM moraines. Palaeosols found stratigraphically bounding moraines and reworked till suggest these glacier fluctuations occurred between 43 and 38 ka and between 36 and 31 ka (Helmens *et al.*, 1997). Major glacier advances during MIS 3 seem to have been caused by cool and humid climatic conditions. The LGM consisted of two glacial maxima just before 19.5 and 15.5 ka, separated by an interval of glacier retreat. The moraines of complex 2 (early stade of LGM) partially enclose the moraines of complex 3 but reach ca. 100–150 m farther down-valley. Morainic complex 2, however, has been distinctly more affected by erosional and denudational processes, is generally incomplete, and displays more subdued ridges than the sharp crests of morainic complex 3. The youngest of the two LGM morainic complexes, complex 3 (late stade of LGM), shows the most impressive morainic morphology of the different morainic complexes recognised in the Bogotá mountains. The arcuate, multiple ridge system rises tens of metres above the valley floors and the related maximum ice extent can be continuously traced throughout the mountain ranges studied. The Bogotá mountains were deglaciated at ca. 12.5 ka following a Lateglacial advance of cirque glaciers that left a distinct system of winding morainic ridges standing a few metres high (complex 4) (Helmens, 2004).

Two radiocarbon dates provide minimum ages for the retreat of glaciers from morainic complex 3, i.e. $15\,510 \pm 190$ ^{14}C yr BP (site Colorado 5 in Helmens *et al.*, 1997) and $14\,660 \pm 280$ ^{14}C yr BP (Boca Grande 3; Helmens, 1988) obtained from basal lake sediments enclosed by the moraines. Minimum ages for complex 2 of $19\,190 \pm 120$ and $18\,130 \pm 170$ ^{14}C yr BP, obtained from organic-rich sediments found overlying glacio-fluvial gravel directly behind the moraines of complex 2 (Peña Negra 6; Helmens, 1988), are in accordance with a radiocarbon date of $19\,370 \pm 230$ ^{14}C yr BP for a Palaeosol found on top of the complex (section 9; Khobzi, pers. comm., in Helmens, 1988). The maximum age for morainic complex 2 probably is not older than 23.5 ka (Helmens and Kuhry, 1995). This age corresponds to a period of widespread formation of organic-rich soils in the Bogotá area; the period is not represented in the dated Palaeosol sequence covering the morainic complex, nor is it expressed in the sediment sequence behind the moraines.

Independent chronologies for the glacial and palynological records of the eastern Cordillera suggest a close match between stadials characterised by low upper Andean forest limits and glacier advances in the surrounding high-mountain ranges (Helmens and Kuhry, 1995). A pollen record obtained from a series of lake sediments collected from an ecologically sensitive location on the western slopes of the eastern Cordillera has revealed that cold glacial conditions did not persist throughout the LGM (ca. 21–14 ka), but instead were interrupted by a distinct interval of climate warming around ca. 18 ka (Helmens *et al.*, 1996). The La Laguna pollen record indicates a lowering of the forest limit by 1100 to 900 m beneath its present altitude at ca. 3300 m a.s.l. (mean annual 9 °C isotherm) for the stadial intervals directly preceding and following the 18 ka interval, implying a drop in mean annual temperatures of 8 to 6 °C using a modern thermal lapse rate of 0.66 °C per 100 m (van der Hammen and Gonzalez, 1963; Kuhry, 1988). Glaciers advanced down-valley during the stadial periods to altitudes of ca. 3350 (morainic complex 2) to 3500 m a.s.l. (complex 3), reflecting a lowering of the ice-front by ca. 1200 to 1100 m respectively, compared to present values. Between 19.5 and 17 ka, the ice front retreated, extensive soil formation took place, and the upper Andean forest limit shifted to altitudes similar to that of the La Laguna site (2900 m). At the same time temperatures rose considerably to values only 3 to 4 °C lower than those in the present interglacial period. The interval between 19.5 and 17 ka has been defined as the La Laguna Interstadial (Helmens *et al.*, 1996). Following Kuhry (1988), the preceding stadial period (21 to 19.5 ka) and the following stadial period (17 to 14 ka) are termed the Early Fúquene Stadial and the Late Fúquene Stadial, respectively.

Methods

Given the high-quality geomorphologic mapping of moraines (Helmens, 1988), and data on the valley topography, we were able to estimate ELAs for individual palaeoglaciers in the region using the area–altitude balance ratio (AABR) method (Osmaston, 2005). This method is considered more rigorous because it accommodates divergences from the standard glacier shape and also accounts for non-linear mass balance gradients (Kaser and Osmaston, 2002). Rather than depending on the simple altitude ratio between head wall and terminus (i.e. toe-to-headwall-altitude ratio, or THAR), or mean morphological ratio of accumulation and ablation areas (i.e. accumulation area ratio, or AAR), the AABR method makes use of the palaeoglacier hypsometry and weights the area of surface altitude contour bands by relative altitude above or below the ELA.

To begin, palaeoglacier forms and ice surfaces were reconstructed using digital elevation data in a geographical information system (ArcGIS™). We digitised the original field maps of glacial moraines for moraine complex 2 (early stade of LGM) made by Helmens (1988), drawn on national 1:25 000 and 1:50 000 topographic maps of the IGAC with 25 to 50 m contour intervals. These base maps were geo-referenced and projected to the WGS-84 datum, to be compatible with available 3-arcsecond (90 m) Shuttle Radar Topographic Mission (SRTM) digital elevation data. The 100 m contours were both hand-digitised from the base maps and generated from the SRTM data using ArcGIS. Based on the digitised moraine positions from the field maps overlaid on the contour map, the extent and ice surface elevation were reconstructed digitally for 23 palaeoglaciers according to conventional considerations (e.g. Sissons, 1974; Seltzer, 1992). The hypsometry, or vertical distribution of surface area, of each palaeoglacier was computed for 100 m elevation bands and used in the AABR calculation.

We utilised the iterative routine outlined by Osmaston (2005) to compute palaeo-ELAs with the AABR method for different balance ratios (BRs). The balance ratio is the ratio of mass balance gradients below and above the ELA. For $BR = 1$, there is a linear gradient of mass balance assumed throughout the glacier, a condition that is not often observed on modern tropical glaciers. Tropical mass balance curves vary in detail, but in general display steeper gradients below the ELA than above it, resulting in $BR > 1$ (Kaser and Osmaston, 2002). When the mass balance gradient below the ELA is two times greater than that above the ELA, the $BR = 2$ and the ELA is shifted lower than in the linear case with $BR = 1$. In this study, we considered BRs over a range of values considered for tropical glaciers in other publications: $BR = 1–7$, and $BR = 25$ (following: Lachniet and Vazquez-Selem, 2005).

To explore the nature of intra-regional ELA variability, we considered different groupings of glaciers and computed both mean ELA and standard deviation for the range of BRs. Initially, the glaciers were all considered together in one group. Then, the glaciers were classified *a priori* into three groups by their regional location (by *páramo*), which is assumed to have a first-order control over precipitation. The BR selected to compute the palaeo-ELA for each group was the one that minimised the variation of ELA (standard deviation, or SD) within the respective groups (following Osmaston (2005)). We computed ΔELA as the change in ELA from LGM to modern ($\Delta ELA = ELA_{mod} - ELA_{LGM}$), using a modern ELA of 4800 m for all palaeoglacier localities as a conservative estimate based on the published observations. This necessarily introduces a small error for the valley-specific ΔELA since there is likely to be a gradient in modern ELA as a result of the precipitation gradient. The position of the modern ELA in the Colombian Andes corresponds closely to the 0 °C annual isotherm (Kuhn, 1981). Despite considerable differences in topographic setting, from flat- to steep-topped volcanoes in the central Cordillera and to an alpine topography in the eastern Cordillera, present glacier extent on these mountains is very similar, with differences in precipitation amounting to only a few hundred metres of variation in present ELA (Hoyos-Patiño, 1998).

Results

Palaeoglacier ELA reconstructions

Twenty-three individual palaeoglaciers were reconstructed on the basis of the geomorphologic field data gathered and mapped for moraine complex 2 (early stade of LGM) in the high plain of Bogotá region (Fig. 2). The mapped palaeoglaciers are

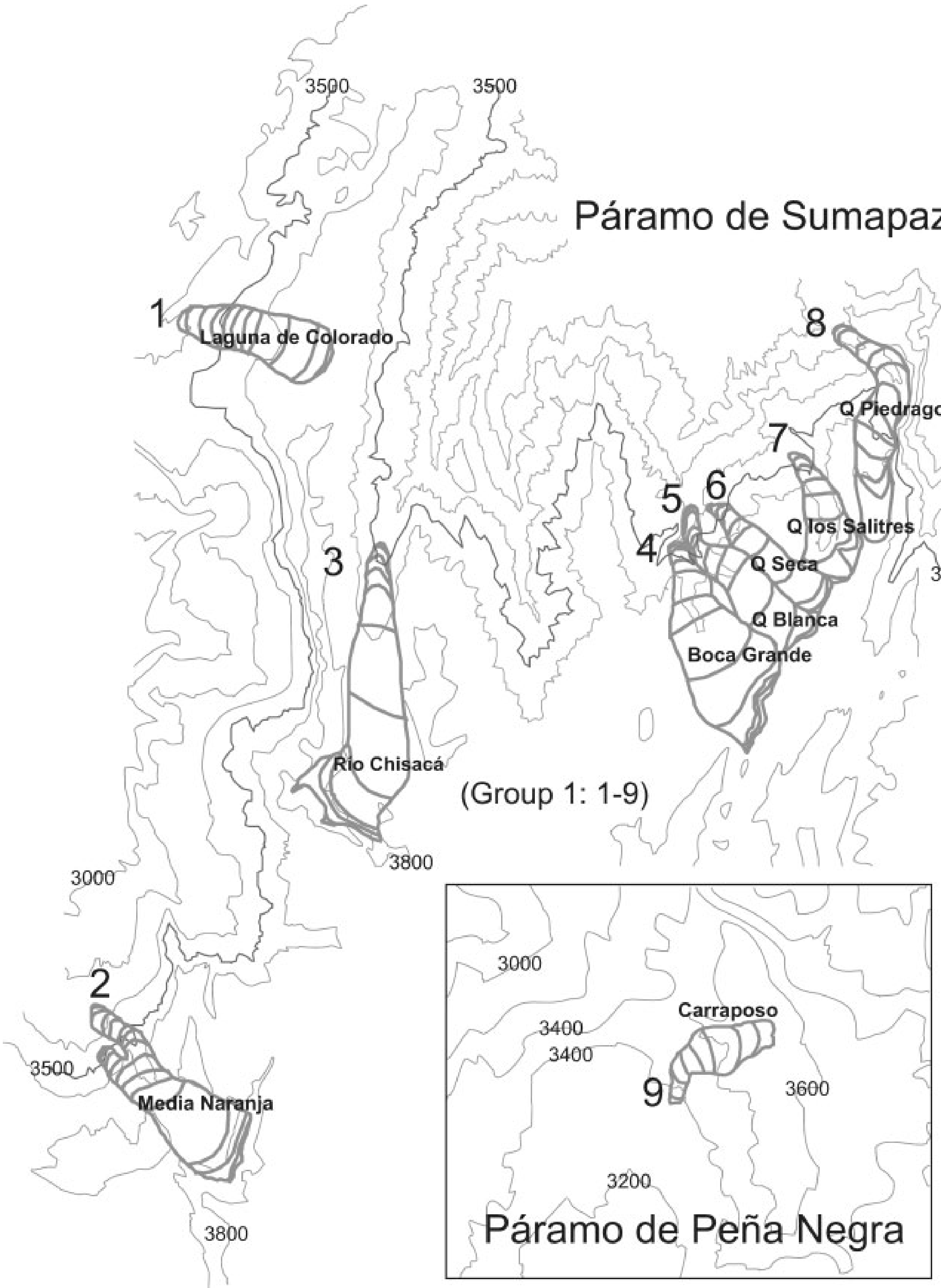

Figure 2 Contour maps showing palaeoglacier surface reconstructions for moraine complex 2 (LGM) glaciers numbered in Table 1. Palaeoglacier contour interval is 50 m, and background map contour interval is 100 m

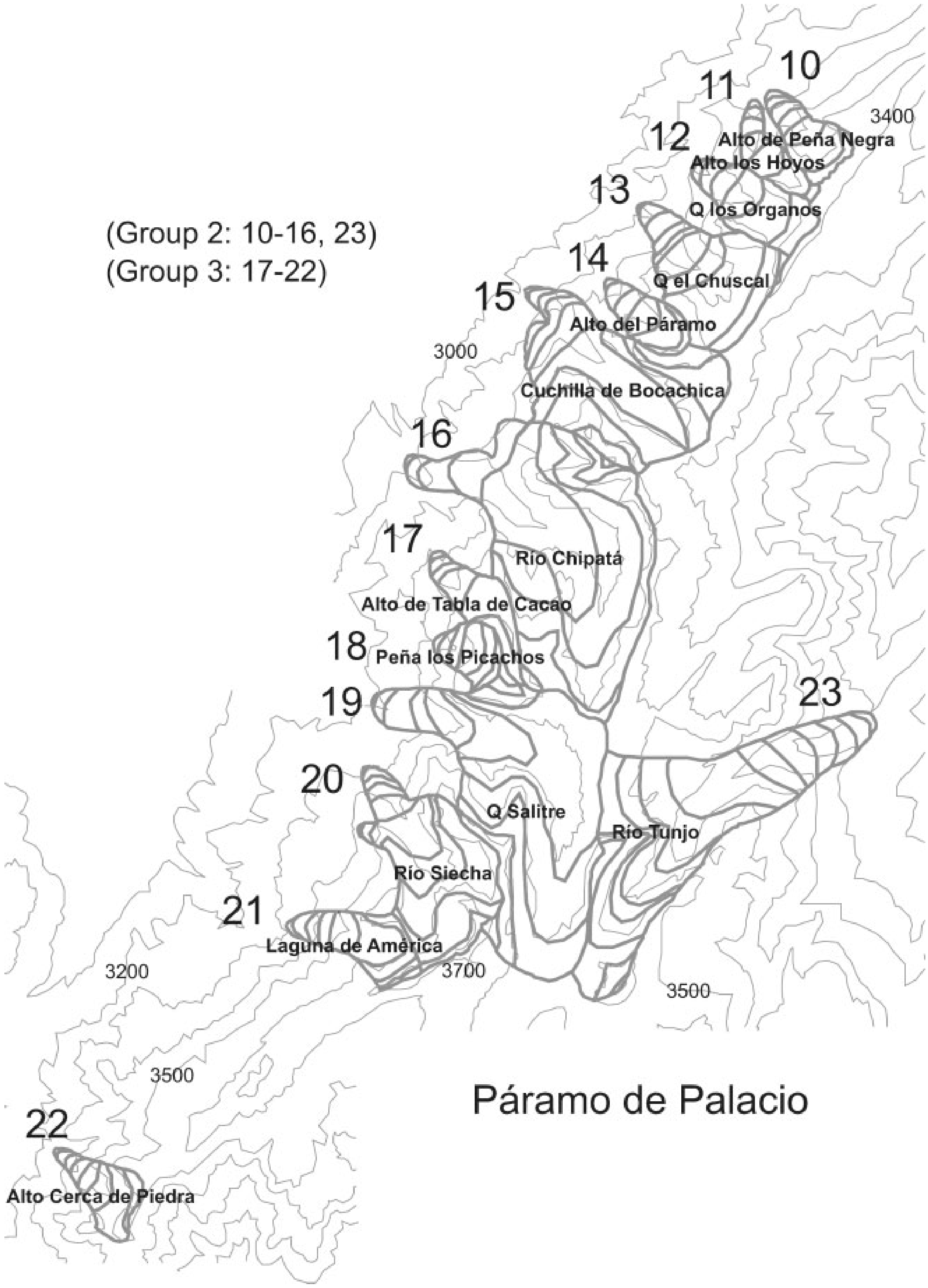

Figure 2 Continued

numbered, and the corresponding dimensions are summarised in Table 1. The glaciers had an average surface area of 3 km^2, and the majority faced towards the NW.

The ELAs for each palaeoglacier were reconstructed using the AABR method, using the same BR for all glaciers. For linear vertical mass balance gradients (BR = 1) the mean ELA = 3526 m, SD = 154 m. However, the variation in ELA is minimised with a BR = 25 (mean ELA = 3399 m, SD = 142 m).

To further refine the analysis of ELA variation, the palaeoglaciers were divided into three different groups according to modern precipitation regimes indicated by the W–E gradient in terminus altitude (Fig. 3(a)). Group 1 consists of the glaciers on the Páramo de Sumapaz and Páramo de Peña Negra. Group 2 glaciers are those along the most northerly ridge of the Páramo de Palacio, with lowest headwalls $\leq$ 3500 m. Group 3 includes the remainder of the glaciers in the Páramo de Palacio that had higher headwalls and more likely experienced a greater rain-shadow effect. Within each group, ELA variance was minimised with different BRs. The ELA results that minimised the group variances are included in Table 1. The general trend is for higher BRs towards the east, following the spatial gradient in the extent of glacier expansion to moraine complex 2 and associated LGM ELA depression. The 23 glaciers show a strong W–E gradient in ELA, with lower ELA to the east (Fig. 3(b)).

Table 1 LGM palaeoglacier data for the Sabana de Bogotá region

LGM palaeoglacier name	Map no.	Páramo	Long.	Lat.	Aspect	Group	Highest summit altitude (m)	Headwall altitude (m)	Terminus altitude (m)	LGM ELA estimate (m)	BR	ΔELA	Total area (km)	Ablation (%)
Laguna de Colorado	1	Sumapaz	−74.23	4.38	W	1	3829	3850	3350	3671	1	1129	3.75	0.37
Media Naranja	2	Sumapaz	−74.25	4.21	NW	1	3990	3925	3275	3743	1	1057	5.64	0.32
Río Chicasá	3	Sumapaz	−74.22	4.28	N	1	3990	3975	3450	3765	1	1035	7.93	0.52
Boca Grande	4	Sumapaz	−74.14	4.30	NNW	1	3900	3850	3400	3711	1	1089	6.56	0.3
Q. Blanca	5	Sumapaz	−74.13	4.32	NW	1	3900	3850	3325	3699	1	1101	3.44	0.34
Q. Seca	6	Sumapaz	−74.13	4.33	NW	1	3900	3850	3400	3696	1	1104	2.3	0.45
Q. Los Salitres	7	Sumapaz	−74.12	4.33	NW	1	3900	3825	3475	3703	1	1097	2.61	0.4
Q. Piedragorda	8	Sumapaz	−74.11	4.35	NW	1	3900	3700	3225	3593	1	1207	3.21	0.54
Carrasposo	9	Peña Negra	−74.10	5.07	SW	1	3750	3725	3375	3596	1	1204	0.94	0.41
Alto de Peña Negra	10	Palacio	−73.80	4.87	NW	2	3500	3450	3150	3306	7	1494	0.76	0.18
Alto los Hoyos	11	Palacio	−73.80	4.86	NNW	2	3500	3450	3150	3295	7	1505	0.58	0.19
Q. de los Organos	12	Palacio	−73.80	4.85	NW	2	3500	3450	3150	3323	7	1477	1.2	0.16
Q. el Chuscal	13	Palacio	−73.81	4.84	NW	2	3500	3450	3075	3279	7	1521	2.23	0.13
Alto del Paramo	14	Palacio	−73.82	4.84	NW	2	3450	3425	3150	3275	7	1525	0.68	0.2
Cuchilla de Bocachica	15	Palacio	−73.82	4.83	NW	2	3500	3450	3025	3247	7	1553	4.17	0.16
Río Chipatá	16	Palacio	−73.83	4.81	NNE	2	3500	3500	3050	3267	7	1533	7.25	0.18
Río Tunjo	23	Palacio	−73.82	4.78	NE	2	3725	3700	3000	3282	7	1518	5.68	0.17
Alto de la Tabla de Cacao	17	Palacio	−73.84	4.81	NW	3	3650	3600	3275	3429	5	1371	0.86	0.15
Peña los Picachos	18	Palacio	−73.84	4.80	WNW	3	3650	3600	3350	3475	5	1325	0.56	0.2
Q. Salitre	19	Palacio	−73.83	4.78	NNE	3	3700	3700	3175	3439	5	1361	6.73	0.17
Río Siecha	20	Palacio	−73.85	4.78	NW	3	3775	3750	3275	3529	5	1271	2.95	0.22
Laguna de América	21	Palacio	−73.85	4.77	WNW	3	3775	3750	3300	3492	5	1308	1.05	0.2
Alto Cerca de Piedra	22	Palacio	−73.88	4.73	NW	3	3650	3600	3225	3415	5	1385	0.75	0.19

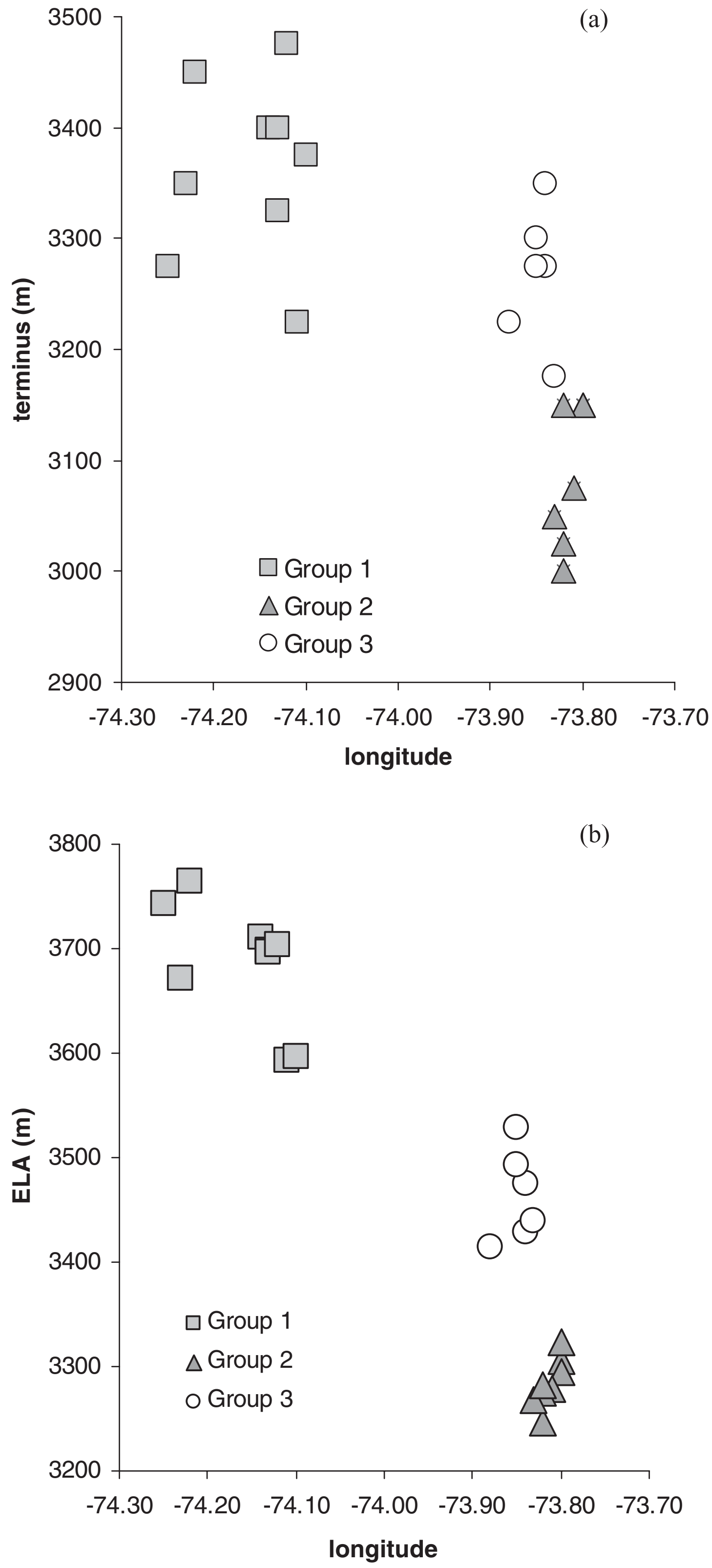

Figure 3 (a) Moraine terminus altitude (m) plotted by longitude for all LGM palaeoglaciers, separated by group. (b) Reconstructed ELA (m) as estimated using the AABR method with BR as listed in Table 1, plotted by longitude for all LGM palaeoglaciers

Intra-regional variability of ELA

There is a considerable amount of intra-regional variance in the ELA, with values ranging over 500 m from a minimum of 3247 m to a maximum of 3765 m. Most of the variance in ELA (87%) can be explained by the altitude of the headwall (Fig. 4). There is a significant correlation ($P < 0.0001$), such that higher headwall altitudes are associated with higher ELAs. To explore the relationship between ELA and catchment area, the relative accumulation area for the palaeoglaciers was computed at the ELA from the hypsometric curve. Accumulation area displays a significant negative correlation ($P < 0.0001$) with both headwall altitude (Fig. 5) and ELA (Fig. 6).

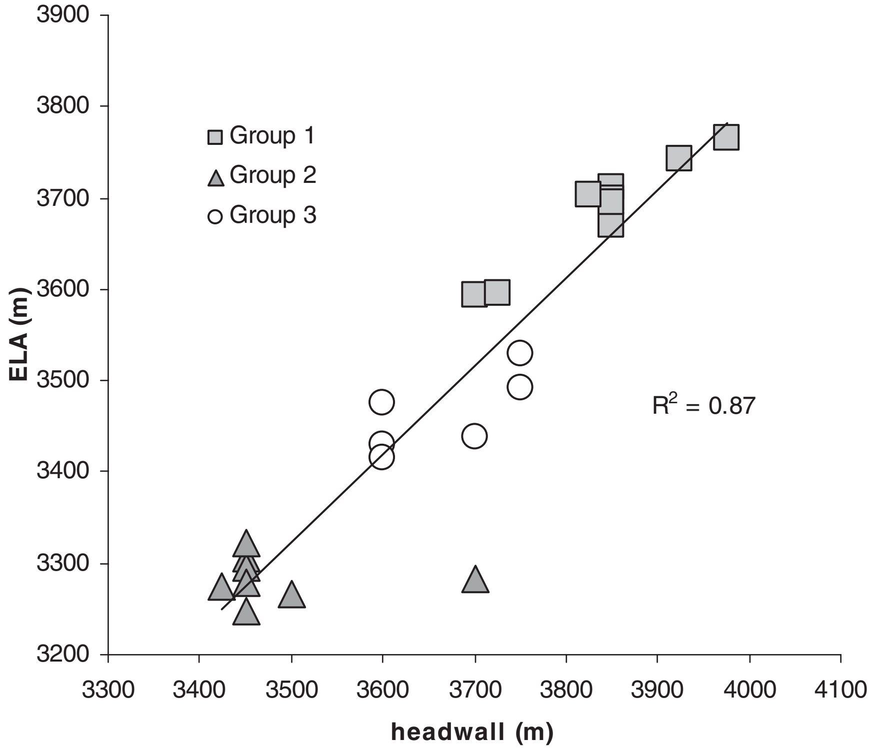

Figure 4 Relationship between headwall altitude (m) and ΔELA (m) for all the LGM palaeoglaciers, coded by group

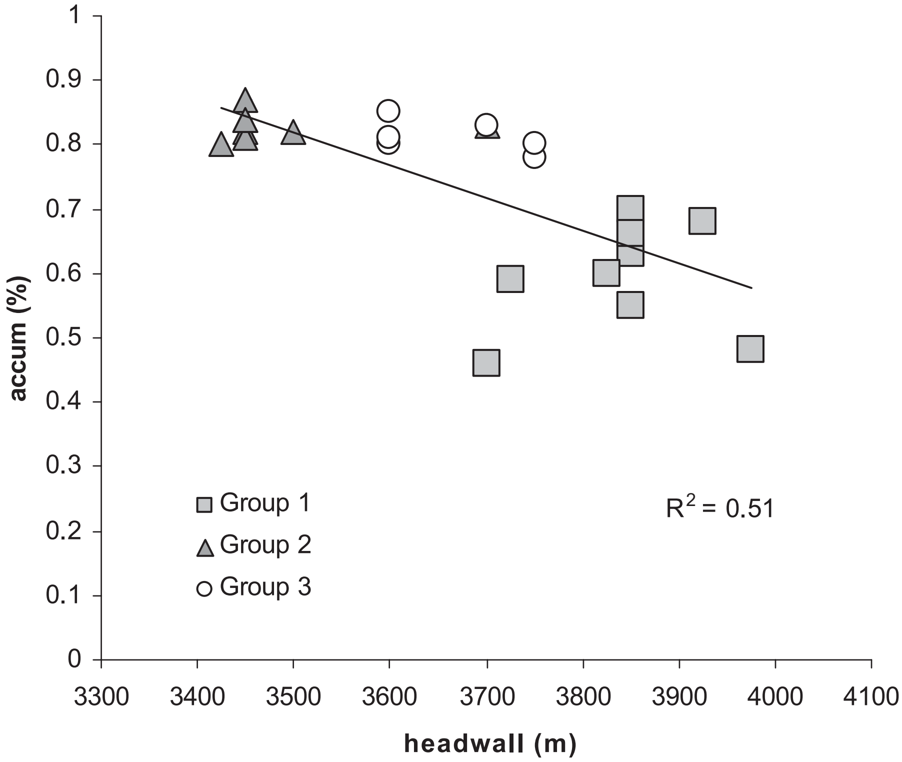

Figure 5 Relationship between headwall altitude (m) and accumulation area (% glacier area) for all the LGM palaeoglaciers, coded by group

Discussion

There are many methods to estimate the ELA of a palaeoglacier, and these have been more thoroughly reviewed elsewhere (i.e. Porter, 2001). The AABR has been noted to be a much more rigorous and precise method of estimating palaeo-ELAs than simple ratio methods such as AAR and THAR (Kaser and Osmaston, 2002). By incorporating the full glacier hypsometry, the AABR method accounts for the area distribution of glacier mass relative to the ELA. The premise is that glacier areas further above or below the ELA should have greater influence on the overall mass balance. This allows non-standard glacier shapes to be accommodated (i.e. Furbish and Andrews, 1984),

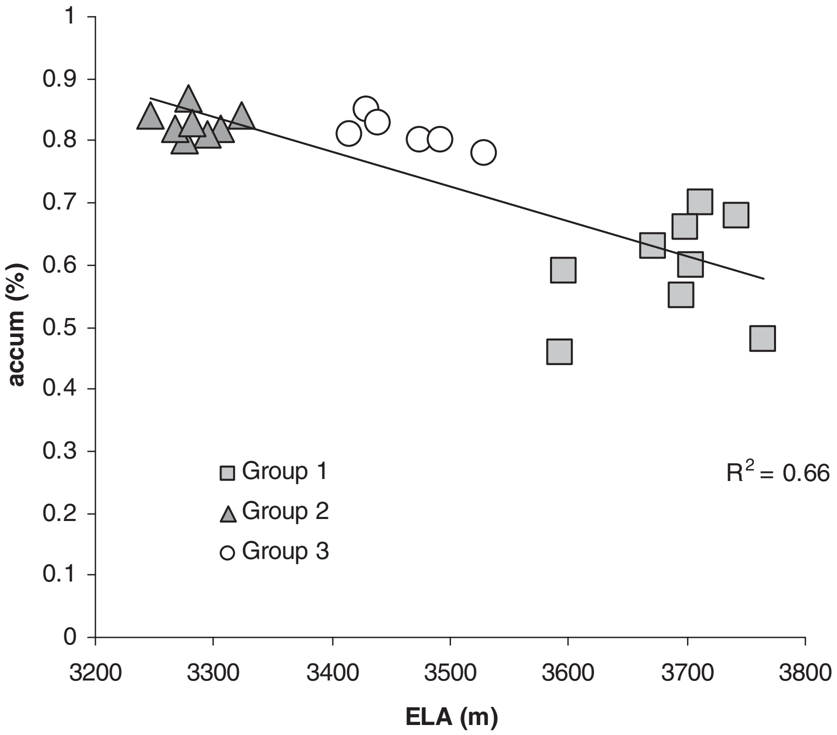

Figure 6 Relationship between ΔELA (m) and accumulation area (% glacier area) for all the LGM palaeoglaciers, coded by group

as well as non-linear vertical mass balance gradients. Simple mean morphological altitude ratios of accumulation and ablation areas treat all areas above and below the ELA equally, a simplification that does not account for the vertical mass balance variation inherent to tropical glaciers (Kaser and Osmaston, 2002). Moreover, the parameters adopted with THAR or AAR methods usually involve assuming a standard alpine glacier form. Tropical glaciers show a stronger vertical mass balance gradient than those in mid-high latitudes (i.e. Kaser, 2001), and both the topography and geomorphologic field evidence suggests that glaciers in the Bogotá region took on non-standard forms. The glaciers are best described as poorly developed valley glaciers or ice caps, rather than distinct U-shaped valley forms.

Selecting the appropriate BR to apply the AABR method requires making assumptions about the palaeoglacier equilibrium mass balance regime for the relevant moraine position. Where no present-day glaciers with mass balance data exist in the same or similar climate, this requires making an implicit inference about the former climate and glacier response. However, local topography and glacier form strongly moderate the climatic sensitivity of glaciers and effectively require each glacier to have a unique BR. Moreover, local topo-climatic conditions of low-latitude glaciers present particular challenges for reconstructing and interpreting former ELAs (Benn *et al.*, 2005). Since precise former climatic environmental information is not available, an effective means of selecting BR is based on the principle that homogeneous groups of glaciers will respond to climate in a similar manner. Thus minimising the statistical variance in ELA estimate for each group provides a self-validating means of assigning BR (Osmaston, 2005). In this case, the *a priori* grouping of palaeoglaciers was affirmed by evidence of precipitation gradients and palaeoglacier form.

The groups of palaeoglaciers defined in this analysis for morainic complex 2 imply that precipitation gradients strongly influenced intra-regional ELA response in a similar manner as seen in the better-preserved moraine complex 3. Glaciers on, and in close proximity to, the easternmost flank of the region were under the influence of moist Llanos winds, and accumulated more mass that subsequently flowed further downslope. The distribution of the most complete moraine complex 3 was used in Helmens (1988) to show that limits of former glacier systems were strongly influenced by local climatological conditions (Helmens, 1988). It appears that glaciers reached farther down-valley in areas with high orographic precipitation, most significantly on the wet, eastern slopes of the Cordillera, where morainic complex 3 was found at the low altitude of 3100 m. In the rainshadow created by large topographic barriers, complex 3 occurs at altitudes as high as 3750 m. Moraine complex 2 terminated as low as 3000 m on the east-facing slope of the Páramo de Palacio along the 'Llanos slope' (palaeoglacier 23), and as high as 3475 m on the northwest-facing, inter-montane slopes of the Palacio de Sumapaz (palaeoglacier 7).

The BRs greater than unity for Groups 2 and 3 also substantiate the strong climate gradients over the region. This signifies that the palaeoglaciers experienced steeper mass balance gradients in the ablation zones, a form that typifies tropical glaciers (Kaser, 2001). The increased ablation on tropical glacier tongues results in a theoretical increase in the equilibrium accumulation area ratio (AAR) for tropical glaciers to a value on the order of 0.82 (Kaser and Osmaston, 2002, section 5.4). This is seen in the smaller ablation areas of groups 2 and 3.

The glacial geomorphology indicates that the palaeoglaciers attained forms that would accommodate broad accumulation areas. Glaciers were flat, and not formed in distinct U-shaped valleys. In several locations, glaciers overtopped adjacent valley walls and spread out laterally on surrounding slopes, and most likely coalesced into an ice cap form. Likewise, the morphology featured broader accumulation areas, with more area in the vertical distance from ELA to headwall than from ELA to terminus.

Evidence of predominant regional gradients notwithstanding, the overall average lowering in ELA from LGM to modern calculated at ca. 1300 m can only be explained by a considerable lowering in temperature. Our calculated average lowering in ELA for the Bogotá glaciers is of similar magnitude to the lowering in forest limit by 1100 m during the early stade of the LGM as inferred from palynological data in the region (Helmens *et al.*, 1996), and implies a drop in mean annual temperature of ca. 8 °C. Similar magnitude temperature changes (6–8 °C) were estimated for the glacier locations based on modern climate data (Mark *et al.*, 2005). This provides multiproxy evidence for significant climate cooling in the LGM in the northern tropical high Andes.

The LGM glaciers of the Bogotá region were relatively small, ranging from 0.5 to 8 km^2, and likewise were strongly asymmetrical, with most preferring the NW aspect. This asymmetric pattern of glaciation conforms to other observations in Colombia. It was noted that 48 glaciers at 11 °N in the Sierra Nevada de Santa Marta, Colombia, have marked asymmetry, with the mean aspect to the NW (Evans, 1977). The asymmetry was proposed to reflect a shading effect on northern slopes, combined with the afternoon convective cloudiness over the west. Furthermore, the glaciers of the more easterly Páramo de Palacio were smaller, had lower headwall altitudes and featured stronger asymmetry than those of the Sumapaz. This pattern conforms to the hypothesised climatic gradient featuring enhanced cloudiness and precipitation in the east that allowed glaciers to form with headwalls up to 500 m lower. Such a differential degree of asymmetry with size was noted by Evans (1977) to occur in many cirque and glaciers globally, inspiring the formulation of a 'law of decreasing asymmetry with increasing glacier cover'. Peaks that were just high enough to be glaciated generally display a greater degree of glacial asymmetry than those regions that had relatively greater glaciated areas.

The strong correlation of ELA and headwall altitude is a robust feature of the dataset. It persists with all BRs, and whether a constant BR is applied to all the palaeoglaciers or when *a priori* groups are defined. Here, the individually reconstructed palaeoglaciers based on the digital topography permit direct comparison of accumulation areas associated with estimated ELAs. Plotting the headwall altitudes with accumulation area as a percentage of total glacier area allows a standard comparison across all palaeoglaciers (Fig. 5). There is a significant negative correlation between accumulation area and headwall, meaning that palaeoglaciers with higher headwall altitudes have relatively smaller catchments. Likewise, ELA is negatively related to accumulation area, and discriminates the *a priori* groupings of palaeoglaciers (Fig. 6). However, the headwall–catchment area relationship is contingent upon the BRs applied. When all the glaciers are given a linear balance gradient BR = 1, the trend is reversed, and a significantly positive correlation is displayed between accumulation area and headwall altitude. Conversely, when the uniformly large BR = 25 is applied to all the glaciers, there is no trend between headwall or ΔELA and accumulation area. Thus, without a means of defining subgroups with different BRs, based on climate gradients, it is not possible to exclude the possibility that the negative correlation between accumulation area and headwall is simply a function of the ELA estimation method.

Other recent work on the hypsometry of glaciated regions lends support to the idea of a topographic limitation to glacier–climate sensitivity. At a mountain range scale, climate has been shown to be one of the primary influences on the form of hypsometric curves (Montgomery, 2004). A close correlation of hypsometry and extent of glaciation in the eastern Himalaya was interpreted as an indication of the fundamental climate control over topographic development via glaciation (Brozovic *et al.*, 1997; Bishop and Schroder, 2000). Other work has shown that while hypsometry can reflect different degrees of glacierisation, unique local circumstances can have profound effects on hypsometry such that similar hypsometric curves can exist for basins with notably different hypsometry (Brocklehurst and Whipple, 2004). Moreover, a close examination of the sensitivity of glacier ELA to local topography across climate gradients in the Rocky Mountains showed that upslope zone morphology such as headwall height has a very strong influence on ELA (Allen, 1998).

Finally, recent chronological work in other parts of the Andes has shown how complexities in local topography can play a predominant role in defining the extent of glaciation. Cosmogenic dating of moraines around Lago Junin in Peru and in the Cordillera Real, Bolivia has revealed a close clustering of similar ages, but for palaeoglaciers of much different sizes (Smith *et al.*, 2005). This was one of the last contributions by Geoff Seltzer and colleagues. It fittingly highlights the importance of focusing on individual valley glacier responses, and advises against generalising regional ELA estimations for interpreting palaeoclimate in the complex Andean topography.

Conclusions

Well-mapped moraines and available contour maps facilitate the reconstruction of 23 individual LGM palaeoglaciers in the Bogotá region of Colombia, and permit an analysis of the climatic and topographic context of glaciation. The drop in ELA of on average ca. 1300 m for the early stade of the LGM is the result of a considerable drop in temperature, probably in the order of 6–8 °C. Spatial variance results from topography and its indirect effect on precipitation, cloudiness, temperature and glacier form.

The glaciation of the Bogotá region at the LGM was characterised by asymmetry, with lower termini and resultant reconstructed ELA to the east. This was probably the result of greater precipitation along the eastern slopes, following the moisture gradients currently observed over the Eastern Cordillera. The palaeoglaciers responded to the strong climate gradients with more active mass balance gradients towards the east. This is affirmed in the statistical minimising of variance in BR by group, showing higher BRs with those glaciers having lower ELAs.

The intra-regional variation in ELA further implies that glacier form moderates the climatic sensitivity of glaciers. The significant negative correlation between palaeoglacier accumulation area and ELA implies that despite having lower headwall altitudes, glaciers with broader accumulation areas reached to lower terminus altitudes and had lower ELAs. Overall, the local topographic context is important to understanding how a glacier responds to regional climatic conditions. Likewise, digital elevation data permit individual palaeoglacier hypsometries to be calculated in order to refine ELA estimation. Careful attention to local topography is therefore required in making palaeoclimate reconstructions from the geomorphologic record, which re-sounds the note of caution articulated in one of the last papers to which Geoff Seltzer contributed (Smith *et al.*, 2005), shedding important new insight into the LGM in South America.

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