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Equilibrium-line altitudes of the Last Glacial Maximum for the Himalaya and Tibet: an assessment and evaluation of results

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Abstract

Contemporary equilibrium-line altitudes (ELAs) for glaciers in the high mountains of the Himalaya and Tibet have considerable variation because the region is influenced by two major climatic systems (the mid-latitude westerlies and the south Asian summer monsoon) and glaciers are subject to strong topographic controls. Reconstructions of past ELAs based on the former extent of glaciers are numerous, but were estimated using a variety of methods some of which are not appropriate for high-altitude glaciers which are strongly influenced by topography and have extensive supraglacial debris cover. Furthermore, few of the reconstructions have adequate chronological control because of the difficulty of dating glacial landforms in high-altitude regions where organic matter for radiocarbon dating is sparse or absent. Cosmogenic radionuclide surface exposure and optically-stimulated luminescence dating are providing data to define the timing of glaciation and these, together with objective analysis of glacier dynamics, may allow better estimates of ELA changes in the future. However, there are currently only two regions (the Hunza Valley and the Khumbu Himal) which are sufficiently well dated to allow reconstructions of Last Glacial Maximum glacier extent and hence ELAs. These two regions cannot provide an adequate assessment of ELA changes for the Himalayan–Tibetan region.

1. Introduction

Proxy data for reconstructing the paleoclimate of Tibet and the Himalaya during the Last Glacial Maximum (LGM) are rare (Shi, 2002; Liu et al., 2002). The most commonly applied technique for reconstructing Late Quaternary paleoclimate involves calculating former equilibrium-line altitudes (ELAs) and ELA depressions relative to the present (ΔELAs) for glaciated catchments. Reconstructed ΔELAs are used to estimate past temperature and/or precipitation changes and to assess regional climatic gradients and variability. Reconstructing former ELAs and ΔELAs for the Himalaya and Tibetan Plateau, however, is problematic. This is essentially because of extreme topographic controls on glacier mass balance, the abundance of supraglacial debris cover, and microclimatic variability, which, in combination, complicate the relationship between climate and glacier extent. These problems are discussed fully in Benn and Lehmkuhl (2000) and Benn et al. (2005). Although numerous papers have published ΔELA data for specific glaciated catchment areas throughout the Himalaya and Tibet, it is difficult to assess the validity of many of the calculations, or to use these data to provide a regional assessment of ΔELAs for the LGM. This is mainly because papers rarely provide a full description of the geologic data used to reconstruct the shape of former glaciers, or adequate discussions of the methods used to reconstruct past and present ELAs. Furthermore, most papers do not present any numerical dating to define the timing of formation of the glacial landforms, and hence the timing of glaciation, that they are using to reconstruct ΔELAs. Therefore, there is not at present an adequate temporal framework to compare ΔELAs across the Himalaya and Tibet. Benn and Lehmkuhl (2000) and Benn et al. (2005) emphasize the need to present comprehensive data on the glacial catchment (including the contemporary

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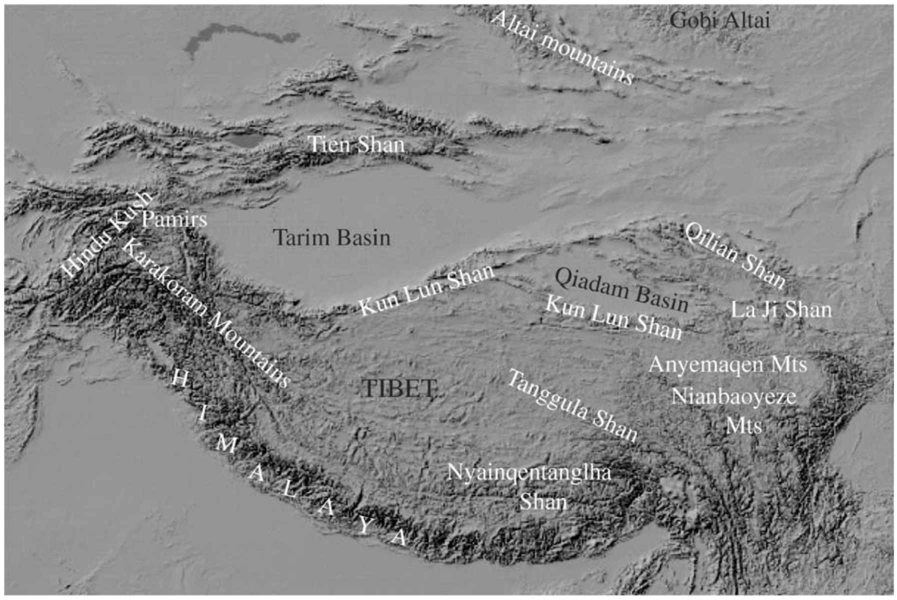

Fig. 1. DEM showing the main regions of the Himalaya and Tibet.

topography, glaciology and climate), the glacial geomorphology and the geochronology used in ELA reconstructions. Furthermore, they underline the importance of employing multiple techniques to test the validity of ELA results (cf. Meierding, 1982; Torsnes et al., 1993; Kaser and Osmaston, 2002). Given these problems, this paper aims to examine the data on ΔELAs throughout the Himalaya and Tibet to provide an objective assessment of the current knowledge and usefulness of published works. As a contribution to LGM Snowline project (Mark et al., 2005) we concentrate on evaluating the ELA data for the global LGM. Two critical sites where LGM moraines have been accurately dated are discussed in detail.

2. Regional climate and glaciology

The Himalaya and Tibetan Plateau stretch E–W for ~2000 km from Burma to Afghanistan and for ~1500 km from India into central China. They comprise several main ranges including the Siwaliks, Lesser Himalaya, Greater Himalaya, Karakoram Mountains, Hindu Kush, Kunlun Shan (Shan = Mountains), Nianqentanglha Shan, Tanggula Shan, Nainbaoyeze Mountains, Qilian Shan and La Ji Mountains (Fig. 1). The average elevation of the Tibetan Plateau is ~5000 m a.s.l. (Fielding et al., 1994). The average elevation in the Himalaya is less, but with numerous peaks that exceed 6000 m a.s.l., and greater relative relief on the land surfaces, commonly more than several kilometres (Fielding et al., 1994). These mountain ranges are influenced by two major climatic systems: the mid-latitude westerlies and the South Asian monsoon (Fig. 2A and B). Furthermore, significant interannual climatic variability in the region is associated with El Nino Southern Oscillation (ENSO). The relative importance of the mid-latitude westerlies and South Asian monsoon varies throughout the region and has varied throughout the Cenozoic (Benn and Owen, 1998). The monsoonal influence is greatest on the eastern and the southern slopes of the Himalaya and eastern Tibet, which experience a pronounced summer maximum in precipitation, which at high-altitude falls as snow (Fig. 2C). In contrast, the more northern and western ranges, such as the Karakoram Mountains, and NW Tibet receive heavy snowfalls during the winter with moisture supplied by the mid-latitude westerlies (Fig. 2D).

The pronounced summer precipitation maximum in the Himalaya reflects moisture advection northwards from the Indian Ocean by the southwest monsoon. Summer precipitation, however, declines sharply from south to north across the main Himalayan chain and is much higher at the eastern end of the mountain chain than in the west. Only in the extreme west (Karakoram and Pamir) is there a winter precipitation maximum, due to the influence of winter westerly winds bringing moisture from the Mediterranean, Black and Caspian Seas. Snow and ice accumulation may also vary across the Himalaya as a consequence of changing lapse rates, related to a reduction in the moisture content of air masses as they are forced over the Himalaya. These air masses have a major influence on glacier formation both regionally and on a local scale. The seasonal distribution of precipitation results in the maximum accumulation and ablation occurring more or less simultaneously during the summer throughout much of the range (Ageta and Higuchi, 1984; Benn and Owen, 1998).

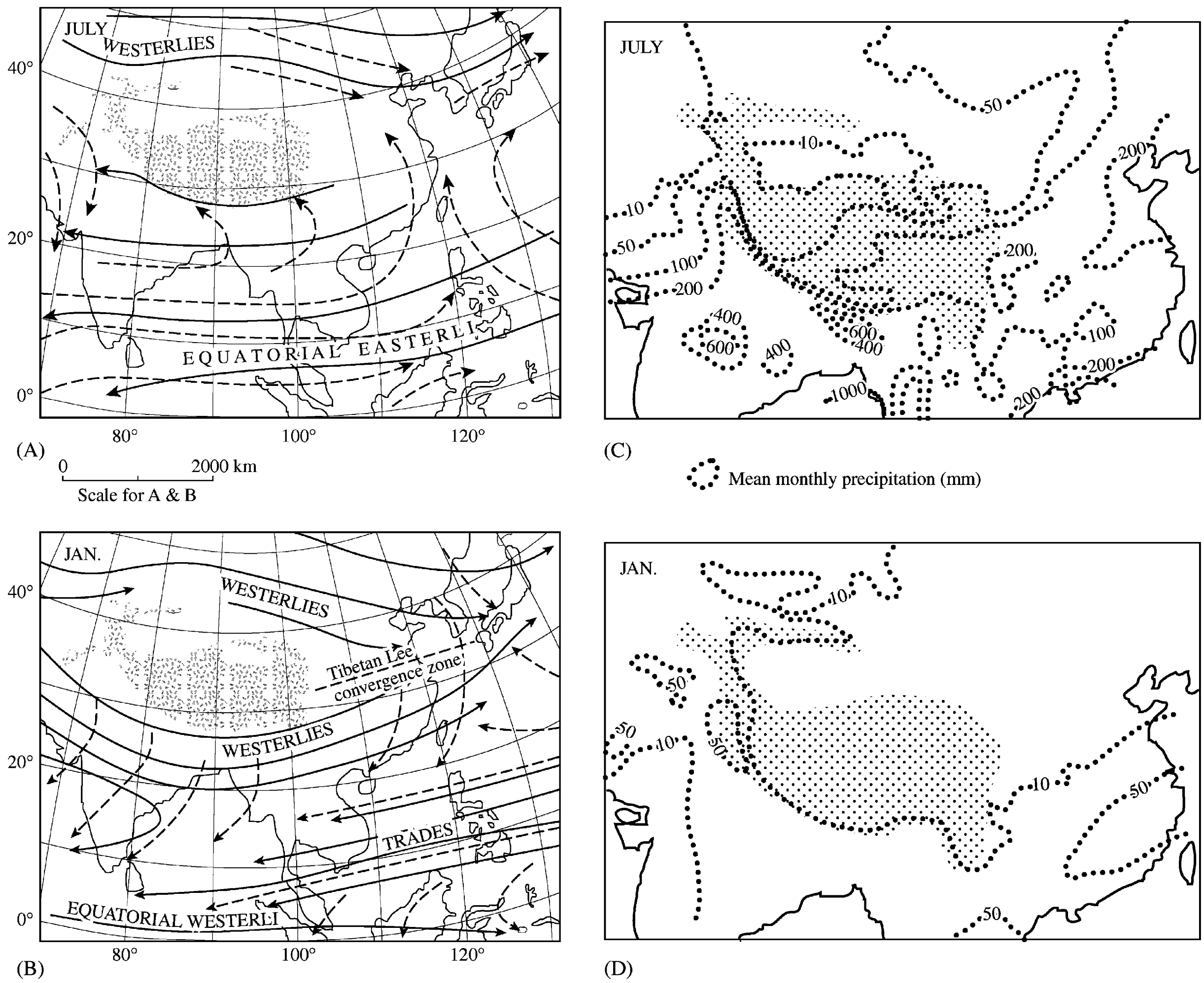

Fig. 2. The characteristic air circulation (A and B) and precipitation (C and D) over southern and central Asia. The Tibetan Plateau and bordering mountains are above 5000 m a.s.l. are shown by the dotted areas. In (A) and (B) the solid lines indicate airflow at about 6000 m in (A) and 3000 m in (B), and the dashed lines indicate airflow at about 600 m. (C) and (D) show the strong N–E and E–W precipitation gradients for July and January, respectively. Adapted from Owen et al. (1998), Benn and Owen (1998), and Lehmkuhl and Owen (2003).

The regional snowline varies in altitude, and is 4600–5600 m and 6000–6200 m a.s.l. south and north of Everest, respectively (Müller, 1958; Shi et al., 1980), 5200–5700 m in the Garhwal (Grinlinton, 1912), 4500–4700 m around Nanga Parbat (Finsterwalder, 1937), 4600–5300 m on the south side of the Karakoram Mountains, and 5200–5700 m on the northern side of the Karakoram (Visser and Visser-Hooft, 1938; Su, 1998) (Fig. 3). It must be highlighted that the snowlines throughout the Himalaya and Tibet may have varied significantly during the last 100 years, but as yet there are no comprehensive studies showing the historical and temporal variability.

Meteorological and topographic characteristics play a large role in influencing the nature of the glaciological systems. Glaciers can be divided into three main regimes: continental interior types in the central and western parts of the Tibet Plateau; maritime monsoonal types in the Himalaya and in southeastern Tibet; and

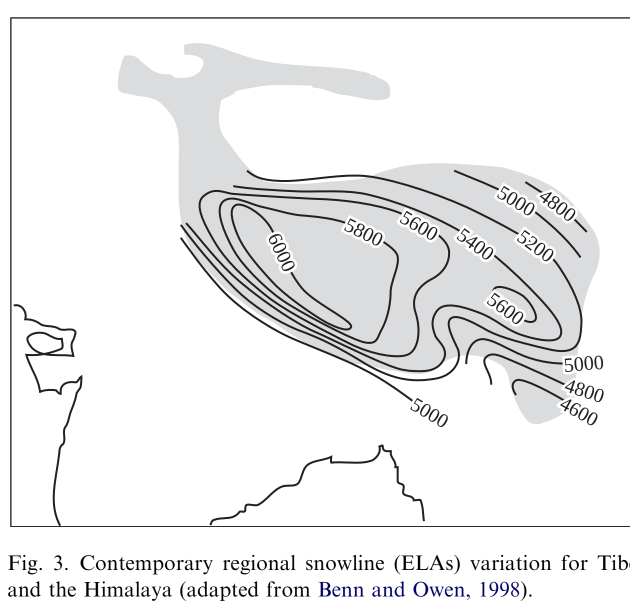

Fig. 3. Contemporary regional snowline (ELAs) variation for Tibet and the Himalaya (adapted from Benn and Owen, 1998).

continental monsoonal types in eastern and north-eastern Tibet (Derbyshire, 1981). Most glaciers within the region occupy steep, high relief catchments, and have cold, high altitude (>5000 m) accumulation areas. Where mass turnover is high, or where debris cover allows glaciers to descend to lower altitudes, ablation zones may be temperate, but in higher altitude, drier environments glaciers are cold throughout. Annual temperature cycles vary with the degree of continentality and altitude. For example, the monthly mean temperatures are $\sim 3^{\circ}$ in summer and $\sim -11^{\circ}$ in winter at 5000 m a.s.l. in the Khumbu Himal, Nepal (Bertolani et al., 2000), and $>20\,^{\circ}$C in the summer and $0\,^{\circ}$C in the winter, at 3000 m a.s.l. in the Karakoram Mountains for the more continental type glaciers (Batura Glacier Investigation Group, 1976, 1979, 1980; Derbyshire, 1981; Goudie et al., 1984; Zheng et al., 2002).

Benn and Lehmkuhl (2000) provide a recent review of the mass balance characteristics of Himalayan glaciers. The maritime and continental monsoonal glaciers of Himalaya and southeast and eastern Tibet are highly diverse, including avalanche- and snowfall-fed cirque and valley glaciers, and very steep hanging glaciers (Benn and Lehmkuhl, 2000; Benn and Owen, 2002). On many glaciers, snow avalanching from precipitous slopes forms an important component of accumulation, and glacier ablation areas can be separated from source areas by steep icefalls or avalanche tracks. Basin topography, therefore, plays an important role in determining glacier geometry and mass balance. As early as 1856, Adolf Schlagintweit (cited in Kick, 1962) used the term "*firnkessel*" for glaciers that are fed predominantly by avalanche from high cliffs in the high mountains of Asia, as opposed to the *firnmulden* types in the Alps that are nourished in firn fields. A comprehensive glacier classification scheme based on mass balance and topographic criteria was proposed by Schneider (1962).

The continental interior glaciers of central and western Tibet have annual precipitation $\leqslant 1000$ mm (above 5000 m a.s.l.) and have basal ice temperature $\leqslant 0\,^{\circ}$C. The Guilya icecap, for example, has an ice temperature of $-18\,^{\circ}$C and has an annual precipitation of 300–400 mm (Shi, 2002). These glaciers are similar to sub-polar types, they have low surface velocities, usually between 2 and 10 m a^{-1}, and may extend outside the permafrost zone (Shi, 2002). These glaciers are less topographically constrained and generally have more limited debris cover than those in the Himalaya.

3. Reconstructing the former extent of glaciers

Since the early work of Penck and Brückner (1909), the standard method for reconstructing the former extent of glaciers has involved geomorphic mapping. More recently, however, remote sensing and digital elevation modeling has been used to define the extent of former glaciers (Isacks et al., 1995; Duncan et al., 1998). Reconstructions based on field mapping rely on the accurate interpretation of the landforms and sediments. Researchers such as Derbyshire (1983, 1996), Derbyshire and Owen (1990, 1997), Fort (1986, 1988, 1989, 1995, 1996), Fort and Derbsyhire (1988), Lehmkuhl and Pörtge (1991), Hewitt (1999) and Benn and Owen (2002), however, have warned of the dangers of misinterpreting glacial and mass movement landforms in the Himalaya and Tibet. This is because intense fluvial and glacial erosion often destroys diagnostic morphologies of glacial and mass movement landforms, making their identification difficult. In addition, the diamictons that comprise mass movement and glacial landforms look very similar and have similar particle size distributions and particle shapes. Misinterpretations of the evidence have led to extreme reconstructions and have helped perpetuate the theory that the Tibetan Plateau was once covered by an ice sheet (Kuhle 1985, 1986, 1987, 1988a, b, 1990a, b, 1991, 1995, 2003). Benn and Owen (2002) reviewed and presented new studies on the sedimentology and geomorphology of Himalayan glaciers and emphasized the need for detailed analysis of landforms using modern sedimentological techniques. Few papers, however, provide detailed descriptions of the landforms and sediments that were used to reconstruct former glaciers.

4. Defining the age of moraines

In the high altitude and high-energy environments of the Himalaya and Tibet, dating glacial successions has long been hampered by the low preservation potential of organic deposits. Where radiocarbon dating has been successfully used it has been limited to dating Holocene landforms (Benedict, 1976; Fushimi, 1978; Müller, 1980; Shiraiwa, 1993; Owen et al., 1997; Lehmkuhl, 1996, 1998). Significant progress has been made in recent years by the application of optically stimulated luminescence (OSL) and cosmogenic radionuclide (CRN) exposure dating, which allow the direct dating of glacial sediments and landforms (e.g. Phillips et al., 2000; Schäfer, 2000; Richards et al., 2000a, b; Schäfer et al., 2002; Owen et al., 2001, 2002a, b; Tsukamoto et al., 2002; Finkel et al., 2003). Both techniques, however, have associated problems (Benn and Owen, 2002). Some of these are inherent to the methods themselves, but others are related to site-specific factors and may be overcome by careful sampling design. Examples include the question of whether the luminescence signal in sedimentary grains has been zeroed prior to burial, and the stability of moraine surfaces chosen for CRN exposure dating. Benn and Owen (2002) argued that the successful application of these techniques and interpretation of the results requires an understanding of glacial debris transport paths and the processes that form and modify glacial landforms. The abundance of supraglacial debris helps produce complex latero-frontal moraines that may accumulate over one or more glacial cycles, and understanding the nature, internal composition and structure of Himalayan glacial landforms is essential for developing strategies to date the timing of glaciation. Dating moraines using CRN surface exposure dating, for example, provides a minimum age on the landform and is essentially an estimate of the onset of deglaciation. In contrast, OSL dating provides an age for the timing of deposition of the sediment and hence when appropriate sections are available it can be used to define the duration of formation of the landforms by collecting samples from different horizons (Richards et al., 2000b). Given the problems inherent in any one numerical dating method, we stress the importance of applying at least two independent dating methods to a moraine to unequivocally assign it to a climatostratigraphic interval. Further discussion on the application of numerical dating techniques to dating glacial landforms in the tropics is given in Mark et al. (2005).

5. Major study regions

Although the Himalaya and Tibet are vast in extent, research has focused on a limited number of key study areas. These are regions that are generally politically stable and logistically accessible, but also ones that have well-preserved glacial geologic evidence. We choose to describe these key regions because of their importance in understanding the Late Quaternary glacial geology of Tibet and the bordering mountains and for reconstructing ELAs. Reviews of Late Quaternary glacial history of the Himalaya and Tibet are provided by Owen et al. (1998), Lehmkuhl et al. (1998), Owen et al. (2002c) and Lehmkuhl and Owen (2005), and a list of the main research papers is provided in the bibliography of Barnard and Owen (2000). Von Wissmann (1959) and Frenzel (1960) provide reviews of the earliest observations, including accounts of the early explorers, on the past and present ELAs throughout Tibet and the bordering mountains. Although these accounts are useful and historically important, the lack of numerical dating makes it difficult to use these data to compare regional variability in ELAs for any particular glacial time.

5.1. Chitral

Owen et al. (2002b) studied glacial landforms and sediments and provided evidence for two Late Quaternary major glaciations in the eastern Hindu Kush. OSL dating was undertaken to define the timing of these glaciations and associated sediment deposition. The oldest glaciation, the Drosh Glacial, was defined by OSL dating to have occurred during Marine Isotope Stage 3 (MIS 3: 40.9 ± 3.5, 40.6 ± 3.7, 52.0 ± 4.6, 36.6 ± 3.0 ka and 27.0–55.2 ka), producing an extensive valley glaciation that extended to an altitude of $\geqslant$1300 m a.s.l. in the main valley, with an ΔELA of ~1200 m. A younger glaciation, the Pret Glacial, produced valley glaciers that extended to an altitude of ~1670 m a.s.l. in the main valley, with an ΔELA of ~1000 m. Owen et al. (2002b) do not describe how the ΔELA was calculated, but refer the reader to Kamp's (2000) Ph.D. thesis and states that he used the toe-to-summit (TSAM; the arithmetical mean of the altitude of the highest peak in the catchment area and the altitude of the terminal moraine) methods of von Höfer (1879) and Louis (1955). The OSL dating, that defines the timing of the advance, suggests that this glacial stage probably represents several glacial advances that occurred during the latter part of the Last Glacial. Moraines representing two minor glacial advances, the Shandur and Barum Glacial stages, were also recognized near the contemporary glaciers. These probably formed during the Middle/Late Holocene and Little Ice Age, respectively. No ΔELA were calculated for these stages.

5.2. Swat Himalaya

On the assumption that the firn line on temperate mountain glaciers in the Swat Himalaya coincides approximately with the present-day ELAs, Porter (1970) suggested that contemporary ELAs in Swat are between ~4000 and 4250 m a.s.l. He recognized three main glacial stages in the region: the Laikot (oldest), Gabral and Kalam. He further subdivided the Gabral and Kalam into two and three stades, respectively. Using an accumulation area ratio (AAR) of 0.6 ± 0.1, Porter (1970) estimated a Pleistocene ΔELA for these glaciations of between ~900 and 1100 m for the Laikot, Gabral and Early and Middle Kalam glaciations, and ~300 m for the Late Kalam glaciation. Unfortunately, Porter (1970) was unable to date these glaciations, but Richards et al. (2000a) provided OSL dates for the Gabral 2 and Kalam I of 77 ± 18 ka (MIS 4-3) and 38 ± 10 ka (MIS 3), respectively. This suggests that the either of the Late Kalam stades (II and III) may be equivalent to a LGM glacial advance. However, since no dated evidence exists for a specific advance at the LGM, the chronology of Richards et al. (2000a) restricts the possible magnitude of ΔELA during the LGM to >300 m.

5.3. Nanga Parbat and middle Indus valley

Contention exists over the extent and timing of glaciation in the Nanga Parbat Himalaya and middle Indus valley of northern Pakistan (Richards et al., 2000a, 2001; Phillips et al., 2000). Using luminescence dating, Richards et al. (2000a) showed that moraines within the main Indus valley date to the global LGM. In contrast, on the basis of the lack of CRN ages that date to the LGM, Phillips et al. (2000) concluded that advances in the Holocene (9.0–5.5 ka) are the only ones up valley from moraines that formed during MIS 3. Richards et al. (2001) critique Phillips et al. (2000) by highlighting the uncertainty involved in using a limited number of CRN dates, and challenge the assumption that a lack of dated LGM evidence unequivocally precludes a glacial advance during the LGM. None of these workers, however, made accurate reconstructions of the former shape of the glaciers that produced the moraines that they dated. Scott (1992) calculated a ΔELA of between 720 and 800 m for the local LGM using toe-to-headwall altitude ratio (THAR) of 0.4. Nevertheless, the numerical dating of both Richards et al. (2000a) and Phillips et al. (2000) suggest that this ΔELA may be for a MIS 3/MIS 2 glacial advance.

5.4. Hunza valley

Evidence for eight glacial advances is present in the upper Hunza valley (Derbyshire et al., 1984; Li et al., 1984). These glacial advances are amongst the best numerically dated successions in the Himalaya (Derbyshire et al., 1984; Owen et al., 2002a; Spencer and Owen 2004). Owen et al. (2002a) showed that the Ghulkin I Glacial Stage is equivalent to the global LGM. However, during this time glaciers only advanced a few kilometers beyond their present positions. This is in contrast the Borit Jheel and Yunz glacial stages that are dated to MIS 3 or earlier when glaciers advanced into diffluence cols and into the main trunk valleys. Despite this being one of the best-dated glacial successions in High Asia, ΔELA calculations have not been adequately undertaken. Haserodt (1989) provides a regional estimate of the ΔELA of 1100–1250 m, but does not describe the techniques used and/or the geologic evidence. Furthermore, he provides no geochronological control on the glacial chronologies. Given the limited advance and very low gradients in the ablation areas of the main glaciers (Batura, Pasu, Ghulkin and Gulmit) in the upper Hunza valley and during the Ghulkin I Glacial Stage (equivalent to the global LGM) it would require a ΔELA of only a few hundred meters to create such an advance. The ΔELA during earlier glacial advances (Yunz and Borit Jheel Stage) would be significantly larger, and would have probably exceeded 1000 m.

5.5. Kashmir

Holmes and Street-Perrott (1989) reassessed the classic work of De Terra and Paterson (1939) on the Quaternary glacial geology of Kashmir and suggested that many of the deposits and landforms mapped as glacial were products of mass movement. Using topographic maps of the modern glaciers, Holmes and Street-Perrott, showed that the ELAs in Kashmir rise in a broadly south-west to north-east direction, from about 3900 to 4700 m a.s.l. (Fig. 4). They proposed that precipitation and topography are the major controls on present day glacierization. They calculated former ELAs by using a THAR of 0.4, where the toe was the terminal moraine and the headwall was the same as the present day glaciers. They argued that the trend in glacierization is apparently reversed for former glaciers, as shown by the magnitude of past ELA depressions (700–800 m) and the trend in cirque-floor altitudes. They also proposed that this difference reflects a change in the moisture gradients during glacial advances and relatively rapid tectonic uplift along the southwest margin of Kashmir during the Middle and Late Quaternary. However, they were not able to date any of the glacial landforms. It is therefore not possible to assess if landforms on either side of the Kashmir basin were formed during the same glaciation, and can be used to reconstruct the past regional ELA gradients.

5.6. Garhwal

In the Garhwal Himalaya, Sharma and Owen (1996) provide one of the most comprehensive sets of modern ELA reconstructions for the Himalaya using a variety of methods, yet the glacial chronology indicates a lack of discernable LGM moraines. The authors calculated modern ELAs using Porter's (1970) AAR of 0.6 and Andrew's (1975) AAR of 1.3. They also calculate the THAR using Meierding's (1982) ratio of 0.4 and 0.5 and Sissons' (1974) area-weighted mean. This study showed large variability in results obtained by different techniques, and the strong topographic controls on ELAs (Fig. 5). Furthermore, they were able to show that the modern ELAs varied significantly between valleys with mean ELAs varying from 4510 to 5390 m a.s.l. Using these different techniques they calculated former ΔELA for the local LGM (~63 ka), Bhagirathi Glacial Advance at ~640 m, the mid Holocene Shivling Glacial Advance at 40–100 m and the Bhujbas Glacial Advance (LIA) at 20–60 m. New CRN dating undertaken by Barnard et al. (2004), however, provides significantly younger ages on the moraines dated by Sharma and Owen (1996). The difference between the CRN and OSL ages may be the results of post-depositional modification of the surfaces that were dated by the CRN methods or that the moraine was reoccupied on more than one occasion. Alternatively, the OSL dates and/or the CRN dates could be erroneous. The difference in CRN and OSL ages still has to be resolved.

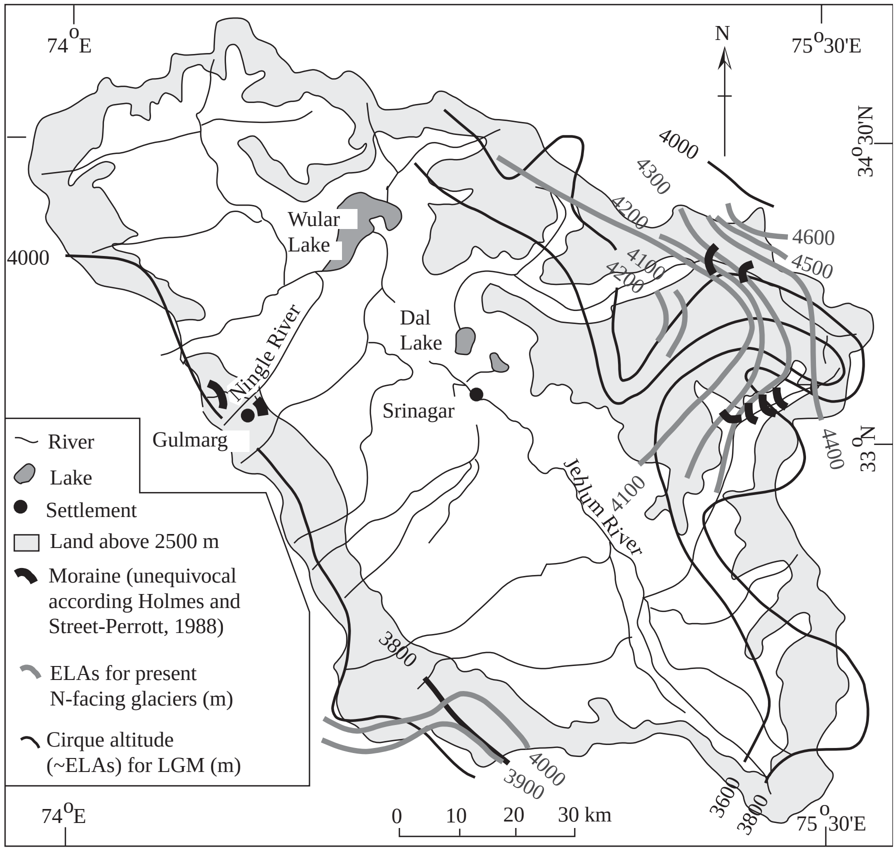

Fig. 4. The location of the main end moraines in Kashmir and past and present ELAs (adapted from Holmes and Street-Perrott, 1989). See the discussion for the validity of the results.

5.7. Lahul

Owen et al. (1996, 1997) provides the first studies of the Late Quaternary glaciation of the Lahul Himalaya in northern India. They showed evidence for five glacial advances, which they called the Chandra, Batal, Kulti and Sonapani I and II. Using CRN dating, Owen et al. (2001) defined the timing of the Batal and Kulti Stages to 12–15.5 and 10–11.4 ka, respectively. Since CRN dates represent minimum ages, they suggested that the CRN ages may be coincident with timing of deglaciation and therefore the Batal Stage might be coincident with the global LGM. Although Owen et al. (1996, 1997, 2001) did not attempt to calculate ELAs, Taylor and Mitchell (2000) provide estimates for present and former ELAs for Lahul and Zanskar. Using a THAR of 0.4 and TSAM of 0.5, they suggest that during the Batal Stage, for a S to N transect across Lahul into Zanskar, the ELAs ranged from ~3960 to 4270 m a.s.l. and 4266 to 4726 m a.s.l., respectively. This as compared with the contemporary ELAs of that range from 4800 to 5500 m a.s.l. across the same mountain ranges.

5.8. Ladakh

Burbank and Fort (1985) state that the contemporary ELAs appear to be between 5200 and 5400 m a.s.l. in the Ladakh and Zanskar Ranges to the north and south of the Indus valley in Northern India. However, they do not state how these contemporary ELAs were determined. Using an AAR of 0.65 and a THAR of 0.40, Burbank and Fort (1985) calculated that the ELAs for a late Pleistocene maximum advance were ~4300 and 4700 m a.s.l. in the Ladakh Range and Zanskar Range, respectively. They attributed this north–south rise in ELA values to be the result of differing topographic controls. In the Zanskar Range, vertical strata of indurated sandstones and conglomerates cut by narrow steep-walled canyons create a bulwark that effectively precluded significant down-valley glacial advance. However, Burbank and Fort were not able to date these moraines, or test whether they formed during the same glaciation on either side of the Indus valley. Recent CRN dating by Bovard (2001) has shown that the late Pleistocene maximum moraines of Burbank and Fort (1985) in the Ladakh Range are >100 ka old. Furthermore, moraines ~5 km from the present ice margins and

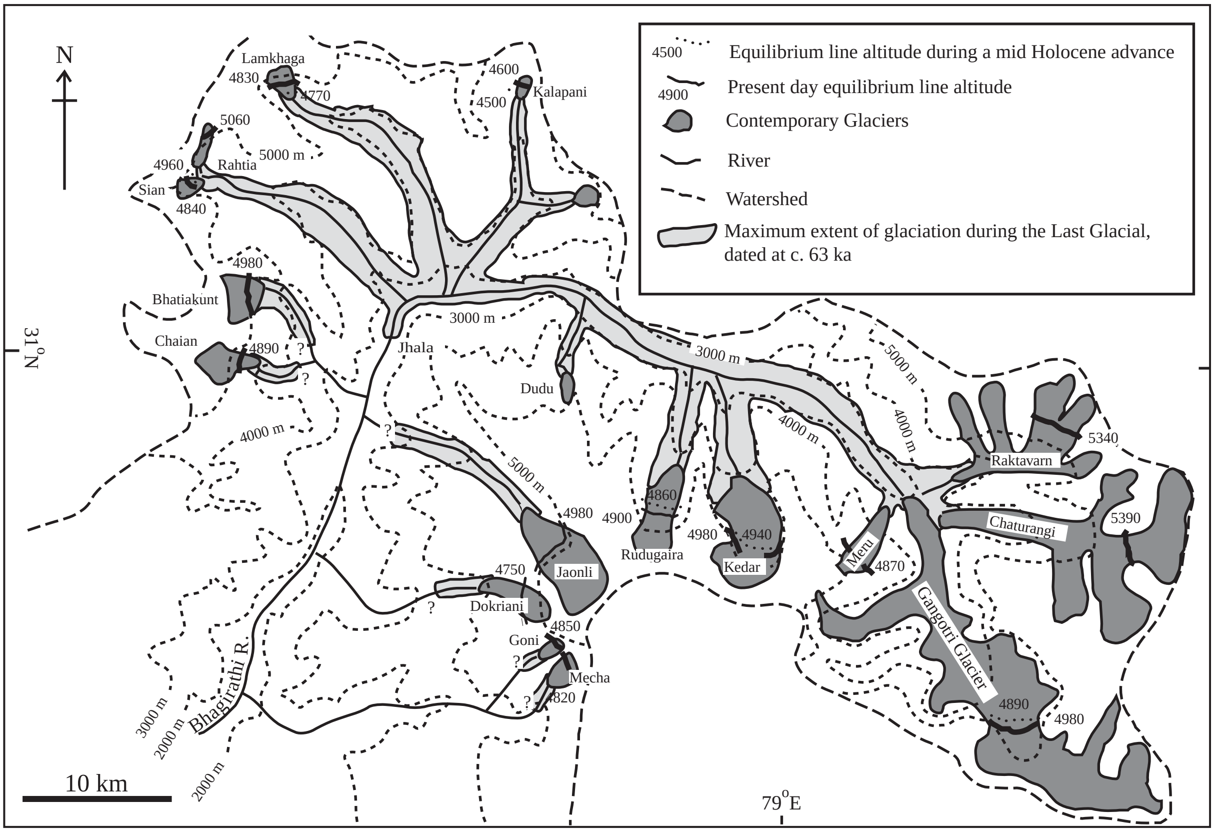

Fig. 5. Reconstruction of glacier extent for the Bhagarathi Glacial Stage in the Garhwal Himalaya (adapted from Sharma and Owen, 1996). The location and results of OSL dating are also shown (adapted from Sharma and Owen, 1996). Note the large differences in ELA values for adjacent valleys and the strong north–south gradient.

mapped as "recessional" by Burbank and Fort (1985) and informally designated as the "Kar Stage" by Fort (1983) have CRN ages of ~70 ka. These dates suggest that if glaciers advanced during the global LGM in the Ladakh Range then they were extremely limited in extent and would have had ΔELAs much less than a few hundred meters. Furthermore, the CRN ages question Burbank and Fort's (1985) assumption that the terminal moraines on both sides of the Indus valley formed during the same glaciation.

5.9. *Zanskar*

Osmaston (1994) provides the first comprehensive study of Zanskar. He calculated that contemporary ELAs in Zanskar range from 5500 to 5800 m and that the ΔELA during Pleistocene glacial maximum was ~600 m, however, he did not describe how the ΔELA was calculated. Moreover, he was unable to define the ages of his Pleistocene glacial maximum advance. Taylor and Mitchell (2000) extended his work and attempted to correlate the glacial successions in Zanskar with those presented by Owen et al. (1996, 1997) in Lahul. Yet as Owen et al. (2001, 2002d) showed, Taylor and Mitchell's (2000) correlation with the Lahul chronologies is problematic. Nevertheless, Taylor and Mitchell's (2000) work provides a framework to be tested, and they provided OSL dates on their "Batal Stage" moraines of between 40 ± 9.3 and 78 ± 12.3 ka. They argued that this was probably the maximum extent of ice during the last glacial cycle, spanning MIS 4 and MIS 3. They also dated their "Kulti Glacial Stage" at between 10 and 16 ka, but suggested that this might be older and may represent the LGM. However it is difficult to assess Taylor and Mitchell's (2000) dating because they do not present a description of the laboratory methods, the dose rate data (radioisotope concentrations) or equivalent dose data that is necessary to evaluate the dating. Their "Kulti Glacial" was restricted to ~10 km beyond the present ice margins. Taylor and Mitchell (2000) used THAR of 0.4 and 0.5 as an estimate of former ELAs, and concluded that ΔELA for their "Batal" and "Kulti" glaciations were ~500 and 300 m, respectively.

In the Markha Valley, Damm (1997) used the TSAM method of Louis (1955) to calculate former ELAs. He recognized two old glacial stages and that had ΔELAs of >1000 m, and three advances that he believed that dated from the LGM to the Lateglacial, which had ΔELAs of between 700 and 300 m. Furthermore, he showed a strong topographic control on glaciation with glaciers predominantly on north slopes having ELAs of 5700–5800 m a.s.l. compared to those on south faces having ELAs between 6100 and 6200 m a.s.l. However, he could not substantiate the ages of his moraines using numerical dating.

5.10. Langtang

Shiraiwa (1993) recognized six glacial stages in the Langtang valley of Nepal: Lama Stage (equivalent to the penultimate or early last glaciation); the Gora Tabela Stage (equivalent of the global LGM); the Langtang Stage (Holocene Maximum); the Lirung Stage (late Holocene); and the Yala I and II Stages (Little Ice Ages). Only the Langtang, Lirung and Yala stages have been dated by numerical methods (radiocarbon dating). Shiraiwa (1993) constructed a steady-state glacier mass balance model for the Langtang valley and used geologic data to reconstruct the climate during the Gora Tabela Stage. He suggested that the precipitation was reduced to 200 mm with an increase winter balance of ~400 mm and an air temperature decrease of 4 °C. Shiraiwa (1993) argued that glaciers during the Gora Tabela Stage were supported by non-monsoonal precipitation. Using chronology and relative dating of Shiraiwa and Watanabe (1991), Fort (1995) estimated a ΔELA of 510 m for the Gora Tabela Stage, which she considered was equivalent to the global LGM. She used a combination of methods including the altitude of medial and lateral moraines, and an AAR of 0.6. However, the age of the Gora Tabela Stage moraines still has to be defined by numerical dating.

5.11. Mount Everest

The first detailed study of the geomorphology of the northern flank of Mt. Everest was produced following the 1966–1968 Chinese scientific expeditions (Zheng, 1988). They recognized evidence for at least three Pleistocene glaciations, and assigned tentative ages to these by correlating moraines in the region with numerically dated moraine successions in other parts of the Tibetan Plateau and to the timing of global glaciations. The glacial successions in other parts of Tibet, however, were poorly constrained by dating (Osmaston, 1989) and, as Benn and Owen (1998) highlighted, regional correlation are complicated by of microclimatic and regional climatic effects. Kuhle (1986, 1987) alluded to the glacial succession in the region, reinterpreting the evidence to fit into his Tibetan ice sheet hypothesis. However, the existence of an ice sheet across Tibet and covering this region is disputed by most workers (Burbank and Kang, 1991; Lehmkuhl et al., 1998). The glacial succession in the Rongbuk valley was reinterpreted by Burbank and Kang (1991) who assigned ages to the moraines on the basis of relative weathering studies. This study was reassessed by Mann et al. (1996) who used boulder weathering studies, lichenometry and radiocarbon dating of calcium carbonate coatings in moraine soils to assign new ages to the moraine successions. All the radiocarbon dates were Holocene and provided minimum ages on the moraines.

All these researchers recognized evidence for multiple local glaciations on the northern slopes of Everest, but their interpretation of the ages of moraines and hence the timing of glaciation and climate change varies considerably. The oldest glaciation is represented by till remnants that comprise bouldery deposits. Zheng (1988) attributed these to pre-late Pleistocene glaciations which he calls the Xixibangma and Nyanyaxungla Glaciations. Till remnants that are probably equivalent to the Nyanayaxungla Glaciation are present in the Rongbuk valley >10 km from the snouts of the contemporary glaciers (Zheng, 1988; Burbank and Kang, 1991). Zheng (1988) believed these formed during the middle Pleistocene while Burbank and Kang suggest that they may have formed during MIS 10 or 8. A second set of moraines is present in the Rongbuk valley, ~3–4 km north of the present Rongbuk glacier. Zheng (1988) and Burbank and Kang (1991) subdivided these into two stages assigned to the Qomolangma I and II Glaciations. Zheng (1988) believed that the Qomolangma I moraines formed during the early Late Pleistocene and Burbank and Kang (1991) suggests that they formed during MIS 6 or younger. Burbank and Kang (1991) believed that the Qomolangma I formed during MIS 2 (equivalent to the LGM) while Zheng (1988) suggests that it is >11,000 years BP. In contrast, Mann et al. (1996) believes that these moraines have minimum ages of 9500 years BP. A further set of moraines flank the modern glaciers and are believed by Zheng (1988) and Burbank and Kang (1991) to have formed during the Neoglacial and this is supported radiocarbon dating undertaken by Mann et al. (1996) that gives a minimum age of 1900 years BP.

On the basis of the height of the highest lateral and medial moraines, changes in the curvature of contours on the glaciers and an AAR = 0.65, the modern ELAs are lowest on the glaciers lying to the northwest of Mt. Everest (Rongbuk, Pumori and West Rongbuk glaciers) where the ELAs average between 5800 and 5900 m a.s.l. (Burbank and Kang, 1991). Due north of Everest, the East Rongbuk and Changtze glaciers have ELAs of ~6400 and 6200 m a.s.l., respectively (Burbank and Kang, 1991). Burbank and Kang (1991) calculated the Neoglacial and Late Pleistocene ΔELA to be ~50–100 and 350–450 m, respectively. Burbank and Kang (1991) used an AAR of 0.65 and they supported the use of this value by calculating the ELA based on the height of the highest lateral and medial moraines; and changes in glacier surface contours from convex to concave forms.

Using at AAR of 0.6, Williams (1983) showed that ELAs varied across Mount Everest ranging from 5200 to 5800 m from south to north. There is, however, considerable variation in ELAs between neighboring glaciers, for example, the Ama Dablam Glacier and its neighbor the Chhukung glacier on the southern side of Everest (Khumbu Himal) have ELAs of 5000 and 5200 m a.s.l., respectively (Benn and Lehmkuhl, 2000). Williams (1983) suggested that ELAs during the maximum Late Pleistocene glaciation rose in altitude from 4300 to 5500 m across the range. The ELA depression to the south of the range (950 m) was about twice as great as to the north (400 m). During four distinct Holocene glaciations ELAs were depressed about 30% as much as the late Pleistocene maximum depression. Gradient changed over 85 km from 7.1 m km^{-1} to maximum late Pleistocene glaciation of 11 m km^{-1} reflecting increase aridity during glacial times. However, it is not known if the moraines used to make such a conclusion are of the same age. Sufficient numerical dating has only been undertaken on the southern side of the Everest (Richards et al., 2000b; Finkel et al., 2003), and is reviewed in detail in Section 6.1 below.

5.12. Kanchenjunga

Using OSL methods, Tsukamoto et al. (2002) were able to define three glacial advances in the Kanchenjunga Himal to 5–6, 8–10 and 20–21 ka. The oldest advance, and coincident with the global LGM was represented by valley glaciers that extended to Gyabla at an altitude of ~2700 m a.s.l. some 15 km from the present ice margin. Asahi and Watanabe (2000) calculated contemporary ELAs for glaciers in the Ghunsa Khola watershed in the Kanchenjunga Himal and showed that they increase from ~5000 m a.s.l. at a latitude of 27°35′N to ~6000 m a.s.l. at 27°50′N. However, they did not calculate the former ELAs. Asahi et al. (personal communications) used a maximum elevation of lateral moraines, relative height of cirques, and mean elevation of glaciers (MEG) to reconstruct the ELA for the LGM across the Kanchenjunga Himal and showed that they rose from ~4200–4500 m a.s.l. to ~5200 m a.s.l. at a latitude from 27°33′N to 27°49′N. This compares with present ELA of between ~5300 and >6000 m a.s.l. over the same latitudinal range.

5.13. Bhutan

Isacks et al. (1995) used satellite imagery and digital topographic data to map modern and maximum ice extents. Assuming that the maximum extent of glaciation throughout the region was simultaneous and was equivalent of the LGM, both the modern and LGM ELAs are shown to rise from south to north in response to decreasing precipitation across the mountains front, with the ΔELA during the LGM being between 700 and 1300 m. Isacks et al. suggested that this represented a temperature depression of 4–8 °C. However, as Owen et al. (1998) and Iwata et al. (2002) highlighted, their work is not supported by any field observations and no glacial landforms have been dated.

5.14. Nyainqentanglha Shan

Lehmkuhl et al. (2002) recognized two glacial advances in the Nyainqentanglha Shan and adjacent region. They used morphological criteria including moraines, terraces and alluvial fans, and luminescence dating, to distinguish two main glacier advances, which they believed occurred during the Middle and Late Pleistocene. They assume the ΔELA during the Pleistocene glacial periods to have been 300–500 m in the drier regions and 600–800 m in the wetter parts on the southern and southwestern slopes (Fig. 6, section C–C′). The ELAs were calculated using the TSAM method of Louis (1955). However, Lehmkuhl et al. (2002) do not provide details on the timing of the glaciation that is associated with their stated ΔELA, except to state that the ΔELA is for the Pleistocene glacial periods.

5.15. Tanggula Shan

The ΔELA across the Tanggula Shan was calculated using the TSAM method of Louis (1955) by Lehmkuhl (1995a, b, 1998) (Fig. 6, section A–A′). However, he was unable to define the timing of glaciation. Using CRN methods, Schäfer et al. (2002) dated four boulders on the Tanggula Pass. Although the data set is small, it suggests that glaciation in this region was extremely limited over the last 200 ka, with an advance at about 180 ka and another at about 70 ka. Therefore, it is likely that the ΔELA of Lehmkuhl (1995a, 1998) is for an advance that is significantly older than the global LGM.

5.16. Kunlun Shan

Lehmkuhl (1995a, b) used Louis's (1955) TSAM method to calculate the past and present ELAs across the Kunlun Shan. He showed a significant rise in present and past ELAs from the NE to SW across the Kunlun Mountains (Fig. 6, section A–A′). Unfortunately, none of the moraines and landforms that he used in his calculations has been dated and it is therefore not possible to test whether glaciation and the landforms they used in their reconstruction were contemporaneous across the Kunlun Shan. Derbyshire (1996) cites ΔELA for the LGM in the Kunlun Mountains of between 400

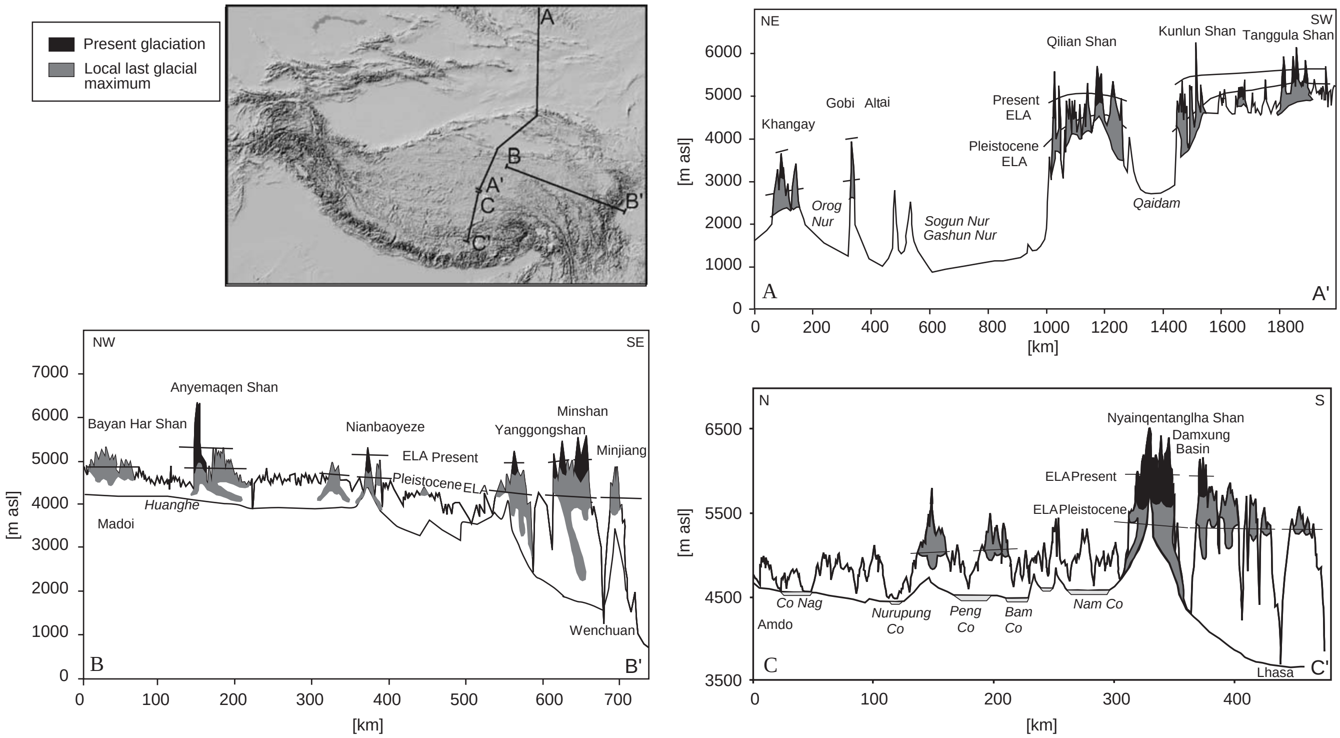

Fig. 6. Map including selected cross sections across Tibet including the distribution of present and local LGM-ice and the ELAs (from Lehmkuhl and Owen, 2003). Cross-section A–A′ including the distribution of present and local LGM-ice and the ELAs the Khangay and Gobi Altai to the Tibetan Plateau (according to Lehmkuhl 1995a, 1995b, 1997b). Cross-section B–B′ including the distribution of present and local LGM-ice and the ELAs from the Eastern fringe of the Tibetan Plateau (according to Lehmkuhl 1995a, b, 1998). Cross-section C–C′ including the distribution of present and local LGM-ice and the ELAs from Southern Tibet (according to Lehmkuhl et al., 2000, 2002).

and 1000 m, but no descriptions of the methods used on chronological control are provided.

5.17. Nianbaoyeze Mountains

Lehmkuhl and Liu (1994) estimate that the present ELA is about 5100 m a.s.l. and that during the last glaciation the ELA was at about 4350–4400 m a.s.l. (Fig. 6, section B–B′). Their methods are not fully described, except that they allude to the use of "accumulations" (presumably moraines) and cirque altitudes. Furthermore, they do not define what is meant by the last glaciation, but it is likely that they are referring to glaciation that produced the "*Ximencuo moraine*" in the Ximencuo valley and which has a TL date of on lacustrine sediment within this moraine date to 32±3 ka, while aeolian sediments on top of the moraine have an older date of 54±4 ka. Lehmkhul (pers. comm.), therefore, suggest that this may have formed during MIS 4. Owen et al. (2003b) refer to the glaciation that produced this moraine as the Ximencuo Glacial Stage and they have dated this using CRN and OSL to MIS 2. This moraine is probably equivalent to a LGM advance, although it may represent an early Lateglacial advance. However, Owen et al. (2003b) did not recalculate the ELAs of Lehmkuhl and Liu (1994).

5.18. Anyemaqen Mountains

Using the TSAM method of Louis (1955), Lehmkuhl (1995a, b, 1998) calculated the Pleistocene ΔELA for the Anyemaqen Mountains to be ~500 m (Fig. 6, section B–B′). However, he was unable to define the timing of glaciation. Owen et al. (2003b) recognized three sets of glacial moraines along the northern slopes of the Anyemaqen Mountains. Using CRN dating, they were able to show that the local LGM probably occurred in MIS 3, with advances restricted to within a few kilometers of the present ice margins during MIS 2 and the early Holocene. Owen et al. (2003b) were unable to resolve whether the moraines that formed during MIS 2 were produced during the LGM or during the Lateglacial. Nevertheless, the glacial advance during MIS 2 was very restricted in extent and the ΔELA calculated by Lehmkuhl (1995a, b, 1998) are not representative of the global LGM.

5.19. La Ji Mountains

Owen et al. (2003a) mapped and dated moraine successions along the northern margins of the La Ji Mountains at the northeastern edge of the Tibetan plateau using CRNs. The glacial geologic evidence and

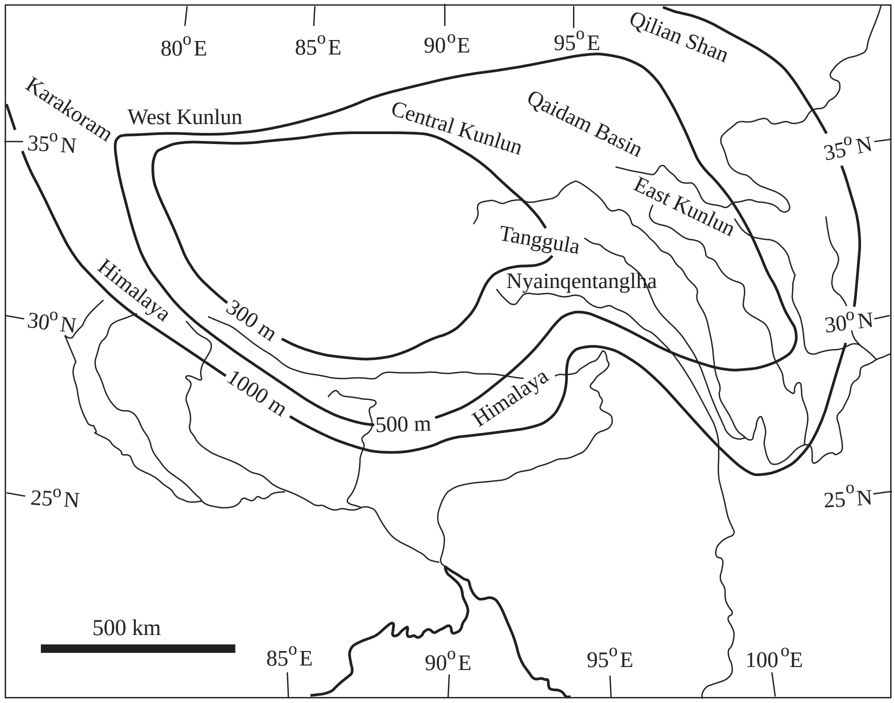

Fig. 7. Equilibrium-line altitude depression for the global LGM across Tibet and the bordering mountains according to Shi et al. (2000) and Shi (2002). Note that although the ΔELA values are probably not correct (see text for a discussion on the validity of the ΔELA values) the general pattern of variation is probably realistic. Adapted from Shi et al. (2000) and Shi (2002).

the CRN dating (between ~20 and 10 ka) show that glaciers existed in this marginal region of Tibet during the latter part of the global LGM and during the Lateglacial. These data suggest that temperatures were low enough and/or monsoon precipitation was sufficiently high to support one or more limited glacial advances between ~20 and 10 ka. Nevertheless glaciation was limited to glaciers of <10 km in length and the estimated ΔELA was ~600–800 m using an AAR of 0.5 to 0.8.

5.20. Qilian Shan

Lehmkuhl and Rost (1993) calculated ELAs for the northern slopes of the Qilian Shan. They showed that in catchments that have heights of >5500 m a.s.l., glaciers advanced to the valley mouths at altitudes of ~2000 m a.s.l. during the last glaciation. The present snowlines are between 4400 and 4700 m a.s.l. and the last glacial snowline was between 3400 and 3800. This is a ΔELA of >1000 m. However, Lehmkuhl and Rost (1993) do not fully describe the methods they used in calculating the ΔELAs nor is there any numerical dating to define the time of glaciation in this region. Their maximum extent of glaciation, may be earlier than the global LGM. Lehmkuhl (1995a, b) shows that within the inner ranges of the Qilian Shan the ΔELA is less, between 600 and 800 m, for the LGM. However, as in the study of Lehmkuhl and Rost (1993) no detailed descriptions of the ELA methods are provided although reference to the Louis' (1955) TSAM method is made in the paper. Furthermore, no numerical dating was undertaken to define the ages of the glaciation. If the ΔELA calculations of Lehmkuhl (1995a, b) are correct and they all date from the same glaciation then there is a clear regional gradient in ELAs across the Qilian Shan (Fig. 6, section A–A′). Using electron spin resonance (ESR) and thermoluminescence (TL) dating, Zhou et al. (2002) showed that moraines within the Bailang valley on the northern side of the Qilian Shan, are old and date to MIS 12, MIS 6 and MIS 4-2. In contrast, Owen et al. (2003b) used CRN to date moraines on the southwestern slopes the Qilian Shan and showed that glaciers extended 5–10 km beyond their present positions during the global LGM and where probably maintained at their maximum extent until the Lateglacial. The landforms that Lehmkuhl and Rost (1993) and Lehmkuhl (1995a, b) used in their ELA reconstructions could, however, be are significantly older than the LGM.

6. Critical study areas

The above discussion highlights the problems and uncertainty in reconstructing and using published ELA data for the Himalaya and Tibet. Maps quantifying ΔELA for Himalaya and Tibet, such as shown in Fig. 7, therefore have to be regarded with great caution.

Table 1
Summary of the locations, geologic evidence and methods used to calculate ELAs for areas throughout the Himalaya and Tibet

LGM snowline locality[b]	Country	Lat °N	Lon °E	Aspect	Summit altitude (m)	Headwall altitude (m)	Lowest terminal moraine (m)	Glacier type	ELA at LGM (m)	LGM method	Modern ELA (m)	Modern ELA method	Modern ELA date	LGM date	No. dates	DMC	DC	References
Chitral	Pakistan	36	72	S	6871	Not known	1300	Valley	4050	TSAM	5050	Observations	2000	No MIS-2 dates	9	5	N/A	Kamp, 2000; Owen et al., 2002b
Swat Himalaya	Pakistan	35	72	S	5936	Not known	1770	Valley	~3200	AAR = 0.6±0.1	4000–4250	Regional extrapolation	1968	No MIS-2 dates	2	5	N/A	Porter, 1970; Richards et al., 2000a
Nanga Parbat and Middle Indus valley	Pakistan	35	74	N, S, E	8126	Not known	1500	Valley	3000–3700	THAR = 0.4	3750–5200	THAR = 0.4	1990	27±4 ka	1	1	6	Scott, 1992; Richards et al., 2000a, 2001; Phillips et al., 2000
Hunza Valley[a]	Pakistan	36	74	E	7885	Not known	2450	Valley	3700[a]	Not described	4800	Not described		~22–18 ka	>10	1	1	Schneider, 1962; Derbyshire et al., 1984; Li et al., 1984; Owen et al., 2002a; Spencer and Owen, 2004
Kashmir	India	34	74	N, S	5500	Not known	<2500	Valley	3200–3500	THAR = 0.4, cirque altitudes	3900–4700	THAR = 0.4	1985–1987	No MIS-2 dates	0	5	N/A	De Terra and Paterson, 1939; Holmes and Street-Perrott, 1989
Garhwal Himalaya	India	31	79	N, S, W	7050	5200–6400	2300	Valley	4300	AAR = 0.6 and 1.3, THAR = 0.4 and 0.5	4500–5400	AAR = 0.6 and 1.3,THAR = 0.4 and 0.5	1994	No MIS-2 dates	>10	1	7	Sharma and Owen, 1996; Barnard et al., 2004
Lahul	India	32	77	N,S, E, W	6400	5790–6100	2740	Valley	3960–4726	THAR = 0.4, TSAM	4800–5500	THAR = 0.4, TSAM	1999	15.5–12 ka	>10	1	6	Owen et al., 1996, 1997, 2001; Taylor and Mitchell, 2000
Ladakh	India	34	77	S	~6100	Not known	~3600	Valley	4300–4700	AAR = 0.65, THAR = 0.4	5200–5400	Not described	Early 1980s	No MIS-2 dates	>20	1	7	Fort, 1983; Burbank and Fort, 1985; Bovard, 2001
Zanskar	India	33	77	N,S, E, W	~6400	Not known	4140	Valley	4700	AAR = 0.65, THAR = 0.4 and 0.5	5200–6200	Not described	1990s	10–16 ka	2	1	4	Damm, 1997; Taylor and Mitchell, 2000; Owen et al., 2002d
Langtang	Nepal	28	85	N,S	7239	Not known	2600	Valley	5920	AAR = 0.6	5320	AAR = 0.4, changes in surface contours, highest lateral moraines	1980s	No MIS-2 dates	0	5	N/A	Shiraiwa and Watanabe, 1991; Shiraiwa, 1993; Fort, 1995
Mount Everest (northern side)	Nepal, Tibet and China	28	86	N	8848	Not known	5300	Valley	5750	AAR = 0.65, highest altitude of lateral moraines	6200	AAR = 0.65, highest altitude of lateral moraines, changes in surface contour shape	1980s	No MIS-2 dates	1	1	3	Williams, 1983; Zheng, 1988; Burbank and Kang, 1991; Mann et al., 1996
Mount Everest (southern side)[a]	Nepal, Tibet and China	28	86	N,S, E, W	8848	Not known		Valley	4250[a]	AAR = 0.6	5200	Cirque glaciers, highest lateral moraine, changes in surface contours	1980s	22–18 ka	>10	1	1	Williams, 1983; Zheng, 1988; Richards et al., 2000b; Finkel et al., 2003
Kanchenjunga	Nepal	27	88	N,S	8586	Not known	~2700	Valley	4200–5200	Not known	5000–6000	Not described	1992	20–21	2	1	1	Asahi and Watanabe, 2000; Tsukamoto et al., 2002
Bhutan Himalaya	Bhutan	27	90	N,S, E, W	7554	Not known	Not known	Valley	Not known	THAR = 0.5	Not known	THAR = 0.5	1990s	No MIS-2 dates	0	5	N/A	Isacks et al., 1995
Nyainqentanglha	Tibet, China	30	90	N,S	7162	Not known	4300	Valley	5300–5400	TSAM	5900–6000	TSAM	1990s	No MIS-2 dates	0	5	N/A	Lehmkuhl et al., 2002

Table 1 (*continued*)

LGM snowline locality[b]	Country	Lat °N	Lon °E	Aspect	Summit altitude (m)	Headwall altitude (m)	Lowest terminal moraine (m)	Glacier type	ELA at LGM (m)	LGM method	Modern ELA (m)	Modern ELA method	Modern ELA date	LGM date	No. dates	DMC	DC	References
Tanggula Shan	Tibet, China	33	91	N,S, E, W	6525	Not known	~5100	Valley	4900–5000	TSAM	5200–5300	TSAM	1990s	No MIS-2 dates	2	1	N/A	Lehmkuhl, 1995a, b, 1998
Kunlun Shan	Tibet, China	35	94	N,S	6860	Not known	Not described	Valley	~4000–5000	TSAM	5000–5100	TSAM	1990s	No MIS-2 dates	0	5	N/A	Lehmkuhl, 1995a, b, 1998; Derbyshire, 1996
Nianbaoyeze Mountains	Tibet, China	33	101	N,S, E, W	5369	Not known	~4050	Ice cap and valley	~4800–4900	TSAM	~5100	TSAM	1990s	21–16 ka	17	1	4	Lehmkuhl and Lui, 1994; Owen et al., 2003b
Anyemaqen Mountains	Tibet, China	34	99	N,S, E, W	6282	Not known	~3900	Ice cap and valley	~4900	TSAM	~5200	TSAM	1990s	26–9 ka	20	1	7	Lehmkuhl, 1995a, b, 1998; Owen et al., 2003b
La Ji Mountains	Tibet, China	36	101	N,E	4469	Not known	3850	Valley	4300–4400	AAR = 0.5–0.8	5000	Regional extrapolation	2003	~25–9 ka	17	1	7	Owen et al., 2003a
Qilian Shan[b]	Tibet, China	37	101	N,S	5650	Not known	~3200	Valley	4000–4300	TSAM	4900–5000	TSAM	1990s	22–7 ka	>20	1	7	Lehmkuhl and Rost, 1993; Lehmkuhl, 1995a, b; Zhou et al., 2002; Owen et al., 2003c

The dating control method (DMC) is graded as follows: 1 = Radiometric date exists for actual terminus position (e.g. cosmogenic date taken from boulder on a terminal moraine); 2 = Lithologic correlation made with radiometrically dated feature within the glacier valley (e.g. dated glacial outwash can be traced to terminus position; minimum date taken from a lake immediately upvalley of terminal moraine); 3 = Lithologic correlation made with radiometrically dated feature within region, where region maybe an individual mountain (volcano) or mountain range, in which case the dated feature is within 50 km, or (in case of range) within 200 km and with same aspect relative to climate gradients; 4 = Correlation with a radiometrically-dated regional sequence; 5 = Correlation with a generalized or global glaciation scheme. The dating control (DC) is graded as follows: 1 = Date is less than 500 years from LGM; 2 = Date is less than 1000 years from LGM; 3 = Date is less than 2000 years from LGM; 4 = Date is less than 3000 years from LGM; 5 = Date is less than 4000 years from LGM; 6 = Date is less than 5000 years from LGM; 7 = Date is greater than 5000 years from LGM. By definition, any site with DMC = 5 will not be associated with a radiometric date, and will thus not merit a DC; for the sake of mapping, any site with DMC = 5 is given DC = 8. More details on the DMC and DC are described in detail in Mark et al. (this volume).

[a]See new calculation described in this study.

[b]Where possible this is the global LGM, but in many regions this is probably the local LGM during the last glacial cycle.

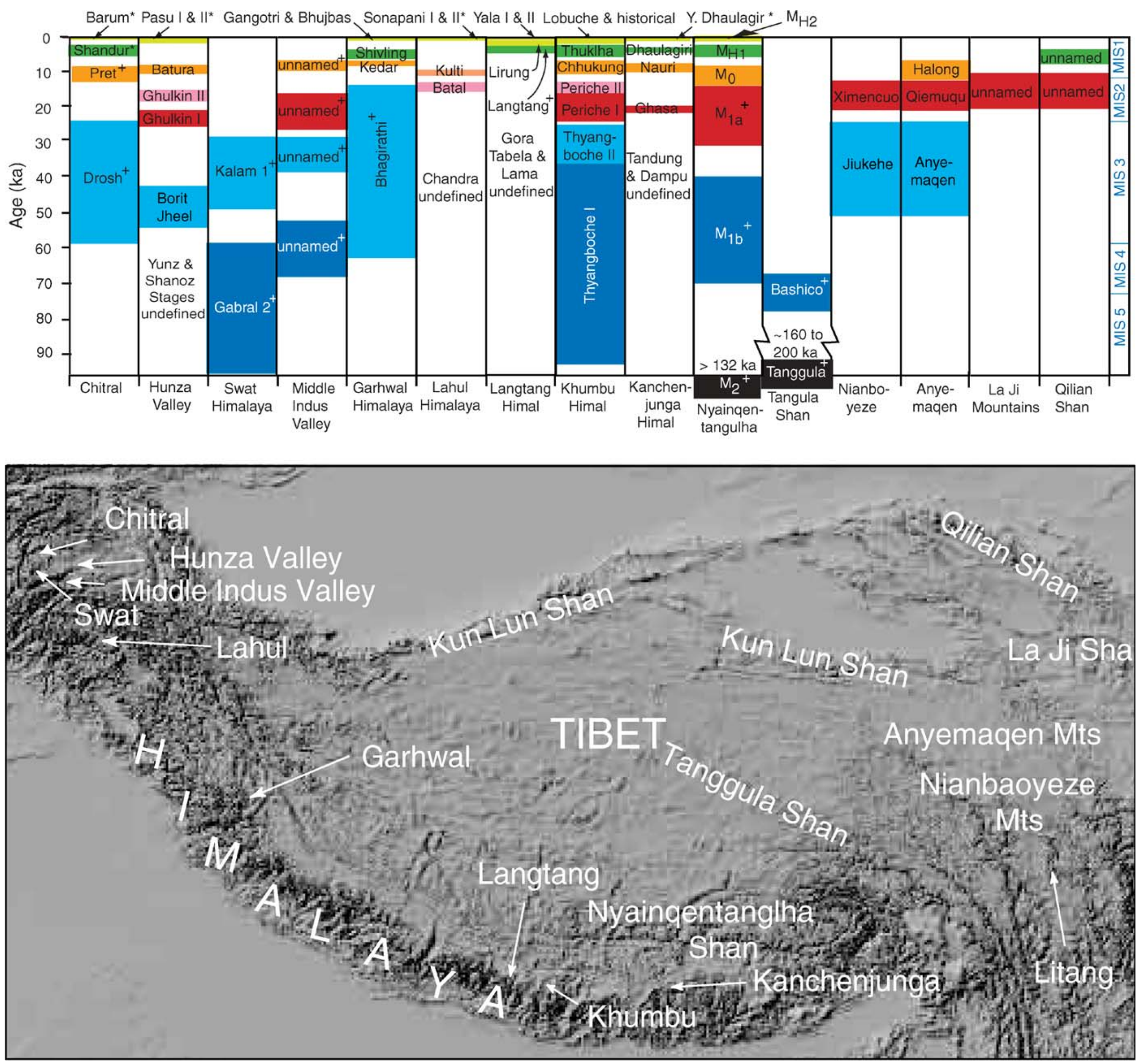

Fig. 8. Summary of the glacial chronologies that have been numerically dated in the Himalaya and Tibet. The color bars indicate the duration of each glaciation. The different colors are used to highlight probable correlations between regions. Data taken from Shiraiwa (1993), Sharma and Owen (1996), Richards et al. (2000a, b), Phillips et al. (2000), Schäfer et al. (2002) Owen et al. (2001, 2002a, b, 2003a, b), Tsukamoto et al. (2002), Lehmkuhl et al. (2002), Finkel et al. (2003) and Barnard et al. (2004). An asterisk and cross after each name indicates that no numerical dating has been undertaken to confirm an age and the duration of the glacial is poorly defined, respectively.

Nevertheless the regional pattern of variation is probably realistic. Table 1 and Fig. 8 summarize these data and it is apparent that the chronologies and reconstructions in these regions are tentative. We believe that the best areas to reconstruct ELAs for the LGM are the Khumbu Himal and Hunza valley in the Karakoram Mountains. These are the only regions where moraines have been dated using two independent methods and where there are detailed topographic maps that allow us to undertake ELA calculations. We have therefore recalculated former ELAs in these areas.

6.1. Khumbu Himal

The extent and age of former glaciers in the Khumbu Himal were established by Richards et al. (2000b) and Finkel et al. (2003) using OSL and CRN dating, respectively (Fig. 9). Eight glacial advances were defined: the Thyangboche I (>30 ka), Thyangboche II (MIS-3, 35±3 ka), Periche I (global LGM, 23±3 ka), Periche II (~16±2 ka), Chhukhung (early Holocene, 9.2±0.2 ka), Thuklha (Neoglacial, 3.6±0.3 ka), Lobuche (Little Climatic Optimum) and historical (~500 yr BP to present).

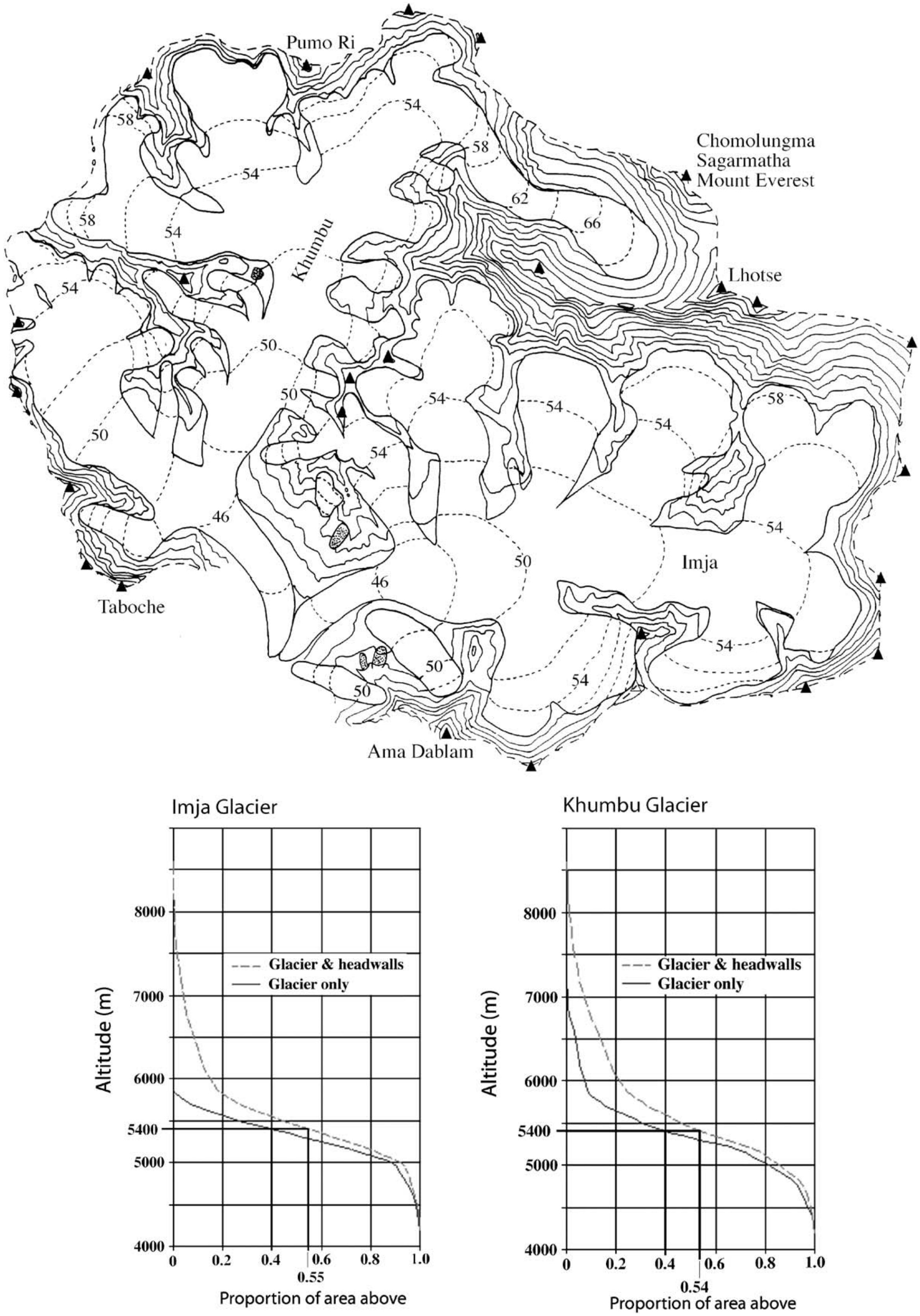

Fig. 9. Reconstruction of the Khumbu and Imja glacier systems in the Khumbu Himal for the LGM. The hyposometric curves are for the contemporary glaciers.

Table 2
Correlations, relative chronologies and numerical dates for the glacial successions in the Khumbu Himal

Radiocarbon dating of Müller (1980), Benedict (1976), Fushimi (1978)	OSL dating of Richards et al. (2000a)	CRN dating of Finkel et al. (2003)
Outer moraine (Pumore)		Historical
(~410–550 °C-14 years BP)		(<300 years BP)
Outer moraine (Tsola)	Lobuche (~1–2 ka)	Lobuche (~1 ka)
(~1150–1200 °C-14 years BP)	$n = 1$	$n = 5$
		Thuklha (3.5 ± 0.3 ka)
		$n = 3$
	Chhunkung (~10 ka)	Chhunkung (9.0 ± 0.2)
	$n = 4$	$n = 6$
	Periche (18–25 ka)	Periche II (15 ± 1 ka)
	$n = 4$	Periche I (20 ± 3 ka)
		$n = 11$
		Thyangboche II (30 ± 3 ka)
		$n = 3$
		Thyangboche I (68 ± 5 ka)
		$n = 6$

The number (n) of OSL and CRN dates are shown.

These dating results are summarized in Table 2. Richards et al. (2000b) and Finkel et al. (2003) showed that glaciers only advanced about 5 km from their present positions during the LGM.

Contemporary ELAs in the Khumbu region have been determined by Asahi et al. (personal communication), using the maximum elevation of lateral moraines and relative height of cirques, and range between 5250 and 5800 m a.s.l. The ELAs of many contemporary glaciers, however, are located on or near avalanche cones, and are thus strongly influenced by topography and do not simply reflect climatic snowlines. A trendline through the data indicates a mean value of 5600 m at 27°55′–28°00′N, the latitude of the Imja-Khumbu catchments, in close agreement with the ELA of 5600–5700 determined for the (partially snowfall-fed) Khumbu Glacier by Inoue (1977) and Müller (1980).

Reconstructed LGM (23 ± 3 ka) glaciers for the Khumbu and Imja catchments are shown in Fig. 9. Below ~5400 m, the glacier limits are constrained by extensive lateral and medial moraines, which are well preserved in both catchments. Above this altitude, the former glacier surfaces were determined from trimlines on spurs, the upper limit of ice-scoured bedrock, and interpolation. Large medial moraine complexes below spurs in both catchments occur up to 5400 m, providing a minimum altitude for the LGM glacier ELAs. Reconstruction of ELAs using other methods relies on appropriate choice of indices such as AAR and THAR. However, these are difficult to determine a priori, because of the influence of topography on glacier form and mass balance. The former glaciers (like their modern counterparts) received substantial inputs of snow and ice from high-altitude headwalls, as well as direct snowfall. Furthermore, evidence from glacier landsystems indicates that the glacier ablation zones were covered by debris, and would have exerted a strong influence on ice melt, and therefore the size of ablation area required to balance inputs. It is useful, therefore, to calculate AARs for the glaciers based for an ELA of 5400 m, because the maximum elevation of lateral moraines show that the LGM ELA cannot have been below that altitude. AARs were calculated in two ways (1) using the hypsometry of the whole catchments (glacier surface and possible avalanche source areas), and (2) the glacier surfaces alone. Values from the two catchments are closely similar: 0.54 and 0.55 for the Khumbu and Imja, respectively (glaciers and headwalls) and 0.40 for both glaciers (glaciers only). Kaser and Osmaston (2002) have argued that the true AAR for a group of former glaciers will be the value yielding minimum variation in reconstructed ELAs. If true, this supports the conclusion that the LGM ELAs of the Imja and Khumbu glaciers were close to 5400 m, because this value implies almost identical AARs for the two glacier systems. Interestingly, the AARs for the whole catchments are within the range recommended by Porter (1975) for mid-latitude snowfall-fed glaciers, suggesting that, if glacier headwalls are considered as part of the glacier system, the influence of avalanche inputs can be taken into account using standard AAR values. Further research is required to test this idea.

A similar exercise was not attempted for THARs, because headwall altitudes in both catchments are very variable, contain extremely high summits (8848 and 8501 m for the Khumbu and Imja, respectively), making representative headwall altitudes difficult to determine.

LGM ELAs of 5400 m indicate that ΔELA is 200–300 m. This figure is much lower than that for most other parts of the Himalaya, indicating that monsoon snowfall in the Khumbu at the LGM may have been much less than that of today, permitting only limited glacier expansion (Richards et al., 2000a). Testing and quantification of this conclusion is the subject of ongoing research.

6.2. Batura Glacier, Karakoram Mountains

The Batura Glacier in the upper Hunza valley of the Karakoram Mountains in Northern Pakistan is one of the longest (59.2 km) outside of the Polar regions. On its southern side, 12 glaciers advance into the main ice stream, whereas on its northern side only four glaciers contribute to the main trunk glacier. Presently, the Batura glacier has an area of $\sim$285 km^2 (Shi and Zhang, 1984; and this study). Based on glaciological studies, Shi and Zhang (1984) showed that the firn line is between 4700 and 5300 m a.s.l. with an accumulation area of 144 km^2 and an ablation area of 141 km^2. Hence the AAR is therefore approximately 0.5. Examination of changes in the surface morphology of the glacier and the highest modern lateral moraines, the apparent contemporary ELA varies between 4300–4700 (contours) and 4600–5100 m a.s.l. (lateral moraines). We also calculated the TSAM for each of the main glaciers that flow into to the trunk valley glacier. These values range from 3575 to 5315 m for individual catchments for the main ice flow. Using these estimates of the range of contemporary ELA of between 4300 and 5100 we can estimate the present accumulation area to be between 40% and 60% of the total glacier area. This is consistent with Shi and Zhang's (1984) estimate of 50%.

The reconstruction of the Batura Glacier for the LGM is based on the field mapping and CRN and OSL dating of Derbyshire et al. (1984), Owen et al. (2002a) and Spencer and Owen (2004) and the 1:60,000 scale map of the Institute of Glaciology (1978). The dating studies of Owen et al. (2002a) and Spencer and Owen (2004) provide consistent ages for the moraines and help substantiate some of the earlier dating by Derbyshire et al. (1984) (Table 3). The reconstruction for the LGM is shown in Fig. 10. The reconstruction for the ablation zone is more accurate than that for the accumulation zone. This is because moraines could be easily traced up glacier within the ablation zone, but trimlines within the accumulation zone are more difficult to identify. Our trimlines are estimated based on changes in valley form and where reconstructions are uncertain we have essentially used the area of the present ice field. This will slightly underestimate the size of the ablation area, but given the extremely steep valley size, this underestimation is negligible. Hypsometric curves for the contemporary and LGM glacier are presented in Fig. 10. The difference between these curves is small and taking any value for the AAR the ΔELA is $\leqslant$100 m. If we apply Shi and Zhang AAR of 0.5 then the ΔELA for the LGM is $<$100 m. Similarly, using an AAR of 0.4–0.6 based on our estimated ranges of the contemporary ELA, the LGM ΔELA is $\leqslant$100 m. It has also to borne in mind that our glacial reconstruction probably underestimates the size of the LGM accumulation area, which would further reduce the ΔELA.

7. Conclusions

Clearly, since studies began there has been considerable variability in methods used in reconstructing former equilibrium line altitudes (ELAs) (Table 1). Many of the methods used are not applicable for the high-altitude glaciers, which are largely controlled by topography and are strongly influenced by supraglacial debris cover, or glaciers within the semi-arid interior of Tibet. Where ΔELA have been calculated it is presumed that the calculations are based on accurate reconstructions of former glaciers and that the age of the reconstructed glacier is known. With the exception of Sharma and Owen (1996) and Lehmkuhl (1995b) there are no publications that provide accurate reconstructions of the former glaciers. Furthermore, there are only two regions within the Himalaya and Tibet where numerical dating can adequately define moraines to the LGM. These are the upper Hunza valley and the Khumbu Himal where both OSL and CRN dating confirms an LGM age on moraines. New reconstructions and calculations in these regions using a variety of techniques show that the ΔELA during the LGM was $\sim$200–300 m in the Khumbu and $\sim\leqslant$100 m for the Batura Glacier. We consider these two regions as critical sites for understanding ELAs in the Himalaya–Tibetan region and they help support the view that the ΔELA in the Greater Himalaya and Transhimalaya was small during the LGM. LGM moraines have been recognized in the Kanchenjunga Himal, but they have only been dated by OSL methods and these results need to be confirmed by an independent dating technique. There are other regions where CRN and OSL dating has been applied, but the associated uncertainty cannot unequivocally place them at LGM and they may be Lateglacial. These include the La Ji, and Qilian Shan. Cosmogenic radionuclide surface exposure and optically stimulated luminescence dating techniques are providing data to constrain the timing of glaciation.

Although ΔELAs are calculated in many papers, few attempt to convert ΔELA into changes in paleoclimate. This is probably because of the uncertainty in the relative importance of precipitation and temperature changes in creating a ΔELA. Early workers including Klute (1930) and von Wissmann (1959) and more recently Owen et al. (2002a, b) and Finkel et al. (2003) have highlighted the importance of precipitation changes in forcing glacial advances along the Himalaya.

Table 3
Quaternary glacial stages in the upper Hunza Valley, showing all the TL dates of Derbyshire et al. (1984) in plain type, the CRN dates of Owen et al.'s (2002a) in bold type, and the OSL dates of Spencer and Owen (2004) in italics

Till unit	Stage name	Dates	Description of the glaciation and the main glacial geologic evidence
t8	Pasu II	Historical (19th and 20th Centuries)	A minor advance of a few hundred meters—sharp-crested, unstable and sometimes ice-cored moraines with fresh boulders that have no rock varnish
t7	Pasu I	830±80 years[a,g] 325±60 years[a,g] *4.3±0.4 ka = t7* *8.4±0.9 ka = t7*	A minor advance of ~1 km, restricted to tributary valleys—high sharp-crested moraines with boulders that have a light yellow surface color
t6	Batura	No dates[b] **~8.5–10.3 ka** ***n* = 14** *7.8±0.7 ka <t6* *22.8±3.6 ka>t6* *28.5±2.7 ka>t6*	A glacial advance of 1–2 km—well defined moraines with strong varnished and weathered boulders
t5	Ghulkin II	No dates **~14.9–18.2 ka** ***n* = 14** *8.7±0.8<t5* *12.0±1.1 ka = t5* *18.0±1.7 ka>t5*	Minor glacier advance of several kilometers—a multiple series of rounded moraine ridges with strong to post-maximal rock varnish, deeply weathered boulders with a weak carbonate development beneath boulders
t4	Ghulkin I	47.0±2.35 ka[c,g] **~21.9–23.5 ka** ***n* = 12** *18.4±1.6 ka<t4* *31.5±3.0 ka = t4* *31.6±3.3 ka = t3/t4* *44.4±5.2 ka>t4*	Minor valley glaciation, expanding into diffluence cols and into the main Hunza valley—well defined moraines with very strong to post-maximal rock varnish, deeply weathered boulders deeply with cavernous weathering and incipient calcrete and pendant growth of carbonate
t3	Borit Jheel	65.0±3.3 ka[d,g] 50.0±2.5 ka[e,g] **~52.8–60.4 ka** ***n* = 14**	Main valley glaciations with tributary valley glaciers filling and locally overtopping diffluence cols—highly eroded moraines with deeply weathered boulders that have a very strong to post maximal shiny black rock varnish, and an extensive underlying calcrete
t2	Yunz	139.0±12.5 ka[f,g] **>60 ka** ***n* = 10**	Extensive main valley glaciation—deeply weathered till remnants on benches in the main Hunza valley at an altitudes of ~3900 m a.s.l. on the upper western slopes of the Pasu–Ghulkin diffluence col
T1	Shanoz	No Dates	Extensive broad valley glaciation—deeply weathered erratics on summit surfaces >4150 m a.s.l.

Table adapted from Owen et al. (2002a). The number (*n*) of CRN dates is shown. (= t – OSL date equivalent to stage age; >t – OSL date older then stage age; <t – OSL date younger than stage age).

[a]Uncalibrated radiocarbon dates on wood from a moraine of the Minapin Glacier, middle Hunza valley.
[b]Derbyshire et al. (1984) considered this glacial stage to be mid-Holocene in age.
[c]TL date on lacustrine silts intercalated with Ghulkin I till of moraines from the Pisan Glacier in the middle Hunza valley.
[d]TL date on glaciolacustrine sediments, that were probably deposited during the Borit Jheel Stage, ~ 1.5 km north of the present snout of the Pasu glacier.
[e]TL date on glaciotectonized glaciolacustrine sediments within Borit Jheel till ~2.5 km south of the snout of the Batura glacier.
[f]TL date of on glaciolacustrine silts beneath Borit Jheel till southeast of the Ghulkin Glacier.
[g]No details of the laboratories and the laboratories procedures are provided in Derbyshire et al. (1984).

This is exemplified by the local LGM coinciding with times of insolation maximum when the monsoon would have been more effective in supplying precipitation (as snowfall) to glacial catchments resulting in positive glacier mass balances.

Regional reconstructions of former ΔELA for the LGM, like those presented in von Wissmann (1959), Frenzel (1960), Shi et al. (2000), Shi (2002), and Lehmkuhl and Owen (2005), need to be considered with great care. Although the patterns presented in these reconstructions are probably realistic, the magnitude of ΔELA needs to be substantiated with geologic data (Figs. 6 and 7). In particular, these reconstructions presume that the landforms and sediments that are used for the reconstructions were produced during the same glaciation. In fact, the numerical data are so sparse and

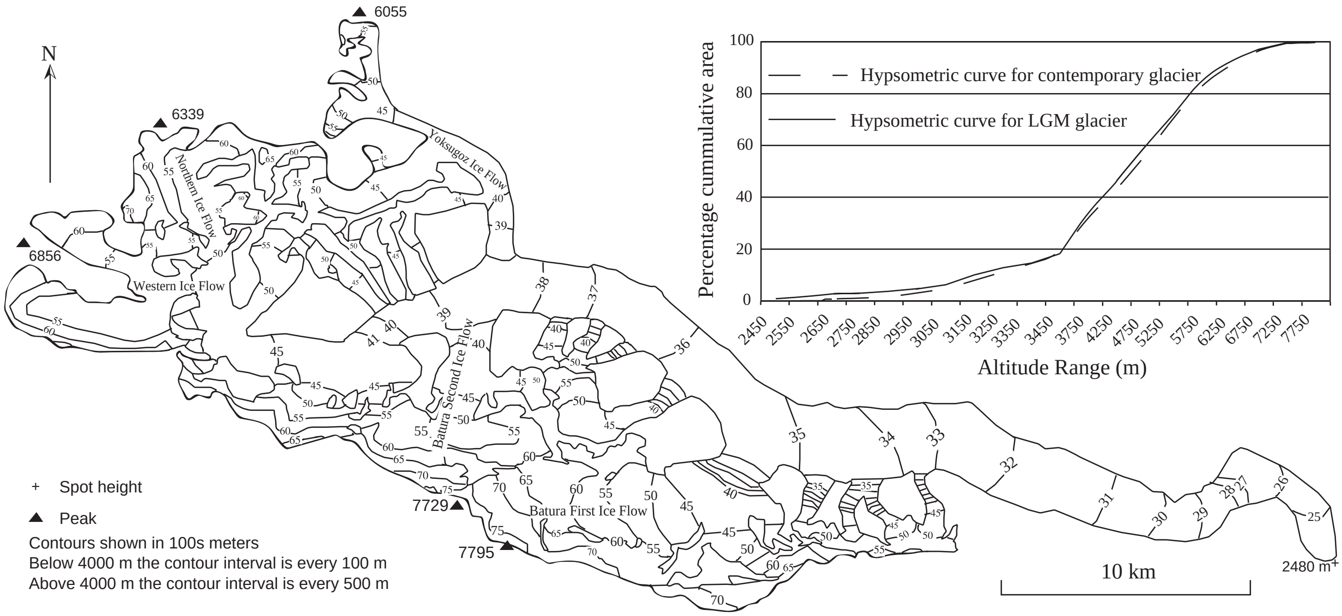

Fig. 10. Reconstruction of the Batura glacier in the Hunza valley for the LGM.

regional correlations based on relative dating are extremely problematic (cf. Owen et al., 1997). Contemporary ELAs for glaciers throughout the high mountains of the Himalaya and Tibet show a similar regional pattern (Fig. 3). Given, therefore, that the past and present regional ELA pattern is clearly a consequence of the mid-latitude westerlies and the south Asian summer monsoon, it is likely that the region ΔELA variability is similar to that presented by Shi et al. (2000), Shi (2002) and Lehmkuhl and Owen (2005). The relative importance and influence of these two climatic systems, however, needs to be quantified to substantiate that the regional variability is similar to today. Of course the magnitude of ΔELA clearly needs to be calculated. Furthermore, there are strong topographic controls on ELAs, often producing contrasting ELA values between adjacent glaciers.

In summary, we present a compilation of the published and new ELA data in Table 1. The ELA data is evaluated using the methods established by Mark et al. (2003), where the dating method control (DMC) is ranked 1–5 (1 is the best control) and dating control (DC) is ranked 1–7 (1 is the most precise). Any site with a DMC of 5 is given a DC of 8. We emphasize that for all the data available there are only 6 sites that are even vaguely acceptable from a dating point of view, and we believe that the Hunza and Khumbu sites are the only credible sites at present for valid ELA reconstructions for the LGM because two different sets of different dating technique studies have been undertaken for each region. It is therefore not reasonable at this stage to argue or discuss regional variability in the ΔELAs for the LGM for Tibet and the bordering mountains. We hope, however, that this summary and review will provide a framework for further studies of present and past ELAs throughout the Himalaya and Tibet that will ultimately aid in accurate paleoenvironmental reconstructions.

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