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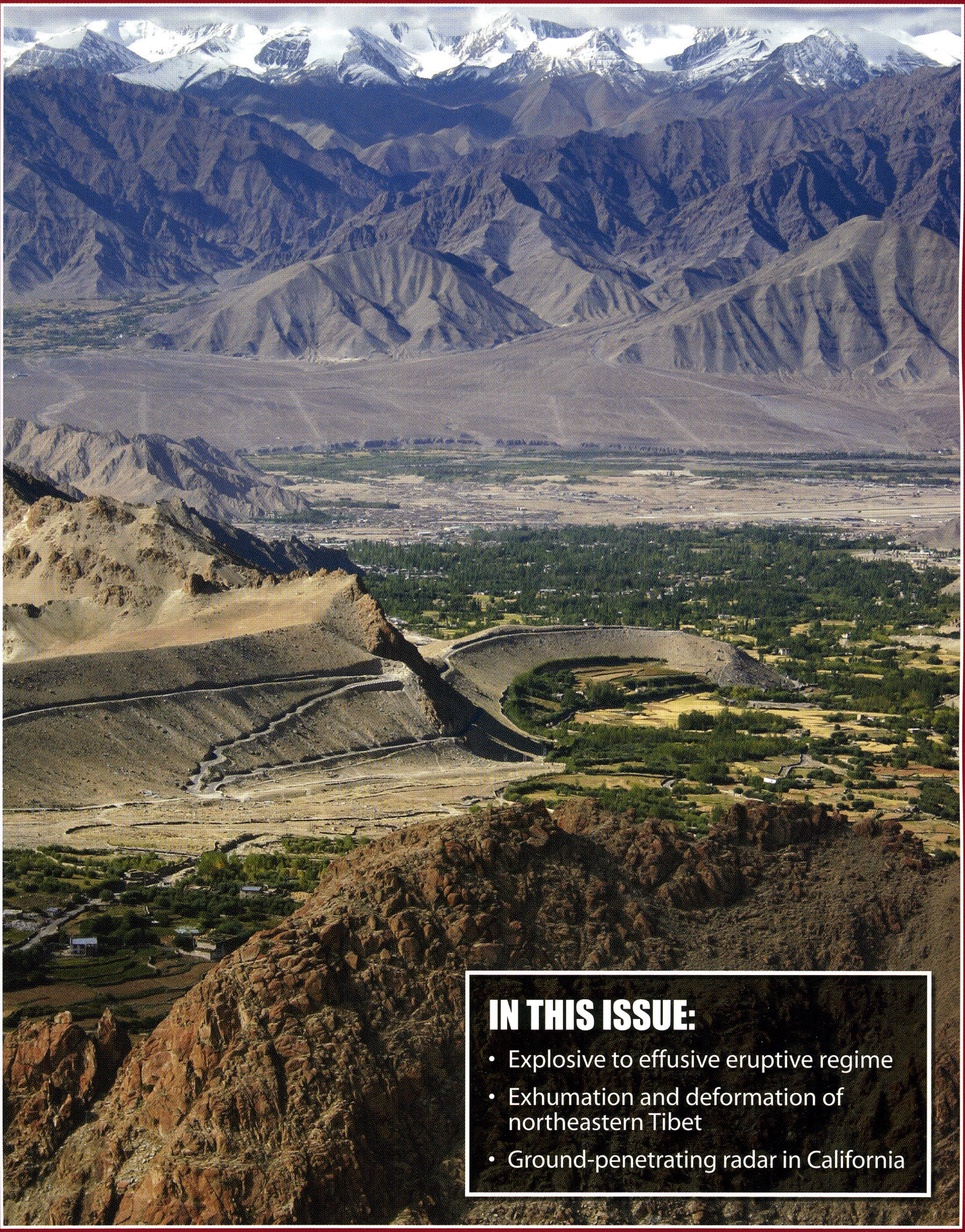

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Terrestrial cosmogenic nuclide surface exposure dating of the oldest glacial successions in the Himalayan orogen: Ladakh Range, northern India

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ABSTRACT

Terrestrial cosmogenic nuclide surface exposure dating of moraine boulders and alluvial fan sediments define the timing of five glacial advances over at least the last five glacial cycles in the Ladakh Range of the Transhimalaya. The glacial stages that have been identified are: the Indus Valley glacial stage, dated at older than 430 ka; the Leh glacial stage occurring in the penultimate glacial cycle or older; the Kar glacial stage, occurring during the early part of the last glacial cycle; the Bazgo glacial stage, at its maximum during the middle of the last glacial cycle; and the early Holocene Khalling glacial stage. The exposure ages of the Indus Valley moraines are the oldest observed to date throughout the Himalayan orogen. We observe a pattern of progressively more restricted glaciation during the last five glacial cycles, likely indicating a progressive reduction in the moisture supply necessary to sustain glaciation. A possible explanation is that uplift of Himalayan ranges to the south and/or of the Karakoram Mountains to the west of the region may have effectively blocked moisture supply by the south Asian summer monsoon and mid-latitude westerlies, respectively. Alternatively, this pattern of glaciation may reflect a trend of progressively less extensive glaciation in mountain regions that has been observed globally throughout the Pleistocene.

Keywords: terrestrial cosmogenic nuclide, surface exposure dating, Transhimalaya, glacial geology, geochronology, Ladakh.

INTRODUCTION

Many of the late Quaternary glacial successions within the Himalaya and Tibet can now be successfully dated using terrestrial cosmogenic radionuclide (TCN) surface exposure and optically stimulated luminescence (OSL) techniques. These dates are beginning to allow a quantitative investigation of the relative timing between glacial cycles in different regions and an examination of the connections between climatic oscillations and glaciation (Sharma and Owen, 1996; Richards et al., 2000a, 2000b; Phillips et al., 2000; Owen et al., 2001, 2002a, 2002b, 2002c, 2003a, 2003b; Tsukamoto et al., 2002; Finkel et al., 2003). Taken together, these results suggest that the timing of glacial advances within monsoon-influenced regions is broadly synchronous and that glaciation is controlled primarily by the south Asian summer monsoon, and secondarily by the cooling that is associated with oscillations in the Northern Hemisphere ice sheets and oceans. This finding may reflect the choice of study areas, most of which are within the monsoon-influenced regions of the Himalaya and Tibet, or it may only be apparent because the current precision of the dating (mainly terrestrial cosmogenic nuclide methods) does not allow regional differences to be resolved within better than a few thousand years during the late Quaternary. It may well be that the scale of the asynchroneity of glacial advances across the Tibetan Plateau is less than a few thousand years and hence cannot be resolved by the dating techniques. Nevertheless, prior to the recent application of numerical dating, defining the ages of moraines in this region was highly problematic, and as this study will illustrate, initial attempts to define ages of moraines based on relative weathering, soils, and/or correlation to other proxy data sets could lead to estimates that are wrong by as much as an order of magnitude.

With the exception of polar regions, the high mountains of Asia contain the greatest concentration of glaciers. A proper understanding of the dynamics of these glacial systems is essential for modeling the response of geomorphic systems, particularly the hydrologic component, to global warming (Thompson, 2000). Research on glacial systems in the semiarid (annual precipitation <200 mm) areas of the Himalayan mountain belt has been confined to only a few studies that were not able to adequately define the timing of glaciation (Fort, 1978, 1981, 1983; Burbank and Fort, 1985; Damm, 1997; Taylor and Mitchell, 2000). A recent study in the semiarid region of the Tanggula Shan of central Tibet, however, found that glaciation during the last glacial was extremely restricted (Schäfer et al., 2002). A critical test for the influence of

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monsoonal forcing on glaciation is the extent and timing of glaciation in the semiarid regions of the Himalaya. To examine whether glaciation has been synchronous between the monsoon-influenced and semiarid regions of the Himalaya, we studied the glacial successions in the Ladakh Range of the semiarid Transhimalaya of India. This work included an extensive program of terrestrial cosmogenic nuclide dating of moraines and fan sediments. Developing a chronology of glaciation in this region is a necessary first step toward examining climatic and possible tectonic links to glaciation. Here we present a detailed chronostratigraphy for glaciation and speculate on the factors that forced the timing and style of glaciation in Ladakh.

STUDY AREA

The Ladakh Range of the Transhimalaya trends WNW, reaching altitudes in excess of 6000 m above sea level (asl). It is a granodiorite batholith bounded to the south by the Indus suture zone, which parallels the Indus Valley in this segment of the Transhimalaya (Searle, 1986). At Leh (34°09′N, 77°34′E, 3514 m asl), the mean annual precipitation is ~115 mm, ~41% as rain between July and August, the result of the south Asian summer monsoon, and ~35% occurring mainly as snow between December and March, the result of mid-latitude westerlies; the remaining precipitation is mainly rain (Taylor and Mitchell, 2000; Osmaston, 1994). The average maximum diurnal temperature range is −2.8 °C to −14 °C in January and 24.7 °C to 10.2 °C in July (Osmaston, 1994). Climatic data for elevations higher than Leh (3514 m asl) are not available, and although precipitation is likely to increase with altitude, most of our sampling sites are within a few hundred meters of the altitude of Leh and probably experienced environmental conditions similar to those at Leh. The exceptions, however, are sampling sites on the younger moraines that are at altitudes up to 5185 m asl. At such sites, mean annual precipitation is likely higher, albeit not by much (probably <500 mm), and mean annual temperatures are significantly lower.

Drew (1873), Dainelli (1922), Fort (1978, 1981, 1983), Burbank and Fort (1985), Osmaston (1994), and Brown et al. (2002) provide basic descriptions of the glacial geomorphology in Ladakh. With the exception of the study by Brown et al. (2002), in which terrestrial cosmogenic nuclide dating methods were employed to date a glacial moraine near Leh to 90 ± 15 ka, previous work in this area established only relative chronologies. Fort (1983) recognized evidence for at least four glacial advances. Burbank and Fort (1985) described the physical characteristics of the moraines, including brief descriptions of their weathering characteristics. They suggested that younger but altitudinally higher moraines are more weathered than the older moraines at lower altitudes. They attributed this phenomenon to an apparent reversal of weathering due to bioclimatic zoning. Morphostratigraphically, however, the moraines clearly decrease in age with altitude up each valley.

The oldest glacial deposits comprise a large moraine complex, stretching several kilometers along the main Indus Valley almost 1 km wide. The moraine complex is composed of elongated smooth ridges that rise to >100 m above the valley floor and are parallel to the valley axis (Fort, 1983; Figs. 1 and 2). The moraine surfaces slope 20° to 25° rising to several broad crests tens of meters wide. Boulders reaching several meters in diameter are scattered on slopes and crests of the moraines; the moraines have a surface density of ~10–20 boulders (>1 m in diameter) per 100 m^2. The boulders exhibit granular and cavernous weathering, and some are weathered level to the surface of the moraine. Fort (1983) assigned these landforms to the "Indus Valley stage" glaciation. Valley-parallel striations are present in the Nimmu Gorge (Figs. 1, 2A, and 2B), which is likely the result of this glaciation. Dainelli (1922) took this as evidence to suggest that a large trunk glacier, many tens of kilometers in extent, once extended into the Indus Valley.

Dainelli (1922), Fort (1978, 1983), and Burbank and Fort (1985) described a set of moraines ~50 m thick at an altitude of 3300–3600 m asl in almost every valley between Bazgo and Leh (Figs. 2C and 2D). The moraines have steep slopes (>25°) with sharp crests. Boulders are scattered on the moraine slopes; the crests have a density of 20–30 boulders (>1 m in diameter) per every 100 m^2. Some of the surface boulders exceed 5 m in diameter and exhibit granular and occasional cavernous weathering. Rock varnish is present on most boulders, and in places it is a dark black-brown color. The rock varnish development, however, is not uniform. This suggests that the boulders are eroding. There is very limited soil development on the crests of the moraines. Fort (1983) ascribed these moraines to the maximum extent of glaciation during the last glacial; she assigned them to the "Leh stage." Using ^{10}Be terrestrial cosmogenic nuclides, Brown et al. (2002) dated four boulders (69.5 ± 4.0, 93.2 ± 5.4, 86 ± 5.7, and 66.5 ± 6.9 ka) on the crest of a Leh stage moraine north of the town of Leh and suggested that the moraine formed at 90 ± 15 ka.

Fort (1983) identified younger sets of moraines up-valley from the Leh stage and designated these moraines, which are ~20–40 m thick and occur at altitudes ranging between 4300 and 4600 m asl, as the "Kar stage" (Figs. 2E and 2F). These moraines have steep slopes (>25°) with abundant gelifluction lobes and debris. The crests are subrounded and armored with boulders that have a density that exceeds 40 boulders (>1 m in diameter) per 100 m^2. The boulders have well-developed rock varnish that is commonly dark black-brown in color. Many of the boulders are pitted (>5 mm deep) and exhibit cavernous weathering. There is little soil development on the crests of these moraines. Burbank and Fort (1985) attributed the formation of these moraines to a recessional stage during deglaciation of the global Last Glacial Maximum (LGM).

Hummocky moraines are present in the Bazgo valley at an altitude of 4600–4800 m asl (Fig. 2G). The hummocks are several tens of meters in diameter and rise 5–15 m in height. Boulders are scattered on these hummocks; the boulder density is ~20–30 boulders (>1 m in diameter) per 100 m^2. Cavernous weathering is rare, but the boulders have small pits that are occasionally >10 mm deep. A dark brown rock varnish is well developed on most of the boulders. Although soil development is poor, a thin (<20 cm) organic-rich solum layer caps many of the moraines and is most abundant within the hummock depressions. These moraines have not been previously described; we refer to them as the "Bazgo glacial stage" moraines.

A still younger set of moraines, 150 m thick and typically sharp- and multicrested, are present near the termini of the contemporary glaciers at around 4950–5200 m asl (Fort, 1983). These moraines have steep sides, often at the angle of repose, and their crests are armored with boulders, the boulder density exceeding 30–40 boulders (>1 m in diameter) per 100 m^2. The boulders exhibit little evidence of weathering and in places have a poorly developed light yellow rock varnish. There is no soil development on these moraines. Burbank and Fort (1985) believed these to be late Holocene and called them "Neoglacial moraines" (Fig. 2H). We refer to these moraines as the "Khalling glacial stage" moraines, so as not to imply a specific age of formation.

METHODS

Building on the glacial geologic framework, we aimed to develop a quantitative chronology of glaciation in the Ladakh region. The morphostratigraphic relationships between different landforms of different glacial stages were examined together with the surface weathering characteristics of the moraines.

Samples for terrestrial cosmogenic nuclide surface exposure dating were collected along the Indus Valley and within four tributary valleys (Leh, Phyang, Tharu, and Bazgo) in the Ladakh Range (Fig. 1). Samples of quartz-bearing lithologies, usually granodiorite, were collected from the surfaces of boulders residing along moraine crests at locations where there was no apparent evidence of exhumation or slope instability. Many factors can affect the terrestrial cosmogenic nuclide concentration in a sample, including weathering, exhumation, prior exposure, etc. The net result of these processes would be a large spread in apparent exposure ages of individual boulders. Accordingly, multiple boulders from each moraine ridge were collected and analyzed. The presence of multiple boulders having similar apparent ages is taken as evidence that most of the terrestrial cosmogenic nuclide inventory accrued during exposure to cosmic rays in their existing geometry. Each boulder was carefully examined to determine the factors that affect terrestrial cosmogenic nuclide production and to assess the likelihood that it was exposed in a geometry other than its present position. Each boulder was photographed for future reference. The degree of weathering of the boulders and the moraines and the physical characteristics of the moraines were recorded. The inclination from the boulder site to the tops of the surrounding mountain ridges and peaks was measured to determine the topographic shielding. In addition to boulders on moraines, a succession of sediment samples, forming a depth profile, was collected from an alluvial fan that lies between the moraines of the Indus Valley and Leh glacial stages (Figs. 1 and 3). Each sediment sample was composed of ~1 kg of 2–5-cm-diameter quartz-rich pebbles that were collected along a 5-cm-thick horizon within a section excavated into the alluvial fan.

The samples were crushed, sieved, and quartz was separated from the 250–500 µm size fraction following the methods of Kohl and Nishiizumi (1992). After addition of ^{9}Be carrier, Be was separated and purified by ion exchange chromatography and precipitation as $Be(OH)_2$ at pH > 7. The $Be(OH)_2$ was oxidized by ignition in quartz crucibles. BeO was then mixed with Nb metal prior to determination of the $^{10}Be/^{9}Be$ ratio by accelerator mass spectrometry at the Center for Accelerator Mass Spectrometry in the Lawrence Livermore National Laboratory (samples with the prefix 'I') and the Purdue Rare Isotope Measurement (PRIME) Laboratory at Purdue University (samples with the prefix 'ZK'). Isotope ratios were normalized to ICN Pharmaceutical, Inc. ^{10}Be prepared by K. Nishiizumi (1995, personal commun.).

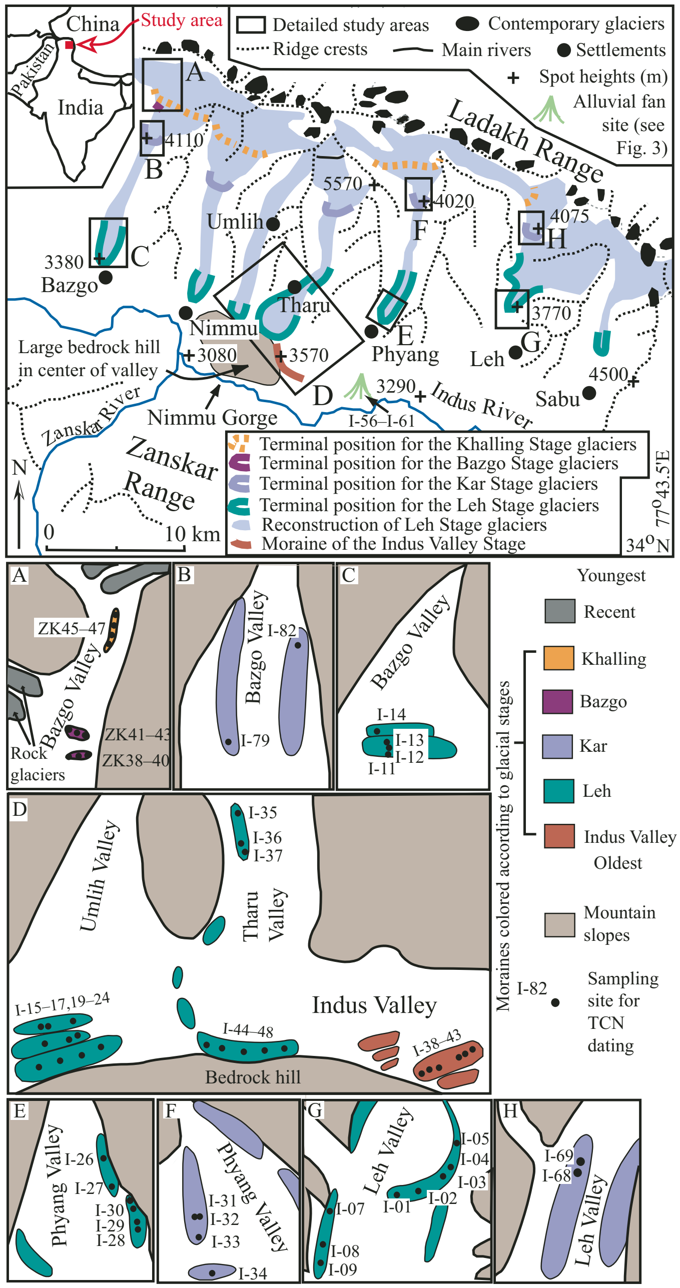

Figure 1. Glacial landforms and sampling locations in the Indus Valley and the Ladakh Range. TCN—terrestrial cosmogenic nuclide.

The measured isotope ratios were converted to ^{10}Be concentrations in quartz using the total Be in the samples and the sample weights. Production rates were scaled to the latitude and elevation of the Ladakh sampling sites using the star scaling factors of Stone (2000) and an assumed 2.2% sea-level–high-latitude (SLHL) production muon contribution. These scaling factors reproduce the star scaling factors of Lal (1991). ^{10}Be concentrations were then converted to zero-erosion exposure ages using a SLHL ^{10}Be production rate of 5.4 ± 0.3 atoms/g quartz per yr (cf. Nishiizumi et al., 1989; Ivy-Ochs et al., 1998; Kubik and Ivy-Ochs, 2004). The impacts of topographic and depth corrections were determined by numeric integration of the flux corrected for the dip and topography at all azimuth directions (Nishiizumi et al., 1989).

There is as yet debate regarding the correct scaling method and SLHL rate. The choice of an absolute production rate and scaling technique does impact the model exposure ages. When

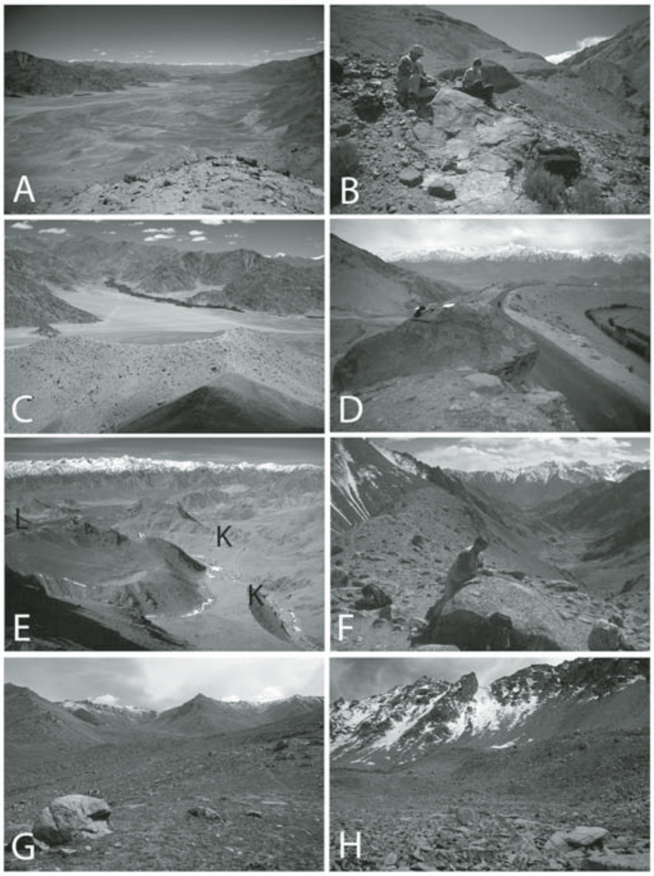

Figure 2. Landforms of different glacial ages in the Indus Valley and Ladakh Range. (A) View looking east along the Indus Valley at moraines produced during the Indus Valley glaciation. (B) Glacially polished and striated bedrock in the Nimmu Valley that was produced by a main trunk valley glacier during the Indus Valley glaciation. (C) View looking north, toward Tharu in the main Indus Valley, at a terminal moraine formed during the Leh glacial stage. (D) View looking south along the crest of a Leh glacial stage moraine north of Leh. The boulder in the foreground is typical of those sampled in this study. (E) View looking south toward Leh, the Indus Valley, and Zanskar Range. The moraines labeled L and K were formed during the Leh and Kar glaciations, respectively. (F) View south along the crest of a Kar moraine in the Bazgo Valley. (G) Moraines of the Bazgo glacial stage in the Bazgo Valley. (H) View of a lateral moraine formed during the Khalling glacial stage in the Bazgo Valley.

exposure ages are then compared to global temperature records, it is important to recognize the inherent uncertainty of exposure dating. However, uncertainties in absolute production rates and scaling factors have far less impact on relative chronologies for events in a limited geographic area.

Bearing these factors in mind, we examined a number of scenarios that encompass changes in production rates, erosion rates, and possible effects of magnetic field changes (Tables DR1 to DR4).[1] This provides an assessment of the sensitivity of our conclusions to uncertainties in the parameters that influence production rates and highlights some issues worth noting. A reduction of the SLHL ^{10}Be production rate to 5.2 ± 0.3 atoms/g quartz per yr, for example, increases exposure ages ~3%–5% relative to the ^{10}Be production rate of 5.4 ± 0.3 atoms/g quartz per yr that we use in calculating the ^{10}Be ages (Table DR4 [see footnote 1]). It is worth emphasizing that there are numerous ongoing scientific investigations with the aim of resolving production rate uncertainties. However, it is not our goal with this work to attempt any resolution of this matter. The production rate of 5.4 ± 0.3 that we use encompasses, within 1σ, most production rates used by other researchers (cf. Nishiizumi et al., 1989, 1996; Ivy-Ochs et al., 1998; Kubik and Ivy-Ochs, 2004). Furthermore, a production rate of 5.4 ± 0.3 atoms/g quartz per yr makes our model ages lower limits. As will be clear in the discussion, our conclusions are not sensitive to the production rate uncertainties, and our data are best viewed as lower limits to their true age.

Incorporating changes in the paleomagnetic field would change the apparent exposure ages by up to 15% (Table DR2; see footnote 1). There is, however, considerable debate regarding the technique and magnitude of this correction (cf. Nishiizumi et al., 1996; Masarik and Wieler, 2003; Pigati and Lifton, 2004). While this effect does change the apparent ages, even given the low erosion rates in Ladakh, erosion of boulders is a more serious consideration.

Terrestrial cosmogenic nuclide concentrations can be interpreted as either minimum exposure ages (assuming zero erosion) or maximum erosion rates (erosional equilibrium). For estimating steady-state erosion rates, the samples with older apparent exposure ages provide limits on erosion rates. For many of the samples collected in Ladakh, maximum erosion rates of 1.0–3 m/m.y. are modeled. Given that these are maximum erosion rates and that it is unlikely that they are in erosional equilibrium, a reasonable estimation of erosion for boulders in Ladakh is ~1.0 m/m.y. Indeed, for the older samples, even an erosion rate of 3 m/m.y. cannot be sustained (cf. Table DR1 [see footnote 1]). These data allow us to reasonably constrain the steady-state erosion rate in Ladakh to ~1.0 m/m.y. Assuming that all boulders weather at this rate, we can assess the impact of this amount of erosion on our chronologies. For samples having a 1 m/m.y. erosion rate, an age of 10 ka would increase uncertainty 1%; an age of 50 ka, by 4%; an age of 100 ka, by 10%; an age of 200 ka, by 22%; and an age of 300 ka, 42% (Table DR3 [see footnote 1]). The erosion rate of 1 m/m.y., consistent with the terrestrial cosmogenic nuclide data, is also consistent with field observations for this area and studies in other semiarid regions (Small et al., 1997; Zehfuss et al., 2001).

Even with the added uncertainty of erosion, variations in the geomagnetic fields and production rates, our conclusions are not altered for samples with ages younger than 100 ka. Furthermore, for old samples (older than 100 ka), geologic processes prevent tight distributions of boulder ages, and the data are best utilized

[1]GSA Data Repository item 2006044, calculations for Ladakh samples, Table DR1, Table DR2, Table DR3, and Table DR4, and production rate data for terrestrial cosmogenic nuclide ages presented in this paper using different variables, is available on the Web at http://www.geosociety.org/pubs/ft2006.htm. Requests may also be sent to editing@geosociety.org.

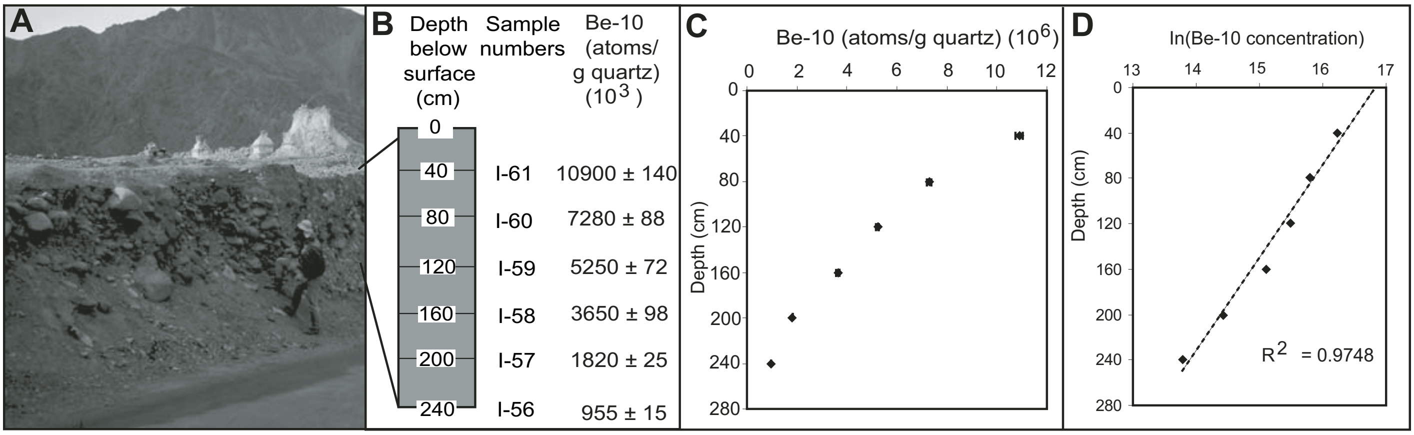

Figure 3. Alluvial fan south of Phyang showing (A) the sampling positions for ^{10}Be dating, and (B) the sampling locations and ^{10}Be concentrations for each sample. The alluvial fan comprises decimeter-size crudely stratified beds of pebbly sands and gravels that dip subparallel to the contemporary fan surface. There is little evidence of soil development, but the fan is capped by a 5–10-cm-thick layer of aeolian silt. (C) Plot of the absolute ^{10}Be concentrations. (D) Plot of $\log_e$^{10}Be concentrations for each sample with depth.

to recognize trends and to set limits on the timing of features that were previously without any chronologic control. We, therefore, present model zero-erosion rate exposure ages in this paper (Table 1).

The concentration of ^{10}Be as a function of depth in samples collected from a depositional surface in the alluvial fan that is stratigraphically associated with the Indus Valley moraines is shown in Figure 3. The concentration of ^{10}Be at any depth is a function of its exposure age, the erosion rate, and its prior history. There are several possible ways to place reasonable and geologically meaningful constraints on the exposure histories of samples residing in this depositional surface. The first, and simplest, scenario is one in which exposure dominates the terrestrial cosmogenic nuclide inventory. In this scenario, the concentration of ^{10}Be would exponentially decrease as a function of depth. Figure 3 shows a semilog plot of the concentration along with a best-fit line to the data. The slope of the line equals Λ/ρ, where $\Lambda = 165$ g/cm^2, the interaction mean free path, and ρ is the density of the material in the depositional surface. The slope of the best fit line is −81.7, yielding a density of 2.02 g/cm^3, consistent with the density of alluvium (Granger and Smith, 2000). These data are consistent with a simple exposure of material in this surface to cosmic rays since the emplacement of the surface. Accordingly, extrapolation of the best-fit line to the surface yields a zero-erosion exposure age of 497 ± 66 ka. The terrestrial cosmogenic nuclide inventory can be alternatively interpreted as implying a surface in erosional equilibrium. However, using ^{10}Be alone, we cannot rigorously distinguish between these two scenarios. In any event, a limiting minimum age is 431 ka (497–66 ka). A possible complicating factor is the possibility that the materials comprising this surface were exposed to cosmic rays before the consolidation of this landform. To maintain the observed exponential drop in concentration, each sample must have inherited the same ^{10}Be concentration. While this possibility cannot be ruled out, we can limit the magnitude of this inheritance. The lowest sample in the depth profile has ~9% of the ^{10}Be concentration found in the uppermost sample. If we attribute all this ^{10}Be to inheritance then the ^{10}Be derived from exposure decreases proportionately. Inheritance at this level does not substantially alter our geologic interpretation of this feature. Other complications are of course possible. Depositional surfaces created by glacial outwash are time-transgressive surfaces, however, the apparent simple behavior of these terrestrial cosmogenic nuclides indicates that the deposition of this feature occurred rather quickly, and its exposure to cosmic rays occurred in its present geometry.

DATING RESULTS

Our field observations are consistent with those of Fort (1978, 1983) and Burbank and Fort (1985) for all the landforms examined in their study. In addition, we recognized a distinct suite of moraines in the Bazgo valley that are morphostratigraphically intermediate between the Kar glacial stage moraines and the "Neoglacial" moraines of Burbank and Fort (1985), which we refer to as the Bazgo glacial stage moraines. We retain Fort's (1978, 1983) and Burbank and Fort's (1985) nomenclature for the moraines they identified and modify the glacial successions based on our new observations, so that from oldest to youngest, these are called the Indus Valley, Leh, Kar, Bazgo, and Khalling glacial stages.

The ^{10}Be dates for each boulder are listed in Table 1 and plotted in Figure 4 by relative age, moraine, and glacial stage. There is no systematic relationship between boulder size and ^{10}Be age for boulders on individual moraines, as might be expected if erosion is controlling the age distribution (Table 1). The data show that most of the samples dated for the Bazgo, Kar, Leh, and Indus Valley glacial stages are of considerable antiquity (many tens to hundreds of thousands of years) and that there is a large spread of ages for boulders within each of the glacial stages. Given the antiquity of the boulders and the stochastic nature of weathering processes, a wide spread of ages is not surprising. As mentioned above, it has to be borne in mind that erosion rates in the order of 1 m/m.y. would increase the true age of the sample and that this effect is more significant for older samples (»100 ka). Keeping this effect in mind, we have broadly assigned an age to each of the glacial stages. An additional geologic factor that may influence cosmic ray exposure ages is production of small boulders by the deep weathering of larger boulders. Samples collected from existing boulders may therefore have spent a large fraction of their exposure history in a much larger boulder. Both of these effects conspire to make the measured ages lower limit ages. It is therefore not surprising that the oldest surfaces display boulder ages ranging from perhaps near the time of deposition to relatively recent times. Dating large numbers of samples from individual and morphostratigraphically equivalent

TABLE 1. SAMPLING LOCATIONS FOR BOULDERS, BOULDER SIZE, TOPOGRAPHIC SHIELDING FACTORS, ^{10}Be CONCENTRATIONS, AND MINIMUM ^{10}Be AGES

Sample number	Location		Altitude (m asl)	Boulder size long axis/height (m/m)	Topographic shield factor†	^{10}Be‡ (10^6 atoms/g)	Minimum ^{10}Be exposure age (ka)
	Latitude (°N)	Longitude (°E)					
I-01	34°11.270′	77°35.681′	3770	2.5/1	0.9975	8.600 ± 0.118	144.8 ± 8.5
I-02	34°11.272′	77°35.700′	3770	2.0/1.5	0.9974	4.770 ± 0.071	79.1 ± 4.6
I-03	34°11.272′	77°35.700′	3740	3.0/1.5	0.9974	8.370 ± 0.141	143.1 ± 8.6
I-04	34°11.465′	77°35.912′	3904	5.0/1.6	0.9971	6.220 ± 0.143	96.6 ± 5.9
I-05	34°11.553′	77°35.991′	3765	2.5/3.5	0.9945	13.000 ± 0.178	225.2 ± 13.5
I-07	34°11.145′	77°35.156′	3645	1.5/1.5	0.9950	1.890 ± 0.043	33.3 ± 2.0
I-08a	34°11.055′	77°35.180′	3704	3.0/0.7	0.9949	8.430 ± 0.144	147.9 ± 8.8
I-08b	34°11.055′	77°35.180′	3704	3.0/0.7	0.9949	11.000 ± 0.179	195.2 ± 11.8
I-09	34°11.046′	77°35.175′	3620	1.5/0.8	0.9949	8.660 ± 0.116	159.4 ± 9.4
I-11	34°14.310′	77°16.973′	3485	5.0/3.0	0.9941	5.110 ± 0.071	99.9 ± 5.8
I-12	34°14.347′	77°16.966′	3402	1.5/2.0	0.9930	9.670 ± 0.236	203.2 ± 12.8
	34°14.341′	77°16.977′	3380	1.5/2.0	0.9944	9.590 ± 0.179	218.1 ± 4.1
I-14	34°14.348′	77°16.947′	3450	6.0/4.0	0.9941	7.810 ± 0.106	157.8 ± 9.3
I-15	34°11.384′	77°22.635′	3575	1.5/0.4	1.0000	12.300 ± 0.291	240.8 ± 14.8
I-16	34°11.384′	77°22.635′	3575	0.9/0.6	1.0000	9.500 ± 0.109	184.3 ± 10.5
I-17	34°11.397′	77°22.665′	3550	0.6/1.1	1.0000	13.300 ± 0.158	266.3 ± 15.5
I-19	34°11.359′	77°22.713′	3553	2.3/0.8	1.0000	9.060 ± 0.133	177.7 ± 10.3
I-20	34°11.359′	77°22.713′	3553	1.0/0.8	1.0000	8.870 ± 0.131	173.8 ± 10.0
I-21	34°11.197′	77°22.373′	3550	1.0/0.8	0.9969	7.610 ± 0.109	144.4 ± 8.5
I-22	34°11.166′	77°22.317′	3550	1.0/0.9	0.9965	7.290 ± 0.104	138.3 ± 8.2
I-23	34°11.158′	77°22.289′	3520	4.5/1.3	0.9965	7.810 ± 0.116	151.2 ± 9.0
I-24	34°11.182′	77°22.267′	3500	2.0/0.5	0.9969	6.880 ± 0.102	133.9 ± 7.9
I-26	34°11.225′	77°29.897′	3525	2.0/2.0	0.9879	5.740 ± 0.093	111.7 ± 6.6
I-27	34°11.209′	77°29.860′	3530	2.5/2.5	0.9883	7.290 ± 0.108	142.3 ± 8.4
I-28	34°11.020′	77°29.673′	3550	1.5/1.2	0.9907	10.400 ± 0.155	202.9 ± 12.1
I-29	34°11.028′	77°29.701′	3592	1.0/1.0	0.9913	5.420 ± 0.079	100.7 ± 5.9
I-30	34°10.993′	77°29.631′	3550	1.0/0.5	0.9914	7.810 ± 0.188	150.1 ± 9.3
I-31	34°13.128′	77°30.725′	3975	1.5/1.1	0.9879	4.020 ± 0.071	60.6 ± 3.5
I-32	34°13.128′	77°30.725′	3965	1.2/0.8	0.9879	6.260 ± 0.102	95.8 ± 5.6
I-33	34°13.101′	77°30.724′	4000	1.1/0.5	0.9899	8.400 ± 0.108	126.7 ± 7.4
I-34	34°13.049′	77°30.739′	3985	1.3/0.4	0.9846	2.660 ± 0.041	40.0 ± 2.3
I-35	34°12.172′	77°25.381′	3700	1.6/0.6	0.9884	19.700 ± 0.309	369.4 ± 22.9
I-36	34°12.126′	77°25.328′	3600	3.0/0.5	0.9870	10.600 ± 0.142	202.0 ± 11.9
I-37	34°12.087′	77°25.328′	3650	3.0/0.8	0.9879	7.910 ± 0.136	144.7 ± 8.6
I-38	34°9.569′	77°24.689′	3525	5.0/1.2	1.0000	15.100 ± 0.203	310.1 ± 18.3
I-39	34°9.557′	77°24.711′	3500	2.0/2.0	1.0000	11.800 ± 0.159	241.9 ± 14.1
I-40	34°9.557′	77°24.711′	3500	2.0/1.2	1.0000	10.500 ± 0.179	213.7 ± 12.6
I-41	34°9.634′	77°24.692′	3500	6.0/2.0	1.0000	14.700 ± 0.198	308.2 ± 18.1
I-42	34°9.611′	77°24.688′	3500	2.0/1.6	1.0000	18.200 ± 0.266	385.5 ± 23.3
I-43	34°9.620′	77°24.690′	3550	1–3 cm clasts	1.0000	6.690 ± 0.090	130.0 ± 7.4
I-44	34°10.516′	77°23.456′	3750	3.0/0.8	0.9959	6.710 ± 0.090	113.7 ± 6.6
I-45	34°10.48′	77°23.475′	3784	1.3/0.5	0.9971	7.460 ± 0.115	124.2 ± 7.3
I-46	34°10.480′	77°23.475′	3784	1.4/0.5	0.9971	8.070 ± 0.136	134.7 ± 8.0
I-47	34°10.482′	77°23.456′	3750	1.5/1.7	0.9974	14.400 ± 0.225	250.1 ± 15.3
I-48	34°10.563′	77°23.428′	3760	1.2/0.8	0.9977	6.450 ± 0.110	108.2 ± 6.4
I-56	34°8.733′	77°26.268′	3258	2–5 cm clasts	See Figure 3	0.955 ± 0.015	See Figure 3
I-57	34°8.733′	77°26.268′	3258	2–5 cm clasts	See Figure 3	1.820 ± 0.025	See Figure 3
I-58	34°8.733′	77°26.268′	3258	2–5 cm clasts	See Figure 3	3.650 ± 0.098	See Figure 3
I-59	34°8.733′	77°26.268′	3258	2–5 cm clasts	See Figure 3	5.250 ± 0.072	See Figure 3
I-60	34°8.733′	77°26.268′	3258	2–5 cm clasts	See Figure 3	7.280 ± 0.088	See Figure 3
I-61	34°8.733′	77°26.268′	3258	2–5 cm clasts	See Figure 3	10.900 ± 0.140	See Figure 3
I-68	34°13.263′	77°36.676′	4125	2.0/1.0	0.9904	4.130 ± 0.101	57.3 ± 3.5
I-69	34°13.276′	77°36.755′	4150	2.0/1.2	0.9900	6.950 ± 0.140	96.3 ± 5.8
	34°18.669′	77°19.884′	4342	1.5/1.3	0.9834	4.100 ± 0.075	51.6 ± 3.0
I-82	34°18.728′	77°20.022′	4150	1.2/0.8	0.9757	1.350 ± 0.021	18.9 ± 1.1
ZK38	34°20.025′	77°21.000′	4700	1.0/0.8	0.9800	7.005 ± 0.196	74.3 ± 4.6
ZK39	34°20.024′	77°20.993′	4698	3.0/1.0	0.9810	4.731 ± 0.116	49.6 ± 2.9
ZK40	34°20.033′	77°21.017′	4702	1.2/1.3	0.9800	4.164 ± 1.256	43.5 ± 2.7
ZK42	34°20.294′	77°21.201′	4784	1.0/0.8	1.0000	5.881 ± 0.167	59.4 ± 3.7
ZK43	34°20.295′	77°21.200′	4790	0.6/0.3	0.9860	4.087 ± 0.113	40.9 ± 2.5
ZK45	34°19.559′	77°23.130′	5178	4.0/1.5	0.9540	0.765 ± 0.328	6.3 ± 0.4
ZK46	34°19.578′	77°23.132′	5184	3.5/1.2	0.9510	1.274 ± 0.44	10.6 ± 0.7
ZK47	34°19.579′	77°23.133′	5185	2.5/0.7	1.0000	1.022 ± 0.032	8.3 ± 0.5

Note: Minimum ^{10}Be ages were calculated using Stone (2000) scaling factors; sea-level high-latitude (SLHL) production rate = 5.4 ^{10}Be atoms/g quartz per yr; zero erosion rate; and sample thickness of 5 cm. Samples I-56 to I-61 are sediment samples collected from the alluvial fan shown in Figures 1 and 3; asl—above sea level.

†Shielding factor as calculated to correct for topographic barriers using the methods of Nishiizumi et al. (1989).

‡Atoms of ^{10}Be per gram of quartz before application of shielding correction factor.

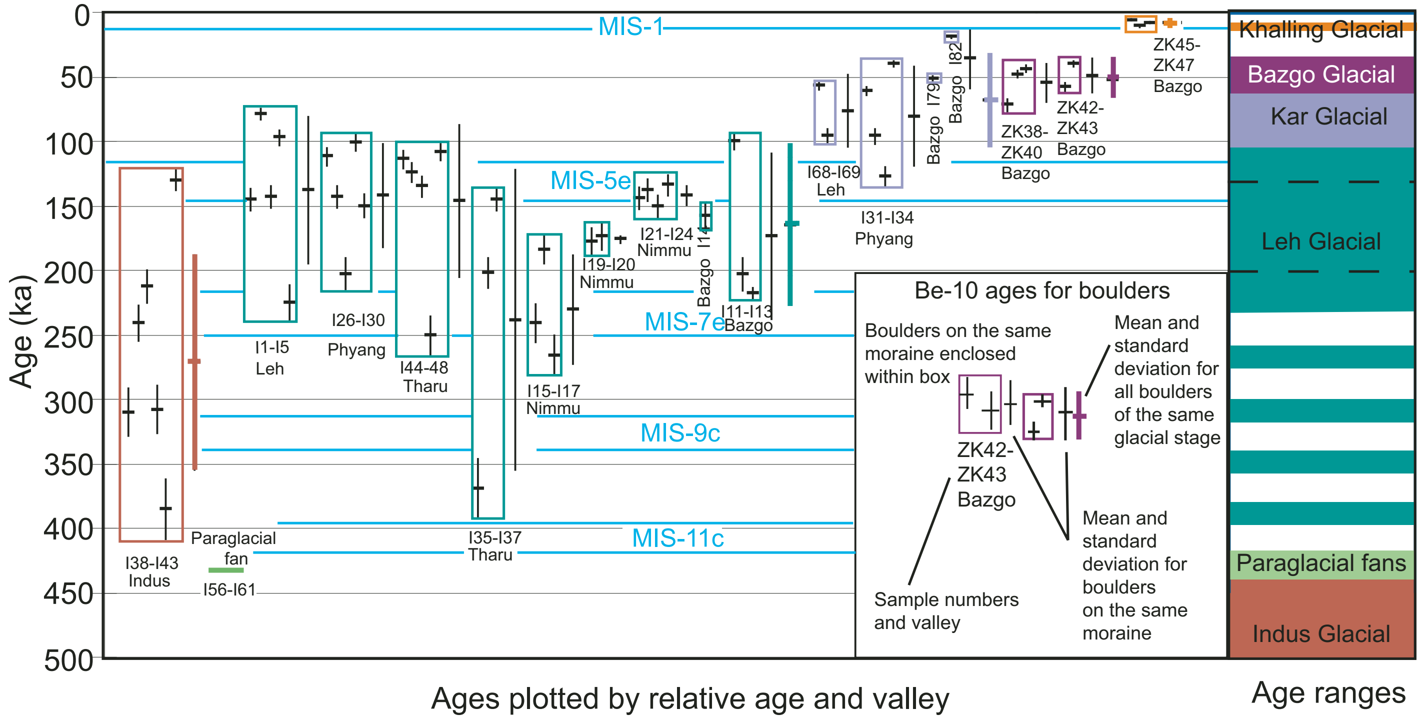

Figure 4. Plot of ^{10}Be dates for boulders on moraines. Colored boxes surround data obtained from moraines of a given stage in the same valley. The minimum age of the alluvial fan is also plotted. The durations of interglacials, bracketed by blue lines, are taken from Winograd et al. (1997). The column on the right illustrates our best estimates for ages for each glacial stage. The uncertainty in assigning an age to the Leh glacial stage is highlighted by the stripes, and the dashed black lines show our best estimate for the age of this glacial stage.

moraines, together with placing each stage in its morphostratigraphic sequence, enables an estimate of the ages of the moraines to be determined, even when the age spread is relatively large (Putkonen and Swanson, 2003). The use of morphostratigraphic relationships is essential for interpreting the surface exposure ages of landforms when there is considerable scatter and overlap of dates for different landforms. This is well illustrated by comparing the age distributions on the Indus and Leh moraines, which show considerable overlap (Fig. 4). In this case we used the dated alluvial fan, which is within and beyond the glacial limits of the Indus and Leh glacial stages, respectively. This allowed us to define the maximum and minimum ages on the Indus and Leh glacial stage moraines, respectively (Figs. 1 and 3).

The young ages for boulders on the youngest moraines (Khalling glacial stage) suggest that inheritance of terrestrial cosmogenic nuclides in the older moraine may not be significant, at least not more than 10,000 yr. In addition, the low concentrations of terrestrial cosmogenic nuclides in the deepest sample (<10% of the shallowest sample) for the alluvial fan depth profile also support the view that inheritance is not large. Brown et al. (2002) reinforced this view by providing near-zero ages for boulders in active streambeds, indicating that, in general, inheritance of terrestrial cosmogenic nuclides from reworking of existing boulders is not a dominant process in these valleys. However, care must be taken when comparing ages on glacial and fluvial boulders that clearly have different transportation and depositional histories. We therefore consider inheritance not to be significant given the dynamic nature of the geomorphic setting, which is one in which the denudation rates are high and rapidly produce fresh surface by mass movement, and glacial and fluvial erosion.

Indus Valley Glacial Stage Moraines

Our data show that boulders on the Indus Valley glacial stage moraines range in age from 130 to 385 ka, suggesting that this moraine is of great antiquity. The youngest sample (I-43, ca. 130 ka) is a composite of surface quartz pebbles. Its young age is likely a minimum age as a consequence of the processes discussed above, i.e., these pebbles were likely initially exposed under conditions of higher shielding. The boulders on this moraine were intensively weathered, and therefore these ^{10}Be ages must be considered to be truly minimum ages.

The ^{10}Be age for the alluvial fan that stratigraphically postdates the Indus Valley stage, is ≥430 ka (Fig. 3). There was little evidence of soil development on this fan, suggesting that there may have been some exhumation of the fan surface and stripping of any soil that may have begun to form. This would suggest that the fan might be considerably older than 430 ka. Therefore, it follows that the Indus moraine must be much older than 430 ka and that the younger ^{10}Be ages on the Indus glacial stage moraine reflect prolonged erosion.

Leh Glacial Stage Moraines

The moraines formed during the Leh glacial stage have ages that range from 79 ka to 369 ka, but that clearly cluster between ca. 100 to 200 ka, suggesting that this glaciation occurred during the penultimate glacial cycle in marine isotope stage 6 (MIS 6). Brown et al. (2002) obtained exposure ages on four boulders from the Leh moraine. They used the scaling factors of Dunai (2000). We recalculated the terrestrial cosmogenic nuclide dates of Brown et al. (2002) using the same method that we applied to our samples. The recalculated ages are 86.2 ± 7.1 ka, 116.0 ± 9.6 ka, 106.9 ± 9.4 ka, and 82.6 ± 9.9 ka, compared with Brown's published ages of 69.5 ± 4.0 ka, 93.2 ± 5.4 ka, 86 ± 5.7 ka, and 66.5 ± 6.9 ka, respectively. Coincidentally, five of the boulders from Leh glacial stage moraines (I-1 to I-5) that we dated were collected on the same moraine dated by Brown et al. (2002). These five ages range between 79 and 225 ka. The dates produced by Brown et al. (2002) cluster on the low end of this range. The boulders on this moraine are deeply weathered, so it is plausible that heavy weathering has resulted in low minimum ages for many of the boulders on this moraine. However, there are several boulders within this data set that are significantly older than 200 ka (I-5, I-13, I-15, I-17, I-35, I-47). These boulders might be explained as having derived from older moraines that were reworked when Leh stage glaciers advanced over older deposits. Given that inheritance in this environment is likely to be relatively small, it is surprising that we have so many old ages on boulders from this glacial stage. Given our extensive data set, with clustering of ages between ca. 130 and 200 ka, it seems reasonable that the Leh glacial stage moraines formed during MIS 6. However, we are cognizant that the Leh glacial stage moraines may have formed earlier.

Kar Glacial Stage Moraines

Our ^{10}Be data for the Kar glacial stage moraines show formation during the last glacial cycle. The wide scatter of data does not allow us to assign these moraines to a specific marine oxygen isotopic stage. The ages on the younger Bazgo glacial stage moraines (see next section), however, suggest that the Kar glacier advance occurred during the early part of the last glacial, most likely during MIS 5.

Bazgo Glacial Stage Moraines

The Bazgo moraines are clearly morphostratigraphically younger than the Kar glacial stage. Although the boulder ages overlap with many of the Bazgo glacial stage boulders, the ages cluster at values that are generally younger than the mean ranges for the Kar glacial stage moraines, providing an age between ca. 41 and 74 ka. This suggests that these formed during the middle part of the last glacial cycle during MIS 3 and/or MIS 4.

Khalling Glacial Stage Moraines

Boulders on Khalling glacial stage moraines have early Holocene ages, if inheritance of terrestrial cosmogenic nuclides is considered insignificant. Unfortunately, we only have three dates on this moraine; clearly more sampling is needed to unequivocally assign this stage to the early Holocene. Owen et al. (2001, 2002a), Finkel et al. (2003), and Barnard et al. (2004a, 2004b) also identified an early Holocene advance in other regions of the Himalaya, suggesting that this is a regional signal, perhaps due to increased monsoon precipitation during the early Holocene insolation maximum.

DISCUSSION

Our study provides the oldest dates for moraines in the Himalayan-Tibetan orogen. The lack of old dates in other study areas may be the result of poor preservation, particularly in the more monsoon-influenced regions where fluvial and mass-movement processes are more dominant. Alternatively, it may be that in other regions, glaciation during the last glacial cycle was extensive enough to have destroyed evidence of early glaciations.

Our data show that the extent of glaciation decreased progressively over at least the last five glacial cycles. The maximum extent of glaciation in Ladakh during the penultimate and last glacial cycles was less extensive than the extent of the local Last Glacial Maximum that has been defined for other regions of the Himalaya (Owen et al., 2002b). The contrast between Ladakh and other Himalayan regions is even more apparent when Ladakh is compared to the glacial records of Nanda Devi and the Garhwal Himalaya, both regions that are heavily influenced by the monsoon and that only have Holocene glacial records preserved (Barnard et al., 2004a, 2004b). The restricted glacial extent in Ladakh is, however, comparable to the record presented by Schäfer et al. (2002) and Owen et al. (2005) for the Tanggula Shan in central Tibet. These comparisons suggest that the strong topographic and hence climatic gradients existing across the Himalaya-Tibetan region control the style of glaciation and preservation.

The progressive restriction in the extent of glaciation in Ladakh over at least the last five glacial cycles is striking. Changes in global ice volume and/or in insolation during the last three or four glacial cycles are not significant enough to explain this trend (Martinson et al., 1987; Berger and Loutre, 1991; Raymo 1992). We therefore suggest that a reduction in moisture flux significant enough to starve the glacier accumulation zones may have occurred over this time. The likely cause of such a reduction in moisture is the tectonic uplift of the Himalayan ranges to the south of Ladakh, which, over time, has reduced the northward penetration of the Indian summer monsoon into Ladakh and/or the uplift of the high Karakoram to the west, thus reducing the influence of the mid-latitude westerlies that bring moisture from the Mediterranean, Caspian, and Black Seas.

Unfortunately, there is no independent assessment of the topographic uplift of the peaks and ridges in the ranges immediately south of Ladakh and to the west in the Karakoram Mountains for the last 500 k.y. The high exhumation rates and anomalously high topography of the Karakoram Mountains highlighted by Foster et al. (1994) and Bishop and Shroder (2003), however, support the view that substantial uplift of the Karakoram Mountains has occurred over the last few million years. Such uplift would have likely helped increase precipitation, supplied by the mid-latitude westerlies, in the Karakoram Mountains, reducing the moisture flux to, and hence precipitation in, Ladakh. Therefore, it is possible that uplift of the high Karakoram is primarily responsible for this change in style of glaciation.

It may be significant that the glacial advances observed during the last 100 k.y. (Kar, Bazgo, and Khalling glacial stages) in the Ladakh Range possibly occurred during insolation maxima, when the Indian summer monsoon influence would have increased northward into the Himalaya, as suggested by Prell and Kutzbach (1987). It must be emphasized that the precision of our ages for the moraines is not precise enough to unequivocally assign them to times during strengthened monsoon activity. Furthermore, multiple marine proxy data from deep-sea cores in the Arabian Sea suggest that insolation maxima may not directly indicate a wetter Indian summer monsoon (Schulz et al., 1998; Kudrass et al., 2001; Altabet et al., 2002). Kudrass et al. (2001) suggested that the insolation maximum centered at ca. 60 ka, for example, did not result in a strengthened Indian summer monsoon, and indicates drier conditions. Nevertheless, it is particularly noteworthy that our data show that there is no evidence in the Ladakh Range for a glacial advance during the global LGM. This was a time when insolation was low and any Indian summer monsoon influence would have been greatly reduced in Ladakh. Hence, glaciers would have been starved of moisture from the Indian summer monsoon. However, if the mid-latitude westerlies were the important forcing factor for glaciation in this region, we would expect them to force glaciation during the global LGM. Since we do not see evidence for glacial advance during the global LGM, we suggest that the Indian summer monsoon has been most important in forcing glaciation over the last 100 k.y. During the early and middle Pleistocene, however, the mid-latitude westerlies may have been more important than the Indian summer monsoon in forcing glaciation. It is possible, however, that uplift of Himalayan ranges south of Ladakh over the last 500 k.y.

was primarily responsible for the changes in extent of glaciation in Ladakh as the influence of the Indian summer monsoon was progressively reduced in Ladakh.

Although we favor a tectonic explanation for the change in extent of glaciation in Ladakh over at least the last five glacial cycles, it is possible that this pattern might reflect a global trend of progressively less extensive mountain glaciation throughout at least the last 500 k.y. Other mountain regions of the world exhibit similar trends, e.g., Tasmania, the Sierra Nevada, Alaska, Peruvian Andes, Patagonia, and the Chilean Lake District (Denton et al., 1999; Barrows et al., 2002; Smith et al., 2002; Kaufman et al., 2004; Singer et al., 2004). Although tectonic influences have also been suggested for the change in glacial extent for some of these regions (e.g., Smith et al., 2002; Singer et al., 2004), there is no unequivocal evidence to directly connect tectonics to these patterns. There is a noticeable qualitative difference between the moraines in Ladakh and those in most other ranges in the world. In Ladakh, differences in the extent of glaciations are large, for example, 5–10 km between the penultimate and last glacial moraines, and many tens of km between the oldest and youngest moraines. Other ranges do not display such large differences. In the Sierra Nevada Mountains, for example, the Tioga, Tenaya, younger and older Tahoe, Mono Basin and Sherwin tills (ranging from ca. 820 ka to 20 ka) differ by only a few kilometers (Kaufman et al., 2004). In comparison to other ranges throughout the world, the glacial extent in Ladakh has decreased dramatically over the past 500 k.y., supporting our hypothesis for progressive drying of the region.

Our study highlights the strong regional contrast in the timing of glaciation across the Himalaya, the Transhimalaya, and Tibet. The contrasting timing of glaciation in this relatively restricted area emphasizes the need for extensive programs of quantitative dating on moraines in the Himalaya to allow accurate correlation of climate changes and landform evolution. The work that Burbank and Fort (1985) undertook to study equilibrium line altitudes (ELAs) between the Ladakh and Zanskar Ranges provides a powerful example. These authors assumed the Leh moraines to be LGM features. Although this was a plausible interpretation, exposure age dating now indicates a much earlier deposition. Furthermore, we have shown that moraines that were thought to have formed during the Neoglacial (the Khalling glacial stage) are actually considerably older (early Holocene), hence illustrating the problems of assigning glacial landforms to specific climatostratigraphic times on the basis of correlation alone.

CONCLUSIONS

Our ^{10}Be surface exposure dates define the timing of glaciation for Indus Valley, Leh, Kar, Bazgo, and Khalling glacial stages of the Ladakh Range to older than 430 ka, the penultimate glacial cycle or older, early last glacial cycle, middle last glacial cycle, and early Holocene, respectively. Each successive glaciation in the Ladakh Range was progressively less extensive. This suggests that there was a significant reduction of the moisture flux to the region that is necessary to support positive glacier mass balances and allow glaciers to advance. This reduction may have been the consequence of the progressive surface uplift of the Himalayan ranges to the south of Ladakh and/or the uplift of the high Karakoram to the west, which would have reduced the influence of the south Asian summer monsoon and mid-latitude westerlies, respectively. This pattern of glaciation, however, may alternatively reflect a global trend of progressively less extensive glaciation in mountain regions throughout the Pleistocene.

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