

The timing and magnitude of mountain glaciation in the tropical Andes

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ABSTRACT: The Andes of Ecuador, Peru and Bolivia host the majority of the world's tropical glaciers. In the tropical Andes, glaciers accumulate during the wet season (austral summer) and ablate year-round. Precipitation is delivered mainly by easterlies, and decreases both N–S and E–W. Chronological control for the timing of glacial advances in the tropical Andes varies. In Ecuador, six to seven advances have been identified; dating is based on radiocarbon ages. Timing of the local Last Glacial Maximum (LGM) and the existence of Younger Dryas advances remain controversial. In Peru, local variability in glaciation patterns is apparent. Surface exposure dating in the Cordillera Blanca and Junin Plain suggests that the local LGM may have been early (~30 ka), although uncertainties in age calculations remain; the local LGM was followed by a Lateglacial readvance/stillstand and preceded by larger glaciations. In contrast, preliminary data from an intervening massif indicate that the largest moraines are Lateglacial. Chronologies from Bolivia also suggest local variability. In leeward Milluni and San Francisco Valleys, local LGM moraines descend to ~4300 m above sea level (a.s.l.), whereas in windward Zongo Valley Lateglacial moraines reach ~3400 m a.s.l. Atlantic and Pacific sea surface temperatures, El Niño–Southern Oscillation and insolation changes all likely play roles in mediating tropical Andean glacial cycles. Copyright © 2008 John Wiley & Sons, Ltd.



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KEYWORDS: Andes; glaciation; mountain glacier; tropical glacier; cosmogenic radionuclides.

Introduction

The snow-capped peaks of the Andes are an icon of South America. Although Ecuador, Peru and Bolivia lie within the tropics, the high altitudes attained by Andean peaks allow glaciers to exist today and to have been far more extensive in the past. Those same high altitudes have perhaps been partly responsible for the relatively slow accumulation of data about glacial fluctuations in the tropical Andes. Two recent overviews examined glaciation in the tropical Andes from different perspectives. Mark *et al.* (2004) summarised the state of knowledge about the glacial record for Ecuador, Peru and Bolivia to provide a framework for ArcView-formatted maps of glacial limits in those countries. Smith *et al.* (2005a) presented an evaluation of snowlines in the tropical Andes at the global Last Glacial Maximum (LGM; designated therein as 21 000 years before present, or 21 ka).

This overview follows the general format, and includes many of the same sites, as Smith *et al.* (2005a), but with a different focus: the record of mountain glaciation in the tropical Andes

from Marine Isotope Stage (MIS) 5e (ca. 130–117 ka) to the Younger Dryas (YD) climate reversal (ca. 12.8–11.6 ka). We limit our discussion to sites where numerical dating has constrained the glacial chronology and highlight studies in which surface exposure dating using cosmogenic radionuclides (CRNs) has increased the detail of glacial chronologies that extend beyond the reach of radiocarbon dating.

Geographic setting

The Andes mountain chain, the longest in the world, is arguably the dominant landform of the South American continent (Clapperton, 1993). The Andes form a continuous topographic barrier along the west coast of South America, typically in the form of a double chain separated by high-altitude plateaus (Anders *et al.*, 2002). The tropical Andes extend from Colombia and Venezuela north of the Equator, through Ecuador on the Equator and Peru and Bolivia south of the Equator, to the Tropic of Capricorn (23.5° S). Because the climatic regimes differ markedly north and south of the Equator, this review will concentrate on mountain glaciation in the tropical Andes south of the Equator only, specifically in Ecuador, Peru, and Bolivia

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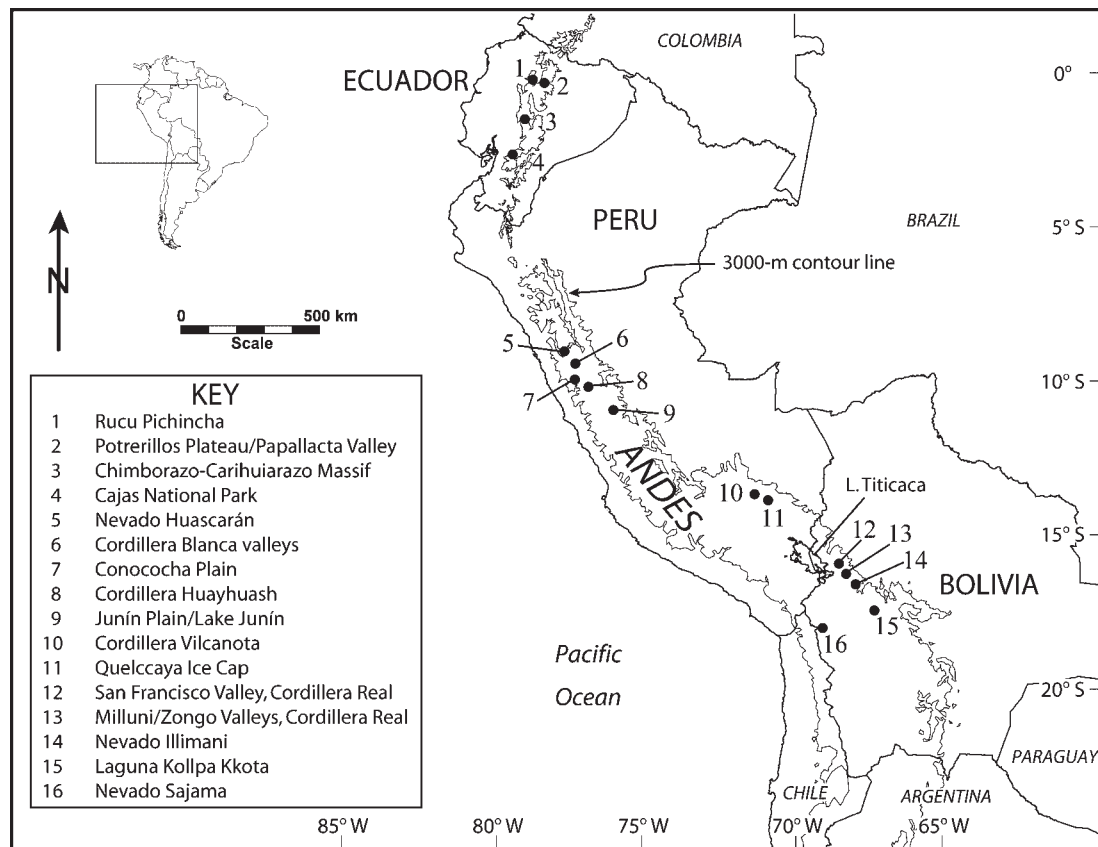


Figure 1 Site location map showing the general location of the tropical Andes, Lake Titicaca, and the sites discussed in the text. Black lines are national boundaries. The grey line is the 3000 m contour line, which encompasses the double chain of the Andes and the intervening high plateaus

(Fig. 1). In the Southern Hemisphere, the tropical Andes cover a north–south distance of more than 2500 km.

In cross-section, the Andes consist of five topographic sections (west to east): the western slope, the Western Cordillera, the high-elevation Central Andean Plateau, the Eastern Cordillera, and the Subandean Zone (Isacks, 1988; Dewey and Lamb, 1992; Gubbels *et al.*, 1993). The Central Andean Plateau, an internally drained region with a relatively constant altitude (~ 3500 – 4000 m above sea level (a.s.l.)), lies between the Western and Eastern Cordillera and stretches from about 11° S to 27° S (Isacks, 1988; Kennan, 2000). The Central Andean Plateau is widest (up to about 500 km) between about 15° S and 25° S in the Central Andes, where it is known as the Altiplano in Peru and Bolivia and the Puna in Argentina (Dewey and Lamb, 1992).

Peak altitudes in the Andes are generally lower in Ecuador than in Peru and Bolivia. Although the Ecuadorian Andes include seven volcanic peaks above 5000 m a.s.l. (e.g. Carihuairazo, Antisana) and one above 6000 m a.s.l. (Chimborazo), non-volcanic peaks typically fall in the range of 4000–4400 m a.s.l. in northern and central Ecuador, falling to 2000 m a.s.l. in southern Ecuador and northern Peru. In contrast, peaks above 6000 m a.s.l. are not uncommon in Peru and Bolivia. The central Peruvian Andes include the Cordillera Blanca and Nevado Huascarán Sur (6768 m a.s.l.), the highest peak in Peru, while the southern Peruvian Andes include the Cordillera Vilcanota, Nevado Ausangate (6372 m a.s.l.), and the Quelccaya Ice Cap. Lake Titicaca (3812 m a.s.l.) lies between the Eastern and Western Cordilleras; the Peru–Bolivia border passes through the lake. The twin chains of the Bolivian Andes are separated by the Altiplano, which reaches its widest point in central Bolivia and hosts vast salars (salt pans) and both saline and ephemeral lakes.

Ice caps on volcanoes Nevado Illimani ($16^\circ 37' S$, $67^\circ 46' W$, 6350 m a.s.l.) in the Eastern Cordillera and Nevado Sajama ($18^\circ 06' S$, $68^\circ 53' W$, 6542 m a.s.l.) in the Western Cordillera have been cored for palaeoclimate research (Thompson *et al.*, 1998; Ramirez *et al.*, 2003).

Geological setting

The Andes vary considerably in structure, composition, volcanic activity and glacial history along their 9000 km length. They have developed as oceanic lithosphere underlying the eastern Pacific Ocean has been subducted beneath continental lithosphere of the South American Plate (e.g. Jordan *et al.*, 1983). Recent models for the evolution of the Andes have generally called upon crustal thickening by structural shortening to provide the majority of the elevation (e.g. Pope and Willett, 1998). The Subandean Zone that bounds the eastern side of the Andes is an active thin-skinned fold and thrust belt (Isacks, 1988; Dewey and Lamb, 1992; Gubbels *et al.*, 1993). A narrow fault system separates the Eastern Cordillera and the Subandean Zone, which is being underthrust by the Brazilian Shield to the east (Dorbath *et al.*, 1991).

The Andes consist of a variety of sedimentary, metasedimentary, plutonic and extrusive rocks ranging in age from Precambrian to Recent (Clapperton, 1993; Kley, 1999). Active volcanism is localised into three main zones: northern ($\sim 3^\circ$ N to 2° S), central ($\sim 14^\circ$ S to 27° S) and southern ($\sim 32^\circ$ S to 42° S). Products of volcanism, though highly visible, have been estimated to make up only 0.5 km of the total thickness of the

Andes, effectively forming an extrusive veneer over the sedimentary and intrusive rocks making up the bulk of the chain (Isacks, 1988). A magmatic gap related to the angle of subduction occurs in the intervening regions. Seismic data suggest that the dip of the subducting slab beneath the South American Plate is shallow ($0\text{--}10^\circ$) below a depth of ~ 100 km in the regions without active volcanism ('flat-slab' regions) and steep ($\sim 30\text{--}45^\circ$) in regions with active volcanism (e.g. Jordan *et al.*, 1983; Kennan, 2000).

One key aspect of the geological setting of the Andes is that uplift is continually moving orogenic mass into the cryosphere. Although the details of the uplift history are complicated, the overall framework is known. Uplift of the Andes has apparently proceeded in stages. Gregory-Wodzicki (2000) estimated that in the Peruvian and Bolivian Andes the Altiplano and Eastern Cordillera were both at no more than 50% of their current altitude 10 Ma. The Altiplano and Eastern Cordillera have undergone $\sim 2000\text{--}3500$ m of surface uplift since 10 Ma (Gregory-Wodzicki, 2000).

In central Peru, uplift and erosional exhumation in the Cordillera Blanca and Cordillera Huayhuash are rapid and ongoing, which is one reason that the peaks are so high and glaciation is extensive. Both cordilleras are part of the Western Cordillera (Cordillera Occidental) of the Andes. In the Cordillera Blanca, uplift has occurred along the Cordillera Blanca Normal Fault (CBNF), which runs along the western side of the range (Garver *et al.*, 2005). Apatite fission-track dating in the Cordillera Blanca ($\sim 9^\circ 30' \text{ S}$, $\sim 77^\circ 30' \text{ W}$) showed that slip on the CBNF has averaged about 2 mm a^{-1} over the past 1.8 Ma, at an overall exhumation rate of approximately 1.1 km Ma^{-1} (Montario, 2001). Garver *et al.* (2005) estimated that ~ 5 km of unroofing has occurred in the Cordillera Huayhuash ($\sim 10^\circ 16' \text{ S}$, $76^\circ 54' \text{ W}$) since 5–6 Ma, suggesting that the central Peruvian Andes have attained much of their present elevation in that time.

The Cordillera Real in the Eastern Cordillera of the Bolivian Andes has been the focus of recent studies on erosion and exhumation rates (Safran *et al.*, 2005, 2006). Safran *et al.* (2005) used CRN concentrations in sediments to calculate basin-averaged erosion rates in the Upper Beni River region, which is bounded on its western edge by the Eastern Cordillera and includes the Zongo Valley. Erosion rates were highest ($0.2\text{--}1.35 \text{ mm a}^{-1}$) toward the crest of the Eastern Cordillera where channels were steepest, leading Safran *et al.* (2005) to conclude that tectonically induced variations in channel steepness appear to be the main control on basin-averaged erosion rates, with a secondary effect from lithological differences and no appreciable influence from climate. Safran *et al.* (2006) used apatite fission-track dating to estimate exhumation rates in the Cordillera Real of $\sim 0.2\text{--}0.6 \text{ mm a}^{-1}$ for the entire dataset and $0.18\text{--}0.48 \text{ mm a}^{-1}$ for the Zongo Valley. Safran *et al.* (2006) estimated two- to threefold variations in exhumation rates within the study area, leading them to caution against assuming spatial uniformity in exhumation rates or patterns even over distances as short as tens of kilometres.

Climatic setting

Climate of the tropical Andes

The main source of precipitation for the tropical Andes of the Southern Hemisphere lies to the east in the Atlantic Ocean and the Amazon Basin, and the primary transport mechanism is seasonal easterly winds (Johnson, 1976; Vuille and Keimig, 2004). Precipitation is concentrated in one (Peru and Bolivia) or two (Ecuador) wet seasons, typically during November–April

and encompassing the austral summer (Johnson, 1976; Vuille *et al.*, 2000). The summer maximum in precipitation and easterly moisture flux over the Andes is driven by the upper-air easterly winds following the solar insolation maximum associated with the upper atmospheric anticyclone to the south-east of the central Andes (Lenters and Cook, 1997, 1999; Garreaud, 2000). Interannual variability in precipitation over the central Andes is strongly tied to the El Niño–Southern Oscillation (ENSO; Garreaud and Aceituno, 2001), which modulates the flux of moisture from the Atlantic (Vuille *et al.*, 2000).

Precipitation gradients are pronounced across the crest of the Andes. The highest mean annual precipitation (MAP) amounts are typically found on the first slopes that intersect the moisture-bearing winds (usually the easternmost east-facing slopes), and the lowest are found in deserts along the Pacific coast (Vuille and Keimig, 2004). Precipitation decreases southward in the Andes from relatively wet Ecuador, where MAP in Quito/Mariscal Sucre (0.15° S , 78.40° W , 2811 m a.s.l.) was 1205 mm for 1891–1990 (World Climate, 2007), through intermediate Peru, where the Junín Plain (11° S) received $\sim 800\text{--}900 \text{ mm a}^{-1}$ in the early 1960s (NOAA Data Rescue, 2007), to the arid Altiplano of southwestern Bolivia, where MAP is < 200 mm (Vuille *et al.*, 2000). Moreover, recent analyses using instrumental records and satellite cloud climatology have challenged long-assumed spatial coherence to regional precipitation on interannual and longer timescales (Vuille and Keimig, 2004). Multiple forcing mechanisms have been identified that differentially influence upper-air wind anomalies along the span of the tropical central Andes. This geographic complexity in regional precipitation could have significant implications for interpreting the climate forcing of palaeoglaciation along the Andes.

Mean annual temperature in the tropical Andes fluctuates within a narrow range, while daily temperature fluctuations are typically much greater (Johnson, 1976; Kaser *et al.*, 1990). For example, in Cerro de Pasco, Peru ($11^\circ 04' \text{ S}$, $77^\circ 38' \text{ W}$, 4500 m a.s.l.), the average daily temperature in 1963 ranged from 2.4° C in July to 4.6° C in December, with an average daily temperature of $3.8 \pm 0.7^\circ \text{ C}$ for the year. The average range between maximum and minimum daily temperatures during a month, however, was $13.3 \pm 3.1^\circ \text{ C}$ (NOAA Data Rescue, 2007).

Tropical glaciers

Glaciers of the tropical Andes are sensitive climate indicators because of their peculiar mass balance seasonality (Kaser and Osmaston, 2002). The relative constancy of temperatures throughout the year compared to pronounced wet–dry seasonality makes tropical glaciers distinct from those at higher latitudes in that accumulation occurs primarily during the wet season (austral summer), whereas ablation occurs year round, with melting reaching a maximum during the accumulation season (Kaser *et al.*, 1990; Wagnon *et al.*, 1999). The constancy of temperature also enhances a steep vertical mass balance gradient such that ablation occurs year round below the equilibrium line altitude (ELA) and accumulation is concentrated above the rain–snowline. Modern ELAs in the tropical Andes have been estimated at $\sim 5000\text{--}5200$ m a.s.l. (e.g. Wagnon *et al.*, 1999). The inner tropics feature particularly steep mass balance gradients, and have smaller ablation zones than those towards the subtropics, where drier conditions force a predominance of sublimation. Generally, given this more negative balance in the ablation zone, tropical

glaciers feature a faster and more pronounced terminus response to climate (Kaser, 1999, 2001; Kaser and Osmaston, 2002).

The steep vertical mass balance profiles of tropical glaciers result in different climate sensitivities of ELA positions along and across the Andes. Glaciers in the inner tropics have ELAs close to the 0°C isotherm, and respond directly to temperature changes. Those in the outer tropics and subtropics (including glaciers in the arid Western Cordillera of southern Peru and Bolivia) commonly have ELAs well above 0°C isotherms, and thus are not as temperature-sensitive. Rather, they are more responsive to changes in precipitation and humidity that alter the sublimation/melt regime (Kaser, 2001). This gradient in temperature–humidity sensitivity of glaciers has also been recognised along the east–west transect across the Andes, with implications for interpreting palaeoclimate forcing (Hastenrath, 1971; Klein *et al.*, 1999).

Recent increases in glacier recession are occurring all along the Andes (e.g. Thompson *et al.*, 2005), and the trend most closely correlates to temperature changes even though the climate mechanisms affecting the surface energy–mass budget differ between the inner (Ecuador) and outer (Bolivia) tropics. Monitoring of the mass and energy balances of Zongo and Chacaltaya Glaciers in Bolivia over more than 15 years has shown that the annual mass balance largely reflects the accumulation variability from just the wet summer months (DJF; Francou *et al.*, 1995, 2003). Yet at Antisana Glacier in Ecuador the mass balance relationship with temperature is more direct, as temperature strongly controls the elevation of the rain-snow line. Measurements from 1995 to 2002 show more seasonally constant ablation all year, with increased variability from February to May and during September due to the ENSO (Favier *et al.*, 2004a,b; Francou *et al.*, 2004). In both locations, however, interannual trends in mass balance are closely (inversely) related to temperature despite the fact that sensible heat flux at the glacier surface is not controlling the mass budget. The key is that temperature is strongly correlated to all the fluxes (humidity, cloudiness and precipitation) that have predominant impact on surface energy/mass balance, even in the drier outer tropical Bolivian Andes.

Given the predominance of ENSO on interannual Andean climate variability, there is a strong control over Andean glacier mass balance by Pacific sea surface temperature (SST) anomalies, via a precipitation linkage (Vuille *et al.*, 2000; Garreaud *et al.*, 2003). Francou *et al.* (2003) and Wagnon *et al.* (2001) have examined the ENSO control over glacier mass balance. El Niño features strongly negative glacier mass balances, while La Niña years are closer to balanced or even feature positive mass balance. El Niño is typically warmer as well as drier in the outer tropical highlands. The greatest effect, however, comes from the weaker precipitation, with lower albedo surfaces exposed to more radiation (less cloud cover). In this context, an increase in ENSO activity after the 1970s has been associated with an increase in recession rates of Central Andean glaciers (Francou *et al.*, 2000). Likewise, small glacier advances in the Cordillera Blanca during the 1960s and 1970s have been attributed to increases in precipitation, while an acceleration in glacier recession in the Central Andes since the 1980s has been associated with increased temperatures and atmospheric vapour (Kaser, 1999; Georges, 2004). ENSO indices and glacier mass balance measures in the Central Andes have been shown to be closely correlated through time (Francou *et al.*, 2003, 2004), but not in all instances (Vuille *et al.*, 2008). There is need for further research to distinguish the relative roles of Pacific and Atlantic SSTs in affecting the thermal and humidity advection regimes that ultimately control the glacier surface energy–mass budget.

The interpretation of Andean palaeoclimates based on palaeoglacier ELAs thus requires an understanding of multiple climate variables varying over space and time. Seltzer (1994a) advocated accounting for the relative influence of both precipitation and temperature when interpreting the difference between modern ELAs and ELAs reconstructed from geomorphological evidence. Klein *et al.* (1999) mapped regional snowlines to demonstrate the spatial variability in ELAs resulting from regional climatic gradients, and modelled some likely 'LGM' changes to precipitation and temperature. More recent efforts have combined mass–energy balance and glacier flow models with digital elevation models to simulate palaeoglacier sensitivity to combinations of climate variables in valley-specific topography (e.g. Kull and Grosjean, 2000; Kull *et al.*, 2002, 2003; Fairman, 2006).

Methods

Radiocarbon and CRN dating both involve post-measurement adjustments that introduce additional uncertainties into the final ages. Radiocarbon ages are calibrated, typically using an open-source computer program (e.g., CALIB; Stuiver *et al.*, 2005). CRN ages are calculated using derived isotope production rates and, if the sample area is not at sea level and high latitude, altitudinal and latitudinal scaling factors, to account for differences in atmospheric density and geomagnetic shielding (Gosse and Phillips, 2001). Favoured radiocarbon calibrations and CRN production rates and scaling methods have changed over time, with the younger CRN procedure arguably in the greater state of flux.

Currently several CRN scaling methods are in use (e.g. Lal, 1991; Stone, 2000; Lifton *et al.*, 2005) and an unequivocal best choice has yet to emerge. As an illustration of the state of the art, the CRONUS-Earth Online Calculator (henceforth CRONUS Calculator; http://hess.ess.washington.edu/math/index_dev.html; Balco *et al.*, 2008) calculates CRN ages with five scaling methods. The calculated CRN age for a local LGM sample from the tropical Andes, for example, may range from ~26 ka to ~35 ka, depending on the scaling method used. CRN ages that appear to support contradictory conclusions about the timing of glacial advances may in fact be in excellent agreement when calculated with the same scaling method.

In this paper, calibrated radiocarbon ages are presented in units of cal. a BP or cal. ka BP and uncalibrated ages are presented in units of ^{14}C a BP or ^{14}C ka BP. CRN ages are presented as originally published, except as noted, with additional discussion based on ages recalculated using the CRONUS Calculator. Readers who wish to recalculate the published CRN ages will find raw data suitable for the CRONUS Calculator included as supporting information Tables 1–4. Recalculated ages for the Junin valleys in Peru and Milluni and Zongo Valleys in Bolivia (Smith *et al.*, 2005b,c) are included as supporting information Table 5.

Chronological data

Chronological data are grouped by country and presented in north–south order (Fig. 1). With our focus on studies with numerical chronologies, we have included many that used surface exposure dating with CRNs to extend the glacial history beyond the reach of radiocarbon dating. As of this publication,

Table 1 Geographical and chronological data for the 16 sites discussed in the text

# on map	Site	Country	Latitude	Longitude	Summit altitude (m a.s.l.)	Sample altitude (m a.s.l.)	Calendar age (ka)	Radiocarbon age (^{14}C a BP)	95.4% (2 σ) cal. age ranges (cal. a BP) ^c	Feature dated	Sample description	Method	Lab ID	Interpreted significance	References
1	Rucu Pichincha	Ecuador	-0.21	-78.58	4784	3975	> 49 500 ^{14}C			Moraine-dammed palaeolake	Peat in lacustrine sediments enclosed by M2 moraines	^{14}C	Hv 17055	Minimum-limiting age for M2 moraines	Heine (1995)
						4100–4200	13 010 \pm 45 ^{14}C	15 096–15 687		Moraine-dammed Palaeolake/bog	Lower peat from palaeolake/bog enclosed	^{14}C	Hv 18076	Minimum-limiting age for M5 moraines	Heine (1995)
						4100–4200	11 155 \pm 100 ^{14}C	12 898–13 225		Moraine-dammed palaeolake/bog	Upper peat from palaeolake/bog enclosed by M5 moraines	^{14}C	Hv 17059	Maximum-limiting age for M6 moraines	Heine (1995)
						4100–4200 ?	8.2–9.0			HL-4 tephra	Tephra overlying M6 moraines (from Rosi, 1989)	?	NA	Minimum-limiting age for M6 moraines	Heine (1995)
2	Papallacta Valley, Potrerillos Plateau, Sucus site	Ecuador	-0.33	-78.17	4170	~3850	13 070 ^{14}C (ave. of 7)	ca. 15.5		Peat overlying till	Plant material and peaty organic matter in sediments overlying Sucus till	^{14}C	Various	Minimum age for Sucus advance	Clapperton <i>et al.</i> (1997)
						~3850	11 720 ^{14}C (ave. of 6)	ca. 13.7		Gytja	Organic matter in sediments overlying Sucus till	^{14}C	Various	Minimum age for deglaciation after Sucus advance	Clapperton <i>et al.</i> (1997)
						~3950	10 855 ^{14}C (ave. of 11)	ca. 12.8		Peat and plant macrofossils	Sediments underlying Potrerillos advance	^{14}C	Various	Maximum age of Potrerillos advance	Clapperton <i>et al.</i> (1997)
						~3950	10 035 ^{14}C (ave. of 5)	ca. 11.3		Peat and plant macrofossils	Sediments underlying Potrerillos advance	^{14}C	Various	Minimum age for deglaciation after Potrerillos advance	Clapperton <i>et al.</i> (1997)
						3870	12 250 \pm 130 ^{14}C	13 814–14 771		Peat overlying till	Upper layer of peat overlying lower peat layer, tephra, and till	^{14}C	Hv 18069	Minimum-limiting age for M5 moraines	Heine (1995); Heine and Heine (1996)
						4055	10 505 \pm 75 ^{14}C	12 151–12 214	12 224–12 753	Peat underlying till	Upper layer of peat underlying M6 moraines	^{14}C	Hv 17063	Maximum-limiting age for M6 moraines	Heine (1995); Heine and Heine (1996)
						4055	7 880 \pm 85 ^{14}C	8483–8490	8519–8530	fAh	Organic material overlying M6 moraines	^{14}C	Hv 17530	Minimum-limiting age for M6 moraines	Heine (1995); Heine and Heine (1996)
3	Chimborazo–Carthuirazo Massif, Site 1	Ecuador	-1.33	-78.78	5202	3770	33 290 \pm 300 ^{14}C	NA		Unit 5 peat below upper till	Peat underlying till	^{14}C	SRR-3028	Minimum-limiting age for M6 moraines	Clapperton (1987)
	Chimborazo–Carthuirazo Massif, Site 2	Ecuador	-1.34	-78.78	5202	3870	38 520 \pm 580/540 ^{14}C	NA		Unit 5 peat below upper till	Peat underlying till	^{14}C	SRR-3029	Minimum-limiting age for M6 moraines	Clapperton (1987)
	Chimborazo–Carthuirazo Massif, Site 3a, N flanks Carthuirazo	Ecuador	-1.36	-78.82	5202	3750	35 440 \pm 680/630 ^{14}C	NA		Unit 5 peat below upper till	Compacted peat underlying till	^{14}C	SRR-2583	Minimum-limiting age for Group 3 (full-glacial) moraines	Clapperton and McEwan (1985); Clapperton (1987)
	Chimborazo–Carthuirazo Massif, Site 3a, N flanks Carthuirazo	Ecuador	-1.36	-78.82	5202	3725	>40 330 ^{14}C	NA		Lower peat layer	Compacted peat underlying till	^{14}C	SRR-2584	Minimum-limiting age for Group 3 (full-glacial) moraines	Clapperton (1987)

Table 1 (Continued)

# on map	Site	Country	Latitude	Longitude	Summit altitude (m a.s.l.)	Sample altitude (m a.s.l.)	Calendar age (ka)	Radiocarbon age (^{14}C a BP)	95.4% (2σ) cal. age ranges (cal. a BP) ^c	Feature dated	Sample description	Method	Lab ID	Interpreted significance	References
	Chimborazo-Carhuairazo Massif, Site 4a, N/NE flanks Carhuairazo	Ecuador	-1.36	-78.82	5202	3725-3870		11 380 ± 50 ^{14}C	13 139-13 342	Upper peat layer	Peat overlying till	^{14}C	SRR-3030	Minimum-limiting age for uppermost till layer	Clapperton (1987)
	Chimborazo-Carhuairazo Massif, Site 4b, N/NE flanks Carhuairazo	Ecuador	-1.36	-78.82	5202	3800		14 770 ± 60 ^{14}C	17 561-18 208 18 310-18 475	Lower peat layer	Peat overlying till	^{14}C	SRR-3031	Minimum-limiting age for uppermost till layer	Clapperton (1987)
	Chimborazo-Carhuairazo Massif, drained lake basin, Río Mocha Valley	Ecuador	-1.45	-78.75	5202	3900-4000		11 370 ± 60 ^{14}C	13 120-13 345	Lower peat in drained glacial lake	Peat in laminated fine-grained sediments underlying till	^{14}C	SRR-2581	Minimum-limiting age for Group 2 (Lateglacial) moraines	Clapperton and McEwan (1985)
	Chimborazo-Carhuairazo Massif, drained lake basin, Río Mocha Valley	Ecuador	-1.45	-78.75	5202	3900-4000		10 975 ± 60 ^{14}C	ave. 13 085	Lower peat in drained glacial lake	Peat in laminated fine-grained sediments underlying till	^{14}C	Hv 18067	Minimum-limiting age for Group 2 (Lateglacial) moraines	Heine (1993)
	Chimborazo-Carhuairazo Massif, drained lake basin, Río Mocha Valley	Ecuador	-1.45	-78.75	5202	3900-4000		10 650 ± 60 ^{14}C	12 404-12 462 12 590-12 830	Upper peat in drained glacial lake	Peat in laminated fine-grained sediments underlying till	^{14}C	SRR-2582	Minimum-limiting age for Group 2 (Lateglacial) moraines	Clapperton and McEwan (1985)
	Chimborazo-Carhuairazo Massif, drained lake basin, Río Mocha Valley	Ecuador	-1.45	-78.75	5202	3900-4000		10 620 ± 85 ^{14}C	ave. 12 565	Upper peat in drained glacial lake	Peat in laminated fine-grained sediments underlying till	^{14}C	Hv 18066	Minimum-limiting age for Group 2 (Lateglacial) moraines	Heine (1993)
4	Cajas National Park	Ecuador	-2.75	-79.17	~4500	3700		12 470 ± 80 ^{14}C	14 260-15 400	Laguna Chorreras sediment	Plant macrofossils in lacustrine sediment core	^{14}C	CAMS-11958	Previous minimum-limiting age for deglaciation of L. Chorreras basin	Seltzer <i>et al.</i> (1995); Rodbell <i>et al.</i> (1999, 2002); Hansen <i>et al.</i> (2003)
						3700		13 160 ± 80 ^{14}C	15 540-16 030	Laguna Chorreras sediment	Plant macrofossils in lacustrine sediment core	^{14}C	CAMS-34965	Current minimum-limiting age for deglaciation of L. Chorreras basin	Seltzer <i>et al.</i> (1995); Rodbell <i>et al.</i> (1999, 2002); Hansen <i>et al.</i> (2003)
						4060		11 770 ± 70 ^{14}C	13 510-13 850	Laguna Pallacocha sediment	Plant macrofossils in lacustrine sediment core	^{14}C	CAMS-11967	Minimum-limiting age for deglaciation of L. Pallacocha basin	Seltzer <i>et al.</i> (1995); Rodbell <i>et al.</i> (1999, 2002); Hansen <i>et al.</i> (2003)
5	Nevado Huascarán (col)	Peru	-9.11	-77.61	6048	5884	ca. 19			Ice cap - basal age	Dip in $\delta^{18}\text{O}$ curve in core C-2 matched to GRIP/GISP2 records	Wiggle-matching		Estimated age for ice cap on Nevado Huascarán	Hansen <i>et al.</i> (2003); Thompson <i>et al.</i> (1995)

6	Quebradas Uquian, Cojup, Llaqa, Queshque, Cordillera Blanca ^a	Peru	-9.65	-77.36	6395	4045	12.7 ± 0.4 to 10.4 ± 0.4	Manchaque moraines	Boulders on moraines	¹⁰ Be	Various	Second Lateglacial/early Holocene readvance or stillstand? Lateglacial readvance or stillstand ca. 16.5 ka	Farber <i>et al.</i> (2005); Rodbell (1993a,b) Farber <i>et al.</i> (2005); Rodbell (1993a,b)
			-9.48	-77.47	3963-4008	16.1 ± 0.9 to 14.2 ± 0.7	Laguna Baja moraines (a)	Boulders on moraines	¹⁰ Be	Various	Various	Lateglacial readvance or stillstand ca. 16.5 ka	Farber <i>et al.</i> (2005); Rodbell (1993a,b)
			-9.83	-77.31	4157-4212	19.5 ± 0.5 to 11.9 ± 0.8	Laguna Baja moraines (b)	Boulders on moraines	¹⁰ Be	Various	Various	Lateglacial readvance or stillstand	Farber <i>et al.</i> (2005); Rodbell (1993a,b)
			-9.49	-77.47	3729-4016	29.3 ± 1.2 to 16.7 ± 1.7	Rurec moraines	Boulders on moraines	¹⁰ Be	Various	Various	ca. 16.5 ka Local LGM	Farber <i>et al.</i> (2005); Rodbell (1993a,b)
			-9.50	-77.46	3627-3890	439 ± 13 to 76 ± 2.1	Cojup moraines (a)	Boulders on moraines	¹⁰ Be	Various	Various	Multiple older glaciations, more extensive than local LGM	Farber <i>et al.</i> (2005); Rodbell (1993a,b)
			-9.82	-77.37	4060	441 ± 13 to 176 ± 4.2	Cojup moraines (b)	Boulders on moraines	¹⁰ Be	Various	Various	Multiple older glaciations, more extensive than local LGM	Farber <i>et al.</i> (2005); Rodbell (1993a,b)
7	Nevado Jeulla Rajo, Conococha Plain	Peru	-10.00	-77.25	5682	No published dates	Moraines	Moraines, bedrock	¹⁰ Be				Smith <i>et al.</i> (2007)
8	Cordillera Huayhuash	Peru	-10.25	-76.83	6617	No published dates			¹⁰ Be				Hall <i>et al.</i> (2004, 2006); Ramage <i>et al.</i> (2004)
9	Junin Plain, Eastern Cordillera (four valleys) ^b	Peru	-11.00	-76.00	4858	17.8 ± 0.6 to 12.2 ± 0.5	Ground moraine, bedrock (Group A)	Boulders and bedrock (Alcachocho Valley only)	¹⁰ Be, ²⁶ Al	Various	Various	Rapid deglaciation following Group B deposition	Smith <i>et al.</i> (2005b,c)
					4252-4391	21.3 ± 0.6 to 13.6 ± 0.6	Laterofrontal moraines (Group B)	Boulders on moraines	¹⁰ Be, ²⁶ Al	Various	Various	Lateglacial readvance or stillstand	Smith <i>et al.</i> (2005b,c)
					4159-4388	31.3 ± 1.4 to 17.3 ± 0.6	Laterofrontal moraines (Group C)	Boulders on moraines	¹⁰ Be, ²⁶ Al	Various	Various	ca. 18-15 ka Local LGM	Smith <i>et al.</i> (2005b,c)
					4168-4464	1606 ± 11.6 to 51 ± 1.1	Lateral moraines (Group D)	Boulders on moraines	¹⁰ Be, ²⁶ Al	Various	Various	Multiple older glaciations, more extensive than local LGM	Smith <i>et al.</i> (2005b,c)
10	Upismayo Valley, Cordillera Vilcanota	Peru	-13.76	-71.25	6384	41 520 ± 4 430 ¹⁴ C	Bog	Bottom of 10 m thick peat layer	¹⁴ C	GX-23726	Various	Maximum-limiting age for outermost moraines	Goodman <i>et al.</i> (2001); Mark <i>et al.</i> (2002)
					4450	13 880 ± 150 ¹⁴ C	Bog	Top of buried, distorted 10 m thick peat layer	¹⁴ C	GX-23725	Various	Minimum-limiting age for group of seven nested moraines upvalley	Goodman <i>et al.</i> (2001); Mark <i>et al.</i> (2002)
					4010	15 640 ± 100 ¹⁴ C	Lake	Basal lacustrine organics	¹⁴ C	AA-27027	Various	Minimum-limiting age for deglaciation from local LGM	Goodman <i>et al.</i> (2001); Mark <i>et al.</i> (2002)
					4580	14 500 ± 220 ¹⁴ C	Lake	Basal lacustrine organics	¹⁴ C	AA-27024	Various	Minimum-limiting age for deglaciation from local LGM	Goodman <i>et al.</i> (2001); Mark <i>et al.</i> (2002)

(Continues)

Table 1 (Continued)

# on map	Site	Country	Latitude	Longitude	Summit altitude (m a.s.l.)	Sample altitude (m a.s.l.)	Calendar age (ka)	Radiocarbon age (^{14}C a BP)	95.4% (2 σ) cal. age ranges (cal. a BP) ^c	Feature dated	Sample description	Method	Lab ID	Interpreted significance	References
11	Huancane Valley, Queiccaya Ice Cap	Peru	-13.97	-70.88	5645			12 240 \pm 170 ^{14}C	13 189–14 855	Peat	Bottom of peat behind Huancané III moraine	^{14}C	I-8443	Minimum-limiting age for outermost moraine belt (Huancané III)	Mercer and Palacios (1977)
12	San Francisco Valley, Cordillera Real ^b	Bolivia	-16.00	-68.32	6427	4470–4750	24.1 \pm 0.9 to 9.4 \pm 0.6			Lateral moraines (6)	Boulders on moraines	^{10}Be	Various	Local LCM and Lateglacial readvance	Zech <i>et al.</i> (2007)
13	Zongo Valley, Cordillera Real ^a	Bolivia	-16.20	-68.12	6088	3806–4105	16.1 \pm 0.7 to 10.4 \pm 0.7			Lateral moraines (Group A)	Boulders on moraines	^{10}Be , ^{26}Al	Various	Lateglacial readvance or stillstand	Smith <i>et al.</i> (2005b)
					6088	3383–3503	17.8 \pm 0.9 to 10.2 \pm 0.3			Lateral moraines (Group B)	Boulders on moraines	^{10}Be , ^{26}Al	Various	Lateglacial readvance or stillstand	Smith <i>et al.</i> (2005b)
					6088	4643	16.2 \pm 0.5 to 8.2 \pm 0.6			Lateral moraines (Group B)	Boulders on moraines	^{10}Be , ^{26}Al	Various	Lateglacial readvance or stillstand	Smith <i>et al.</i> (2005b)
					6350	4591–4596	31.8 \pm 1.1 to 14.5 \pm 0.4			Lateral moraines (Group C)	Boulders on moraines	^{10}Be , ^{26}Al	Various	Lateglacial readvance or stillstand	Smith <i>et al.</i> (2005b)
14	Milluni Valley, Cordillera Real ^a	Bolivia	-16.62	-67.77	6350	6213.3	ca. 18			Base of ice core	ECM; layer counting; wiggle-matching to Huascarán core; $\delta^{18}\text{O}$	ECM, $\delta^{18}\text{O}$	Various	Local LCM ca. 32–28 ka	Smith <i>et al.</i> (2003); Ramirez <i>et al.</i> (2003)
15	Laguna Kollpa Kkota	Bolivia	-17.43	-67.13	4560	4400		17 670 \pm 120	20 747–21 244	Lacustrine sediments	Humic acid extract from lacustrine sediment core C	^{14}C	CAMS 2392	Minimum-limiting age for deglaciation from local LCM	Seltzer (1994b); Seltzer <i>et al.</i> (1995)
					4560	4400		17 580 \pm 170	20 570–21 173	Lacustrine sediments	Humic acid extract from lacustrine sediment core C	^{14}C	CAMS 2393	Minimum-limiting age for deglaciation from local LCM	Seltzer (1994b); Seltzer <i>et al.</i> (1995)
					6411.2	4400		17 690 \pm 780	19 907–22 036	Lacustrine sediments	Bulk organic matter from lacustrine sediment core A	^{14}C	BETA 48225	Minimum-limiting age for deglaciation from local LCM	Seltzer (1994b); Seltzer <i>et al.</i> (1995)
16	Nevado Sajama	Bolivia	-18.10	-68.88	6542	6411.2		21 200 \pm 370	24 520–25 380 (24 950 \pm 430)	Ice cap	Wood at 130.8 m in core C-1 (132.4 total depth)	^{14}C	LLNL	Minimum age for base of ice cap	Thompson <i>et al.</i> (1998)
					6411.2			20 400 \pm 120	23 880–24 160 (24 020 \pm 140)	Ice cap	Wood at 130.8 m in core C-1 (132.4 total depth)	^{14}C	WHNOS	Minimum age for base of ice cap	Thompson <i>et al.</i> (1998)

^aLal (1991)/Stone (2000) altitude/latitude scaling method; zero erosion; geomagnetic correction (ages < 800 ka).

^bLifton (2005) altitude/latitude scaling method; zero erosion.

^cReferences for calibrated radiocarbon ages: Stuiver and Reimer (1993); Reimer *et al.* (2004); Stuiver *et al.* (2005).

however, no surface exposure ages have been published for Ecuador, and thus we rely on published radiocarbon ages to limit the timing of glacier expansion and retreat in Ecuador. Table 1 provides geographical and chronological data for all of the sites discussed in the text. The temporal patterns of glaciation in the tropical Andes are summarised graphically by use of time–distance diagrams (Fig. 2). One important limitation of most such time–distance datasets is the paucity of information on the extent of ice retreat between glacial maxima, and consequently there is considerable uncertainty in the details of ice retreat as portrayed in Fig. 2.

Ecuador

Although Ecuador straddles the Equator, glaciers exist on high peaks (many of them volcanic) that rise above the regional snowline in the Eastern and Western Cordillera of the Ecuadorian Andes (Fig. 3). Jordan and Hastenrath (1998) identified four glacierised mountains in the drier Western Cordillera and 13 in the wetter Eastern Cordillera, altogether hosting more than 100 small glaciers. Jordan and Hastenrath (1998) estimated that glaciers covered nearly 22 km² in the Western Cordillera and about 75 km² in the Eastern Cordillera. The distribution of glacial erosion features and glacial deposits indicates that glacier coverage has been greater in the past (Clapperton, 1990).

In the Ecuadorian Andes, the task of deciphering the glacial history is complicated by the abundance of tephra and windblown volcanic sediments (*cangagua*) on the high volcanic peaks that contain the most complete glacial records. Clapperton (1990) pointed out, for example, that an older till deposit on the volcano Chimborazo was covered by at least 57 layers of tephra. Published studies based on surface exposure dating with CRNs are not yet available for Ecuadorian glacial deposits.

Mark *et al.* (2004) provided an overview of chronologies based on radiocarbon dating and presented ArcView-formatted digital maps of glacier limits in Ecuador. Smith *et al.* (2005a), focusing on the LGM record, summarised studies based primarily on radiocarbon chronologies for three locations in Ecuador: Rucu Pichincha volcano, the Papallacta Valley on the Potrerillos Plateau and the Chimborazo–Carihuairazo Massif. These three sites are discussed here, along with a fourth site, Cajas National Park in the southern Ecuadorian Andes (Fig. 3, Table 1).

Rucu Pichincha (Fig. 1, Site 1)

At Rucu Pichincha (4784 m a.s.l.) in the Western Cordillera (0° 12.5' S, 78° 35' W) Heine (1995) and Heine and Heine (1996) identified six moraine groups (M1–M6, where M1 is oldest). M1 moraines descend to 3550–3600 m a.s.l. and are described as being oxidised and deeply weathered (to depths of 3.5 m); they overlie a deeply weathered lava flow dated at >0.9 Ma (Rosi, 1989, reported in Heine and Heine, 1996). M2 moraines are located 'right next to' M1 moraines and are similarly deeply weathered (Heine, 1995). Lake sediments enclosed by M2 moraines at 3975 m a.s.l. were beyond the range of radiocarbon dating (>49.5 ¹⁴C ka BP). M3 moraines are found in only two places (down to ~3700 m a.s.l.) and are less oxidised and weathered than M2 moraines.

M4 lateral and terminal moraines are typically narrow and only slightly weathered, and enclose 'humpy tills' that Heine

(1995) interpreted as former ice-cored moraines. Heine (1995) mapped M4 moraines almost to 3700 m a.s.l. Peat from a palaeolake/bog enclosed by M5 moraines at 4100–4200 m a.s.l. gave minimum-limiting radiocarbon dates of 11 155 ± 100 ¹⁴C a BP (ca. 13.0 cal. ka BP) and 13 010 ± 45 ¹⁴C a BP (ca. 15.5 cal. ka BP) for the M5 moraines. Heine (1995) proposed that the M4 moraines were deposited during the global LGM and that the M5 moraines were deposited as recessional moraines.

M6 moraines are found at 4200–4400 m a.s.l., and Heine and Heine (1996) bracketed the age of the M6 moraines between 11 155 ± 100 ¹⁴C a BP (age of underlying peat, ca. 13.0 cal. ka BP) and 8.2–9.0 ka (age of overlying HL-4 tephra from Rosi, 1989). Heine and Heine (1996) interpreted their findings to indicate that glaciers at Rucu Pichincha did not advance during the YD climate reversal.

Papallacta Valley on the Potrerillos Plateau (Fig. 1, Site 2)

Heine (1995), Heine and Heine (1996) and Clapperton *et al.* (1997) studied the Papallacta Valley on the Potrerillos Plateau in the Eastern Cordillera (0° 20' S, 78° 12' W). The plateau is located approximately 15–30 km north of the peak of Volcán Antisana (5704 m a.s.l.); the Papallacta Valley lies south of Papallacta Pass on the southern edge of the Plateau. Most of the Plateau lies in the range 3900–4200 m a.s.l., with some ridges exceeding 4400 m a.s.l. and the highest peak reaching 4502 m a.s.l. Glaciers and an ice cap are present on Antisana but glacial ice is absent on the plateau. The limits of the most extensive ice cover are indicated by (weathered) moraines at ~3000 m a.s.l. on the leeward west side and ~2700 m a.s.l. on the windward east side of the Plateau (Clapperton *et al.*, 1997). Clapperton *et al.* (1997) estimated the ELA on Antisana as 4970 ± 50 m a.s.l., without specifying how the estimate was made.

Heine (1995) and Heine and Heine (1996) distinguished seven groups of moraines (M1–M7, where M1 is oldest and M7 is Neoglacial) in the Papallacta Valley (Fig. 2). Heine (1995) credited glaciers that deposited M1 and M2 moraines with carving the U-shaped Mullumica Valley (located approximately 11 km north of Lago de Sucus on the Potrerillos Plateau, based on the map in Heine, 1995). The Mullumica Valley is partially filled by a lava flow that was dated to >150–180 ka by fission-track dating (Salazar, 1985). Heine (1995) proposed that the age of the lava flow in Mullumica Valley provided a minimum-limiting age for the M1 and M2 moraines. Heine (1995) assigned pre-LGM status to M3 moraines (shown descending to ~3400 m a.s.l. on the map), based largely on the degree of weathering. Heine (1995) and Heine and Heine (1996) interpreted the M4 moraines found at 3800–3900 m a.s.l. as the deposits marking the maximum extent of MIS 2 glaciation (i.e., LGM moraines), M5 moraines as older than ca. 12.2 ¹⁴C ka BP but younger than the LGM, and M6 moraines found at 4055 m a.s.l. as Lateglacial (bracketed by radiocarbon ages of ca. 10.5 ¹⁴C ka BP below and ca. 8 ¹⁴C ka BP above).

Clapperton *et al.* (1997) disagreed with the interpretation that the M4 moraines found at 3800–3900 m a.s.l. marked the LGM glacial extent (Heine, 1995; Heine and Heine, 1996) and named two younger advances: the Sucus and the Potrerillos. Clapperton *et al.* (1997) reported that an unpublished radiocarbon age from organic material found at 3500 m a.s.l. beneath two tills separated by a lava flow indicated that 'glaciers advanced past this point at least twice after ca. 30,000 yr B.P.' (presumably ¹⁴C a BP; data were not included). The sampling site was at a lower altitude (3500 m a.s.l.) than the Sucus termination (~3850 m a.s.l.; see below), suggesting that the Sucus advance

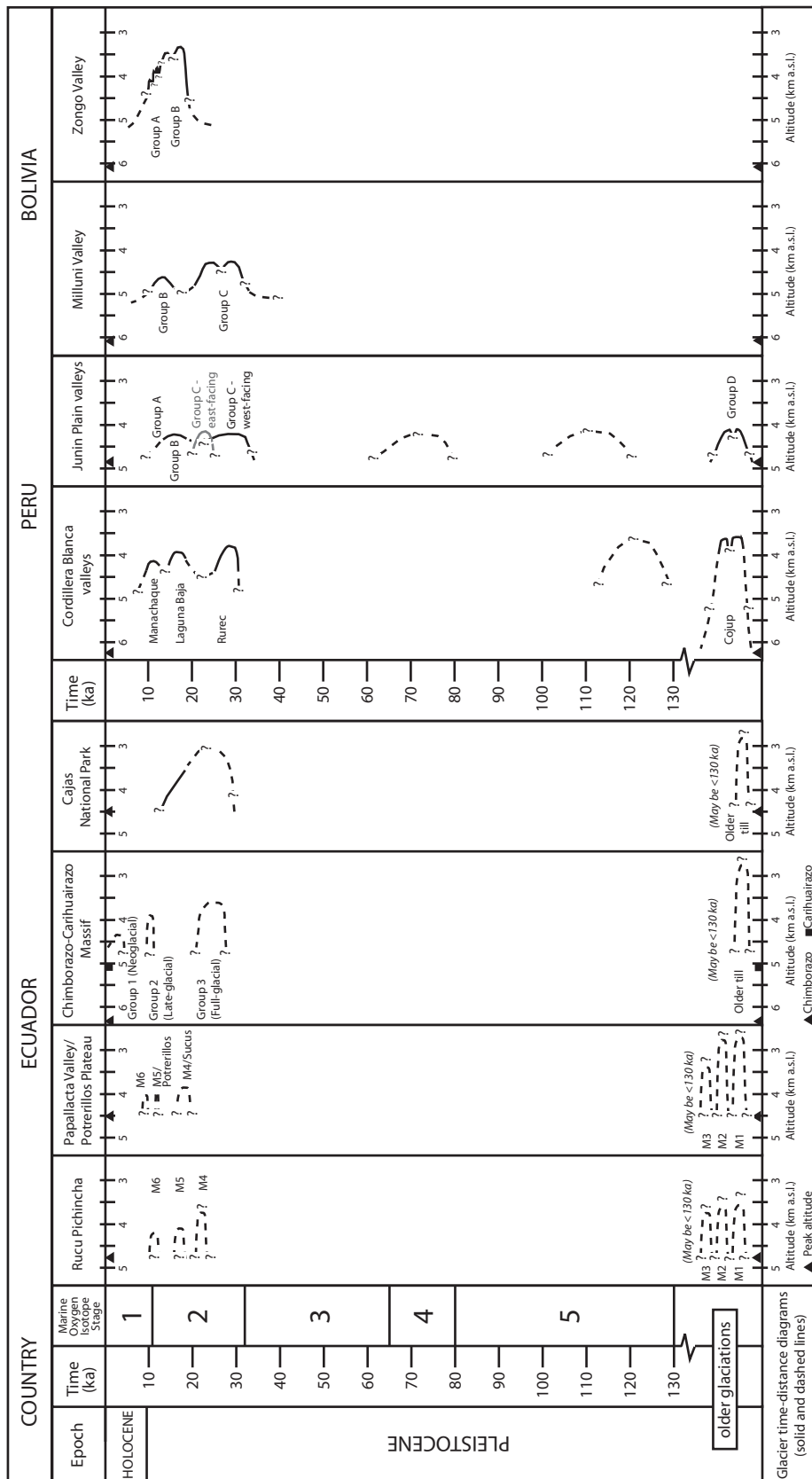


Figure 2 Time-distance diagram for glacial advances at eight sites discussed in the text. For San Francisco Valley, Bolivia, see Zech et al. (2008)

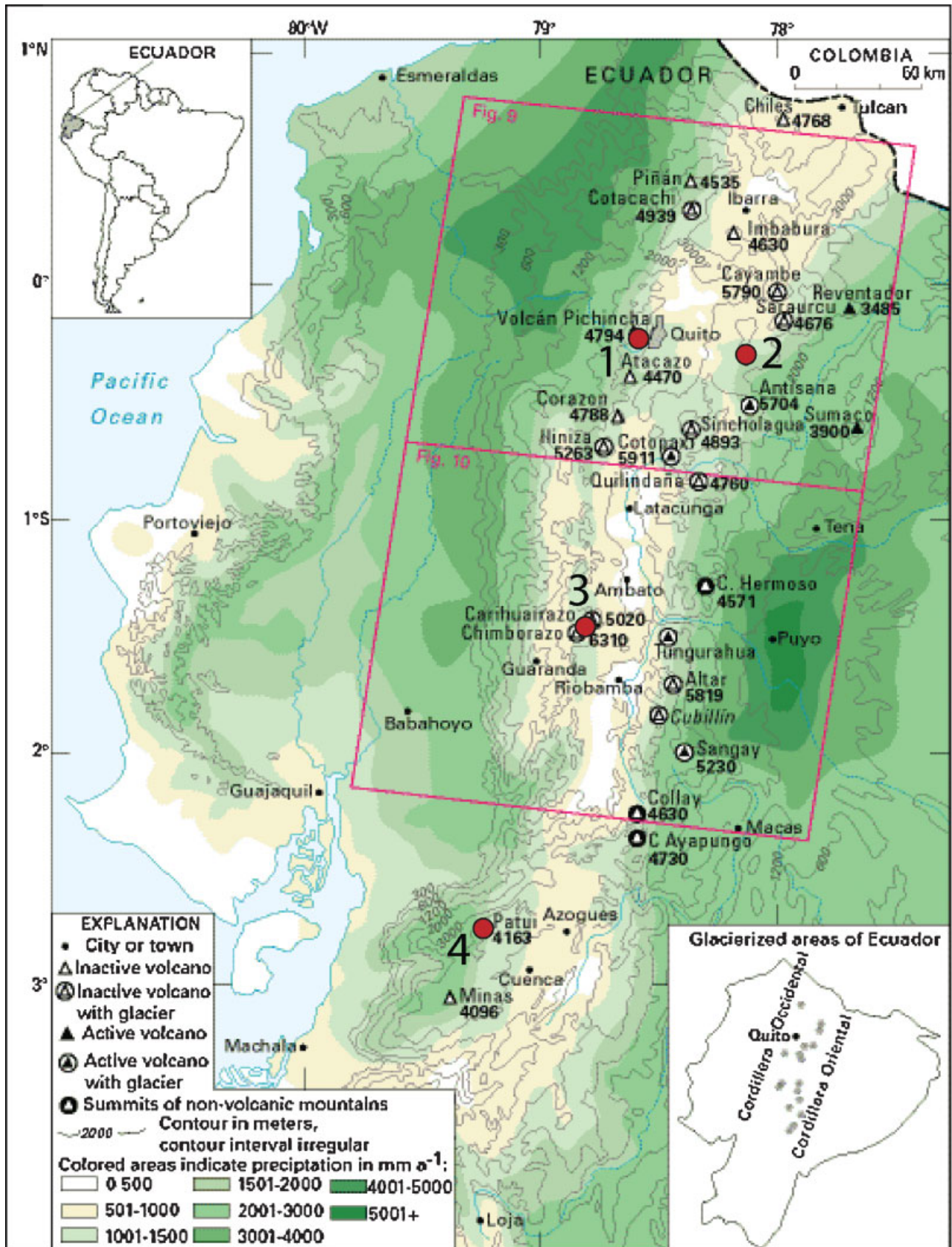


Figure 3 Sites in Ecuador: (1) Rucu Pichincha; (2) Papallacta Valley and Potrerillos Plateau; (3) Chimborazo–Carhuairazo massif; (4) Cajas National Park. Base map showing geographical features, political boundaries and annual precipitation amounts from Jordan and Hastenrath (1998); used by permission of US Geological Survey (open-file report)

was younger than both of the tills, which in turn were younger than ca. 30 ¹⁴C ka BP (Clapperton *et al.*, 1997).

The Sucus advance descended to ~3850 m a.s.l. Clapperton *et al.* (1997) reported seven minimum-limiting radiocarbon

ages on plant material and peaty organic matter in sediments overlying till between two Sucus lateral moraines. Based on the radiocarbon dating, the Sucus advance occurred before ca. 13 070 ¹⁴C a BP (ca. 15.6 cal. ka BP), the average of the

seven ages. The average of six minimum-limiting radiocarbon ages on sediments underlying the younger Potrerillos advance date deglaciation after the Sucus advance to ca. 11 720 ¹⁴C a BP (ca. 13.7 cal. ka BP).

The Potrerillos advance descended to ~3950 m a.s.l. Potrerillos moraines are bracketed between 10 855 ¹⁴C a BP (ca. 12.8 cal. ka BP), the average of 11 maximum-limiting radiocarbon ages, and 10 035 ¹⁴C a BP (ca. 11.3 cal. ka BP), the average of five minimum-limiting radiocarbon ages for the Potrerillos advance. On the basis of the radiocarbon dating, Clapperton *et al.* (1997) interpreted the Potrerillos advance as contemporaneous with the YD climate reversal.

Chimborazo–Carihuairazo Massif (Fig. 1, Site 3)

Volcanoes Chimborazo (6310 m a.s.l.) and Carihuairazo (5020 or 5102 m a.s.l.) are part of a volcanic massif some 20–30 km in diameter located in the Western Cordillera (1° 30' S, 78° 50' W) of the Ecuadorian Andes. Both volcanoes are inactive and glaciated. Though generally andesitic, the massif includes high-silica andesite from early eruptive stages, later-stage dacite–rhyolite and basic andesite erupted from flank fissures during the waning stages of volcanism. The final eruption is thought to have occurred prior to 11 ka BP (Clapperton, 1990).

Clapperton and McEwan (1985) identified three groups of moraines in the Río Mocha Valley between Chimborazo and Carihuairazo and assigned them general ages (Fig. 2): Neoglacial (Group 1), Lateglacial (Group 2), and full-glacial (Group 3). Group 3 moraines extend to just below 3600 m a.s.l., Group 2 to 4050 and 3900 m a.s.l., and Group 1 to 4300–4400 m a.s.l. Clapperton and McEwan (1985) reported radiocarbon ages of 10 650 ± 60 and 11 370 ± 60 ¹⁴C a BP (ca. 12.7 and 13.4 cal. ka BP, respectively) on the upper and lower peat layers within laminated fine-grained sediments underlying till in a drained glacial lake basin located upvalley of Group 2 moraines between the two peaks (~3900–4000 m a.s.l.). Clapperton and McEwan (1985) correlated till overlying compacted peat on the northern flanks of Carihuairazo with Group 3 moraines in Río Mocha Valley on the south-west side of the volcano. The upper layer of peat was dated at 35 440 ± 680/630 ¹⁴C a BP; the lower peat layer was beyond the limit of radiocarbon (>40 ¹⁴C ka BP).

Clapperton and McEwan (1985) noted that deposits of highly oxidised and weathered till were present beyond the limits of the Group 3 moraines (down to altitudes of 3350 m a.s.l.), and commented that larger glaciations older than those that deposited the Group 3 moraines may have been responsible for eroding the deep glacial valleys. Clapperton (1986) revised downward the lower limit of till deposits older than Group 3 to 2750 m a.s.l., emphasising that no glacial deposits had been found at lower altitudes and thus 2750 m a.s.l. may represent the limit of glaciation in the Ecuadorian Andes.

Clapperton (1986) extended the classification of glacial deposits (Neoglacial, Lateglacial and full-glacial moraines, and older till deposits) to include the entire Chimborazo–Carihuairazo Massif, as well as other volcanic peaks in Ecuador (e.g. El Altar). Clapperton (1986) estimated changes in ELA for the Neoglacial (–200 to –360 m), Lateglacial (–440 to –600 m), and full-glacial (–600 to –960 m) advances on Chimborazo, Carihuairazo and other volcanoes. He noted the asymmetry of past and present glacier extents and ELAs, with both lying at lower altitudes on the eastern slopes than on the western slopes of the peaks. He attributed the pattern to the easterly source of precipitation.

Clapperton (1987) reported six radiocarbon dates from peat deposits at altitudes of 3725–3870 m a.s.l. on the north and north-east flanks of Carihuairazo. The four radiocarbon dates from peat lying between the uppermost and middle two till layers (of three) fell between ca. 33 ¹⁴C ka BP (uppermost peat) and >40 ¹⁴C ka BP (basal peat), which suggested to Clapperton that the most recent glacial advance may have culminated 'no more than a few thousand years' after the uppermost peat formed and thus earlier than the global LGM. The two radiocarbon dates from peat overlying the upper till were ca. 11.4 and 14.8 ¹⁴C ka BP, indicating recession from the location by ca. 15 ¹⁴C ka BP.

Clapperton (1990) summarised his view of the history of glaciation on the massif as follows:

- During the Neoglacial (specified as 5 ka to present) glaciers built massive moraines close to present ice margins (4300–4400 m a.s.l.); stratigraphic relationships within the moraines suggest that moraines were constructed by repeated advances. [Group 1]
- Lateglacial advances (ca. 12–10 ka, encompassing the YD climate reversal) built a group of three to four arcuate terminal moraines (4050 and 3900 m a.s.l.) several kilometres downvalley from the Neoglacial moraines. Lateglacial moraines are dated by the peat layers within the glacial lake sediments in Río Mocha Valley. [Group 2]
- Full-glacial advances built a group of three to four relatively small moraines (5–10 m high) within 100–200 m high outer moraines (~3600 m a.s.l.) between ca. 30 and ca. 14 ka. Maximum-limiting ages are provided by peat layers underlying full-glacial till on the north-east slope of Carihuairazo (ca. 33 ¹⁴C ka BP). [Group 3]
- Older tills are present beyond the limits of the full-glacial moraines (to ~2750 m a.s.l.); differing degrees of weathering suggest that at least two generations of older tills exist, but chronological control is lacking.

J. Heine (1993) examined the stratigraphy of exposed glacial lake sediments at the Río Mocha site and agreed with Clapperton and McEwan (1985) that the valley was at one time dammed by a glacier that deposited the Group 2 moraine that crosses the valley, although he challenged Clapperton's interpretation that the aforementioned peat layers necessarily represent intervals of ice recession. In 1993 G. Seltzer and D. Rodbell visited the site and concurred with Heine (1993) that the connection between the radiocarbon-dated peat deposits and ice margin positions was ambiguous (Rodbell and Seltzer, 2000). Heine's radiocarbon dates of 10 620 ± 85 ¹⁴C a BP (upper peat layer) and 10 975 ± 85 ¹⁴C a BP (lower peat layer) were comparable to those of Clapperton and McEwan (1985), though his date on the lower layer was younger by a statistically significant amount (>4σ). Heine used average calibrated ages (12 565 and 13 085 cal. a BP, respectively) to estimate a sedimentation rate, and from this he extrapolated a minimum age of 15 535 cal. a BP for the lake sediments beneath the lower peat. Heine thus concluded that the Group 2 moraine that originally dammed the lake pre-dated the YD by at least 2500 a.

Cajas National Park (Fig. 1, Site 4)

Cajas National Park (2° 40'–3° 00' S, 79° 00'–79° 25' W) is located on the continental divide in the southern Ecuadorian Andes, ~25 km west of the city of Cuenca and ~150 km south-south-west of Chimborazo (Fig. 1). The park encompasses a broad plateau with altitudes ranging from ~3100 to ~4500 m

a.s.l. but typically ~4000 m a.s.l., except in deep valleys. The park is dotted with numerous glacial lakes, U-shaped valleys and moraines, although no glaciers are currently present (Rodbell *et al.*, 2002; Hansen *et al.*, 2003).

Clapperton (1986) described deeply weathered till exposed at 2800 m a.s.l. in the Rio Tomebamba Valley west of Cuenca as being similar to weathered tills located beyond the limits of the full-glacial (Group 3) moraines on Carihuairazo (Fig. 2). Soil overlying 20 m of the Cuenca Plateau till was beyond the limit of radiocarbon dating (>40 ¹⁴C ka BP). Noting that the weathered till at 2800 m a.s.l. roughly coincided with the limits of glacial troughs, Clapperton (1986) concluded that 2800 m a.s.l. marked 'the absolute limits of Quaternary glaciation in the Ecuadorian Andes'. Clapperton (1986) estimated the full-glacial glacier limit for the Cuenca Plateau at 3800 m a.s.l. in the north-west and 3200 m a.s.l. in the south-east, representing ELA lowerings of 1100 m and 1480 m, respectively.

Geomorphic evidence in the park indicates that a region of ~400 km² above ~2800 m a.s.l. was covered by an ice cap during the LGM (Rodbell *et al.*, 2002). Soil catena studies (cf. Birkeland, 1994) by Goodman (1996) on moraines at 3760, 3360 and 3080 m a.s.l. in the Tomebamba drainage reveal A/Bw/Cox profiles and <2 weighted mean % pedogenic iron (CBD (Citrate-Bicarbonate-Dithionite) extractable) iron, which strongly suggests that these moraines are relatively young and were likely deposited during MIS 2; no older moraines were identified (Rodbell *et al.*, 1996, 2002).

Minimum ages for deglaciation of the park are provided by radiocarbon dates from lacustrine sediment cores. Moraine-dammed Laguna Chorreras is located at 3700 m a.s.l. in a narrow tributary valley to the Tomebamba River Valley and Laguna Pallcacocha is located in a cirque at ~4060 m a.s.l. at the head of the Tomebamba River Valley in the park (Hansen *et al.*, 2003). Hansen *et al.* (2003) reported a radiocarbon age of 13 160 ± 80 ¹⁴C a BP from the basal organics in a core from Laguna Chorreras, and they estimated that the base of the core dates to ca. 17 000 cal. a BP, the time when ice retreated from the lake basin. The oldest radiocarbon age in the Pallcacocha core was 11 770 ± 70 ¹⁴C a BP, and the base of the core was estimated to date from ca. 14 500 cal. a BP (Hansen *et al.*, 2003). These radiocarbon ages indicate deglaciation from ~3700 m a.s.l. by ca. 17 000 cal. a BP, with deglaciation of ~4050 m a.s.l. by ca. 14 500 cal. a BP. Pollen analyses from the Chorreras and Pallcacocha cores indicate a wetter and cooler climate than today from ca. 17 000 to ca. 11 000 cal. a BP, followed by an expansion of moist montane forests and increased fire activity during the Holocene (Hansen *et al.*, 2003).

Peru

The Peruvian Andes are the world's most glacierised tropical region (Kaser and Osmaston, 2002). Morales-Arno and Hastenrath (1998) estimated that glaciers covered 2600 km² on 20 distinct cordilleras in the Peruvian Andes, with the greatest concentrations in the Cordillera Blanca and the Cordillera de Vilcanota (Fig. 2). Georges (2004) used satellite data to estimate ice coverage in the Cordillera Blanca in 1990 at 620 km². The presence of moraines and other glacial deposits in regions of the Peruvian Andes that currently have either no glaciers or glaciers terminating far upvalley from moraines indicate that glacier coverage has been considerably greater in the past (e.g., Rodbell, 1993a; Mark *et al.*, 2004; Smith *et al.*, 2005c).

Seltzer *et al.* (2000, 2002) interpreted sedimentological, palaeobiotic and isotopic changes in a sediment core from Lake Junín (11° S, 76° W, ~4080 m a.s.l.) to indicate that the local LGM in the central Peruvian Andes occurred ca. 30–22.5 ka, followed by deglaciation ca. 22–21 ka and a minor readvance ca. 21–16 ka, all during wet climatic conditions, and then rapid glacial retreat as the climate became drier after 16 ka. Wet conditions returned after ca. 10 ka (Seltzer *et al.*, 2000). According to Seltzer *et al.* (2002), deglaciation ca. 22–21 ka occurred as a response to increased mean annual temperatures rather than a change in the moisture regime.

Recent studies that include surface exposure dating with CRNs have brought to light an interesting aspect of the glacial history of the Peruvian Andes. In some regions the glacial advances of the local LGM are relatively minor compared to older advances (Smith *et al.*, 2005c), whereas in at least one neighbouring region ~100 km away the local LGM deposits mark the outermost moraines identified (Hall *et al.*, 2006). Differences in valley hypsometry and maximum peak altitudes seem likely to have played a role in these regional variations.

Smith *et al.* (2005a) summarised studies from eight locations in the Peruvian Andes in their analysis of LGM snowlines. Mark *et al.* (2004) provided an overview of chronologies based on radiocarbon dating and presented ArcView-formatted digital maps of glacier limits in Peru. We will not revisit previous summaries (Mark *et al.*, 2004; Smith *et al.*, 2005a) of studies in the Cordillera Oriental (Rodbell, 1991, 1992, b), Cerros Cuchpanga (Wright, 1983, 1984), Nevado Huaytapallana (Seltzer, 1987, 1990) and the Cordillera Ampato (Dornbusch, 2002).

Here we discuss one site from which ice cores have been retrieved, four regions in which surface exposure dating of glacial deposits has been used to develop glacial chronologies, and two sites where radiocarbon dating has provided the framework of the glacial history and where, in one of them, surface exposure dating is in early stages (Fig. 4, Table 1).

Nevado Huascarán (Fig. 1, Site 5)

In 1993 Thompson and colleagues retrieved two ice cores from the ice cap in the col (6048 m a.s.l.) between the two peaks of Peru's highest mountain: Nevado Huascarán (9° 06' 41'' S, 77° 36' 53'' W; 6768 m; Fig. 4). Cores C1 (160.4 m) and C2 (166.1 m) were drilled to bedrock. Thompson *et al.* (1995) used borehole and atmospheric temperatures to calculate that the basal ice was frozen to the bedrock; characteristics such as layering and air bubbles were interpreted to indicate that basal ice had not melted in the past. Basal ice was dated at approximately 19 ka BP by matching the dip in the Huascarán δ¹⁸O curve at 164.1 m in C2 to the midpoint of the YD interval in GRIP and GISP 2 cores from Greenland, then estimating layer thinning with depth (Thompson *et al.*, 1995). Thompson *et al.* (1995, 2003) interpreted the 6 ‰ dip in the δ¹⁸O record from near-basal Huascarán ice as evidence for a cooler and drier last glacial stage in the Peruvian cordillera.

Quebradas Uquian, Cojup, Llaca and Queshque, Cordillera Blanca (Fig. 1, Site 6)

Farber *et al.* (2005) used surface exposure dating with cosmogenic ¹⁰Be to date 44 boulders on moraines in four valleys (Quebradas Uquian, Cojup, Llaca and Queshque) on the western side of the Cordillera Blanca (9° 30' S, 77° 15' W;

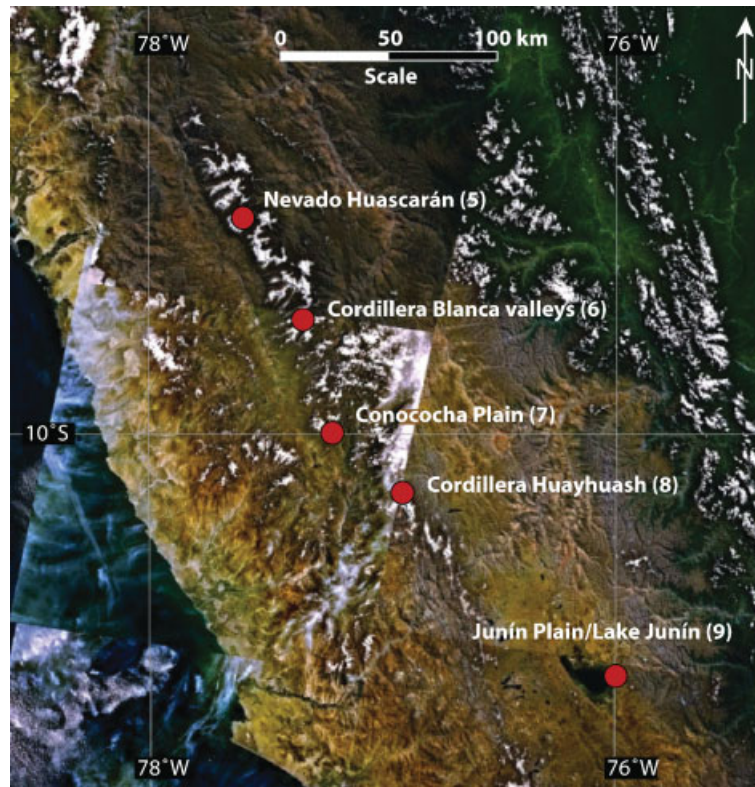


Figure 4 Satellite image of the central Peruvian Andes showing the location of five sites discussed in the text: Nevado Huascarán, valleys in the Cordillera Blanca, valleys bordering the Conococha Plain, the Cordillera Huayhuash, and valleys bordering the Junín Plain and Lake Junín. Numbers in parentheses refer to site numbers in Fig. 1

Fig. 4) in central Peru. Glacierised peaks above the valleys include Nevado Palcaraju (6274 m a.s.l.), Nevado Chinchey (6222 m a.s.l.) and Nevado Huantsán (6395 m a.s.l.). Modern glacier limits above the valleys are ~4700–5000 m a.s.l.

Farber *et al.* (2005) built on earlier work by Rodbell (1993a,b), who mapped and classified moraines into four groups (Manachaque, Laguna Baja, Rurec and Cojup, in order of increasing age) based on morphology, weathering features and minimum ages from radiocarbon dating (Fig. 2). Farber *et al.* (2005) used bracketing radiocarbon dates for a Lateglacial moraine in the nearby Rio Negro Valley as age control for exposure ages from boulders on the same moraine. Ages discussed herein are ^{10}Be ages calculated without erosion, with geomagnetic correction, and using altitude and latitude corrections based on Stone (2000); Farber *et al.* (2005) discussed their age calculation methods at length. The ages reported by Farber *et al.* (2005) are comparable to ages calculated using the time-dependent Lal (1991)/Stone (2000) scaling method of the CRONUS Calculator. Raw data are included as supporting information Table 1.

Remnants of the outermost (Cojup) moraines are preserved down to altitudes below 3400 m a.s.l. Rurec moraines are large and extend down to altitudes between 3400 and 3800 m a.s.l. Laguna Baja moraines lie within the Rurec moraine loops and descend to between 3800 and 4000 m a.s.l. Manachaque moraines are generally found above 4000 m a.s.l.

Ages on the outermost Cojup moraines in Quebradas Cojup and Qeshque ranged from ca. 440 ka to 76 ka, with concentrations of ages around 125 ka, 225 ka and 440 ka (Farber *et al.*, 2005). Farber *et al.* (2005) provided a detailed discussion of considerations related to boulder erosion and concluded that erosion was likely minimal. Given the distribution of ages on the Cojup moraines, they favoured the interpretation that the Cojup moraines represent compound features resulting from multiple advances rather than a degrading feature resulting from a single advance.

Ages on the Rurec and Laguna Baja moraines fell within MIS 3–2: ca. 30–20 ka for the Rurec moraines and ca. 18–16 ka for the Laguna Baja moraines (Farber *et al.*, 2005). Farber *et al.* (2005) argued that the oldest age on each group of moraines was the best representation of the actual deposition age: ca. 30 ka for the Rurec advance and ca. 16.5 ka for the Laguna Baja advance. Farber *et al.* interpreted the Rurec advance as the local LGM and the Laguna Baja advance as a stillstand or readvance.

Ages on the Manachaque moraines in Quebrada Uquian were Lateglacial to earliest Holocene: ca. 13.2–10.4 ka (Farber *et al.*, 2005). The same moraines were dated by radiocarbon to between $13\,170 \pm 100$ and $12\,920 \pm 100$ cal. a BP, and the stratigraphy exposed at this locality suggests rapid retreat immediately after deposition of these moraines (Rodbell and Seltzer, 2000).

Farber *et al.* (2005) correlated the Rurec and Laguna Baja moraines to Group C and B moraines, respectively, in the Junín region (Smith *et al.*, 2005b,c), approximately 200 km to the south-east. Farber *et al.* (2005) interpreted the correlation between the Cordillera Blanca and Junín chronologies as evidence for a synchronous regional local LGM and Lateglacial readvance/stillstand, and suggested that the regional Andean local LGM was synchronous, within error, with some recent exposure ages for the LGM of the Northern Hemisphere (Balco *et al.*, 2002).

Conococha Plain, Nevado Jeulla Rajo, Southern Cordillera Blanca (Fig. 1, Site 7)

The Nevado Jeulla Rajo Massif ($10^{\circ} 00' \text{ S}$, $77^{\circ} 16' \text{ W}$) marks the southern end of the Cordillera Blanca and the Callejon de Huayllas Valley in the central Peruvian Andes (Fig. 4). Lake Conococha and the Conococha Plain (~4050 m a.s.l.) border the western side of the massif, which has peak altitudes of

~5600 m a.s.l. and currently hosts a number of small glaciers. Several sets of large lateral moraines extend onto the Conococha Plain from the Jeullesh Valley. Multiple smaller end moraines lie upvalley, closer to the active ice margin. The largest pair of lateral moraines cross-cuts another pair from Jeullesh Valley and a pair from Quenua Ragra Valley to the east. A diamicton containing striated boulders is exposed in stream channels that cross the Conococha Plain downvalley from the large lateral moraines, suggesting that at least one older glaciation was more extensive than any of the advances that deposited the lateral moraines (Smith *et al.*, 2007).

Preliminary results from surface exposure dating using cosmogenic ^{10}Be indicate that the large lateral moraines that extend onto the Conococha Plain from Jeullesh Valley are local LGM and younger (Smith *et al.*, 2007). Moraines on the western side of the Nevado Jeulla Rajo Massif may thus be contemporaneous with the Rurec and Laguna Baja moraines (local LGM and younger) preserved in valleys in the Cordillera Blanca to the north (Rodbell, 1993a; Farber *et al.*, 2005) and significantly younger than the morphologically similar Group D moraines (pre-local LGM) preserved on the Junín Plain to the south (Smith *et al.*, 2005b,c).

Cordillera Huayhuash (Fig. 1, Site 8)

The Cordillera Huayhuash ($10^{\circ} 15' \text{ S}$, $76^{\circ} 50' \text{ W}$) is a glaciated massif within the Eastern Cordillera which lies between the southern end of the Cordillera Blanca and the northern end of the Junín Plain (Fig. 4). Peak altitudes exceed 6000 m a.s.l. and include Yerupaja (6617 m), Yerupaja Sur (6515 m) and Siula Grande (6344 m). The Cordillera Huayhuash is oriented generally north–south, with an east–west spur off the western side. More than a dozen glaciers are mapped along the crest of the range (Peaks & Places Publishing, 2004) and modern ELAs (1986–2005) have been estimated at 4941–5291 m a.s.l. (McFadden *et al.*, 2006). Valleys on the eastern side of the Cordillera typically have shallower gradients than those on the western side.

An extensive field study of the Cordillera Huayhuash (2003–2005) by Hall *et al.* (2004, 2006) and Ramage *et al.* (2004) involved surface exposure dating using cosmogenic ^{10}Be , radiocarbon dating, sediment coring and estimation of changes in ELAs through time. Preliminary results from surface exposure dating of bedrock and boulders on moraines in three valleys suggest that pre-local LGM moraines have not been preserved in the Cordillera Huayhuash. Moraine ages are predominantly Lateglacial and early Holocene (Hall *et al.*, 2006). Palaeo-ELA reconstructions suggest that valley orientation and morphology exerted a strong influence on glacial extent, resulting in differences east and west of the drainage divide (Ramage *et al.*, 2004).

Junín Plain (Fig. 1, Site 9)

Smith *et al.* (2005b,c) used surface exposure dating with CRNs (^{10}Be and ^{26}Al) to date 140 boulders on moraines in valleys bordering the eastern edge of the Junín Plain (11° S , 76° W) in central Peru (Fig. 4). The resulting chronology spans multiple glacial cycles and includes exposure ages greater than 1 Ma, suggesting that long-term rates of boulder erosion have been very low ($\sim 0.3 \text{ m Ma}^{-1}$). Recalculated ages for the Junín valleys are included as supporting information Table 5.

The Junín Plain lies between the Eastern and Western Cordillera of the Andes at an altitude of $\sim 4100 \text{ m a.s.l.}$ The plain is dominated by Lake Junín, a large ($\sim 300 \text{ km}^2$), shallow ($\sim 15 \text{ m}$) lake that was cored by Seltzer *et al.* (2000) in 1996. The Junín region receives approximately 800–900 mm of precipi-

tation annually, most of it during the austral summer (NOAA Data Rescue, 2007).

Smith *et al.* (2005b,c) worked in four valleys in the Eastern Cordillera, three of which face west toward the Junín Plain and one of which faces east toward the Amazon Basin (Fig. 5). The three west-facing valleys (Alcacochoa, Antacocha and Calcalcocha) all have large lateral moraines at their lower ends, smaller end moraines about halfway up the valley length, and a moraine-dammed lake. The west-facing valleys range in length from about 10 to 14 km and have relatively gentle gradients ($\sim 2\text{--}3\%$). The east-facing valley (Collpa) has a large left-lateral moraine but no lake, and is shorter ($\sim 3 \text{ km}$) and steeper ($\sim 9\%$) than the west-facing valleys. Headwall peak altitudes for the four valleys are $\sim 4600\text{--}4850 \text{ m a.s.l.}$ All of the valleys are currently ice-free.

Based on surface exposure ages and geomorphic setting, Smith *et al.* (2005b,c) divided the moraines into four groups (A through D, in order of increasing age and distance downvalley; Figs. 2 and 5). Ages discussed herein are ^{10}Be ages calculated without erosion and with geomagnetic correction (for ages $< 800 \text{ ka}$), using altitude and latitude corrections based on Stone (2000), in the manner of Farber *et al.* (2005). The ages in Smith *et al.* (2005c), which were recalculated with a revised geomagnetic correction, are comparable to ages calculated using the time-dependent Lal (1991)/Stone (2000) scaling method of the CRONUS Calculator (Fig. 5). Raw data are included as supporting information Table 2.

The oldest moraines, Group D, are large lateral moraines that are plastered onto the bedrock walls of the west-facing valleys and extend beyond the valley walls onto the Junín Plain. Without including erosion in age calculations, Group D ages are typically $> 150 \text{ ka}$ (up to ca. 1400 ka), with concentrations of ages around 175–225 and 340–440 ka (Smith *et al.*, 2005c).

Group C and B moraines include the end moraines located approximately midway up the lengths of the west-facing valleys and the left-lateral moraine in the east-facing valley. Exposure ages on the lowest end moraines (Group C) typically range from ca. 31 ka to 21 ka (ca. 24–21 ka in the east-facing valley), while ages on the upper end moraines (Group B) typically range from ca. 19 to 15 ka. The youngest ages (Group A; typically ca. 12–14 ka) were obtained from ground moraine and bedrock in the upper reaches of the longest west-facing valley (Alcacochoa).

Smith *et al.* (2005b,c) concluded that Group C moraines represented the local LGM advance in the Junín valleys and postulated an early local LGM relative to global ice volume records. The Group C moraines ages are consistent with the timing of the local LGM as interpreted from the sediment record from Lake Junín (Seltzer *et al.*, 2000, 2002). Smith *et al.* (2005b,c) interpreted the Group B moraines as a Lateglacial readvance or stillstand that was followed by relatively rapid deglaciation, which is also consistent with the Lake Junín sediment record (Seltzer *et al.*, 2000).

Ramage *et al.* (2005) estimated the ΔELAs for the local LGM moraines (Group C) as $\sim -220 \text{ m}$ to -550 m , depending on the method used to calculate ELA. The ELAs of the largest advances (Group D) and the local LGM advances (Group C) were not markedly different, which Ramage *et al.* (2005) attributed to the valley hypsometry (specifically, relatively shallow gradients below the base of the headwall) and termination on a high-altitude plateau.

Cordillera Vilcanota and Quelccaya Ice Cap (Fig. 1, Sites 10 and 11)

The studies of Mercer and Palacios (1977), Mercer (1982, 1984), Goodman *et al.* (2001) and Mark *et al.* (2002) in and

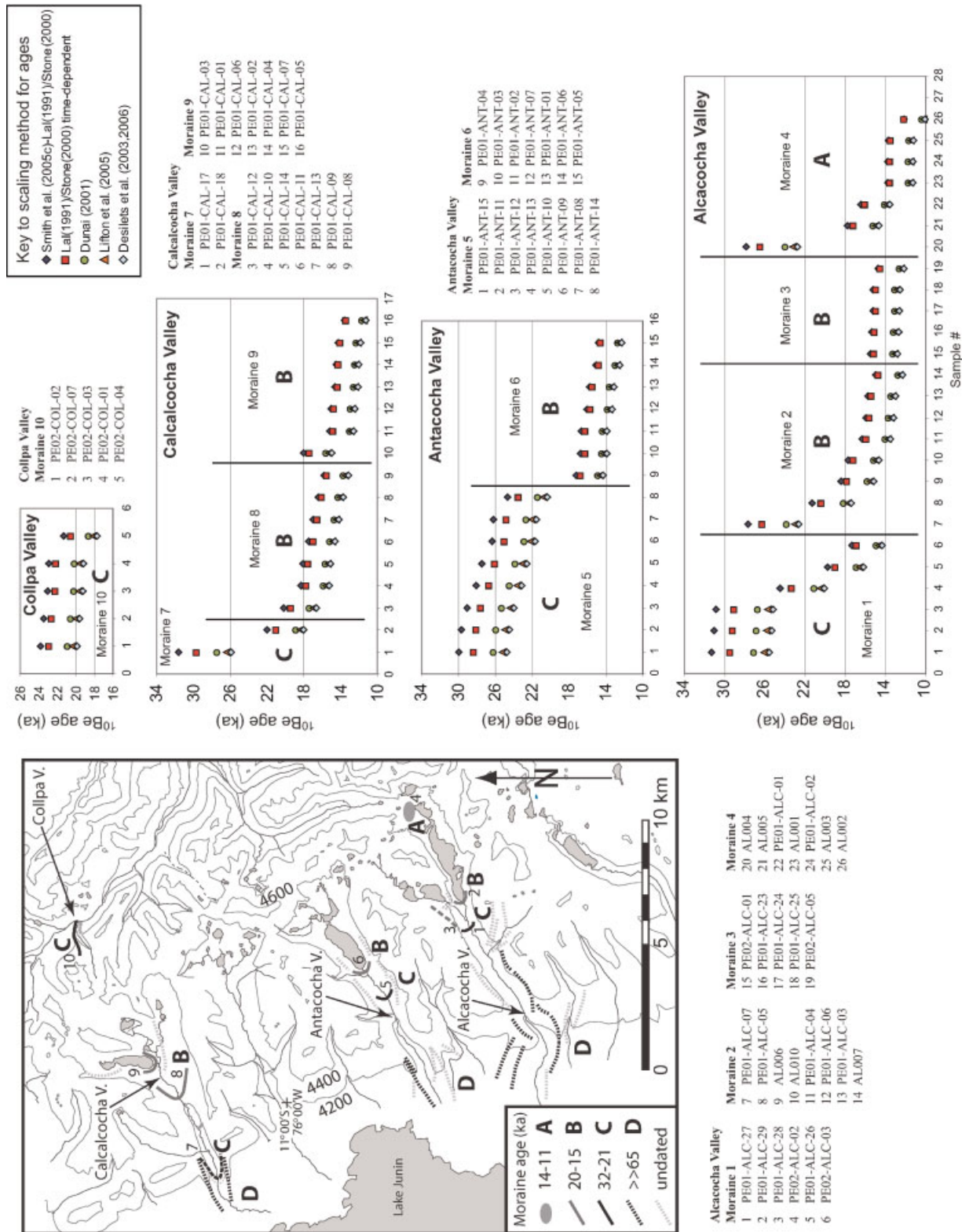


Figure 5 Map of the Junin valleys (Smith *et al.*, 2005b,c) and plots showing the relationship between published CRN ages (Groups C, B and A) and those calculated by the CRONUS Calculator (main calculator v. 2.1) using four different altitude and latitude scaling methods. Ages designated as 'Smith *et al.* (2005c)-La(1991)/Stone(2000)' are the published ages from Smith *et al.* (2005c); a MATLAB subroutine calculated atmospheric pressures and the program used those instead of altitude and standard atmospheric pressure in an effort to account for the temperature inversion in the tropical Andes. The other ages were calculated by the CRONUS Calculator using the sample altitude; ages would be slightly older if calculated using the MATLAB-derived pressure. Recalculated ages for the Junin valleys are included as supporting information Table 5. This figure is available in colour online at www.interscience.wiley.com/journal/jqs

around the Cordillera Vilcanota and the Quelccaya Ice Cap provide considerable radiocarbon age control for late Quaternary glaciation in this region. Several salient points are summarised here.

The outermost moraines at ~3600 m a.s.l. in the Upismayo Valley of the Cordillera Vilcanota are older than $41\,520 \pm 4430$ ^{14}C a BP, which is the basal age of a 10 m thick peat layer located upvalley at 4450 m a.s.l. (Goodman *et al.*, 2001). A sample from the upper part of the same peat layer provided a maximum-limiting age of $13\,880 \pm 150$ ^{14}C a BP (ca. 16.7 cal. ka BP) for a group of seven nested moraines located farther upvalley (Goodman *et al.*, 2001).

Results from sediment coring provide minimum-limiting dates for deglaciation from the local LGM in the Cordillera Vilcanota. In Laguna Casercocha (4010 m a.s.l. on the northwestern side of the Cordillera in a tributary to the Upismayo Valley), organic material overlying glacial silts in a sediment core yielded a radiocarbon age of $15\,640 \pm 100$ ^{14}C a BP (ca. 18.5 cal. ka BP), indicating a transition from glacial to non-glacial sedimentation beginning shortly after 20 cal. ka BP (Goodman *et al.*, 2001). In moraine-dammed Laguna Comercocha (4580 m a.s.l. approximately 6 km east of the Upismayo Valley), basal lacustrine organic material dated to $14\,00 \pm 220$ ^{14}C a BP (ca. 17.4 cal. ka BP; Goodman *et al.*, 2001).

Mercer and Palacios (1977) identified three moraine belts in the Huancané Valley on the west side of the Quelccaya Ice Cap and obtained a minimum-limiting age of $12\,240 \pm 170$ ^{14}C a BP (ca. 14.3 cal. ka BP) at 4750 m a.s.l. for the outermost belt (Huancané III). Additional minimum-limiting ages (e.g., Rodbell and Seltzer, 2000; Goodman *et al.*, 2001; Mark *et al.*, 2002) have been published for the Huancané moraines, but no maximum-limiting ages have been reported. Kelly and Thompson (2004) presented preliminary surface exposure ages (^{10}Be) from moraines along the margins of the Quelccaya Ice Cap. The ages were generally Lateglacial to early Holocene.

Bolivia

Jordan (1998) estimated that glaciers covered more than 560 km² in the section of the Bolivian Andes lying north of latitude 18° 23' S, with most of the glaciers located in the Cordilleras Apolobamba, Real and Quimsa Cruz, and on Nevado Santa Vera Cruz (Fig. 6). The Bolivian Andes become increasingly arid toward the south, with the result that many peaks that exceed the altitude of the 0°C isotherm in southern Bolivia remain unglaciated for lack of precipitation (Jordan, 1998). ELAs in southern Bolivia have been estimated at 5100–5800 m a.s.l. (Klein *et al.*, 1999). As in Peru, the presence of moraines and other glacial deposits in regions of the Bolivian Andes that currently have either no glaciers or glaciers terminating far upvalley of moraines indicate that glacier coverage has been considerably greater in the past (e.g. Mark *et al.*, 2004; Smith *et al.*, 2005b).

Baker *et al.* (2001a) and Seltzer *et al.* (2002) reconstructed a 25 ka precipitation history of the Altiplano from sedimentological, palaeobiotic and isotopic changes in long sediment cores from Lake Titicaca (16–17.5° S, 68.5–70° W, 3810 m a.s.l.). Baker *et al.* (2001a) and Seltzer *et al.* (2002) interpreted changes in the diatom population and magnetic susceptibility of sediments to indicate that the local LGM was a time of wet climatic conditions and that deglaciation occurred ca. 21–19.5 ka. The wet climatic conditions lasted until ca. 15 ka, followed by alternating periods of dry (ca. 15–13 ka, ca. 11.5–10 ka) and wet conditions (ca. 13–11.5 ka) that continued into the Holocene (Baker *et al.*, 2001a).

The repeated filling and emptying of large palaeolakes on the Bolivian Altiplano was recognised early on as evidence of different climatic regimes (e.g. Servant and Fontes, 1978; Minchin, 1882). Coring in the Salar de Uyuni provided a 170 000-year history of alternating pluvial and arid periods in which large lakes filled basins such as the Salar de Uyuni, then dried and were replaced by salt pans (Baker *et al.*, 2001b; Fritz *et al.*, 2004). Most recently, Plazcek *et al.* (2006) refined the history of palaeolakes on the Altiplano in southern Bolivia (18–22° S) using U/Th and radiocarbon dating of lake shoreline deposits.

Deep-lake cycles filled basins on the southern Altiplano from 120–98 ka (Ouki phase) and 18.1–14.1 ka (Tauca phase), with shallow-lake cycles 95–80 ka, ca. 46 ka and 24–20.5 ka, followed by a minor lake cycle (Coipasa) 13–11 ka (Plazcek *et al.*, 2006). The Tauca deep-lake cycle produced the deepest (~140 m) and largest lake of the past 120 ka, with a high stand 16.4–14.1 ka. Plazcek *et al.* (2006) concluded that the previously named 'Minchin' lake phase (ca. 50–28 ka; e.g. Servant and Fontes, 1978) probably included several different lake cycles and proposed that the name be dropped. Plazcek *et al.* (2006) suggested a link between wet phases on the Altiplano and Pacific SST gradients, noting the coincidence of the Tauca deep-lake cycle with periods of intense upwelling in the eastern tropical Pacific that are indicative of strong La Niña conditions of ENSO.

Mark *et al.* (2004) provided an overview of chronologies based on radiocarbon dating and presented ArcView-formatted digital maps of glacier limits in Bolivia. Smith *et al.* (2005a) summarised studies from seven locations in the Bolivian Andes in their analysis of LGM snowlines. We will not revisit previous summaries (Smith *et al.*, 2005a; Mark *et al.*, 2004) of studies in the Cordillera Apolobamba (Lauer and Rafiqpoor, 1986), Río Palcoco (Argollo, 1980, 1982; Seltzer, 1992), Cordillera Quimsa Cruz (Müller, 1985) and Río Kollpaña (Servant *et al.*, 1981; Servant and Fontes, 1984; Gouze *et al.*, 1986).

Here we discuss three valleys for which surface exposure dating of glacial deposits has been used to develop glacial chronologies, one valley in which radiocarbon dating provides a minimum-limiting age for the local LGM, and two sites from which ice cores have been retrieved (Fig. 1, Table 1).

San Francisco Valley, Cordillera Real (Fig. 1, Site 12)

Zech *et al.* (2007) used surface exposure dating with cosmogenic ^{10}Be to date 12 boulders distributed across six moraines in San Francisco Valley in western Bolivia (~16° 00' S, 68° 32' W). San Francisco Valley is located approximately 18 km east of Lake Titicaca in the Cordillera Real (Fig. 5). As calculated using the scaling method of Lifton *et al.* (2005), the oldest exposure ages obtained were 24.1 and 22.6 ka, which came from the outermost moraine that Zech *et al.* (2007) sampled. Based on these results, Zech *et al.* (2007) concluded that the glacial advance that deposited the moraine was synchronous with the global LGM; they noted, however, that recalculation using the scaling method of Stone (2000) yielded ages closer to 30 ka for the same samples. For further discussion of this site and results of calculations using the CRONUS Calculator, see Zech *et al.* (2008). Raw data are included as supporting information Table 3.

Argollo (1980, 1982) reported radiocarbon ages of ca. 33–36 ^{14}C ka BP on samples described as peat reworked by a glacial deposit in the San Francisco Valley. He interpreted the ages as maximum-limiting ages for the glacial advance that disrupted the peat. Smith *et al.* (2005a) discussed the muddled publication history of the ages and suggested that additional

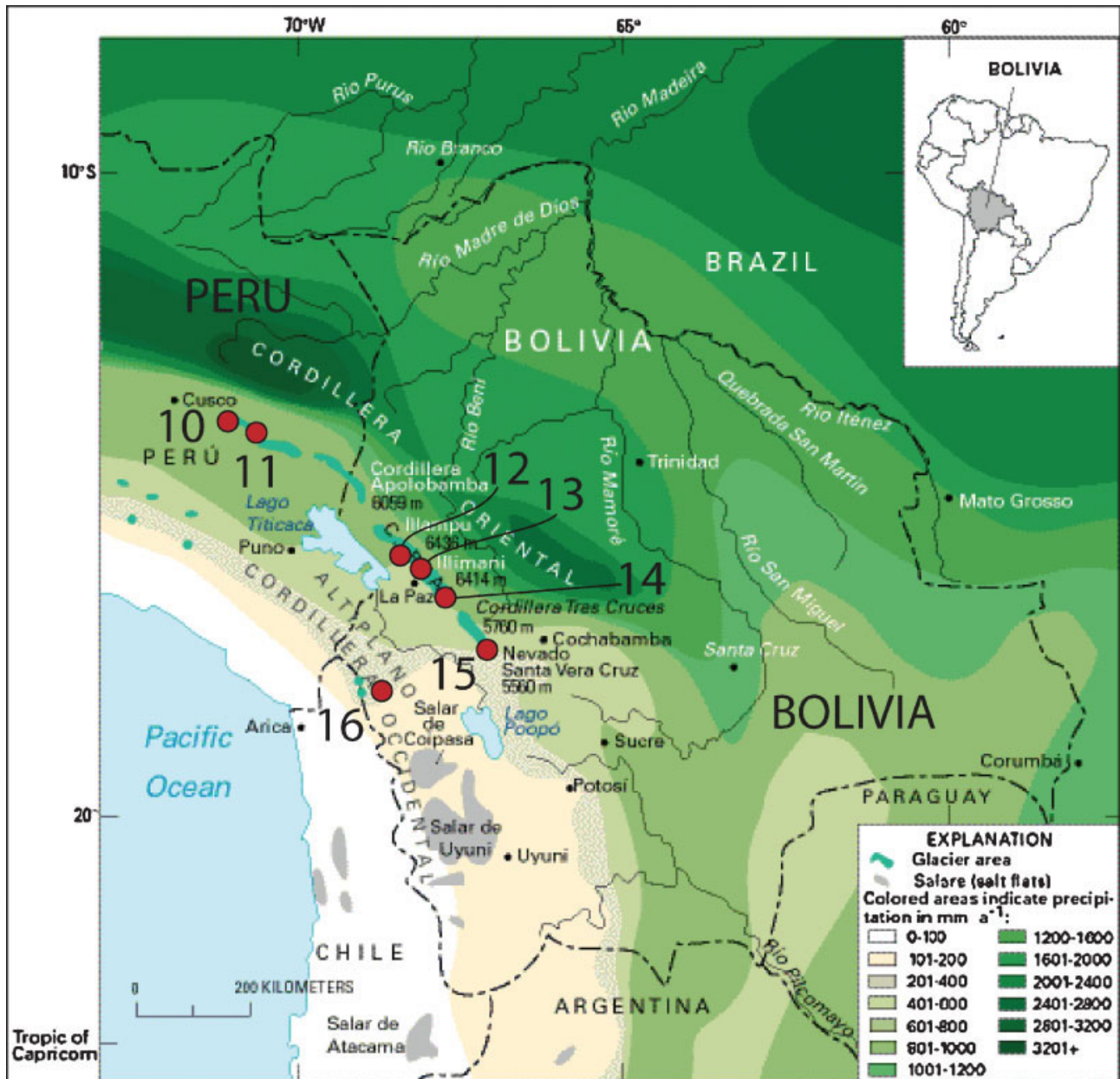


Figure 6 Sites in southern Peru and Bolivia: (10) Cordillera Vilcanota; (11) Quelccaya Ice Cap; (12) San Francisco Valley; (13) Milluni and Zongo Valleys; (14) Nevado Illimani; (15) Laguna Kollpa Kkota; (16) Nevado Sajama. Base map showing geographic features, political boundaries and annual precipitation amounts from Jordan (1998); used by permission of US Geological Survey (open-file report)

sampling of the peat deposits would increase confidence in the dates as definitive maximum-limiting ages.

Milluni and Zongo Valleys, Cordillera Real (Fig. 1, Site 13)

Smith *et al.* (2005b) used surface exposure dating with CRNs (^{10}Be and ^{26}Al) to date 42 boulders on moraines in the Milluni and Zongo Valleys in western Bolivia ($\sim 16^{\circ} 16' \text{ S}$, $68^{\circ} 08' \text{ W}$; Fig. 7). The valleys are located in the Cordillera Real $\sim 35\text{--}40$ km east of Lake Titicaca, $\sim 55\text{--}60$ km south-east of San Francisco Valley, and between ~ 10 and 30 km north of La Paz. Nevado Huayna Potosí (6088 m a.s.l.) forms the headwall for both valleys and hosts the Zongo Glacier (Wagnon *et al.*, 1999). The divide separating the two valleys has an altitude of ~ 4800 m a.s.l. The Milluni Valley is south- and west-facing and

terminates on the Altiplano (~ 3800 m a.s.l.), whereas the Zongo Valley runs approximately north from the divide and descends into the Amazon Basin. A precipitation gradient exists along the lengths of the valleys, with mean annual precipitation decreasing from the northerly end of Zongo Valley to the southerly end of Milluni Valley (Kessler and Monheim, 1968; Safran *et al.*, 2005).

The valleys are morphologically distinct from each other: Milluni Valley is relatively broad and gently sloping, while Zongo Valley is narrower, deeply incised and steeper (Safran *et al.*, 2005; Smith *et al.*, 2005b). Milluni Valley has large double-crested lateral moraines at its lower end and multiple end moraines and moraine-dammed lakes in its upper reaches. The double-crested moraines descend to ~ 4300 m a.s.l. Boulders on Milluni Valley moraines are typically < 1 m high. Zongo Valley has numerous rocky moraine loops in its axis and at the intersections with tributary valleys; large (> 2 m high)

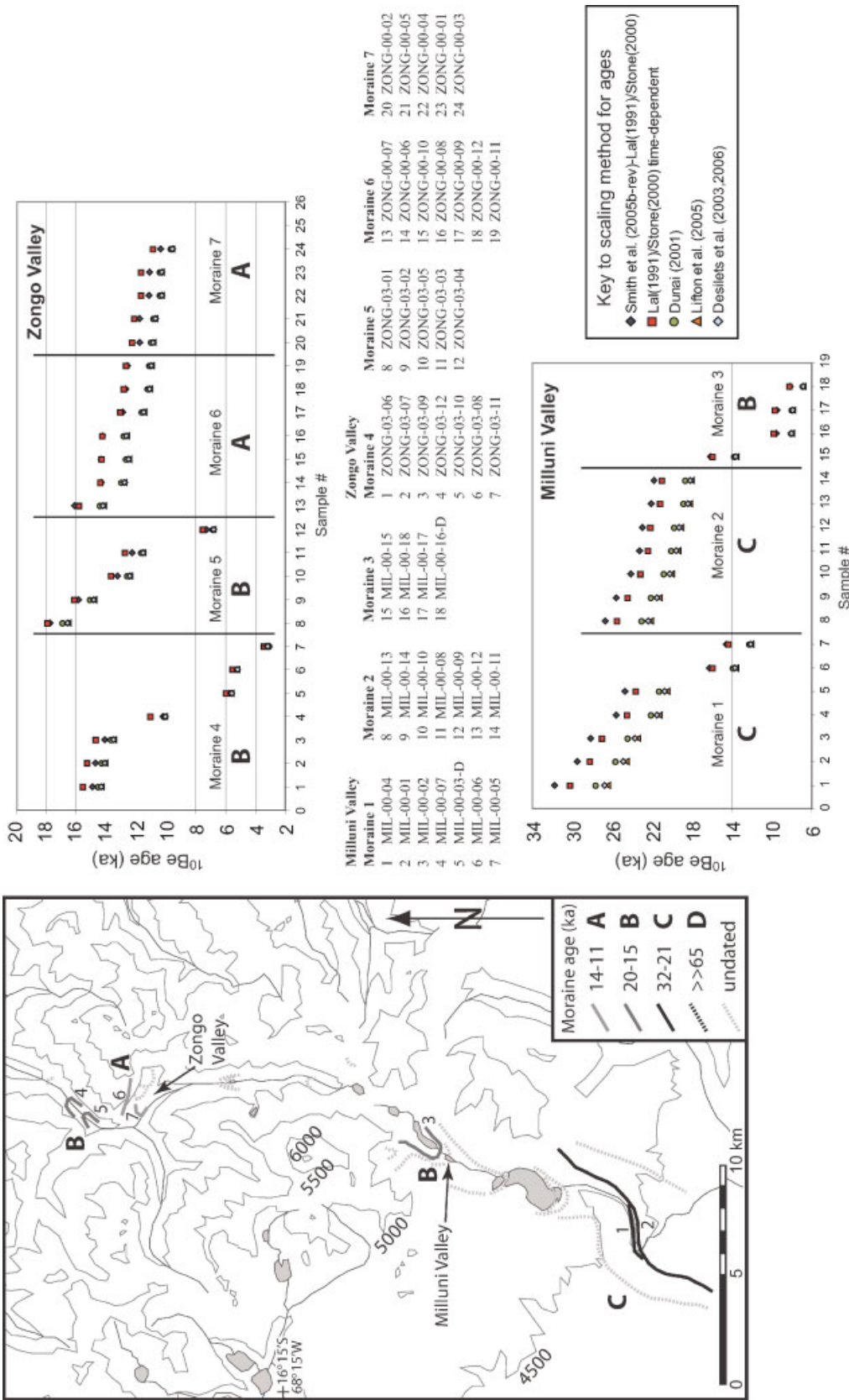


Figure 7 Map of Milluni and Zongo Valleys (Smith et al., 2005b) and plots showing the relationship between CRN ages (Groups C, B and A) calculated using the MATLAB program of Farber et al. (2005) and those calculated by the CRONUS Calculator (main calculator v. 2.1) using four different altitude and latitude scaling methods. Ages designated as 'Smith et al. (2005b-rev)-Lal(1991)/Stone(2000)' were recalculated using a revised geomagnetic correction (and MATLAB-derived pressures) and are younger than the published ages from Smith et al. (2005b). The other ages were calculated by the CRONUS Calculator using the sample altitude; ages would be slightly older if calculated using the MATLAB-derived pressure. Recalculated ages for Milluni and Zongo Valleys are included as supporting information Table 5. This figure is available in colour online at www.interscience.wiley.com/journal/jqs

boulders are abundant on moraines. Several small moraine-dammed lakes are present along the length of Zongo Valley (Abbott *et al.*, 1997).

The Milluni and Zongo Valleys are also lithologically distinct from each other. Bedrock making up Nevado Huayna Potosí and underlying the Milluni Valley is the Cordillera Real granite, which hosts economic tin deposits (Lehmann, 1985). The Zongo Valley cuts down into metasediments (Safran *et al.*, 2005).

Smith *et al.* (2005b) dated boulders on two moraines in Milluni Valley and four moraines in Zongo Valley (Figs. 2 and 7). Ages discussed herein are ^{10}Be ages calculated without erosion and with geomagnetic correction, using altitude and latitude corrections based on Stone (2000), in the manner of Farber *et al.* (2005); the ages were recalculated with a revised geomagnetic correction and are slightly younger than ages published in Smith *et al.* (2005b). The recalculated ages are comparable to those calculated using the time-dependent Lal (1991)/Stone (2000) scaling method of the CRONUS Calculator. Raw data are included as supporting information Table 4. Recalculated ages for Milluni and Zongo Valleys are included as supporting information Table 5.

Ages on the large double-crested, left-lateral moraine at the lower end of Milluni Valley (~4595 m a.s.l.) ranged from ca. 32 to ca. 14.5 ka, with the seven ages on the inner crest (Fig. 7, Moraine 1) falling between 31.8 and 14.5 ka and the seven ages on the outer crest (Fig. 7, Moraine 2) falling between 26.7 and 21.8 ka. Four ages on a moraine loop farther upvalley (~4640 m a.s.l.) were 16.2, 9.5, 9.5 and 8.2 ka. Smith *et al.* (2005b) correlated the large lateral moraines with Group C moraines (local LGM moraines) and the younger moraine loop with Group B moraines (Lateglacial readvance/stillstand) from the Junín Plain. The Group B correlation is somewhat tenuous, given the number of moraine loops upvalley from the large lateral moraines in Milluni Valley and the bimodal nature of the ages. Unlike the west-facing Junín valleys (Smith *et al.*, 2005b,c), Milluni Valley does not have clear evidence of moraines that predate the local LGM.

Ages on four moraines in Zongo Valley were all younger than 18 ka, even at altitudes of ~3400 m a.s.l. (more than 1 km lower than moraines of comparable age in Milluni Valley). Smith *et al.* (2005b) correlated the upper two moraines with Group A and the lower two with Group B from the Junín region. Smith *et al.* (2005b) speculated that shading by valley walls, debris cover and rockfalls in Zongo Valley combined with greater orographic precipitation to maintain ice to far lower altitudes than in wider, drier Milluni Valley. Smith *et al.* (2005b) estimated the ELA depression for the local LGM at 300–600 m in Milluni Valley and 800–1000 m in Zongo Valley.

Nevado Illimani (Fig. 1, Site 14)

In 1999, a French–Swiss–Bolivian team obtained an ice core drilled to bedrock (136.7 m) from the ice cap atop Nevado Illimani (16° 37' S, 67° 46' W, 6350 m a.s.l.) in the Eastern Cordillera of the Bolivian Andes (Ramirez *et al.*, 2003). Basal ice temperature was -8.4°C and the age of the basal ice was estimated to be ca. 18 ka (Knüsel *et al.*, 2003; Ramirez *et al.*, 2003). Ramirez *et al.* (2003) interpreted isotope depletion at the base of the core as evidence of wet conditions during the last glacial stage (LGS). They largely dismissed palaeoclimatic explanations for elevated dust concentrations as indicators of dry conditions during the LGS in favour of mechanical explanations involving localised ice–bedrock interactions during glacier formation (Ramirez *et al.*, 2003).

Ramirez *et al.* (2003) cited evidence of wet conditions during the LGS from the palaeolake record on the Altiplano (Baker *et al.*, 2001a,b), the diatom record in Lake Junín (Seltzer *et al.*, 2002) and the pollen record of vegetation changes in the Amazon Basin (Colinvaux *et al.*, 2000) as support for their interpretation of a wet, cold LGS followed by warming ca. 20 ka. Ramirez *et al.* (2003) interpreted the change in isotopic values at 17–15 ka in the Illimani ice core as resulting from a shift to drier conditions, rather than a response to warming that began some 4000 a earlier. Ramirez *et al.* (2003) used a modern analogue approach to estimate that the Amazon and Altiplano may have been as much as 20% wetter during the LGS than today, if one were to assume that the entire isotopic shift could be attributed to precipitation increase.

Laguna Kollpa Kkota (Fig. 1, Site 15)

Laguna Kollpa Kkota (17° 26' S, 67° 08' W, Fig. 1) is a moraine-dammed cirque lake located at 4400 m a.s.l. on the western slope of the Eastern Cordillera of the Bolivian Andes (Seltzer, 1994b). No moraines are present between the headwall of the Kollpa Kkota Valley (4560 m a.s.l.) and the lake, or below the lake-damming moraine (<4400 m a.s.l.). Late Pleistocene (ca. 12–14 ka BP) snowline, as extrapolated from the Cordillera Quimsa Cruz (60 km to the north), was ~4620 m a.s.l. (Müller, 1985), which is above the headwall of the Kollpa Kkota Valley. Modern snowline, also extrapolated from the Cordillera Quimsa Cruz (Müller, 1985), is ~5100 m a.s.l.

Seltzer (1994b) dated basal lacustrine sediments from cores collected in Laguna Kollpa Kkota. He interpreted radiocarbon ages of ca. 17.6 to ca. 17.7 ^{14}C ka BP (>20 cal. ka BP) as minimum dates for deglaciation from the glacial advance that deposited the lake-damming moraine. Seltzer (1994b) concluded that the coring site had not been glaciated since at least 20 cal. ka BP, which may be the closest limiting age for the local LGM in Bolivia. Seltzer calculated an ELA depression of ~600 m for the glaciation that produced the lake-damming moraine in Kollpa Kkota Valley.

Nevado Sajama (Fig. 1, Site 16)

In 1997, Thompson and colleagues retrieved four ice cores from the ice cap atop the extinct volcano Nevado Sajama (6542 m a.s.l.) located at the western edge of the Bolivian Altiplano (18° 06' S, 68° 53' W). Cores C-1 (132.4 m) and C-2 (132.8 m) were drilled to bedrock 3 m apart; the other two cores were shallow (40 m and 4 m). The temperature of the ice at the ice–bedrock contact was -9.5°C , indicating that the basal ice was frozen to the bedrock. Organic material trapped in the ice at 130.8 m in C-1 was radiocarbon dated; calibrated ages from separate labs were $24\,950 \pm 430$ cal. a BP and $24\,020 \pm 140$ cal. a BP (Thompson *et al.*, 1998). Mid to late Holocene radiocarbon ages (<5 cal. ka BP) were obtained above 101 m in C-1. To date the glacial portion of C-1 better, the $\delta^{18}\text{O}$ curve for the 30 m of ice between the 5 cal. ka BP and 25 cal. ka BP ages was compared to the $\delta^{18}\text{O}$ curve from the GISP 2 ice core (Thompson *et al.*, 1998).

Thompson *et al.* (1998) interpreted the multiproxy record from the Sajama ice cores to indicate cold, wet conditions and the presence of palaeolakes on the Altiplano from 25 to 22 ka, a period of drying ca. 22–21 ka, a return to moister conditions and lake-refilling ca. 21–15 ka, warm and dry conditions ca. 15.5–14.3 ka, and a climatic reversal that began at 14 ka and

lasted until ca. 11.5 ka (the Deglacial Climate Reversal, or DCR). Increased dust and other changes in Holocene ice from ca. 9 to ca. 3.4 ka were interpreted to indicate warmer, drier conditions than present (Thompson *et al.*, 1998).

Summary and discussion

Regional patterns in chronologies

The tropical Andes, even in the limited sense in which the term is employed here, encompass a distance of more than 2500 km. Some north–south and east–west variations in glacial history should be expected, given the differences in climate from the relatively moist equatorial Ecuadorian Andes south to the arid Andes of southern Bolivia, and the effects of orographic precipitation gradients from east to west across the crests and intervening plateaus of the range. The many cordilleras of the Andes are not uniform, either: those with the highest peaks typically have the most detailed, and youngest, assemblages of glacial depositional features.

We can make some basic observations about glaciation in the tropical Andes, bearing in mind that the results of four Peruvian studies that used surface exposure dating with CRNs illustrate the difficulty in forming general statements about tropical Andean glaciation.

- Glaciation has been more extensive in the past than it is today, as abundant moraines lying beyond present ice limits testify. In this sense the mountain glaciers of the Andes are like those of much, if not all, of the world.
- In locations in the tropical Andes where numerical dating has been completed, the local LGM was apparently a relatively modest event when compared to older glaciations. Glaciation during MIS 4 does not appear to be well represented in the moraine record.
- Local variability is superimposed on any regional patterns, such that sites separated by ~100 km can have very different glacial records (e.g. Junín Plain and Cordillera Huayhuash, Peru). Local factors such as valley hypsometry and maximum peak elevations appear to control ice extent to a large degree.
- The idea of a quasi-uniform ELA lowering of 900–1000 m throughout the Andes (Porter, 2001) is not borne out by detailed studies; rather, the presence of high plateaus between the Eastern and Western Cordillera limit the altitude to which glaciers in bordering valleys can descend and thus may constrain the drop in ELA to relatively low values (e.g. Junín Plain, ca. 300–600 m). Conversely, valley orientation and morphological characteristics can combine to increase glacier length and probably lower ELA beyond the typical regional value (e.g. Zongo Valley).

An early local LGM (ca. 30 ka) has been inferred for Ecuador (Clapperton *et al.*, 1997), but interpretation of the timing of the local LGM in the Peru and Bolivia depends to some extent on the manner in which CRN ages are calculated. Studies that have used the time-dependent Lal (1991)/Stone (2000) scaling method suggest that the local LGM in Peru (Farber *et al.*, 2005; Smith *et al.* (2005b,c) and Bolivia (Smith 2005b) occurred closer to ca. 30 ka than 21 ka (that is, before the global LGM as inferred from the marine isotope record), whereas a study in Bolivia using an alternative scaling method (Lifton *et al.*, 2005) suggests that the local LGM was closer to ca. 24 ka (Zech *et al.*, 2007). When CRN ages from all of the studies are calculated using the same scaling method, however, the

disparity disappears; the convergence of chronologies suggests regionally coherent climate forcings. Most of the studies indicate that the local LGM was followed by a Lateglacial stillstand or readvance (ca. 18–15 ka in several studies). Evidence for larger pre-local LGM glaciations has been identified in Ecuador (e.g. Clapperton, 1986) and Peru (Farber *et al.*, 2005; Smith *et al.*, 2005b,c), but the timing of those glaciations is not yet sufficiently well constrained to infer regionally coherent signals.

There is little firm evidence for glacial advances in the tropical Andes during the YD climate reversal (Rodbell and Seltzer, 2000). The short duration of this event coupled with the estimated accuracy of CRN ages ($\pm 10\%$) preclude the widespread use of CRN to date moraines definitively to this interval. Radiocarbon-dated evidence for YD glacial advances in Ecuador is controversial (Clapperton and McEwan, 1985; Heine, 1993, 1995; Heine and Heine, 1996; Clapperton *et al.*, 1997), and in Peru the YD appears to correspond to an interval of rapid ice retreat based on radiocarbon-dated ice margin positions in the Cordillera Blanca and in the Cordillera Vilcanota (Rodbell and Seltzer, 2000).

Palaeoclimatic inferences that can be drawn from the chronologies

Glaciers in the tropical Andes, like glaciers everywhere, exist only under certain conditions of temperature, precipitation and solar radiation receipt (Seltzer, 2001). Unlike glaciers at higher latitudes, however, tropical glaciers typically experience both the most ablation and the most accumulation during the wet austral summer (Kaser and Osmaston, 2002). Moreover, the feedbacks to the glacier surface energy budget that have the predominant impact on interannual mass balance are tied to the precipitation regime (Francou *et al.*, 2003). As in the Himalayas (Finkel *et al.*, 2003), glaciers in the tropical Andes might be expected to advance when summer insolation is highest.

Annually, easterly flow and moisture transport are concentrated into the austral summer wet season, while ENSO produces larger-wavelength variations interannually. During the El Niño phase of ENSO, warmer SSTs in the eastern tropical Pacific produce generally warmer, drier conditions, reduced cloudiness and westerly winds over the tropical Central Andes (Garreaud *et al.*, 2003). The opposite occurs during the La Niña phase: cooler, wetter, cloudier conditions prevail over the tropical Central Andes. It is important to note that this high-elevation Andean response is opposite from the El Niño flooding documented in sedimentary cores near the coast (Moy *et al.*, 2002; Rein *et al.*, 2005). As noted above ('Tropical glaciers'), ENSO also tends to have a similar impact on modern Andean glaciers ranging from Ecuador to Bolivia, despite differences in the seasonal dependence and physical mechanisms linking ENSO with mass balance variations (Francou *et al.*, 2004). Global climate models have been used to evaluate changes on even longer timescales (ka) by glacial and orbital forcing. Modelling by Garreaud *et al.* (2003) suggested that glacial conditions in the Northern Hemisphere would have little effect on the amount of moisture reaching the Altiplano, whereas precession-driven changes in insolation would have strong effects. Specifically, lower insolation during austral summer would produce cooling, enhanced westerly flow and drying of the Altiplano relative to present, while higher insolation would have the opposite effect. Thompson *et al.* (2005) proposed that the progressively younger basal ages of ice cores from 18° S to 38° N indicate that ice caps began to grow in

response to precession-driven increases in insolation and convection that resulted in more precipitation in the tropics. Various lines of evidence from Peru and Bolivia (e.g. Baker *et al.*, 2001a; Seltzer *et al.*, 2000, 2002), however, support the hypothesis that the tropical Central Andes were wet during the local LGM and until ca. 16–15 ka, and the level of Lake Titicaca appears to have tracked precessionally driven austral summer insolation over the Holocene (Abbott *et al.*, 1997). Modelling studies by Clement *et al.* (1999) also suggest that long-term variability in the ENSO cycle may, in turn, be tied to precessionally driven insolation in the tropics.

The existence of widespread evidence for a stillstand or readvance after the local LGM in the tropical Andes (ca. 18–15 ka as calculated by Smith *et al.*, 2005c) implies a regional-scale forcing. The coincidence with wet conditions in Lake Junín (Seltzer *et al.*, 2000, 2002) and the high stand of palaeolake Tauca on the Altiplano (Placzek *et al.*, 2006) suggests that the cause may have been a precipitation increase. Placzek *et al.* (2006) speculated that the palaeolake high stand may have been linked to Pacific SST gradients, creating a La Niña effect.

An early local LGM (ca. 32–28 ka) in the tropical Andes would have preceded a palaeolake high stand on the Altiplano by some 4–8 ka (Placzek *et al.*, 2006), suggesting a different climate forcing than the ca. 18–15 ka stillstand/readvance. Austral autumn insolation at 10° S reached a minimum ca. 27 ka and January insolation at 15° S reached a minimum ca. 31 ka (Berger and Loutre, 1991). Lower insolation around the time of the local LGM should have diminished convection, cloudiness and precipitation, thereby decreasing glacier accumulation and increasing ablation by decreasing albedo (Garreaud *et al.*, 2003). Evidence for a local LGM ca. 32–28 ka, however, suggests that the ablation-enhancing effects of low insolation were offset by decreased temperature during a lengthy wet period (Seltzer *et al.*, 2000, 2002). Seltzer *et al.* (2002) attributed deglaciation ca. 22–21 ka to increased mean annual temperature rather than decreased precipitation.

An early local LGM in the tropical Andes would have been approximately contemporaneous with Heinrich event H3 (Rashid *et al.*, 2003). Research in Brazil suggests a connection between Heinrich events and increased moisture in the Amazon Basin. Wang *et al.* (2004) identified periods of speleothem growth (wet periods) in eastern Brazil (10° 10' S, 40° 50' W) during several Heinrich events, but not during H3. Jennerjahn *et al.* (2004), however, found evidence for increased precipitation during the YD and Heinrich events H1–H8 in sediment cores obtained from the continental margin off north-east Brazil (~3–5° S, ~36–37° W).

Baker *et al.* (2001a) proposed a link between below-normal SST in the northern equatorial Atlantic and wet conditions in Amazonia and on the Altiplano. Wang *et al.* (2004) found that speleothem growth was most closely correlated to times of high austral autumn insolation at 10° S; they suggested that wet periods were associated with southward displacement of the ITCZ when the Atlantic SST gradient north of the Equator steepened in response to the presence of continental ice sheets.

Farber *et al.* (2005) suggested that the early local LGM in the tropical Andes was synchronous with the global LGM as dated by Balco *et al.* (2002). As surface exposure dating continues to expand our understanding of the details of the continental ice sheet record, the global LGM may emerge as a more nuanced event that was in part synchronous with the tropical Andean local LGM.

The tropical Andes do not present a consistent pattern of glacial response during the YD climate reversal. The Zongo Valley in Bolivia contains evidence of moraine formation during the YD (Smith *et al.*, 2005b), but may be unusual in this

respect. According to Rodbell and Seltzer (2000), the rapid retreat of glaciers in Peru during the YD was driven by an abrupt reduction in the moisture balance of the tropical Andes. Both the Huascarán (central Peru; ~9° S) and Sajama (SW Bolivia; ~18° S) ice cores span the Lateglacial interval (Thompson *et al.*, 1995, 1998). The Sajama core is especially valuable because it is independently dated and contains a far thicker Lateglacial section (~28 m) than does the Huascarán core (~3 m). The high-resolution $\delta^{18}\text{O}$ record in the Sajama core reveals a pronounced cooling midway through the last deglaciation (the DCR). The DCR began ca. 1000 a prior to the onset of the YD and lasted some 2500 a – nearly twice as long as the YD (Thompson *et al.*, 1998). The Peruvian palaeoglaciers reported by Rodbell and Seltzer (2000) apparently began to advance at about the time of the onset of the DCR, but abruptly retreated midway through the DCR. Because the Sajama $\delta^{18}\text{O}$ record does not reveal a substantial and prolonged warming that could have driven the rapid retreat of the glacier termini, the cause of this retreat was probably a reduction in regional precipitation. An interval of reduced effective moisture midway through the DCR is supported by other independent proxy palaeoclimatic parameters. The difference between the $\delta^{18}\text{O}$ of authigenic calcite in Lake Junín, Peru (~11° S) and Huascarán (~9° S) ice reflects the relative degree of evaporative ^{18}O enrichment of Lake Junín (Seltzer *et al.*, 2000). This time series reveals an increase in relative aridity in the Junín region at about the same time that the Peruvian palaeoglaciers began to retreat rapidly.

Conclusions

The tropical Andes of Ecuador, Peru and Bolivia display ample evidence for multiple glaciations. In general, older glaciations were considerably more extensive than those that occurred during MIS 3–2, at least in some parts of the tropical Andes. Evidence that the largest glaciations preceded the local LGM has been found in Peru (Farber *et al.*, 2005; Smith *et al.*, 2005c) and suggestive evidence for the same has been found in Ecuador (e.g. Clapperton, 1987), but no such evidence has been identified as yet in Bolivia (Smith *et al.*, 2005b). Studies that have included surface exposure dating with CRNs suggest that the local LGM may have been early (ca. 32–28 ka) relative to the global LGM as commonly defined (21 ka), or closer (ca. 24 ka) to the global LGM, depending on the manner in which the CRN ages are calculated. Future refinements in the CRN methodology may resolve these uncertainties. A distinct glacial advance during MIS 4 has not been clearly identified thus far. The tropical Andes show a clear widespread signal for a Lateglacial stillstand or readvance (dated ca. 18–15 ka by several studies), but evidence for a YD advance remains equivocal, especially in Ecuador.

Results of the four Peruvian studies that used surface exposure dating with CRNs illustrate the difficulty in forming general statements about the extent of mountain glaciation in the Andes. Local factors such as valley hypsometry and maximum peak elevations appear to control ice extent to a large degree, such that sites separated by ~100 km (e.g. Cordillera Huayhuash and Junín Plain, Peru) can apparently have vastly different glacial chronologies.

Palaeoclimatic influences on the timing and nature of mountain glaciation in the tropical Andes remain a subject of active research. The unresolved quandary posed by apparently larger tropical terrestrial vs. ocean temperature depressions at the LGM motivated initial research into low-latitude snowline changes (e.g. Mark *et al.*, 2005). However, new chronological

insights into palaeoglacier extents have opened up many more questions about climate controls over time. Possible factors include Pacific SSTs, Atlantic SSTs and precession-driven changes in insolation. No clear single explanation for the timing of glaciation has emerged. The scale of the region is large: the tropical Andes of Ecuador, Peru and Bolivia stretch north–south more than 2500 km. We should expect variability in the climate forcing factors from equatorial Ecuador to the arid southern end of Bolivia. Recent analyses into modern climate patterns throughout the region reveal the extreme local-scale variability that can mask any regional forcings (e.g. Vuille and Keimig, 2004). Similarly, palaeoclimatic inferences made from remnant moraines critically depend on local forcings, making it important to evaluate multiple individual palaeoglaciers and not rely on single ELA estimates for regions.

Other possible influences on the glacial record in the tropical Andes include the ongoing uplift of the range, which is progressively raising peaks and increasing the size of accumulation areas. Peak altitudes exert a strong influence on glaciation past and present; were it not for volcanism, for example, Ecuador would have a far more limited glacial record than it does. Conversely, active volcanism in Ecuador introduces preservation bias into the glacial record by covering or removing glacial deposits during eruptions.

Possibilities for future work in the tropical Andes are numerous. Surface exposure dating with CRNs would probably clarify the controversial glacial record of Ecuador. The groundwork has been laid at various sites discussed above, particularly Chimborazo–Carihuairazo and the Potrerillos Plateau. The Cordillera Oriental in northern Peru has been the subject of detailed studies (e.g. Rodbell, 1991), but lacks numerical dating. The Bolivian Andes present a vast array of opportunities for research. Beyond basic chronology, inverse modelling techniques hold promise for elucidating the range and combination of multiple climate forcings for individual palaeoglaciers (e.g. Kull and Grosjean, 2000; Kull *et al.*, 2002, 2003; Plummer and Phillips, 2003; Fairman, 2006).

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