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Quaternary glaciation of the Himalayan–Tibetan orogen

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ABSTRACT: Glacial geological evidence from throughout the Himalayan–Tibetan orogen is examined to determine the timing and extent of late Quaternary glaciation in this region and its relation to similar changes on a global scale. The evidence summarised here supports the existence of expanded ice caps and extensive valley glacier systems throughout the region during the late Quaternary. However, it cannot yet be determined whether the timing of the extent of maximum glaciation was synchronous throughout the entire region or whether the response was more varied. The lack of organic material needed for radiocarbon dating has hindered past progress in glacial reconstruction; however, application of optically stimulated luminescence and terrestrial cosmogenic radionuclide methods has recently expanded the number of chronologies throughout the region. Limits to the precision and accuracy available with these methods and, more importantly, geological uncertainty imposed by processes of moraine formation and alteration both conspire to limit the time resolution on which correlations can be made to Milankovitch timescales (several ka). In order to put all studies on a common scale, well-dated sites have been re-evaluated and all the published terrestrial cosmogenic nuclide ages for moraine boulders and glacially eroded surfaces in the Himalayan–Tibetan orogen have been recalculated. Locally detailed studies indicate that there are considerable variations in the extent of glaciation from one region to the next during a glaciation. Glaciers throughout monsoon-influenced Tibet, the Himalaya and the Transhimalaya are likely synchronous both with climate change resulting from oscillations in the South Asian monsoon and with Northern Hemisphere cooling cycles. In contrast, glaciers in Pamir in the far western regions of the Himalayan–Tibet orogen advanced asynchronously relative to the other regions that are monsoon-influenced regions and appear to be mainly in phase with the Northern Hemisphere cooling cycles. Broad patterns of local and regional variability based on equilibrium-line altitudes have yet to be fully assessed, but have the potential to help define changes in climatic gradients over time. Copyright © 2008 John Wiley & Sons, Ltd.

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KEYWORDS: mountain glaciation; Himalaya; Tibet; geochronology; monsoon.

Introduction

The Himalaya–Tibetan orogen is the most glaciated mountain region outside of the polar realms ($\sim$126 200 km^2; Haeberli *et al.*, 1989). The economies of countries within and bordering the orogen are mostly agriculture based and are thus sensitive to variations in the glacial and hydrological systems that provide much of the water to these regions. The Quaternary geological archive on the Tibetan Plateau can be used to reconstruct and model the effects of environmental change on the regional climate and hydrology of the orogen and adjacent regions. Changes in the extent of glaciers and snow cover on the Tibetan Plateau provide significant feedbacks in climate modelling, which requires understanding the nature of glacial fluctuations (Dey and Bhanu Kumar, 1982; Dey *et al.*, 1985; Bush, 2000). Tibet is also an ideal natural laboratory to examine the relationship between global climate and the processes of continental–continental collisions, tectonic aneurysm (Zeitler *et al.*, 2001) and glacial buzzsaw (Porter, 1989; Brozovic *et al.*, 1997) models.

The extent and timing of glaciation and the associated hydrological and climatic responses in the Himalayan–Tibetan orogen are poorly defined owing to the inaccessibility to the region. Accurate reconstructions of former ice extent have been hindered by the difficulty in distinguishing glacial from

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non-glacial landforms. There are few opportunities to apply numerical dating tools, many of which have large errors, which could define the ages of landforms and that in turn can establish regional correlations (Benn and Owen, 1998, 2002). Reconstructions of the extent of former glaciations vary considerably between researchers and few of these studies have been rigorously evaluated in light of new methodologies. In this paper, we review the current knowledge on the Quaternary glaciation of the Himalaya–Tibetan region and critically evaluate the reconstructions of the timing and extent of Quaternary glaciation. We focus mainly on the late Quaternary and select important regions within the orogen for detailed consideration.

This paper builds on a series of research papers published in three *Quaternary International* volumes (Owen and Lehmkuhl, 2000; Owen and Zhou, 2002; Yi and Owen, 2006), summaries for the Global Mapping Project of INQUA on the glacial geology of each country in central Asia (Ehlers and Gibbard, 2004) and a review by Lehmkuhl and Owen (2005).

Physical setting

Collision of the Indian and Eurasian continental plates, initiated ca. 50 Ma ago (Yin and Harrison, 2000), produced the Himalayan–Tibetan orogen. Geologically, the region is a complex assemblage of rocks of different ages that are still being actively deformed by the continued northward movement of the Indian plate at $\sim$50 mm a^{-1} (DeMets *et al.*, 1994; Yin and Harrison, 2000). The region stretches $\sim$2000 km east–west and $\sim$1500 km north–south, and has an average altitude of $\sim$5000 m above sea level (a.s.l.) (Fielding *et al.*, 1994). A series of approximately east–west trending mountain ranges are present. These include, from south to north, the Siwaliks, Lesser Himalaya, Greater Himalaya, Transhimalaya, Nyainqentanggulha Shan, Tanggula Shan, Bayan Har Shan, Kunlun Shan, Qilian Shan and Pamir. The Tian Shan and Altai Mountains north of the Tibetan Plateau are a distant manifestation of the continental–continental collision and will not be considered. We focus on the contiguous mountain mass to the south and south-west of the great deserts of China and Mongolia (Fig. 1).

Two climatic systems dominate this vast region: the mid-latitude westerlies and the South Asian monsoon (Benn and Owen, 1998; Fig. 2). Long-term variability (thousands of years) of the mid-latitude westerlies and the South Asian summer monsoon is linked to changes in Northern Hemisphere insolation, while shorter-term variability (years to decades) is attributed to changes within the climate system, such as variation in Eurasian snow cover, the El Niño–Southern Oscillation, and tropical sea surface temperatures (Hahn and Shukla, 1976; Dey and Bhanu Khumar, 1982; Dickson, 1984; Sirocko *et al.*, 1991; Prell and Kutzbach, 1992; Soman and Slingo, 1997; Bush, 2000, 2002; Rupper, 2007). In addition, model results suggest that recent anthropogenic forcing (e.g. greenhouse gas, land cover change) is becoming important in controlling variations in the Asian monsoon (Wake *et al.*, 2001).

Most of the southern and eastern region experiences a pronounced summer precipitation maximum, reflecting moisture advected northwards from the Indian Ocean by the south-west monsoon (Murakami, 1987). Summer precipitation declines sharply from south to north across the main Himalayan chain and is low over western Tibet (Fig. 3(A) and (B)). There is a winter precipitation maximum only in the extreme west of the orogen due to the influence of winter westerly winds bringing moisture from the Mediterranean, Black and Caspian seas.

Snow and ice accumulation varies across the orogen as a consequence of changing moisture contents of air masses forced over the Himalayan–Tibetan orogen. The orographic forcing produces drier air that eventually descends and warms the northern slopes of the Transhimalaya and the northern edge of the Tibetan Plateau. These winds may have a major control on glacier formation. A combination of the above factors results in the variation of equilibrium-line altitudes (ELAs) from 4300 m a.s.l. in south-east Tibet to over 6000 m a.s.l. in western Tibet (Owen and Benn, 2005). The seasonal distribution of precipitation creates glaciers of the summer accumulation type, with maxima in accumulation and ablation occurring more or less simultaneously during the summer, in most of the region (Ageta and Higuchi, 1984; Owen and Derbyshire, 1989; Benn and Owen, 1998; Ageta and Kadota, 1992; Kulkarni,

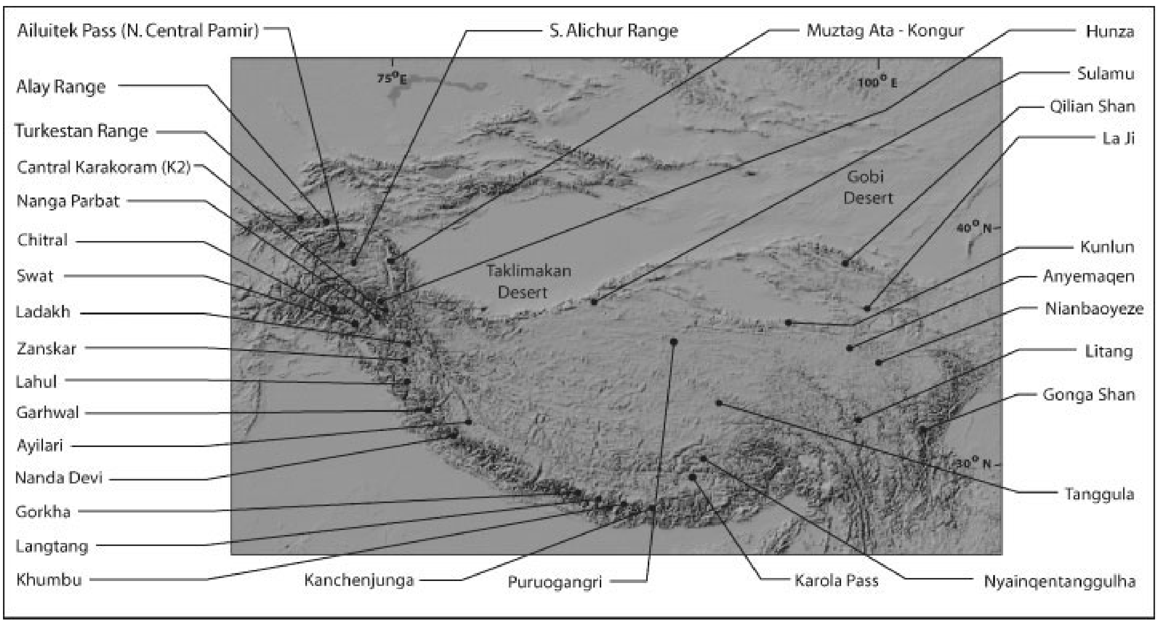

Figure 1 Major study regions in the Himalayan–Tibetan orogen that are referenced in the text

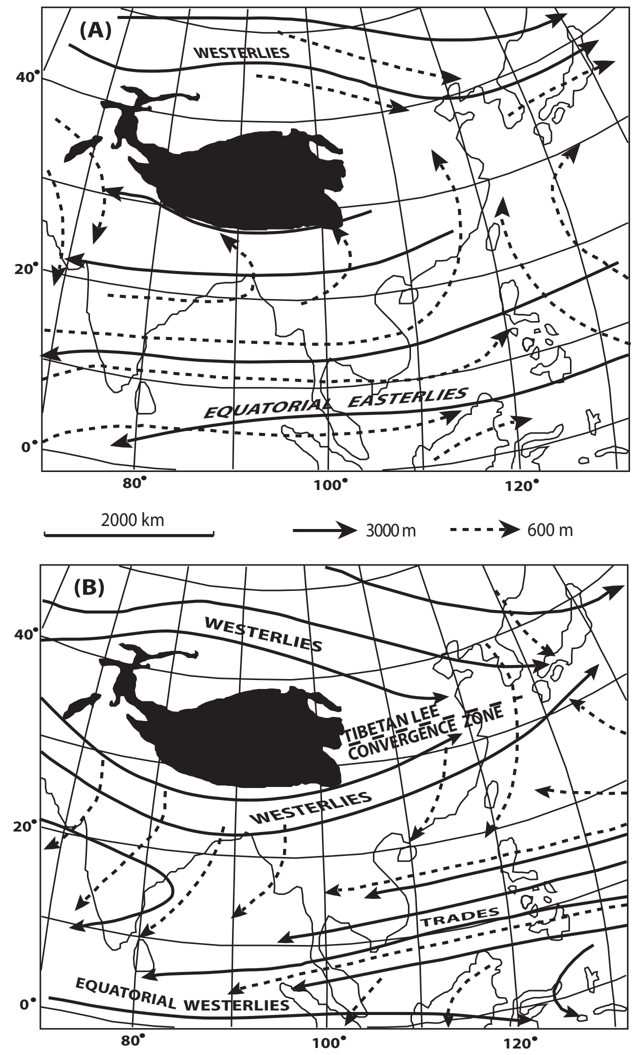

Figure 2 Characteristic air circulation over the Eastern Hemisphere for (A) summer (July) and (B) winter (January). The solid lines indicate airflow at ~3000 m for winter (A) and ~6000 m a.s.l. for summer in (B) and the dashed lines at ~600 m in (A) and (B). Names refer to air systems. The area of Tibet and the Himalaya is shown in black. Adapted from Barry and Chorley (2003)

1992; Nijampurkar and Rao, 1992). The summer accumulation maximum reflects high precipitation during the summer monsoon, when moisture-bearing air masses originating over the Indian Ocean bring snow to high altitudes in the Himalaya and western Tibet (Yasunari and Inoue, 1978; Higuchi *et al.*, 1982; Ageta and Higuchi, 1984). Glacier ablation is also at a maximum during the summer, when high temperatures result from direct solar heating, advected sensible heat, and latent heat released by condensation during deep atmospheric convection (Webster, 1987a,b). In contrast, winters are cold and relatively dry, although heavy snowfalls can occur in association with westerly winds, particularly in the western parts of the Himalayan–Tibetan orogen (Inoue, 1978; Benn and Owen, 1998). Glacier accumulation occurs by direct snowfall, blowing snow and avalanching. The relative importance of these three processes varies considerably between glaciers. Avalanching plays an important role in many catchments due to the widespread occurrence of high, steep mountain sides that capture, but cannot retain snowfall (Benn and Owen, 1998; Benn and Lehmkuhl, 2000). Continuous, thick debris cover insulates the underlying ice, and can significantly reduce ablation (Inoue, 1977).

Derbyshire (1982) recognised three main glacial regimes across the Himalayan–Tibetan region. These included glaciers with a continental regime, which by extent are the most dominant, spanning the Transhimalaya and western and central Tibet; a maritime regime in the Himalaya and eastern Tibet, which can be subdivided into a western group and an eastern (monsoon-dominated) group; and a transitional regime in the northeastern part of Tibet. Derbyshire (1982) highlighted the differences in the geomorphology, lithofacies and land systems types associated with each glacier regime. This has been further expanded by research in specific regions of the Himalaya (e.g. Owen and Derbyshire, 1988, 1989; Benn and Owen, 2002). However, given the vast scale of the region there is generally an absence of detailed geomorphic and sedimentological studies of Himalaya and Tibetan glacial systems.

Extent of glaciation

Reconstructing the extent of glaciation throughout the Himalayan–Tibetan orogen has a long history. Early explorers into the region included Thompson (1849), Cunningham (1854), Godwin-Austin (1864) and Conway (1894). They described the geomorphic and sedimentary features they saw on their expeditions, but commented little on their significance. Early important contributions to the Quaternary geology were presented by Drew (1873), Dainelli, (1922, 1934, 1935), Norin (1925), Trinkler (1930), Misch (1935), de Terra and Paterson (1939) and Paffen *et al.* (1956) for regions throughout the western end of the Himalayan–Tibetan orogen. The first comprehensive map for the entire Himalayan–Tibetan region of the extent of glaciation for the Last Glacial was provided by Klute (1930; Fig. 4(A)). This was later modified by Frenzel (1960), who used the detailed work, based on the observations of the earliest explorers, of von Wissmann (1959) (Fig. 4(B)).

During the late 1980s and early 1990s, Kuhle (1985, 1986, 1987, 1988a,b, 1990a,b, 1991, 1993, 1995) argued that an extensive ice sheet covered most of the Tibetan Plateau during the Last Glacial (Fig. 4(C)). This was disputed by many and it stimulated much research in Tibet, especially as China began to open its borders to western scientists and strengthened its international research collaborations. Over the past few decades, numerous publications have presented the evidence against an extensive ice sheet (Derbyshire, 1987; Zheng, 1989; Shi, 1992; Hövermann *et al.*, 1993a,b; Lehmkuhl, 1995, 1998; Rutter, 1995; Lehmkuhl *et al.*, 1998; Zheng and Rutter, 1998; Schäfer *et al.*, 2002; Owen *et al.*, 2003a). Now it is generally accepted that a large ice sheet did not cover the Tibetan Plateau, at least not during the past few glacial cycles.

Most recently, Chinese workers have compiled synthesis maps to show the probable extent of former glaciation across the Tibetan Plateau (Shi *et al.*, 1986, 1993, 2005; Liu *et al.*, 1988; Shi, 1988, 1992; Li *et al.*, 1991; Fig. 4(D)). Li *et al.* (1991) perhaps provide the best reconstruction of the maximum extent of glaciation for the entire Himalaya and Tibetan Plateau for the last glacial cycle. However, it is unlikely that the reconstructed glaciers reached their maximum extents at the same time during the Last Glacial (see discussion below). Recently, remote sensing combined with field mapping is allowing extensive areas to be mapped in detail (e.g. Stroeven *et al.*, 2008).

The inconsistencies between the various reconstructions of former ice extent stem from three main reasons. Firstly, the immense size and inaccessibility of the region have led many early researchers to extrapolate their former glacier margins through terra incognito lacking strong geological evidence. The

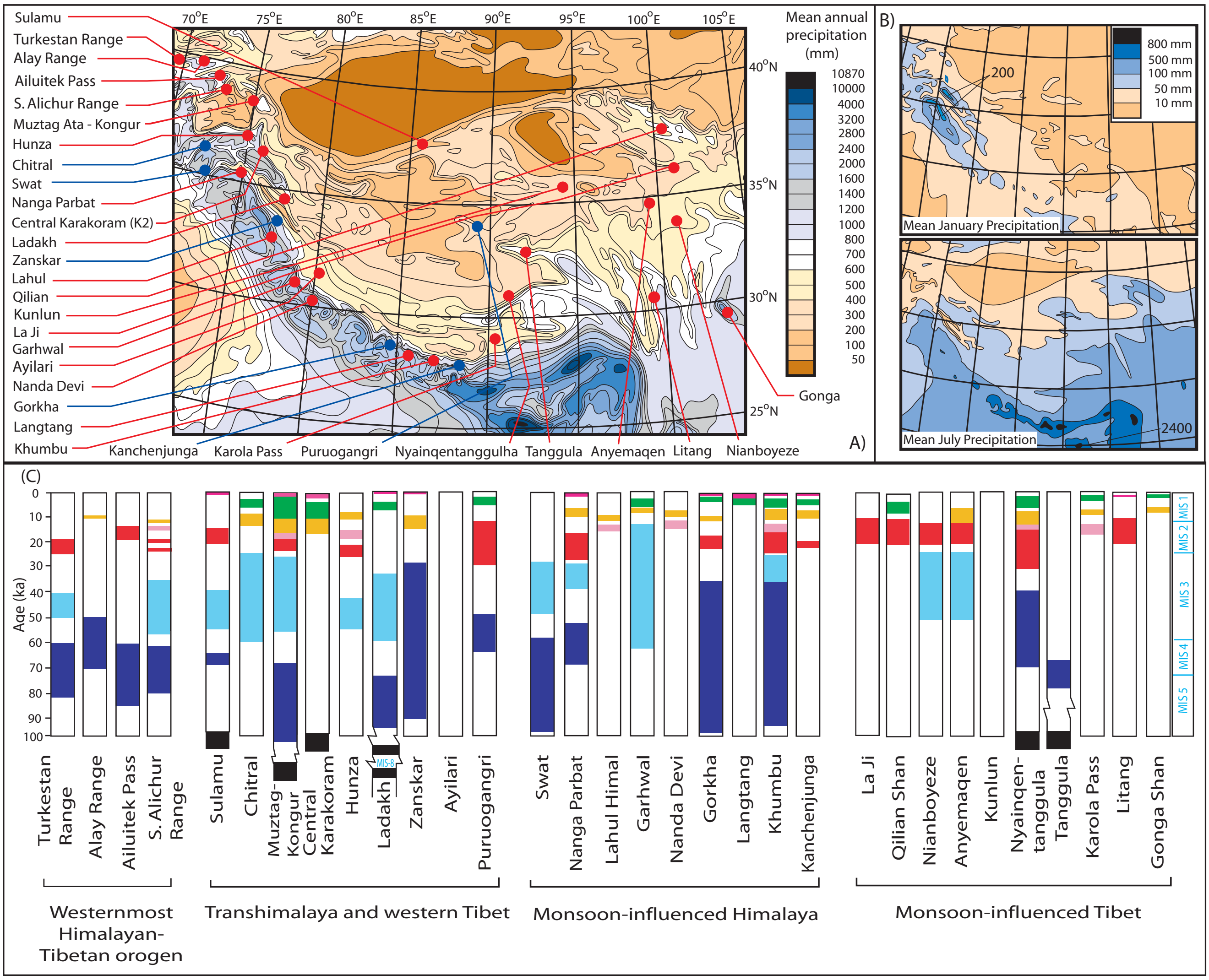

Figure 3 Contemporary mean annual (A) and mean January and July (B) precipitation across Tibet and the bordering regions showing (C) the locations and glacial chronologies that have been numerically dated throughout the Himalaya and Tibet (expanded from Owen *et al.*, 2005, and data from Shiraiwa, 1993; Sharma and Owen, 1996; Phillips *et al.*, 2000; Richards *et al.*, 2000a,b; Owen *et al.*, 2001, 2002a,b, 2003a,b,c, 2005, 2006a,b; Schäfer *et al.*, 2002; Tsukamoto *et al.*, 2002; Yi *et al.*, 2002; Finkel *et al.*, 2003; Zech *et al.*, 2003, 2005; Barnard *et al.*, 2004a,b, 2006a,b; Meriaux *et al.*, 2004; Spencer and Owen, 2004; Chevalier *et al.*, 2005; Abramowski *et al.*, 2006; Colgan *et al.*, 2006; Seong *et al.*, 2007, 2008a). The red dots in part (A) highlight regions where TCN surface exposure dating has been undertaken. The colour bars in part (C) represent the likely duration of each glacial advance based on the best estimate of the ages of moraines presented in the original publications. A tentative correlation is suggested by applying similar colours to the bars. Marine Isotope Stages (MISs) are taken from Martinson *et al.* (1987) and Shackleton *et al.* (1990)

recent increased availability and use of remote-sensing technologies has reduced this problem considerably. Access to many regions has considerably improved since the 1970s after new roads were built across the region.

The second reason stems from the misinterpretation of sediment and landforms. Owen *et al.* (1998) highlighted that much of the landform and sedimentary record within the Himalayan–Tibetan orogen has undergone intense mass wasting, as well as fluvial and glacial erosion. This has often resulted in the destruction of diagnostic morphologies and in rapid reworking of sediment and landforms (Derbyshire, 1983; Owen, 1991; Benn and Owen, 2002). In the case of till, one common result is that they are commonly confused with debris flow or rock avalanche deposits, and vice versa. This is particularly problematic when mass movement deposits contain faceted and striated clasts, which pose the question as to whether the sediments were derived directly or indirectly from glacial ice or from reworked and resedimented tills. The situation may be further complicated when till clasts derived from sediments within the passive flow paths of steep glaciers are left unstriated. Owen *et al.* (1998) termed this the 'diamicton problem'.

Genetic discrimination between landforms and the sediments (notably diamictons) can be made by careful sedimentological and geomorphic analysis for palaeoenvironmental reconstructions (Derbyshire, 1982, 1983; Derbyshire and Owen, 1990; Owen and Derbyshire, 1988, 1989; Li *et al.*, 1984; Fort and Derbyshire, 1988; Hewitt, 1999). Owen (1988, 1991, 1993) and Benn and Owen (2002) presented a set of sedimentological and geomorphic criteria based on contemporary and Quaternary examples from the Karakoram of northern Pakistan and the Himalaya of northern India that provides a useful framework for accurate reconstructions of former glacier extent.

The third reason is a consequence of poor chronological control for the glacial successions. Defining the ages of landforms and sediments throughout much of the Himalayan–Tibetan orogen has been particularly difficult because of the

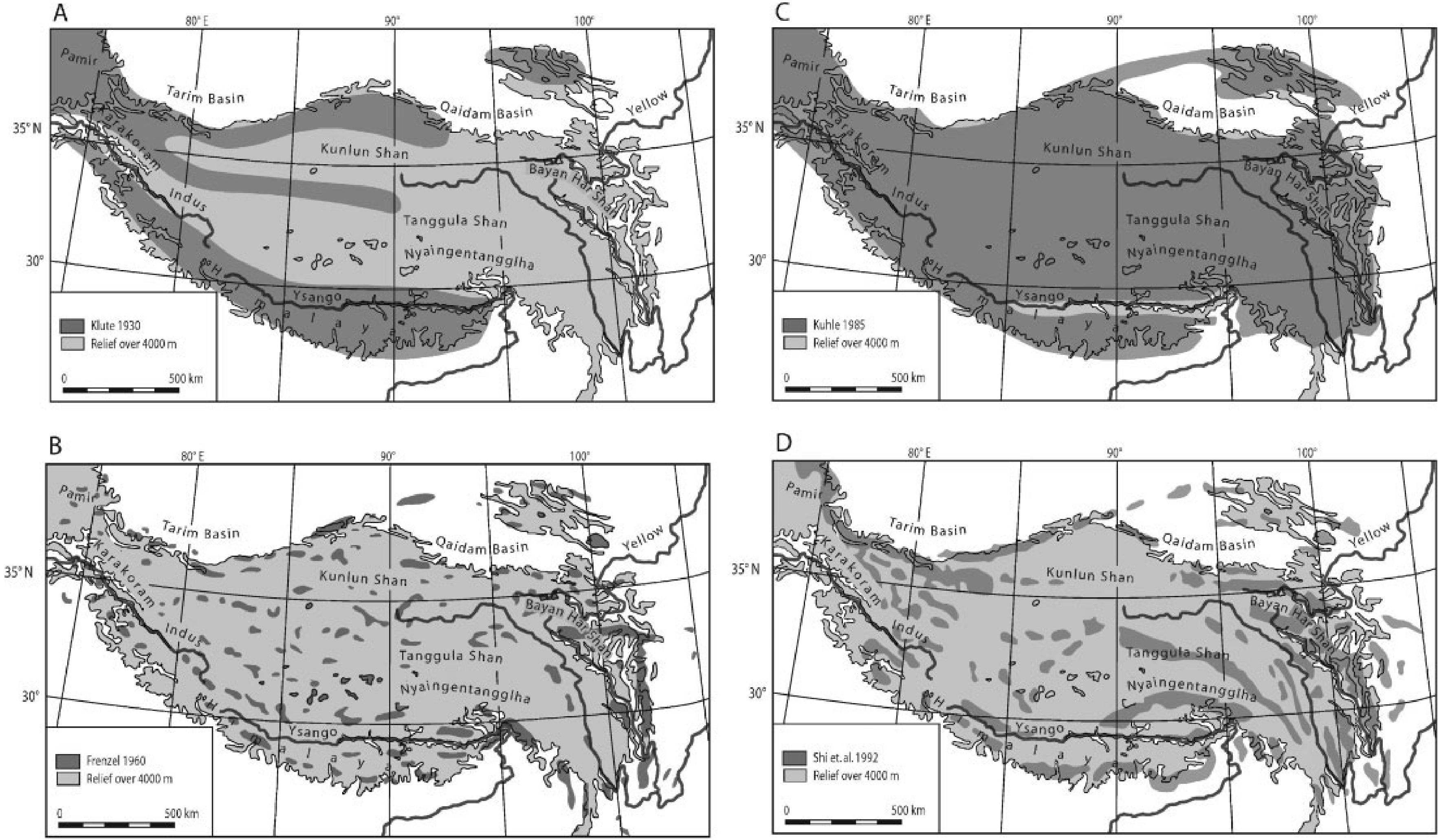

Figure 4 Selected reconstructions for the extent of glaciation (shown in dark grey) across Tibet and the bordering mountains for the maximum extent of glaciation during the Last Glacial. (A) Klute's (1930) reconstruction based on a temperature depression of 4°C, with a shift of climatic zones to the south and an intensification of atmospheric circulation such that precipitation increased towards the dry areas of central Asia. (B) Reconstruction of Frenzel (1960) based on the detailed work of von Wissmann (1959), who evaluated the observations of the earliest explorers. (C) Model of Kuhle (1985) based on field observations and extrapolation of large ELA depressions (>1000 m) from the margins of Tibet into the interior regions. (D) Reconstruction of Li *et al.* (1991) based on detailed field mapping of glacial and associated landforms and sediments

lack of organic material for radiocarbon dating. Radiocarbon dates are available for the wetter parts of the Himalayan–Tibetan region, but most are limited to Holocene proglacial deposits, which poorly define the timing of glaciation (Röthlisberger and Geyh, 1985; Lehmkuhl, 1995, 1997). The newly developing techniques of optically stimulated luminescence (OSL) and terrestrial cosmogenic nuclide (TCN) surface exposure dating are now allowing glacial successions throughout Tibet and the bordering mountains to be dated and correlated more accurately (see next section). Many of the reconstructions are based on correlating landforms, which have not been numerically dated, across vast regions. The resultant reconstructions are spurious when the landforms are not of the same age.

Despite these problems, there is substantial literature that presents detailed maps of the former extent of glaciation for specific areas throughout the Himalayan–Tibetan orogen. These are described in more detail in Ehlers and Gibbard (2004) and summarised in Lehmkuhl and Owen (2005). This shows that former glaciation was dominated by extensive valley glacier systems and expanded ice caps.

The extent of glaciation between adjacent regions within the Himalayan–Tibetan orogen can vary considerably. This is well illustrated across a stretch of northern Pakistan and northern India for glaciation during the Lateglacial at about 14–16 ka (Fig. 3). In the Hunza valley to the north-east, for example, glaciers only advanced a few kilometres from their present position, whereas in the Central Karakoram to the east an extensive valley glacier system extending more than 100 km from the present ice margin occurred (Owen *et al.*, 2002a; Seong *et al.*, 2007). To the south-east of these regions, in Ladakh, there is little evidence of a glacier advance at this time, and when glaciers did advance they were restricted to not more than a few kilometres from their present ice margins (Owen *et al.*, 2006a) with ELA depressions (ΔELAs) of a few hundred metres. To the south of Ladakh, in the Lahul Himalaya, glaciers advanced more than 100 km beyond their present positions at this time (Owen *et al.*, 2001). These contrasts in the extent of glaciation within a relatively small region of the Himalayan–Tibetan orogen highlight the important local climatic gradients and the strong topographic controls.

Timing of glaciation

Early studies of glaciation in the Himalayan–Tibetan orogen concentrated on developing relative glacial chronologies using morphostratigraphy aided by relative weathering studies (e.g. Burbank and Kang, 1991; Hövermann *et al.*, 1993a,b; Hövermann and Lehmkuhl, 1994; Lehmkuhl, 1995, 1997, 1998; Owen *et al.*, 1997; Lehmkuhl *et al.*, 2000) and soil development (e.g. Bäumler *et al.*, 1997; Guggenberger *et al.*, 1998; Zech *et al.*, 2000). These were supplemented by the application of radiocarbon methods. The first significant geochronological study was undertaken by Röthlisberger and Geyh (1985), who obtained 68 radiocarbon dates from the forelands of 16 different glaciers throughout the Himalaya and Karakoram Mountains. The dates showed significant advances at ca. 19 000, 12 700, 7400, 4900–4600, 3700–3100, 2700–2100, 1700–1500, 1200–950, 800, 500 and 400–100 ^{14}C a BP. Subsequent studies present only a few radiocarbon ages (e.g. Derbyshire *et al.*, 1991; Pachur *et al.*, 1994; Mann *et al.*, 1996).

With the development of luminescence dating throughout the 1980s and 1990s, luminescence methods were applied to many of the glacial successions throughout Himalaya and Tibet. Of note was the work by Derbyshire *et al.* (1984), who provided some of the first ages on glaciogenic sediments in their study of the upper Hunza valley in Pakistan. Unfortunately, no descriptions of the methods were provided and it is difficult to evaluate their validity critically. However, later studies by Owen *et al.* (2002a) and Spencer and Owen (2004) support the ages presented in Derbyshire *et al.* (1984).

The most comprehensive application of luminescence dating to glacial successions in the Himalayan–Tibetan orogen was undertaken by Richards (1999). He provided a suite of OSL ages for moraines in northern Pakistan (Richards *et al.*, 2000a) and the Khumbu Himal in Nepal (Richards *et al.*, 2000b). The luminescence dating was applied to proglacial sediments beneath and above tills, and supraglacial sediments intercalated within tills to provide OSL ages that were maximum, minimum and contemporaneous with glacial advances. Significant problems, however, are associated with the application of luminescence dating to glaciogenic and associated sediments. These include partial bleaching, dose rate changes over time (particularly with regard to changing moisture content), and in the case of Himalayan sediments, poor sensitivity and the low brightness of quartz. An incorrect estimate for moisture content, for example, can result in a 1% difference in age for every 1% difference in moisture content. Partial bleaching is probably the greatest problem, however, and can result in an overestimate of the OSL age. As such, OSL ages under most circumstances may be considered as maximum age estimates. These factors are discussed in detail in Richards (2000) and Fuchs and Owen (2008). An assessment of these factors and the uncertainty associated with the method can be made by examining the protocols, methods and calculations used to determine the OSL ages. Unfortunately, many published results do not provide sufficiently detailed data and descriptions to assess and independently calculate the OSL ages. This makes it extremely difficult to assess the validity of many published OSL ages. In the discussions below, therefore, we have not recalculated or corrected any of the published OSL ages.

Dating of glacial successions throughout the Himalayan–Tibetan orogen during the 1990s was driven mainly by the desire to define the timing of glaciation in mountain regions throughout the world to test whether glaciers responded synchronously with the Northern Hemisphere ice sheets and between mountain regions This work was stimulated by Gillespie and Molnar (1995) who, based on evidence throughout the mountainous regions of the world, suggested that regional glaciations were asynchronous with each other, and that glaciers did not all reach their maximum extents at the same times as the high-latitude ice sheets, which was a common assumption at the time. Gillespie and Molnar (1995) further suggested that, even within Asia, advances from one region to another were asynchronous. Some likely reached their maximum extents early in the last glacial cycle, before the global Last Glacial Maximum (LGM) at ca. 18–25 ka). However, in 1995 the data for the Himalayan–Tibetan region were too sparse to demonstrate with confidence the degree of synchrony or asynchrony from region to region within Asia. Benn and Owen (1998), using newly published ages, were able to demonstrate that during the last glacial cycle glacial advances were indeed not synchronous throughout the Himalayan–Tibetan orogen. Rather, in some areas glaciers reached their maxima at the global LGM, whereas in others glaciers were most extensive at ca. 60–30 ka during Marine Isotope Stage (MIS) 3. Benn and Owen (1998) suggested that comparison of these data with palaeoclimatic records from adjacent regions showed that, on millennial timescales, Himalayan glacier fluctuations are controlled by variations in both the South Asian monsoon and the mid-latitude westerlies.

In addition to these two hypotheses, researchers needed a chronological framework to date glacial and associated landforms to test new paradigms in tectonics and geomorphology, such as the tectonic aneurysm and glacial buzzsaw models. This increased the application of geochronological methods to dating glacial successions.

The dating of Richards *et al.* (2000a,b) provided the first real attempt to test the palaeoclimate hypotheses in Asia. Their dating showed that glacial advances in Pakistan and Khumbu were more extensive in the earlier part of the last glacial cycle, but that glaciers did advance during the global LGM. However, their study was not geographically extensive enough to adequately test these hypotheses fully.

The development and application of TCN surface exposure dating throughout the 1990s and early 2000s to glacial successions provided a means to test these hypotheses more adequately. Data from these studies are summarised in Fig. 3. Currently, there are >800 TCN surface exposure ages published in the literature for glacial and associated landforms in the Himalayan–Tibetan orogen. Most of the TCN surface exposure dating use ^{10}Be because it is by far the simplest chemical and physical system to understand and it is formed in quartz, which is extremely common in most Himalayan and Tibetan rocks. The number of TCN ages contrasts markedly with luminescence ($\sim$100) and radiocarbon ($\sim$100) ages for glacial deposits and landforms in the Himalayan–Tibetan orogen.

The first study that applied TCN surface exposure methods in the Himalayan–Tibetan orogen was undertaken by Phillips *et al.* (2000), who attempted to date the moraines around Nanga Parbat in northern Pakistan. Although their dating was somewhat limited (Richards *et al.*, 2001), it suggested that a significant glacial advance occurred between ca. 60 and 30 ka, and that glaciers advanced during the early to middle Holocene. A plethora of papers following this work emerged and they included distant regions throughout the Himalayan–Tibetan orogen (Owen *et al.*, 2001, 2002a,b, 2003a,b,c, 2005, 2006a,b; Schäfer *et al.*, 2002; Finkel *et al.*, 2003; Zech *et al.*, 2003, 2005; Barnard *et al.*, 2004a,b, 2006a,b; Meriaux *et al.*, 2004; Chevalier *et al.*, 2005; Abramowski *et al.*, 2006; Seong *et al.*, 2007, 2008a).

Owen *et al.* (2005) summarised the TCN studies and suggested that the regional patterns and timing of glaciation reflect temporal and spatial variability in the South Asian monsoon and, in particular, in regional precipitation gradients. They argued that in zones of greater aridity, the extent of glaciation has become increasingly restricted throughout the Quaternary, leading to the preservation of old ($\gg$100 ka) glacial landforms. In contrast, in regions that are strongly influenced by the monsoon ($\gg$1600 mm a^{-1}), they argued that the preservation potential of pre-Lateglacial moraine successions is generally extremely poor. This is possibly because Lateglacial and Holocene glacial advances may have been more extensive than earlier glaciations and hence may have destroyed any landform or sedimentary evidence of earlier glaciations. Furthermore, the intense denudation, mainly by fluvial and mass movement processes, which characterise these wetter environments, results in rapid erosion and re-sedimentation of glacial and associated landforms that contributes to their poor preservation potential. They showed that glacial advances could be broadly correlated (on timescales with a coarseness of several millennia) throughout the region, and that glaciers oscillated with a frequency that was similar to Northern Hemisphere millennia cycles.

Despite the vast application of TCN to defining the timing of glaciation, it is not yet possible to adequately determine whether glaciation is truly asynchronous throughout the region for the entire time for which ages are available. Two distinct sets of factors hinder our ability to interpret these chronologies. The first set of factors is geological and includes weathering, exhumation, prior exposure and shielding of the surface by snow and/or sediment. These generally reduce the concentration of TCNs, resulting in an underestimate of their true age. An exception to this general rule is prior exposure, which results in an overestimate of the true age. The net result of these processes can be a large spread in apparent exposure ages of individual boulders on a landform, such as a moraine, or in different positions on an extensive rock surface such as glacially eroded bedrock. To assess these effects, researchers usually collect multiple samples on a surface that they are dating. The presence of multiple boulders or surface samples having similar apparent ages is taken as evidence that the boulders were not derived from older surfaces and/or were not weathered or exhumed and hence the ages representative of the true age of the surface.

The second set of factors involves the calculation of the production rate of TCNs for the sampling location. This is dependent upon the cosmic ray flux, which has varied spatially and temporally in association with variations in the geomagnetic field intensity and atmospheric pressure throughout the Quaternary (Lal, 1991; Gosse and Phillips, 2001). Specifically, it is well documented from many independent proxy records that the Earth's geomagnetic field strength fluctuated considerably throughout the Quaternary (Ohno and Hamano, 1992; Guyodo and Valet, 1999; Yang *et al.*, 2000). There is currently much debate regarding the appropriate scaling models and geomagnetic corrections for TCN production to calculate TCN ages (e.g. Pigati and Lifton, 2004; Staiger *et al.*, 2007).

Unfortunately, the biggest uncertainty in scaling models is for low-latitude and high-altitude regions (Balco *et al.*, 2008). Figure 5 shows the effects of using four different time-dependent scaling models and geomagnetic corrections on ^{10}Be TCN ages for four distant areas in the Himalayan–Tibetan orogen. These were calculated using CRONUS Earth 2 (Balco *et al.*, 2008; http://hess.ess.washington.edu/math/). The figures clearly show that there is up to 30% difference in apparent ages among scaling models over the last 70 ka. The biggest differences are between the time constant (with no correction for geomagnetic field variation) model of Lal (1991) and Stone (2000) and the other models. Differences among models vary by as much as 20%, with the most significant differences occurring after ca. 20 ka. The general trend of these differences between scaling models is the same between the distant regions of Tibet, but the magnitude of the difference is more apparent for the Khumbu Himal, which is at the lowest latitude of any of the regions and reaches the highest elevations.

There is also a large variance in ages among scaling models for changes in altitude (Fig. 6). Differences in ages between scaling models can be as much as 20%, the biggest being between the time constant model of Lal (1991) and Stone (2000) and the other models. This biggest difference is at an altitude of $\sim$2500 m a.s.l. Age differences of up to 12% are apparent between the time-dependent mode1s, and are most apparent above $\sim$4500 m a.s.l.

To complicate matters further, different studies have used different production rates and scaling models to calculate their TCN ages. Table 1 (Supporting Information) lists all the published ^{10}Be TCN ages for boulders on moraines and ice-polished surfaces for the Himalayan–Tibetan orogen. To enable comparisons to be made between regions, we have recalculated all these data for the different scaling models using CRONUS Earth 2 calculator (http://hess.ess.washington.edu/math/); these are listed next to the published ages in Table 1

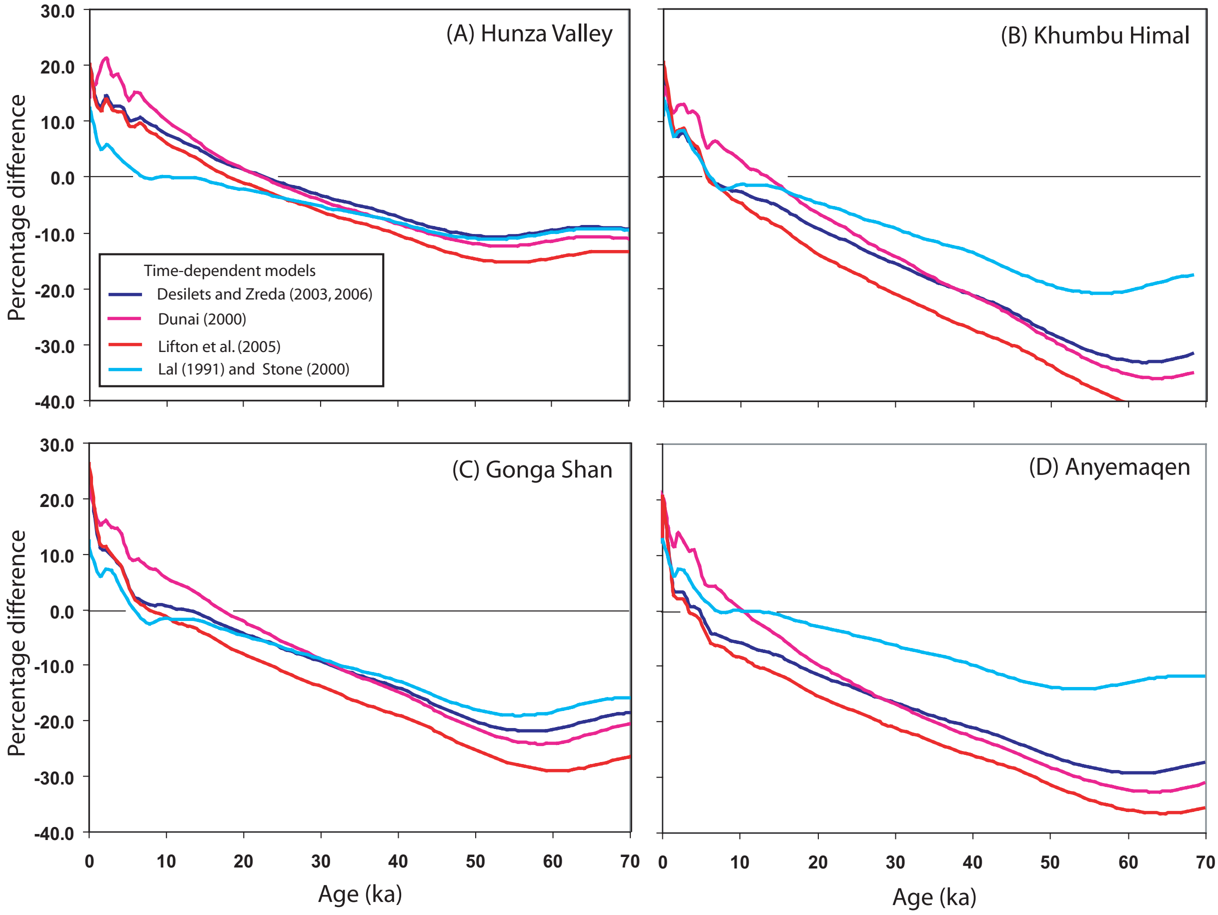

Figure 5 Percentage exposure age difference over the last 70 ka between time-constant (no correction for geomagnetic field variation) scaling method (Lal, 1991, and Stone, 2000; horizontal line, i.e. where $x = 0$) and various time-varying scaling schemes for distant locations in the Himalayan–Tibet orogen (other curves). (A) Hunza valley at 36.46° N, 74.90° E, 2500 m a.s.l. (site of sample KK98-1 of Owen *et al.*, 2002a). (B) Khumbu Himal at 27.89° N, 86.82° E, 4330 m a.s.l. (site of sample E5 of Finkel *et al.*, 2003). (C) Gonga Shan at 29.57°N, 101.99° E, 3174 m a.s.l. (site of sample GS1 of Owen *et al.*, 2005). (D) Anyemaqen at 34.89° N, 99.44° E, 4285 m a.s.l. (site of sample A1 of Owen *et al.*, 2003a). The Lal (1991)–Stone (2000) time-constant production rate at the location for (A) is 27.68 atoms g^{-1} a^{-1} (spallation) and 0.439 atoms g^{-1} a^{-1} (muons); for (B) is 63.05 atoms g^{-1} a^{-1} (spallation) and 0.729 atoms g^{-1} a^{-1} (muons); for (C) is 35.11 atoms g^{-1} a^{-1} (spallation) and 0.534 atoms g^{-1} a^{-1} (muons); and for (D) is 72.95 atoms g^{-1} a^{-1} (spallation) and 0.715 atoms g^{-1} a^{-1} (muons)

(Supporting Information). The differences between the published ages and the newly calculated ages are considerable (up to 40% difference; Fig. 7).

To identify possible times of glacier advances, we have used probability density plots for the ^{10}Be ages for boulders on moraines and glacially polished rock surfaces for selected time slices (Fig. 8). We recognise that pooling all the ages incorporates boulders or glacially eroded rock surfaces that may have ages too great or too small to represent the true age of the glacial advance because of the geological problems highlighted above. However, the large number of ages (many hundreds) help provide an estimate of the pattern of timing of glaciation. Given the uncertainties between the different scaling models to calculate TCN ages we choose the time-constant scaling model of Lal (1991) and Stone (2000). This does not take into account varied production of TCNs due to changes in the Earth's geomagnetic field. This approach allows us to directly compare regions. We recognise that the shape of these probability plots would vary if calculated using different scaling models. This would effectively skew the probability plots after ca. 20 ka towards the older range end of the distribution by between a few to 20% of the age.

The probability plots shown in Fig. 8 demonstrate a decrease in sample frequency with increasing age. This is essentially a function of preservation, with younger landforms being better preserved, more common and less difficult to date. Hence there is more data for the late Quaternary.

Multiple peaks are apparent for most of the probability plots in Fig. 8. This supports the view that multiple glacier oscillations occurred throughout the Holocene (Fig. 8(A)). There is not a dominant peak for the global LGM centred on 21 ka (Fig. 8(A) and (B)). Rather, the dominant peak occurs in the Lateglacial and is centred round ca. 12–14 ka. This suggests that the Lateglacial glacial advance was more dominant than that of the global LGM and/or better preserved. Even given the uncertainties associated with the difference scaling models, a 20% underestimate of the TCN ages would still not allow the data to support a significant global LGM advance.

The probability plots for data for MIS 3 to MIS 5e (inset in Fig. 8(B)) show peaks in the data centred on ca. 26 ka and ca. 47 ka (the ca. 26 ka peak is not apparent when the data are plotted from 0 to 115 ka as the younger ages skew the population). The probability plot is skewed towards the young end of the distribution, which is likely a function of the preferential sampling of younger landforms. The data support the view that significant glacial advances occurred during MIS 3, with the possibility of multiple advances occurring during the earlier part of the last glacial.

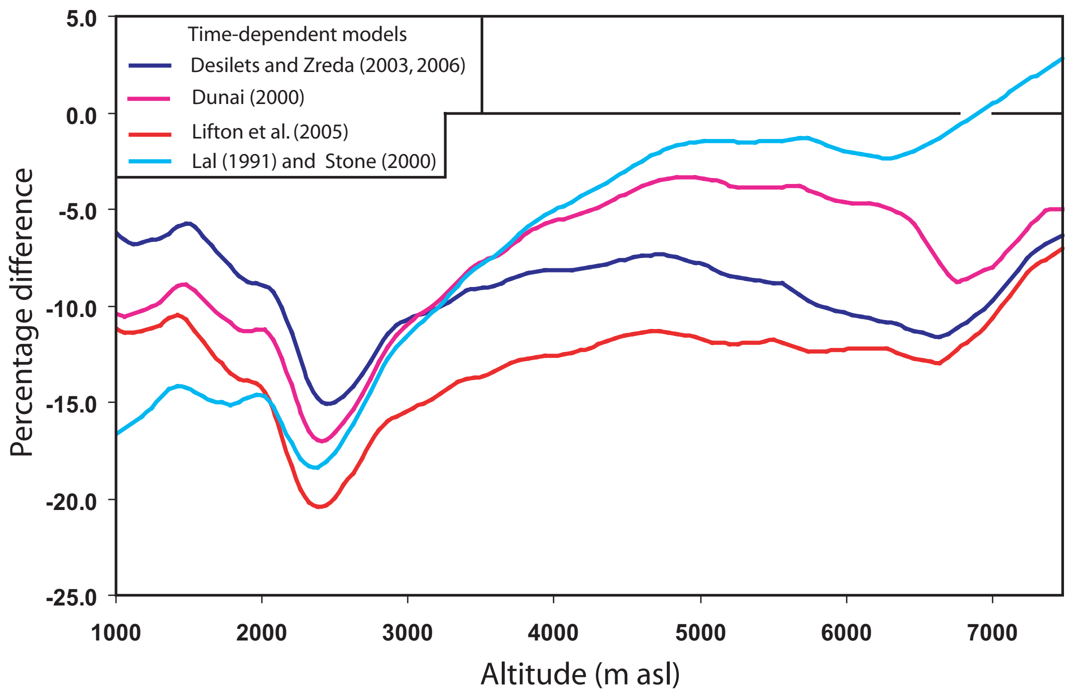

Figure 6 Percentage exposure age difference between time-constant (no correction for geomagnetic field variation) scaling method (Lal, 1991, and Stone, 2000; horizontal line, i.e. where $x=0$) and various time-varying scaling schemes (other curves) for altitudes ranging from 1000 to 7500 m a.s.l. for the Khumbu Himal (at 27.89° N, 86.82° E based on sample E5 of Finkel *et al.*, 2003)

Grouping the data by time intervals does not indicate any regional differences. The above conclusions might be spurious because the pooled ages would not allow us to see whether glaciation was regionally asynchronous. To examine this problem the TCN ages are plotted per major climatic region throughout the Himalayan–Tibetan orogen (Fig. 9). The probability plots support most of the general comments highlighted for the data for the whole region. This includes a decrease in frequency of TCN ages with age, with the exception being for the westernmost Himalaya–Tibet orogen (Turkestan, Alay and South Alichur Ranges, and the Ailuitek Pass), where no Holocene samples have been collected as part of the sampling strategy. The data also show multiple peaks for the Holocene in all regions (except for the westernmost Himalayan–Tibetan orogen; Fig. 9). Furthermore, there are significant Lateglacial peaks in all regions except for the westernmost

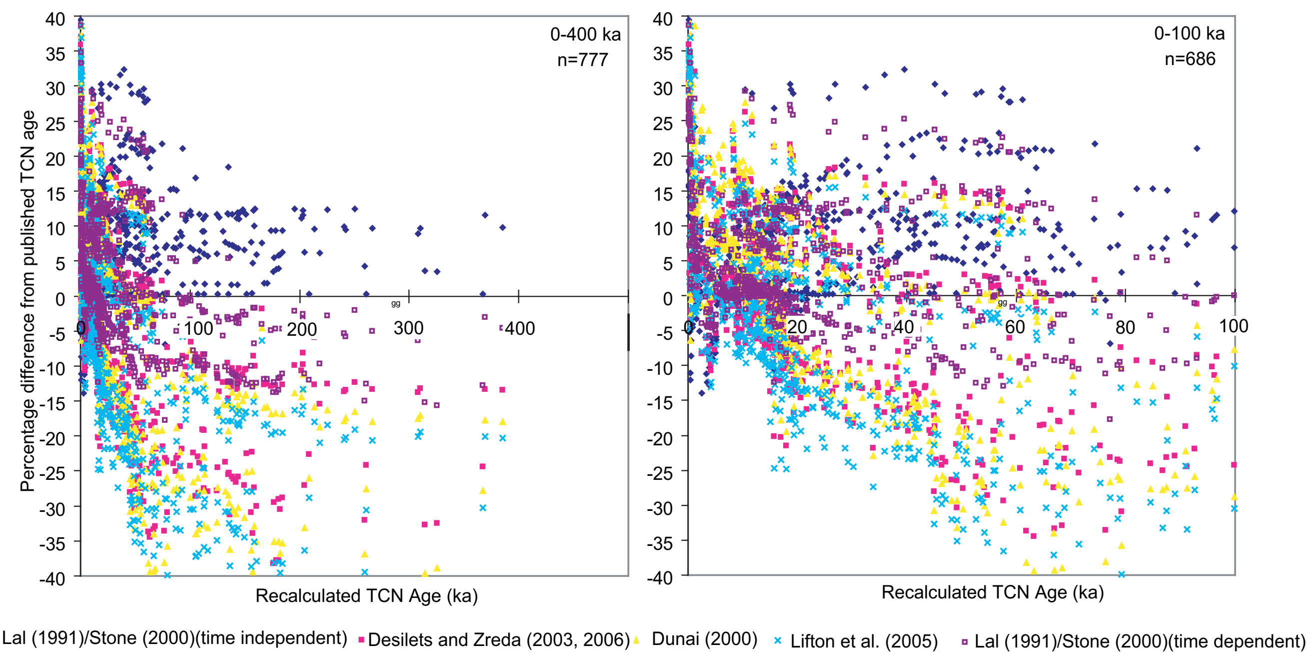

Figure 7 Percentage difference between published ages and ages recalculated using different scaling schemes for (A) all ages (0–500 ka) and (B) the last 100 ka (data from Phillips *et al.*, 2000; Owen *et al.*, 2001, 2002a,c, 2003a,b,c, 2005, 2006a,b; Schäfer *et al.*, 2002; Finkel *et al.*, 2003; Zech *et al.*, 2003, 2005; Barnard *et al.*, 2004a,b, 2006a,b; Meriaux *et al.*, 2004; Chevalier *et al.*, 2005; Abramowski *et al.*, 2006; Colgan *et al.*, 2006; Seong *et al.*, 2007, 2008a)

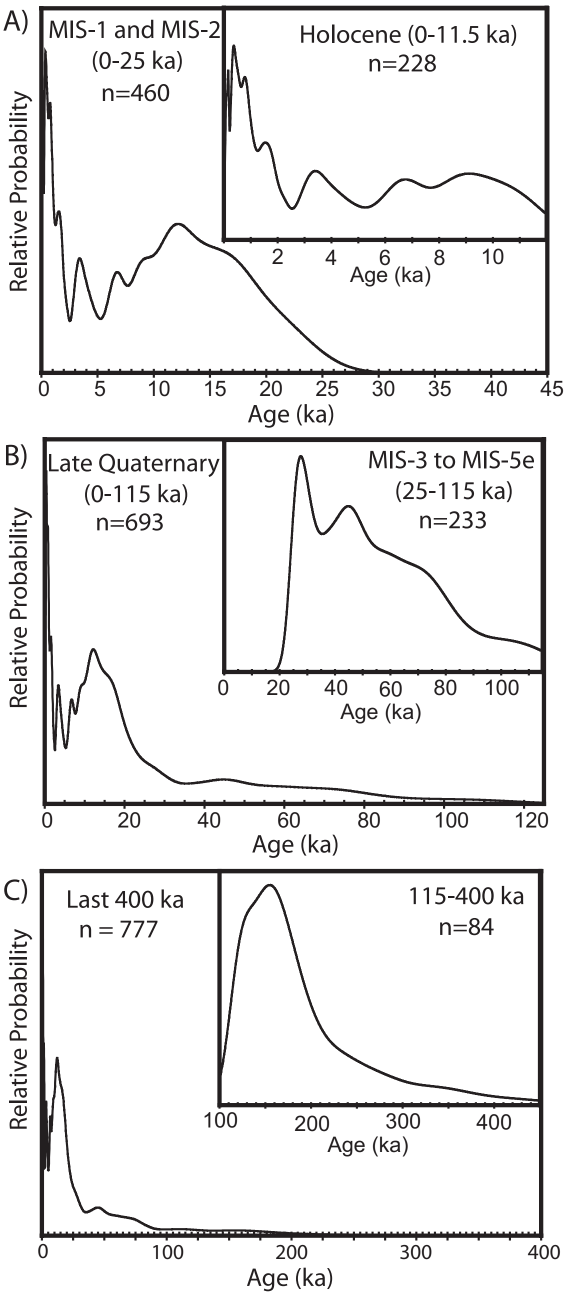

Figure 8 Probability density functions of TCN ^{10}Be ages for all published data from the Himalayan–Tibetan orogen for (A) MIS 1 and MIS 2 and (inset) the Holocene, (B) late Quaternary and (inset) MIS 3 and MIS 5e, and (C) last 400 ka and (inset) 115–450 ka (data from Phillips *et al.*, 2000; Owen *et al.*, 2001, 2002a,c, 2003a,b,c, 2005, 2006a,b; Schäfer *et al.*, 2002; Finkel *et al.*, 2003; Zech *et al.*, 2003, 2005; Barnard *et al.*, 2004a,b, 2006a,b; Meriaux *et al.*, 2004; Chevalier *et al.*, 2005; Abramowski *et al.*, 2006; Colgan *et al.*, 2006; Seong *et al.*, 2007, 2008a). The number of samples used to produce each probability plot is indicated by 'n'

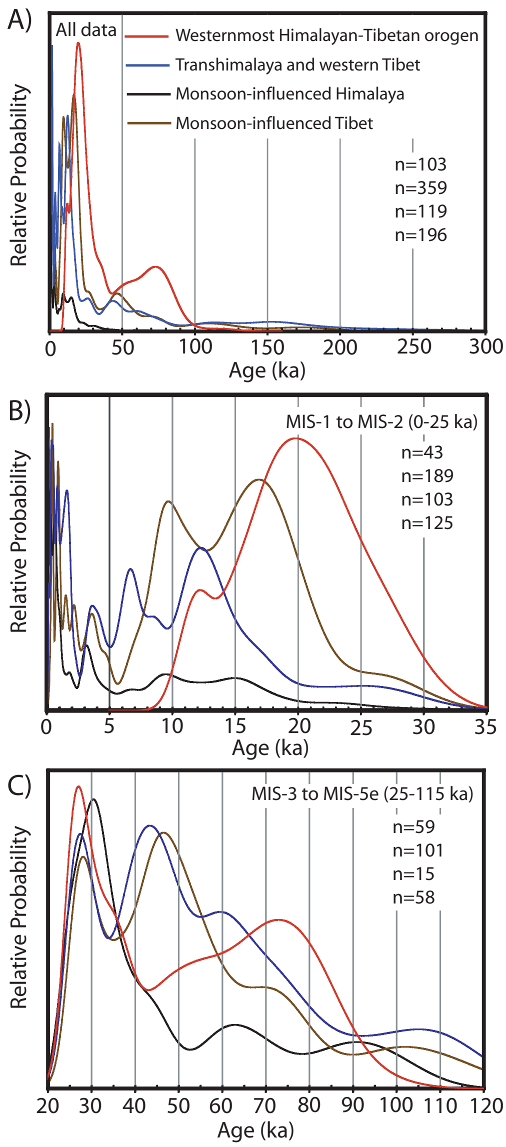

Figure 9 Probability density functions of TCN ^{10}Be ages for all published data from the Himalayan–Tibetan orogen plotted for each major climatic region. (A) All published data; (B) MIS 1 and MIS 2; and (C) MIS 3 to MIS 5e (data from Phillips *et al.*, 2000; Owen *et al.*, 2001, 2002a,b, 2003a,b,c, 2005, 2006a,b; Schäfer *et al.*, 2002; Finkel *et al.*, 2003; Zech *et al.*, 2003, 2005; Barnard *et al.*, 2004a,b, 2006a,b; Meriaux *et al.*, 2004; Chevalier *et al.*, 2005; Abramowski *et al.*, 2006; Colgan *et al.*, 2006; Seong *et al.*, 2007, 2008a). The number of samples used to produce each probability plot is indicated by 'n' in the order of the regions shown in the legend of part (A). Fig. 3 shows the areas included in each region

Himalayan–Tibetan orogen (Fig. 9(B)). The westernmost Himalayan–Tibetan orogen is the only region where there is a notable peak for the global LGM. Similarly, all regions except the westernmost Himalayan–Tibetan orogen have significant peaks during MIS-3 (although the monsoon-influenced Himalaya is subdued by the large number of young ages), whereas the westernmost Himalayan–Tibetan orogen is the only region with a dominant peak during MIS 4. Even with differences in ages due to the different scaling models, the peaks would still fall within MIS 3 and MIS 4 for these respective regions. These data suggest that the westernmost Himalayan–Tibetan orogen is behaving differently from the other regions. This region is most influenced by the mid-latitude westerlies and has little influence from the south Asian monsoon.

There are subtle variations in the data between the other regions, but given the errors associated with pooling data we are reluctant to ascribe much significance to small differences (Fig. 9(B) and (C)). Geochemical data from Tibetan and Himalayan ice cores show that there have been numerous glacier oscillations on millennial timescales throughout the Last Glacial (Thompson *et al.*, 1989, 1997; Thompson, 2000). The multiple peaks in the probability plots for the TCN ages likely reflect many of these glacier oscillations. The possibility of numerous glacier advances also complicates correlations between regions based on TCN dating, especially because the TCN dating method is relatively imprecise (~8% systematic error; Balco *et al.*, 2008). A detailed interpretation of the inter-regional variability therefore requires more comprehensive studies of individual regions using multiple dating methods. Unfortunately, there are few areas where comprehensive geochronological studies have been undertaken. We recognise three main areas – the Hunza valley in the Karakoram, Khumbu Himal and the Anyemaqen and Nianbaoyeze in NE Tibet – that provide the possibility of beginning to define the timing of glaciation and test regional correlations more accurately. These will be described in the next section.

Key areas

Data from three of the most detailed studied regions have been reassessed to evaluate the Quaternary glacial record in the Himalayan–Tibetan orogen more fully. We present the original data and recalculate TCN ages using new production rates for these regions (Figs 10,11 and 12).

The first of these regions, the Hunza valley, has a long history of study, which began in the late 1970s as a consequence of the construction of the Karakoram Highway, which links northern Pakistan with China across the Himalaya and Karakoram. As a consequence, numerous Chinese publications were produced on the glacial geomorphology and history, which culminated in the International Karakoram Project in 1980 involving Pakistani, Chinese and British scientists (Batura Glacier Investigation Group, 1976, 1979; Zhang and Shi, 1980; Miller, 1984). More detailed geochronological studies were undertaken by Owen *et al*. (2002a) and Spencer and Owen (2004), who used TCN and OSL methods, respectively. These data are graphically presented in Fig. 10. The probability plots for the recalculated TCN ages make the glacial stages a little older by a few thousand years as compared to the original data. In combination with the OSL ages of Spencer and Owen (2004), however, the timing of glaciation is similar to that suggested in Owen *et al*. (2002a) (bracketed by the thick colour bars in Fig. 10). Despite the disagreement between the ages presented in these studies, the moraine ages are still assigned to the same glacial times as presented by Owen *et al*. (2002a). They argued, for example, that glaciers were responding both to changes in insolation-controlled monsoon precipitation that allowed glaciers to advance during the early Holocene and MIS 3, and Northern Hemisphere cooling that allowed glaciers to advance during the global LGM.

Figure 10 shows Prell and Kutzbach's (1987) simulated monsoon pressure index for the Indian Ocean and simulated changes in precipitation in southern Asia together with variations in Northern Hemisphere low-latitude solar radiation as a proxy for changing precipitation in the region to assess the likely connections between the monsoon and glaciation. Furthermore, the Heinrich events for the past 70 ka (Bond *et al*., 1992) are plotted as a proxy for changes in the Northern Hemisphere ice sheets and oceans to assess correlations and connections. Comparisons with both datasets are rather tentative at present and the relative roles and connections with the different climate systems are difficult to resolve.

The Khumbu Himal, south of Mount Everest, is the second key area. Numerous studies have been undertaken on the Quaternary glacial and associated landforms in this region (Benedict, 1976; Iwata, 1976; Fushimi, 1977, 1978; Müller, 1980; van Williams, 1983; Richards *et al*., 2000b; Finkel *et al*., 2003; Barnard *et al*., 2006a). The earliest geochronological studies involved radiocarbon dating of late Holocene moraines (Benedict, 1976; Fushimi, 1978; Müller, 1980). This was followed by OSL dating of late Quaternary moraines by Richards *et al*. (2000b), who showed that glaciers advanced during the global LGM (18–25 ka), the early Holocene (~10 ka) and the late Holocene (1–2 ka). These advances were confirmed by Finkel *et al*. (2003) using ^{10}Be TCN dating methods. Finkel *et al*. (2003) were also able to define the ages of a Neoglacial advance at ~3.6 ka, a Lateglacial advance (15–16 ka) and two advances during MIS 3.

The chronology and dating of Finkel *et al*. (2003) and Richards *et al*. (2000b) are plotted on Fig. 11 together with probability plots for the recalculated TCN ages. The OSL and the original TCN ages agree reasonably well, but the TCN are generally younger by a few hundred years. This would be expected since the sediment that was dated using OSL methods should pre-date the overlaying boulders that were dated using TCN methods. The probability plots suggest the TCN ages should be a little older then those published in Finkel *et al*. (2003); however, they still support the view that glaciers advanced during MIS 3, the global LGM, Lateglacial, early Holocene, Neoglacial and late Holocene. A comparison of these data with monsoon intensity (Fig. 11(E)) supports the view that glaciers advance during times of increased insolation. These were times when moisture delivered by an active South Asian summer monsoon was enhanced and provided snow to create positive glacier mass balances allowing glaciers to advance. Furthermore, increased cloudiness during monsoon could also help contribute significantly to positive glacier mass balances (Rupper, 2007). However, glaciers may also have been influenced by cooling cycles during the late Quaternary, for example during the global LGM.

Considerable attention has been paid to the glacial chronologies at the NE margin of Tibet. The Anyemaqen and the Nianbaoyeze are two significant massifs in this region and form the third key study area. These massifs have been studied in detail by Wang (1987) and Lehmkuhl and Lui (1994). TCN and OSL dating was undertaken by Owen *et al*. (2003a) to define the timing of glaciation. They showed that glacial advances occurred during MIS 3 (ca. 30–55 ka), MIS 2 (ca. 12–25 ka) and early Holocene (~10 ka). Recalculated TCN ages are presented as probability plots for the regions in Fig. 12. This shows that there is a significant MIS 3 advance, and Lateglacial and early Holocene advance. The glacial advances would have still dated to the early Holocene, Lateglacial and MIS 3 regardless of the scaling model used to calculate the ages. This indicates and supports the view of Owen *et al*. (2003a) that glaciers in this region are influenced by the South Asian monsoon. The maximum extent of glaciation occurred early in the last glacial cycle (MIS 3) during a time of increased insolation when the monsoon was intensified and supplied abundant precipitation, as snow at high altitude, to feed high-altitude glaciers. This indicates that precipitation, as snow is fundamental in controlling glaciation in these regions.

ELAs

Reconstructions of the timing and extent of glaciation are used as a proxy for palaeoclimate change. The most commonly applied technique that is used involves the reconstruction and use of former ELAs and ELA depressions (ΔELAs) relative to the present. The reconstructed ΔELAs are used to estimate past temperature and/or precipitation changes and to assess regional climatic gradients and variability. Benn and Lehmkuhl (2000), Owen and Benn (2005) and Benn *et al*. (2003) describe how reconstructing former ELAs and ΔELAs can be problematic in high mountain regions, specifically the Himalayan–Tibetan orogen. These problems occur because of extreme topographic

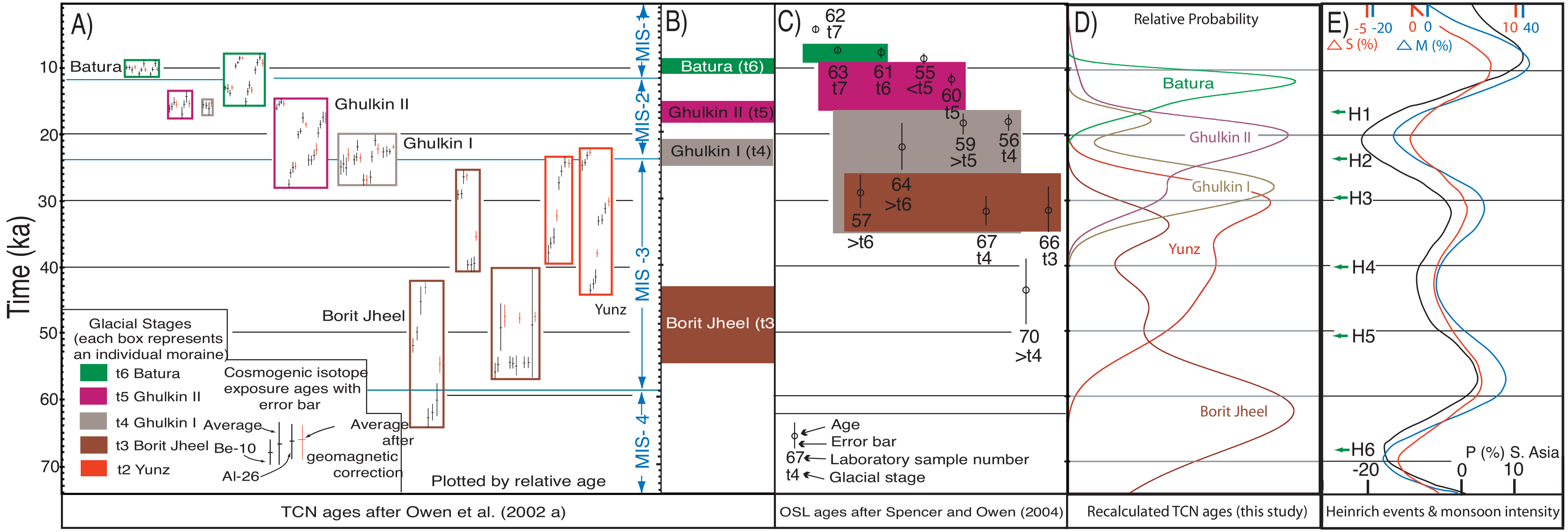

Figure 10 Glacial chronology for the upper Hunza valley. (A) TCN ages for boulders on moraines (after Owen *et al.*, 2002a). The TCN ages are plotted along the *x*-axis according to their geographical locations and their relative ages. Each coloured box encloses samples from the same moraine. The blue horizontal lines represent the boundaries of the Marine Isotope Stages (MISs) (after Martinson *et al.*, 1987, and Shackleton *et al.*, 1990). t2 to t7 refer to the glacial stages. (B) Timing of glacial stages (after Owen *et al.*, 2002a). (C) OSL ages for glaciogenic sediments associated with moraines (after Spencer and Owen, 2004). The relative age is highlighted by 't', such that >t is for sediment below a moraine, <t is for sediment above a moraine, and t is for sediment within the moraines. (D) Probability density functions of TCN ^{10}Be ages for each glacial stage. The ages were recalculated from the data in Owen *et al.* (2002a) using the time-constant (no correction for geomagnetic field variation) scaling method (Lal, 1991; Stone, 2000). (E) Simulated monsoon pressure index (ΔM percentage, solid blue line) for the Indian Ocean and simulated changes in precipitation (ΔP percentage, black line) in southern Asia (Prell and Kutzbach, 1987), and variations in Northern Hemisphere solar radiation (ΔS percentage, red line; Prell and Kutzbach, 1987). H1, H2, H3, H4, H5 and H6 show the timing of Heinrich events during the past 70 ka (Bond *et al.*, 1992)

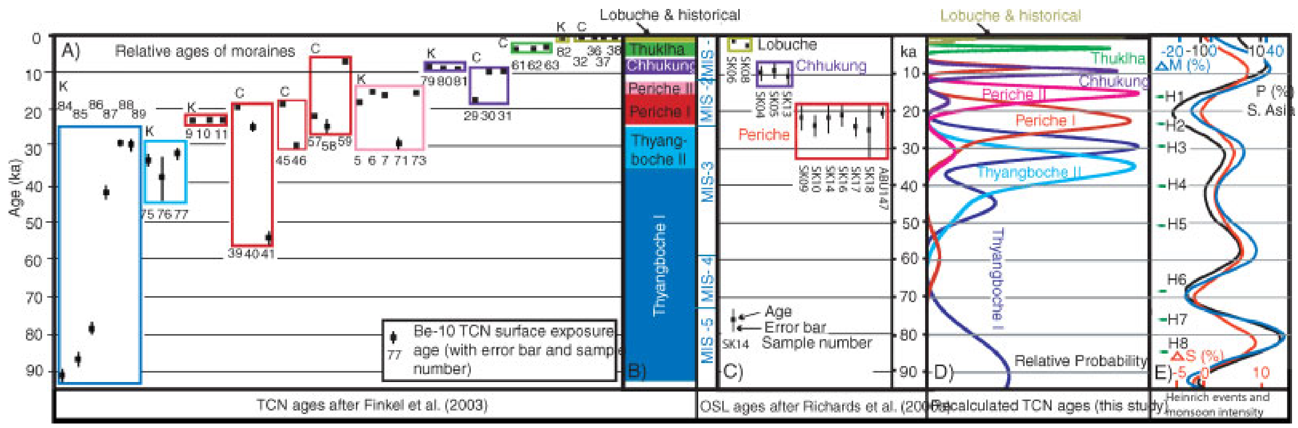

Figure 11 Glacial chronology for the Khumbu Himal. (A) TCN ages for boulders on moraines (after Finkel *et al.*, 2003). The TCN ages are plotted along the *x*-axis according to their relative ages and geographical locations (C, Chhukung; K, Periche to Khumbu Glacier). Each coloured box encloses samples from the same moraine. (B) Timing of glacial stages (after Finkel *et al.*, 2003). The Marine Isotope Stages (MIS) are taken from Martinson *et al.* (1987) and Shackleton *et al.* (1990). (C) OSL ages for glaciogenic sediments associated with moraines (after Richards *et al.*, 2000b). The coloured boxes enclose OSL ages for sediment within the moraines for three glacial stages. (D) Probability density functions of TCN ^{10}Be ages for each glacial stage. The ages were recalculated from the data in Finkel *et al.* (2003) using the time-constant scaling method (Lal, 1991; Stone, 2000). (E) Simulated monsoon pressure index (ΔM percentage, solid blue line) for the Indian Ocean and simulated changes in precipitation (ΔP percentage, black line) in southern Asia (Prell and Kutzbach, 1987), and variations in Northern Hemisphere solar radiation (ΔS percentage, red line; Prell and Kutzbach, 1987). H1, H2, H3, H4, H5, H6, H7 and H8 show the timing of Heinrich events during the past 90 ka (Bond *et al.*, 1992)

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.

Owen and Benn (2005) summarised the current knowledge on contemporary and past ELAs for glaciers throughout the Himalayan–Tibetan orogen. They highlighted the considerable regional variation, which is the result of the influence of the two major climatic systems, the mid-latitude westerlies and the South Asian summer monsoon, plus the influence of strong topographic controls. Furthermore, there may be considerable local variability as highlighted by Sharma and Owen (1996) for the Garhwal Himalaya and Seong *et al.*, (2008b) for Muztag Ata and Kongur Shan.

Owen and Benn (2005) argue that even though there have been numerous studies of ELAs throughout the Himalayan–

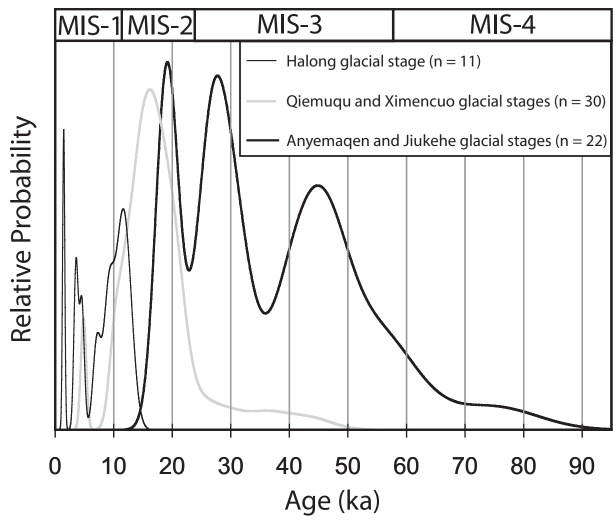

Figure 12 Probability density functions of TCN ^{10}Be ages for glacial stages in the Anyemaqen and Nianbaoyeze of NE Tibet. The ages were recalculated from the data in Owen *et al.* (2003a) using the time-constant scaling method (Lal, 1991; Stone, 2000). The Marine Isotope Stages (MISs) are taken from Martinson *et al.* (1987) and Shackleton *et al.* (1990)

Tibetan orogen, the methods for reconstructing former ELAs vary considerably between studies. Moreover, they emphasised that many of the methods that have been used are not applicable for these high-altitude glaciers. Furthermore, they showed that few of the reconstructions of former glaciers and their ΔELA have any chronological control. Therefore regional comparisons of past ELAs like those presented by Shi (2002) do not fully represent the nature of glaciation because of the poor temporal framework and problematic methodologies. They do, however, provide an overview of the likely regional variation of ΔELAs. Owen and Benn (2005) recognised only two locations, the Hunza valley and the Khumbu Himal, where glaciers of the global LGM have been accurately dated (Fig. 13). They calculated ELAs for these regions and showed that ΔELAs were considerably less (∼200–300 m in the Khumbu and ∼≤100 m for the Batura Glacier; Fig. 13(B) and (C)) than had been previously estimated for the global LGM in these regions (≫500 m).

In summary, present studies do not provide an adequate assessment of ΔELAs for the Quaternary throughout the Himalayan–Tibetan orogen. As a consequence, temperatures and/or precipitation changes have yet to be quantified for the region.

Summary

The timing and extent of glaciation throughout the Himalayan–Tibetan orogen are poorly defined despite the importance of this region for geological studies and the economies of the countries that border its mountains. New technologies, including remote sensing and geochronology, are rapidly helping us to resolve the nature and dynamics of glaciation throughout the region. In particular, the broad patterns of the maximum extent of late Quaternary glaciation are beginning to be resolved. Figure 4(D) provides the best available estimate for the timing of the maximum extent of glaciation across the region, but even after measurement of >1000 TCN, OSL and radiocarbon dates it is still not known conclusively whether the extent of maximum glaciation was reached at the same time.

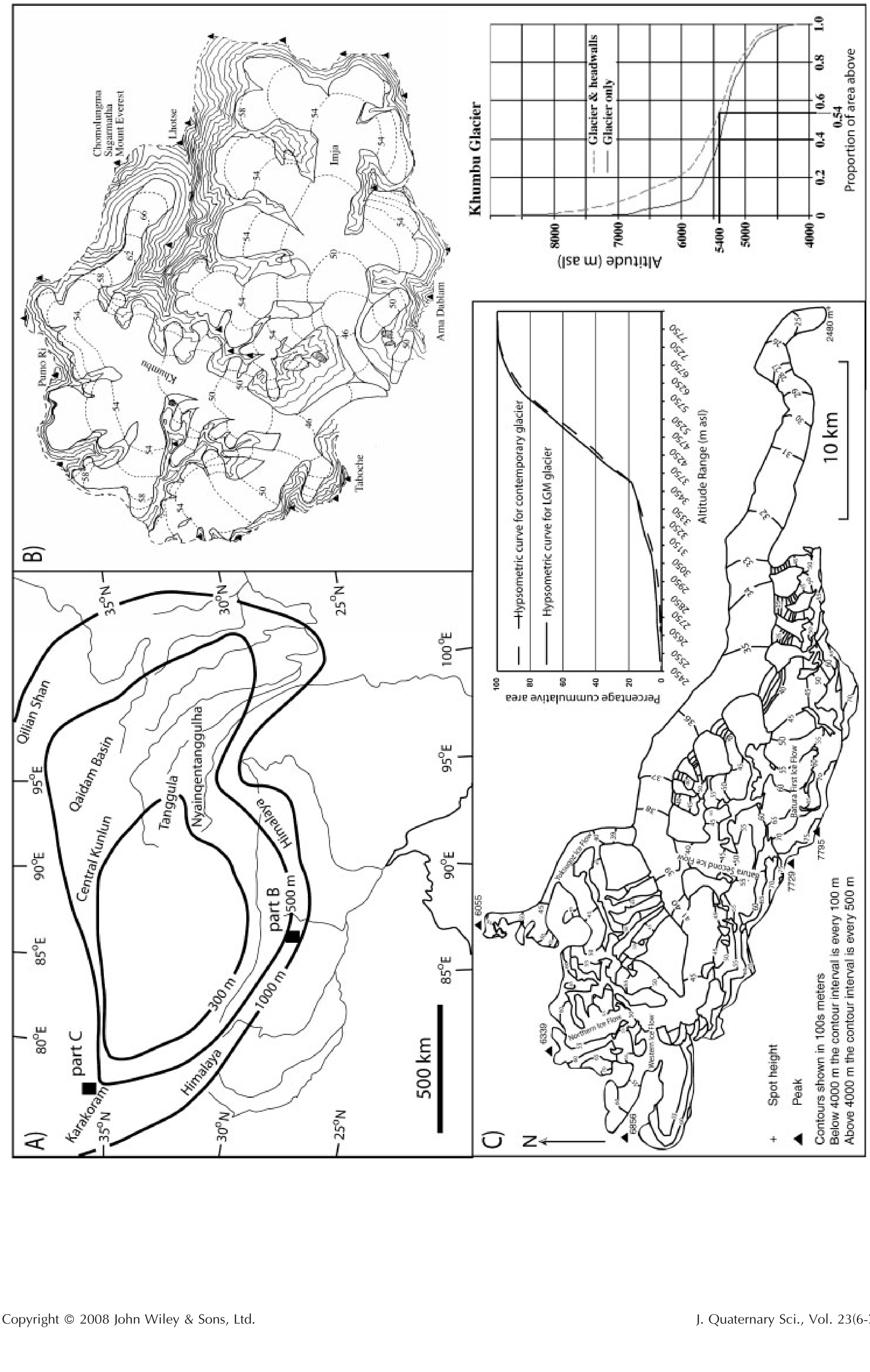

Figure 13 Equilibrium-line altitudes (ELAs) for the Himalaya and Tibet for the LGM. A) ELA depression for the global LGM across Tibet and the bordering mountains according to Shi (2002); Note that the ΔELA values are probably not correct (see text for a discussion on the validity of the ΔELA values) but the general pattern of variation is probably realistic (adapted from Owen and Benn, 2005). Reconstruction of the (B) Khumbu and Imja glacier systems in the Khumbu Himal and (C) the Batura glacier system in Hunza for the global LGM. The hypsometric curves are for the contemporary Khumbu Glacier (B) and Batura Glacier (C) (after Owen and Benn, 2005). The contours in parts (B) and (C) are in hundreds of metres above sea level. The black boxes in part (A) show the locations of parts (B) and (C)

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Detailed study areas provide important insights into testing these reconstructions, but they suggest considerable variations in the extent of glaciation during a specific glacial time between adjacent regions. This is best illustrated in northern Pakistan and northern India, where during the Lateglacial glaciers advanced more than 50 km in the K2 region of the Central Karakoram and Lahul Himalaya, but in the adjacent regions of the Hunza valley and Ladakh the glaciers advanced only a few kilometres beyond their current positions.

The timing of many of these glacial advances is poorly defined because of the lack of organic material needed for radiocarbon dating, the standard geochronological tool of glacial geomorphologists. However, newly developing OSL and TCN methods are helping to expand the chronologies throughout the region. These technologies allow broad correlations on Milankovitch timescales to be made, but are rife with methodological problems. With OSL dating, for example, partial bleaching and changes in dose rates over time can result in large errors. TCN dating can also be problematic, particularly with regard to the uncertainty associated with scaling models to determine ages, and the geological problems associated with preserving stable surfaces in active mountains.

Recalculating all the TCN ages for moraine boulders and glacially eroded surfaces throughout the Himalayan–Tibetan region allows broad statements to be made regarding the synchroneity of glaciation. In essence, glaciers throughout monsoon-influenced Tibet and the Himalaya and the Transhimalaya appear to respond in a similar fashion to changes in monsoon-driven and Northern Hemisphere cooling cycles alone. In contrast, glaciers in the far western regions of the Himalayan–Tibetan orogen are asynchronous with the other regions and appear to be dominantly controlled by the Northern Hemisphere cooling cycles.

Quantifying the magnitude of climate change from glacial geological evidence is challenging. The standard methods using ELAs can be problematic for the steep and debris-mantled glaciers in mountains. Nevertheless, broad patterns of local and regional variability can be established and are helping to examine changes in climatic gradients over time.

In conclusion, we stress the importance of developing a modern framework for the geomorphic and sedimentological analysis of glaciogenic sediments and landforms in high mountain environments for the accurate reconstruction of former glaciers and highlight the problems of dating glacial successions. Furthermore, we encourage programs for numerical dating in critical regions that apply multiple dating methods.

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