Late Quaternary glaciation of Tibet and the bordering mountains: a review

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Abundant glacial geologic evidence present throughout Tibet and the bordering mountains shows that glaciers have oscillated many times throughout the late Quaternary. Yet the timing and extent of glacial advances is still highly debated. Recent studies, however, suggest that glaciation was most extensive prior to the last glacial cycle. Furthermore, these studies show that in many regions of Tibet and the Himalaya glaciation was generally more extensive during the earlier part of the last glacial cycle and was limited in extent during the global Last Glacial Maximum (marine oxygen isotope stage 2). Holocene glacial advances were also limited in extent, with glaciers advancing just a few kilometers from their present ice margins. In the monsoon-influenced regions, glaciation appears to be strongly controlled by changes in insolation that govern the geographical extent of the monsoon and consequently precipitation distribution. Monsoonal precipitation distribution strongly influences glacier mass balances, allowing glaciers in high altitude regions to advance during times of increased precipitation, which are associated with insolation maxima during glacial times. Furthermore, there are strong topographic controls on glaciation, particular in regions where there are rainshadow effects. It is likely that glaciers, influenced by the different climatic systems, behaved differently at different times. However, more detailed geomorphic and geochronological studies are needed to fully explore regional variations. Changes in glacial ice volume in Tibet and the bordering mountains were relatively small after the global LGM as compared to the Northern Hemisphere ice sheets. It is therefore unlikely that meltwater draining from Tibet and the bordering mountains during the Lateglacial and early Holocene would have been sufficient to affect oceanic circulation. However, changes in surface albedo may have influenced the dynamics of monsoonal systems and this may have important implications for global climate change. Drainage development, including lake level changes on the Tibetan plateau and adjacent regions has been strongly controlled by climatic oscillations on centennial, decadal and especially millennial timescales. Since the Little Ice Age, and particularly during this century, glaciers have been progressively retreating. This pattern is likely to continue throughout the 21st century, exacerbated by human-induced global warming.

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The Tibetan Plateau and bordering mountains are the greatest glaciated tracks outside the Polar Region. They have a profound influence on regional and global atmospheric circulation and are therefore important for our understanding the dynamics of global environmental change (Ruddiman & Kutzbach 1989; Molnar & England 1990; Prell & Kutzbach 1992; Owen *et al.* 2002d). Changes in glaciation and hydrology in Tibet and the bordering mountains throughout the late Quaternary may have altered the input of fresh water into the seas and oceans adjacent to the Asian continent. This in turn could have had a major impact on ocean circulation and global climate, an impact analogous to effects of the melting of the Laurentide Ice Sheet on North Atlantic oceanic circulation towards the end of the Last Glacial (cf. Broecker *et al.* 1989). In particular, glaciation throughout Tibet and the bordering mountains was probably far more extensive during the Late Pleistocene than at present, and it is possible that when glaciers began to retreat during the termination of the last glacial cycle substantial amounts of meltwater produced new drainage systems feeding into the adjacent seas. Yet, despite its importance, the timing and extent of glacial advances are still highly debated. To investigate links between glaciation, hydrology and environmental change in the high mountains of Central Asia, and the possible relationship with global climate change, this article aims to synthesis new research and the current knowledge on the late Quaternary glaciation of Tibet and the bordering mountains (Fig. 1).

This paper extends work presented in a comprehensive bibliography produced by Barnard & Owen (2000), research papers (Owen & Lehmkuhl 2000; Owen & Zhou 2002) and summaries of the Global Mapping Project of INQUA on the glacial geology in central Asia (Ehlers & Gibbard 2004). 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. We highlight the problems of dating glacial

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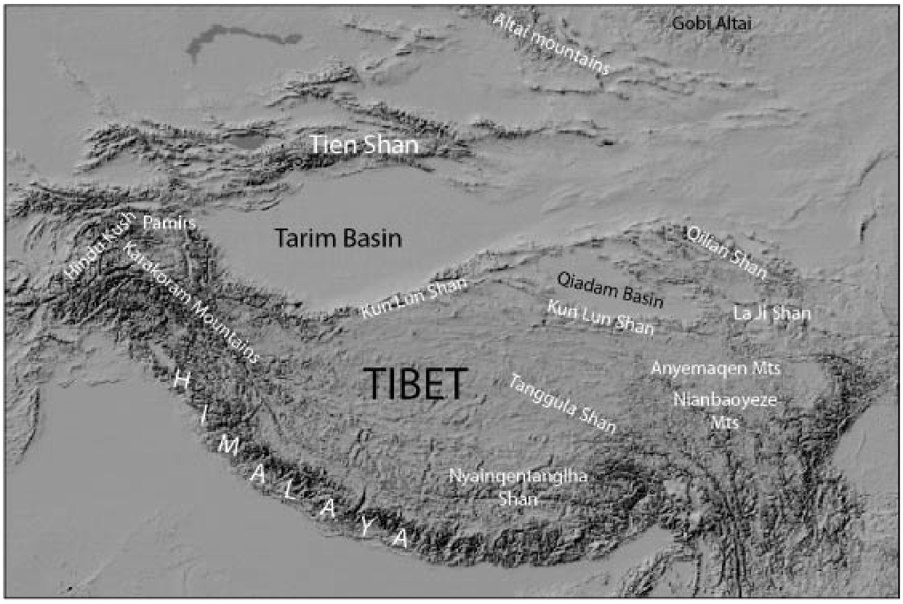

Fig. 1. The extent of the region covered by this article.

successions and encourage programmes for numerical dating in critical regions, as suggested by Benn & Owen (2002). Furthermore, we emphasis the problems of using and calculating equilibrium-line altitudes (ELAs) as a means of quantifying glaciation and reiterate the recommendations provided to help quantify the degree of glaciation using reconstructions of former glacier extent presented by Benn & Lehmkuhl (2000), Benn *et al*. (2005), Owen & Benn (2005). We also highlight the usefulness of other palaeoclimate proxy data, such as the loess, lacustrine, fluvial and palaeobotanical records. In particular, we highlight the importance of the lake record, especially because, during the Pleistocene, lakes in the western Tibetan Plateau covered an area four times the present extent ($\sim$80 000 km^2, vs. $\sim$20 000 km^2; Lehmkuhl & Haselein 2000). Such lake level changes could have profound effects on regional climate because of changes in albedo and precipitation.

Regional setting

Tibet and the bordering mountains formed as a consequence of the collision of the Indian and Eurasian continental plates initiated $\sim$50 million years ago. Geologically, the region is a complex assemblage of rocks of different ages that are still actively being deformed by the continued northward movement of the Indian continental plate at $\sim$50 mm/year (DeMets *et al*. 1994). The average altitude across the Himalayan–Tibetan region is $\sim$5000 m a.s.l. (Fielding 1996). The region stretches $\sim$2000 km and $\sim$1500 km in an east–west and north–south direction, respectively (Fig. 1). This mountain mass comprises a series of approximately east–west trending ranges that include, from south to north, the Siwaliks, Lesser Himalaya, Greater Himalaya, Transhimalaya, Nyaingentanglha Shan, Tanggula Shan, Bayan Har Shan, Kunlun Shan, Altun Shan and Qilian Shan. We include the Pamir, Tian Shan and Altai Mountains in our study region because they broadly border the Tibetan–Himalayan region to the west (Fig. 1).

These mountain ranges are influenced by four major climatic systems: the mid-latitude westerlies, the south Asian monsoon, the Mongolian high-pressure system and the El Nino Southern Oscillation (ENSO). The relative importance of each varies throughout the region, with the eastern end and the southern slopes of the Himalaya being the wettest (Fig. 2). The southern slopes of the Himalaya and the high mountains of eastern Tibet receive snowfall mainly in the summer monsoon season, whereas northern and western Tibet, and ranges such as the Karakoram Mountains, Pamir, Tian Shan and Altai, receive heavy snowfalls during the winter with moisture supplied from the mid-latitude westerlies and the Mongolian high pressure system (e.g. Böhner 1996). This snow supports ice caps and valley glaciers throughout Tibet and the bordering mountains. As a consequence, the region presently has the largest concentration of glaciers outside the Polar Regions ($\sim$126 200 km^2; Haeberli *et al*. 1989).

Major rivers, including the Indus, Ganges, Tsangpo-Bhramaputra, Mekong, Yangtze and Huanghe, drain the region. They are essentially fed by glacial meltwater and monsoon precipitation; they have discharges and they produce sediment loads that are among the highest in the world. Furthermore, these rivers are essential for the agricultural, industrial and domestic needs of approximately three-quarters of the world's population.

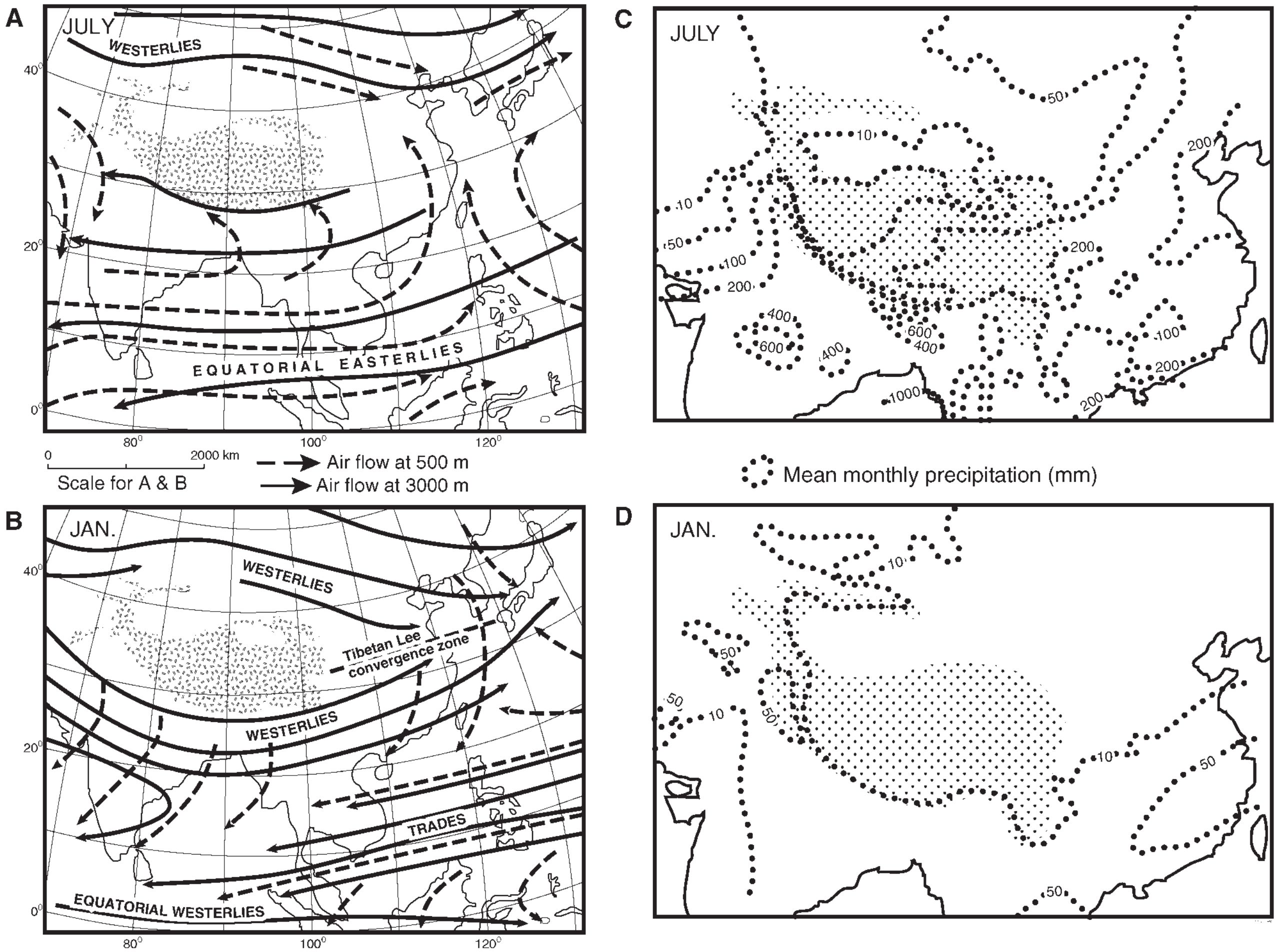

Fig. 2. Characteristic air circulation (A & B) and precipitation (C & D) over southern and central Asia. The Tibetan Plateau and bordering mountains above 5000 m a.s.l. are shown within the dotted areas. The solid lines in (A) and (B) indicate airflow at about 6000 m a.s.l. and 3000 m a.s.l., respectively,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

Timing and extent of glaciation

Relative glacial chronologies have been developed throughout the Himalaya and Tibet using morphostratigraphy aided by relative weathering studies (e.g. Burbank & Kang 1991; Hövermann *et al.* 1993a, b; Hövermann & Lehmkuhl 1994; Lehmkuhl 1995a, b, 1997, 1998b; Owen *et al.* 1997; Lehmkuhl *et al.* 2000, 2002) and soil development (e.g. Bäumler *et al.* 1997; Guggenberger *et al.* 1998; Zech *et al.* 2000). Reconstructing palaeoenvironmental change from glacial geologic evidence in Tibet and the bordering mountains has been difficult because of the lack of organic material for radiocarbon dating, and the problems of correctly identifying the origin of highly dissected landforms. Numerous radiocarbon dates are available for the wetter parts of the Himalayan–Tibetan region, but most are limited to Holocene proglacial deposits that poorly define the timing of glaciation (Röthlisberger & Geyh 1985; Lehmkuhl 1995a, 1997). Newly developing techniques that include optically stimulated luminescence (OSL) and cosmogenic surface exposure dating are now allowing glacial successions throughout Tibet and the bordering mountains to be dated and correlated (Lehmkuhl *et al.* 2000, 2002; Phillips *et al.* 2000; Richards *et al.* 2000a, b; Owen *et al.* 2001, 2002a, b, c, 2003a, b, c, 2005; Schäfer *et al.* 2002; Tsukamoto *et al.* 2002; Finkel *et al.* 2003; Zech *et al.* 2003; Spencer & Owen 2004; fig. 3).

Tibet

Reconstruction of the extent of glaciation across the Tibetan Plateau has a long history (Klute 1930; Frenzel 1960; Kuhle 1985; Frenzel *et al.* 1992; Ono *et al.* 2004; Klinge & Lehmkuhl 2004). More recently, Chinese workers have compiled synthesis maps to show the probable extent of glaciation across the Tibetan Plateau (Shi *et al*. 1986, 1993; Liu *et al*. 1988; Shi 1988; Li *et al*. 1991; Shi 1992). In a series of articles, Kuhle (1985, 1986, 1987, 1988a, b, 1990a, b, 1991, 1993, 1995) hypothesized that an extensive ice sheet covered essentially the whole of the Tibetan Plateau during glacial times. The existence of an extensive ice sheet has been one of the most contentious glacial issues concerning Tibet over the past few decades. Numerous publications, however, discuss the evidence against an extensive ice sheet (Derbyshire 1987; Zheng 1989; Pu 1991; Shi 1992; Hövermann *et al*. 1993a, b; Lehmkuhl 1995a, 1998b; Rutter 1995; Lehmkuhl *et al*. 1998; Zheng & Rutter 1998; Zhou & Li 1998; Schäfer *et al*. 2002; Owen *et al*. 2003c) and it is now generally accepted that a large ice sheet did not cover the Tibetan Plateau, not at least during the past few glacial cycles (see more detailed discussion below).

One of the main reasons for the differing opinions on the distribution and extent of Pleistocene glaciations on and around the Tibetan Plateau is the lack of general agreement on terminology and stratigraphic division of the different end moraine sequences and till deposits in the various mountain areas (Lehmkuhl 1997). Some relative chronologies exist for mountain glaciations, but the timing of glacier oscillations is poorly understood due to the lack of known numerical ages on moraines. Even a common relative stratigraphy based on ELA depressions, such as the one developed for the European Alps, is not available. An evaluation of recent glacial geologic studies suggests that Li *et al*. (1991) provide the best reconstruction of the extent of glaciation for the entire Tibetan Plateau. Their map shows limited glaciation in the interior of the Tibetan Plateau but expanded ice caps and valley glaciers on its margins during the last glacial cycle. They also provide a reconstruction of the former extent of a small ice sheet in northeast Tibet during the penultimate glacial cycle. The lack of numerically dated glacial landforms, however, makes it difficult to test whether the glacial limits that they map are not diachronous. The exact details of the extent of glaciation at particular times during the Quaternary are therefore still highly debated.

In the Chinese literature, the last glaciation is commonly divided into two main stages. These are thought to represent glaciations that occurred during marine oxygen isotope stages (MIS) 2 and 4 and are separated by an interstadial that lasted from about 55 to 32 ka (e.g. Liu *et al*. 1985; Li & Pan 1989; Thompson *et al*. 1989, 1997; Zhang *et al*. 1991). Chinese workers such as Li & Shi (1992) and Li & Pan (1989) argue that the expanded ice caps and valley glaciers on the Tibetan Plateau during the last glacial began substantial retreat between 15 and 13 ka. Much of this work is summarized in Lehmkuhl (1995a), Owen *et al*. (1997) and Benn & Owen (1998). The summary articles emphasize, however, that the timing of glaciation is poorly defined because of the limited number of numerical dates that were undertaken in these studies and state that great care should be taken in making regional generalizations about the timing and extent of glaciation.

During the past decade, however, OSL and cosmogenic surface exposure dating has been providing new insights into the ages of glacial landforms and the timing of glaciation. Studies using these techniques are showing that glaciation was restricted in extent during the last glacial cycle in regions such as central Tibet. In particular, Lehmkuhl *et al*. (2000, 2002) have shown that the morphology of the northern Nyainqêntanglha Shan and Mt. Jaggang, as well as the surroundings of Lakes Siling Co and Dagze Co, has demonstrated that the extent of ice during the late Quaternary was very limited. A luminescence date of 89 ± 10 ka on aeolian silt that overlies the oldest terminal moraines on the northern slope of the Nyainqêntanglha Shan helped define the timing of the penultimate glaciation (Lehmkuhl *et al*. 2002). In addition, Schäfer (2000) and Schäfer *et al*. (2002) presented cosmogenic surface exposure dates from the Tanggula Shan, dating moraines to between 123 ka and 261 ka that were only a few tens of kilometers beyond the present ice margins. These data suggest that there was no extensive plateau glaciation around 20 ka and that there was no ice sheet over the whole of Tibet during the Late or Middle Pleistocene. Owen *et al*. (2005) confirmed the Schäfer *et al*. (2002) study by undertaking a more extensive examination of the Tanggula Shan and extending their work onto the eastern slopes of the Nyainqêntanglha Shan.

Studies on the glacial successions in the Anyemaqen and Nianbaoyeze mountains in NE Tibet using cosmogenic surface exposure dating methods suggest that glaciers in the more monsoon-influenced regions of Tibet advanced during times of increased insolation, such as MIS-3 and the early Holocene (Owen *et al*. 2003c). This suggests that increased moisture flux during these times created higher precipitation in the form of snow at high altitude, which in turn led to positive glacial mass balances and glacial advance. However, although precipitation would have been reduced during the insolation minima of MIS-2 (global Last Glacial Maximum: LGM), temperatures were low enough to lead to positive glacier mass balances, allowing glaciers to advance, albeit not as far as during MIS-3.

Lakes records (sediments and shorelines) in Tibet and the adjacent deserts support the view that higher moisture flux occurred during times of increased insolation. Summaries of fluctuations of late Quaternary lake levels in Tibet and the desert margins of central Asia are provided in Fang (1991), Gasse *et al*. (1991, 1996), Frenzel (1994), Pachur *et al*. (1995), Tarasov *et al*. (1996), Benn & Owen (1998), Qin & Yu (1998), Tarasov & Harrison (1998) and Wünnemann *et al*. (1998). A discussion of the relationship between lake level changes, mountain glacier fluctuations and desert margins, and regional palaeoenvironmental changes in Central Asia is reviewed in Lehmkuhl & Haselein (2000).

There is increasing evidence for limited glacier advances during MIS-2 (LGM) throughout the semi-arid and monsoon-influenced regions of Tibet (Schäfer *et al*. 2002; Owen *et al*. 2003a, b, c, 2005). Schäfer *et al*. (2002), for example, produced cosmogenic surface exposure ages on erratics from the eastern margin of the Tibetan Plateau (close to the city of Litang, 99°33′E, 30°15′E) indicating that valley glaciation was only 10 km away from the present glacier snout. These were dated between 14 ka and 30 ka and suggest that the main glacial advance was at $17\,000 \pm 1000$ yr BP. Similarly, in the Qilian Shan and Li Ji Mountains in NE Tibet, cosmogenic surface exposure and OSL ages support limited glacial advances during the LGM (<10 km long; Owen *et al*. 2003a, b). Likewise, on the Karola Pass in southern Tibet north of the Transhimalya, Owen *et al*. (2005) provide cosmogenic surface exposure ages on moraines that demonstrate a MIS-2 glacial advance <10 km in extent.

There are only a few published studies on the Lateglacial and Holocene fluctuations of mountain glaciers on the Tibetan Plateau and the surrounding areas (Pu 1991; Lehmkuhl 1997; Owen *et al*. 2003a, b, c, 2005). Glacier advances have been dated to about 15 ka in West Kunlun, in the Tian Shan and in the mountain areas surrounding the Qaidam Basin (Kang 1992; Shi 1992; Guo *et al*. 1995; Owen *et al*. 2003a). Several Chinese authors (e.g. Wang & Fan 1987) argue for Holocene glacier advances, but most do not differentiate between Lateglacial end moraines or ice marginal limits and the LGM (MIS-2) terminal moraines. Chronology is mainly based on radiocarbon dating of organic matter that overlies terminal moraines. These dates are minimum ages and consequently only a few Pleistocene and Holocene glacial advances have been dated with a sufficient degree of confidence. Beug (1987), Wang & Fan (1987) and Lehmkuhl (1995a) suggest that Early Holocene glaciers had approximately the same size as modern glaciers. However, recent cosmogenic surface exposure dating of moraines in the Anyemaqen, on the Karola Pass and Gonga Shan suggest that glaciers advanced several kilometers beyond their present positions during the Early Holocene with an ELA depression of approximately 100 m (Owen *et al*. 2003c, 2005). The Little Ice Age and Neoglacial moraines are defined by dates on wood and branches incorporated in moraines in southern Tibet (Arza glacier; Wang & Fan 1987) and, for example, in the Qilian Shan (Pu 1991). Pollen records show a cooler and moister period during the Late Holocene (e.g. Sun & Chen 1991; Schlütz 1999). Historical records suggest that brief cold and wet intervals occurred periodically throughout the Late Holocene. These have been reported from the Taklimagan desert (in the Tarim Basin; Fig. 1) at about 2000 years BP and during the Little Ice Age for example (Yang 1991; Yang *et al*. 2002).

The nature of glacial fluctuations since the Little Ice Age (17th to 19th centuries) is discussed in Su & Shi (2002). They show that the mean temperature of monsoonal temperate glaciers in China has increased by 0.8°C since the Little Ice Age and has resulted in a decrease that amounts to an equivalent of 30% of the modern glacier area, a loss of some 4000 km^2 of glaciated area. They predict that by the year 2100 the temperature in the monsoonal temperate glaciers of China will rise by 2.1°C and that the glacier area will decrease by 75% (~9900 km^2 loss of glaciated area). Furthermore, they predict that precipitation will decrease in the coming decades and that glacier retreat will accelerate, but they argue it is unlikely to exceed a loss of 80% of the total glaciated area. General circulation models for global warming, however, suggest that monsoon precipitation will increase in the coming years, which may lead to increased snowfall and positive glacier mass balances. Therefore, there is considerable uncertainty as to the future of Tibetan glaciers. However, there is little doubt that glaciers have been retreating and that glacier ice has been warming throughout the last century. Such conditions clearly pose a serious threat to the water resources and environment throughout Central Asia.

Himalaya and Transhimalaya

A comprehensive review of the Quaternary glacial history of the Himalaya was presented in Owen *et al*. (1998), and Owen *et al*. (2002a) evaluated studies of the extent of glaciation throughout the Himalaya during the global LGM (~18–24 ka) as part of the EPILOG (Environmental Processes of the Ice Age: Land, Ocean, Glaciers) program of IGBP/PAGES program IMAGES (International Marines Studies of Global Change). Further descriptions of the Quaternary glacial history of each Himalayan region are provided in Ehlers & Gibbard (2004). The new data highlighted in these publications show that the local last glacial maximum for most of the Himalaya occurred during the early part of the last glacial cycle. In most areas, this probably occurred during MIS-3. In contrast, during the global LGM, glaciation was generally restricted throughout most of the Himalaya and Transhimalaya, with glaciers advancing less than 10 km from contemporary ice margins. Furthermore, recent OSL and comsogenic radionuclide surface exposure dating by Richards *et al*. (2000b) and Finkel *et al*. (2003) in the Khumbu Himal supports the view that glacial advances were restricted (<5 km) during the global LGM (MIS-2), with the local last glacial maximum occurring in the earlier part of the last glacial cycle. Correlating numerical dating studies, Finkel *et al*. (2003) suggest that glaciations can

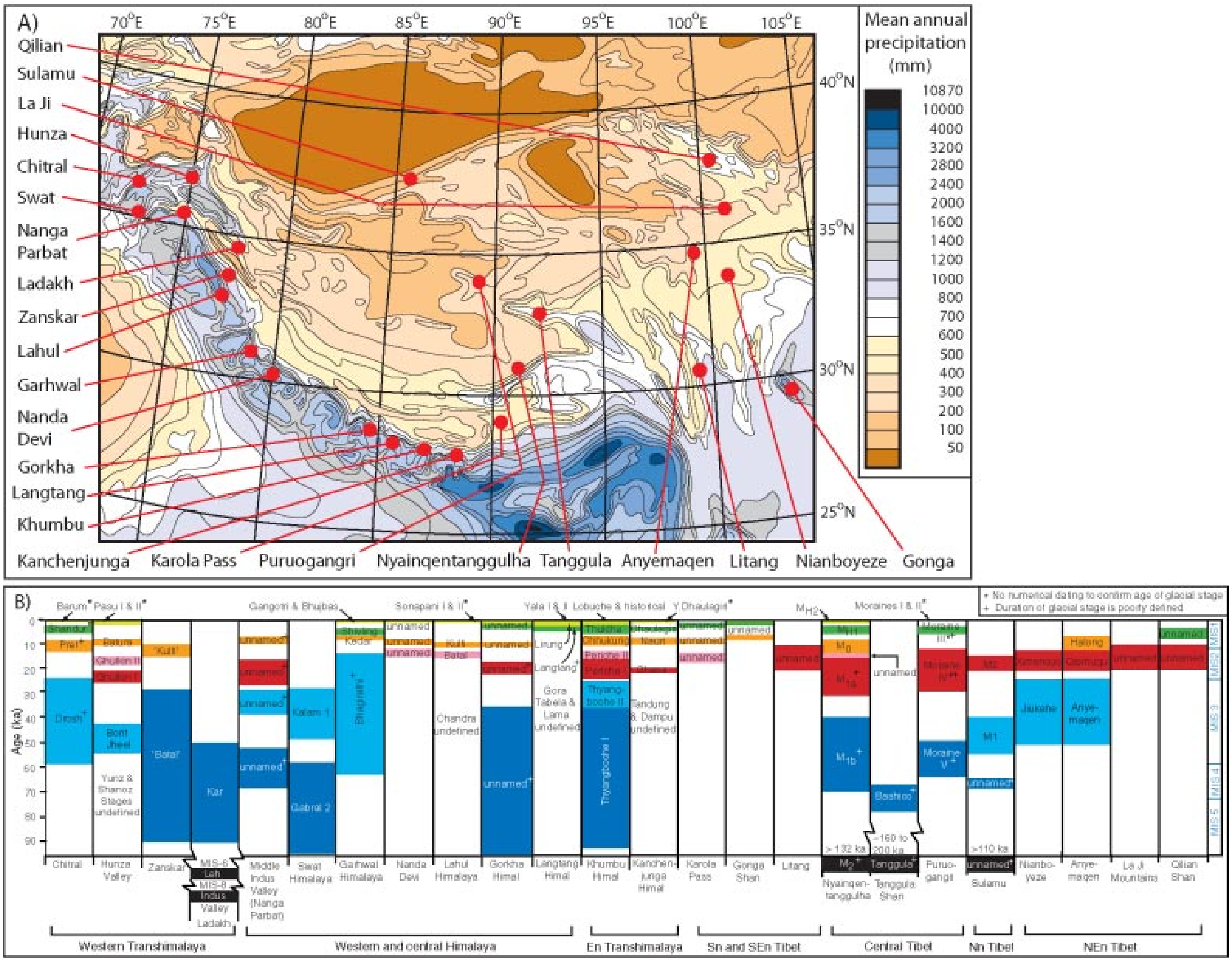

Fig. 3. The contemporary mean annual (A) precipitation across Tibet and the bordering regions showing (B) the locations and glacial chronologies that have reliable numerical dates in the Himalaya and Tibet (adapted from Owen *et al*. 2005). The color bars in (B) represent the likely duration of each glacial advance and the name of each glacial stage has been inserted into the box. An asterisk and cross after each name indicate that no numerical dating has been undertaken to confirm an age and the duration of the glacial is poorly defined, respectively. A tentative correlation is suggested by applying similar colors to the bars of the glacial stages that are likely synchronous. (MIS = marine oxygen isotope stage).

be broadly correlated along the Himalaya. This is summarized in Fig. 3.

The pattern of glaciation, however, appears to be different in the Hindu Kush at the far western end of the Himalayan–Tibetan orogen. Here, Owen *et al*. (2002c) showed that extensive valley glaciers extended to an altitude of ~1670 m a.s.l. during the LGM. Despite this study, most of the data help confirm the view of Benn & Owen (1998) that glaciation through the Himalayan–Tibet region was asynchronous with the Northern Hemisphere ice sheets, with the maximum glacial advances occurring during MIS-3 or MIS-5a to 5d (Owen *et al*. 2005). This asynchroneity is attributed to increased precipitation as snow at high altitudes due to a strengthened monsoon that penetrated further north into the Himalaya during times of increased insolation. In contrast, during times of lower insolation, particularly the global LGM (MIS-2), the influence of the monsoon was reduced, which in turn resulted in lower snowfall and snow accumulation and less extensive glaciation.

It has long been recognized that glaciers throughout the Himalaya and Transhimalayan regions have been retreating throughout the last century (Mayewski & Jeschke 1979; Mayewski *et al*. 1980), but the extent of retreat has not been adequately quantified. As discussed earlier with regard to Tibetan glaciers, it is difficult to assess the likely future trends due to global warming because of the complex feedbacks due to the predicted increase in monsoon precipitation and subsequent increased snowfall leading to possible positive glacier mass balances. Nevertheless, presently retreating glaciers pose serious threats to water resources on the Indian subcontinent as well as hazards such as those from glacial lake outburst floods (GLOFs) that are common as glaciers retreat (Richardson & Reynolds 2000).

Tian Shan and Altai Mountains

There are differing opinions concerning the extent of Late Pleistocene ice in the mountains north of the desert regions of Central Asia. Grosswald *et al*. (1994) and Grosswald & Kuhle (1994) present the view that an extensive ice sheet existed in these regions during the last glacial cycle. They argued that, in the Tian Shan, Late Pleistocene glaciers extended to the foothills, and mountain glaciers south of Lake Baikal terminated in the lake. They estimated LGM ELA depressions to be between 1150 and 1400 m for the Tian Shan, and about 1500 m for the mountains south of Lake Baikal. In contrast, Zech *et al*. (1996) and Heuberger & Sgibnev (1998) show clear evidence that glaciers were restricted to the Tian Shan mountain range and did not reach the foothills and had significantly lower ELA depressions than suggested by Grosswald *et al*. (1994) and Grosswald & Kuhle (1994).

The extent of Pleistocene glaciation in the Russian Altai is debated, but most researchers argue that valley glaciers reached Lake Teleski at 430 m a.s.l. (e.g. Baryshnikov 1992; Budvylovski 1993). The extent of ice was much greater in the western part of the Russian Altai than in the eastern region. In the eastern region, valley glaciers stretched down to the Kuray and Chuya Basins damming the main rivers and forming ice-dammed lakes, which produced the largest mega-floods (GLOFs) in the world (Baker *et al*. 1993; Rudoi 2002). In the eastern part of the Russian Altai and the Mongolian Altai there is evidence of two major Pleistocene glaciations of similar extent (Lehmkuhl 1998a; Klinge 2001; Klinge *et al*. 2003; Lehmkuhl *et al*. 2004). The limited extent of present and Pleistocene glaciers in the western part of the Russian Altai and in the Mongolian Altai is the result of reduced precipitation from west to east, which causes a rise of present and Pleistocene ELAs towards the east. There is an essential lack of numerical dating of glacial sediments in this particular region. Nevertheless, according to the present knowledge, most Late Pleistocene glacier advances in Mongolia and in the Russian Altai took place during MIS 2 and 4 (Grunert *et al*. 2000; Lehmkuhl *et al*. 2004).

Quantifying climate change from glacial geologic data

The most common method for reconstructing climate from glacial data is the use of glacier equilibrium lines. Benn & Lehmkuhl (2000) discuss the mass balance and glacier characteristics of glaciers in the high mountains of Asia and review the methods used to reconstruct former ELAs. They emphasize that methods of ELA reconstruction employed in low-relief environments are not always applicable in high mountains. Benn & Lehmkuhl (2000) argue that some of the methods of ELA estimation (e.g. terminus-to-headwall ratio: THAR) do not give true ELAs, and suggest glacial elevation indices (GEIs) as a more appropriate term, especially in the steep relief of the Himalaya and the Karakoram. Nonetheless, GEIs/ELAs are useful for reconstructing gradients across regions and for regional comparisons. Local and regional variations in ELAs are common in the mountains of Northern India, as shown by the work of Burbank & Fort (1985), Holmes & Street-Perrott (1989) and Sharma & Owen (1996). Nevertheless, these methods show that ELAs during Pleistocene glaciations dropped by between 300 to 500 m in the drier parts, and 600 to more than 1000 m in the wetter parts of the Tibetan Plateau and Himalaya. Figure 4 illustrates examples of present and Pleistocene glaciers and ELAs and can be used to help illustrate and semi-quantify the nature of glaciation in the high mountains of central Asia.

Profile A – A′ in Fig. 4 shows the distribution of present and Pleistocene ice from the Russian Altai in the west towards the Mongolian Altai in the east. The present ELAs in the Russian Altai and Mongolian Altai are between 2600 and 3100 m a.s.l. and between 3100 and 3600 m a.s.l., respectively (Bussemer 2001; Klinge 2001). The Late Pleistocene ELA is calculated to be ~2000 m a.s.l. in the eastern mountain ranges and >2800 m a.s.l. in the western parts of the Russian Altai. The ELA depression was >1000 m in the wettest parts of the western Russian Altai (today's annual precipitation: >1000 mm/a) and between 800 m and 500 m in the eastern part of the Russian Altai and in the Mongolian Altai, respectively (Lehmkuhl *et al*. 2004). This regional gradient of the ELAs was steeper during glacial times. Further to the east, in the Khangay, the ELA depression is again >1000 m (see below, profile C – C′). This may be the consequence of a strong monsoonal influence, which is also evident further east in the mountains of Northern China as the Qinling Shan or Wutai Shan (Lehmkuhl & Rost 1993; Rost 1998, 2000).

Profile B – B′ in Fig. 4 shows the increase in elevation of the ELA from the eastern margin towards the interior of the Tibetan Plateau for the Late Pleistocene (Lehmkuhl 1995a, 1998b). The Late Pleistocene ELA depression varies from between 800 m and 1000 m in the eastern part to 500 m in the western part of Tibet. There is a comparable increase of the modern and Late Pleistocene ELAs towards the plateau on the northern and northeastern flank; for example, from 3300 m a.s.l. in the Qinling Shan (34°15′N, 109°10′E) to 4150 m a.s.l. in the La Ji Shan (36°55′N, 101°E) (Lehmkuhl & Rost 1993; Rost 2000). This general increase in ELA corresponds with the decrease in precipitation from east to west, which was more pronounced during glacial times (cf. Lehmkuhl 1995a).

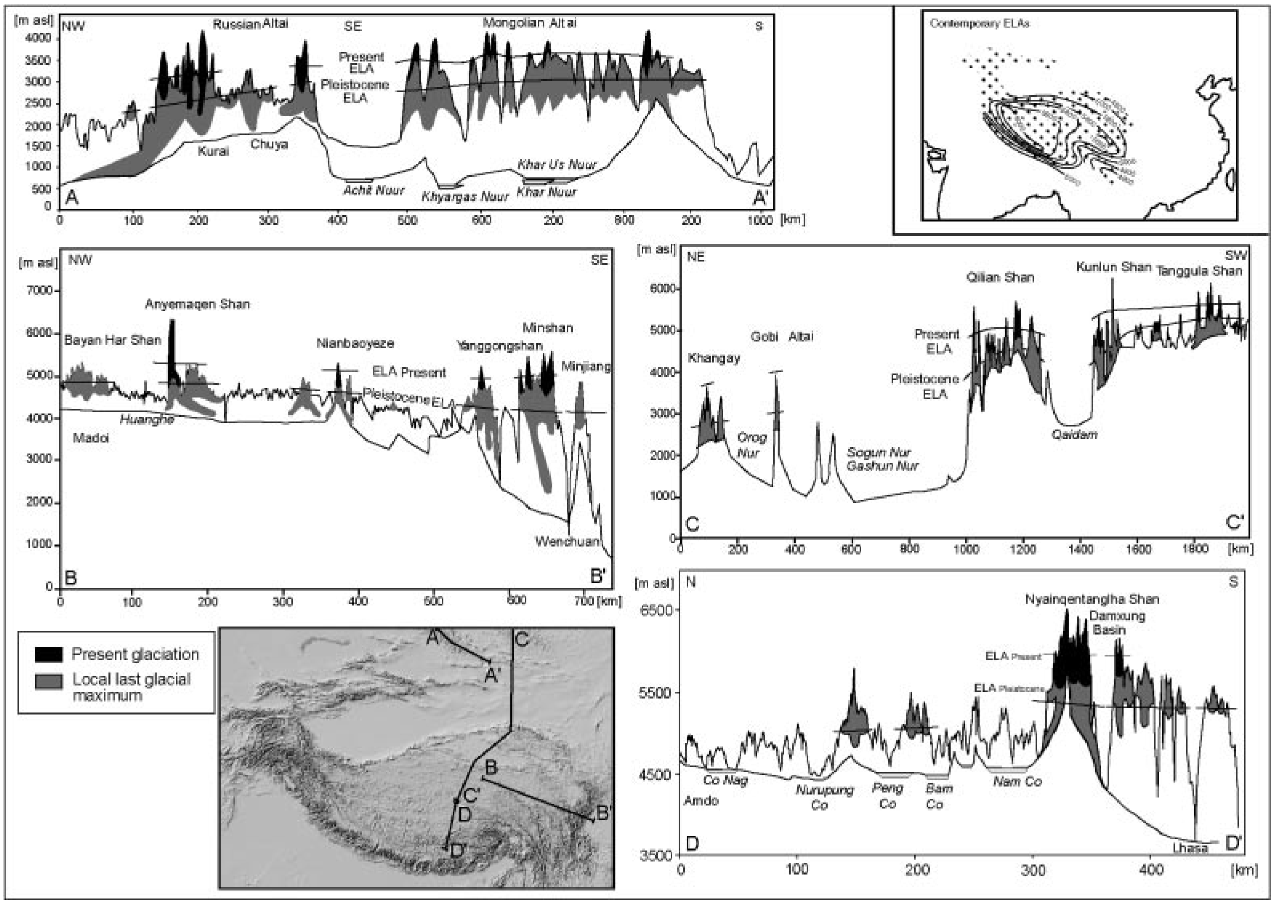

Fig. 4. Map with selected cross-sections of Central and High Asia, including the distribution of present and local last glacial maximum ice and the equilibrium line altitudes (ELAs). Cross-section A – A′ shows the distribution of present and local last glacial maximum ice and the ELAs from the Russian Altai to the Mongolian Altai (according to Klinge 2001 and Lehmkuhl 1999, 2002). Cross-section B – B′ shows the distribution of present and local last glacial maximum ice and the ELAs from the Khangay and Gobi Altai to the Tibetan Plateau (according to Lehmkuhl 1995, 1997b). Cross-section C – C′ shows the distribution of present and local last glacial maximum ice and the ELAs from the eastern fringe of the Tibetan Plateau (according to Lehmkuhl 1995, 1998b). Cross-section D – D′ shows the distribution of present and local last glacial maximum ice and the ELAs from southern Tibet (according to Lehmkuhl *et al*. 2000, 2002). The inset map in the top right corner shows the variation in contemporary regional snowlines across Tibet and the bordering mountains (adapted from Benn & Owen 1998).

In addition, fossil involutions (cryoturbation) and ice-wedge casts provide evidence for LGM permafrost in the Basin of Zoige (3400 to 3500 m a.s.l.) and on the southern shores of Qinghai Lake (Porter *et al*. 2001). Past temperatures can be estimated from these periglacial features and the existence of sand wedges indicates higher aridity – thus supporting the possibility that the ELA depression during the Late Pleistocene was between 600 and 800 m (Lehmkuhl 1998b).

The extent of late Quaternary glaciation from the mountains of Mongolia (Khangay and Gobi Altai) towards the northern, central and southern parts of the Tibetan Plateau is shown in Fig. 4. Presently, there are no glaciers in the central part of the Khangay and Gobi Altai (Profile C – C′ in Fig. 4). However, Lehmkuhl & Lang (2001) calculated the Pleistocene ELA to be between 2700 and 2800 m a.s.l. in the Khangay. Lehmkuhl (1998a) suggested that Quaternary glaciers were present in the Gobi Altai during glacial times. This suggests that the ELA depression in the Gobi Altai was ~1000 m.

On the northern slopes of the Qilian Shan in northeast Tibet, the modern ELAs are between 4600 and 5000 m a.s.l. and the Pleistocene ELA is ~3800 m a.s.l. (an ELA depression of ~1000 m; Lehmkuhl & Rost 1993; Hövermann *et al*. 1998). The modern ELA is 200 to 300 m lower in the outermost ranges of the Qilian Shan than in the innermost ranges (cf. Lehmkuhl 1992, 1995b, 1998b; Shi 1992). This pattern is also seen south of the Qaidam basin, from the Kunlun Shan in the north towards the Tanggula Shan in the south-central part of the Tibetan Plateau (profile C – C′ in Fig. 4). There is a sharp increase in elevation of the ELA from 4550 m a.s.l. in the innermost ranges of the Kunlun

Shan on the northern slope towards the Plateau to 4950 m a.s.l. on the southern slope. This is the only range that is presently glaciated. It has an ELA of ~5300 m a.s.l. and a Pleistocene ELA depression of ~600 m. At the marginal (outermost) northern range, Kuhle (1987) argues that the ELA depression during the Late Pleistocene was ~1000 m a.s.l. However, on the southern slope of the main range of the Kunlun Shan system, which is 70 km wide, glaciers advanced only a few kilometres from the Kunlun Pass to the interior of the Plateau to the south. There is no evidence to support Kuhle's (e.g. 1993) hypothesis that glaciers transported boulders 400 km north from the central Tanggula Shan (cf. Lehmkuhl 1995a, 1998b; Lehmkuhl & Hövermann 1996). In the Tanggula Shan, the modern ELAs are ~5700–5800 m a.s.l. and the Late Pleistocene ELA depression is between 500 and 600 m (Lehmkuhl 1998b). There is little difference between ELAs on the northern and southern sides of the Tanggula Shan. The general increase of modern and Late Pleistocene ELAs towards the plateau corresponds with the decrease in precipitation from north to south, which was more pronounced at the eastern margin during the ice age (cf. Lehmkuhl 1995a).

Profile D – D' (Fig. 4) shows that the present ELA in Southern Tibet is between 5800 and 6000 m a.s.l., with a Pleistocene ELA depression of 300 to 500 m in the drier parts of the Plateau, and 600 to 800 m in the wetter southern and southwestern mountain slopes of the Nyainqêntanglha Shan. Luminescence dates on aeolian silt that overlies the oldest terminal moraines on the northern slope of the Nyainqêntanglha Shan indicate an older glacier advance of the penultimate glaciation (89 ± 10 ka; Lehmkuhl *et al*. 2002). However, the geomorphology of the areas north of the Nyainqêntanglha Shan shows clear evidence that the extent of ice during the Late Pleistocene was limited (Lehmkuhl *et al*. 2000, 2002). In addition, the cosmogenic surface exposure ages presented by Schäfer (2000), Schäfer *et al*. (2002) and Owen *et al*. (2005) in the Tanggula Shan provide evidence for a slightly larger extent of ice during the penultimate glaciation.

Effects of surface uplift on glaciation

Controversy exists over the timing and rates of uplift throughout Tibet and the bordering mountains. The extension exhibited by normal and reverse faulting along Tibet's margins, volcanic activity in northeastern Tibet and palaeobotanical evidence in southern Tibet, however, suggests that much of Tibet probably reached its present elevation by ~13–14 Ma (Colman & Hodges 1995; Edwards & Harrison 1997; Blisniuk *et al*. 2001; Spicer *et al*. 2003). The mountain ranges that surround the Tibetan Plateau are generally much younger than 13 Ma and surface uplift rates are estimated to be a few mm/year (e.g. Zeitler *et al*. 2001). These uplift rates are usually defined using exhumation rates determined by fission track and mineral cooling ages, and incision rates determined from dating strath terraces and calculating contemporary sediment fluxes and volumes of deposited sediments (Collins 1998; Clift *et al*. 2001). Since the mountains produce positive relief, it is generally considered that the exhumation and incision rates must be less than or equal to the surface uplift rates. It is likely that current denudation is keeping pace with uplift and that the net gain in elevation is small. It is therefore highly unlikely that maximum uplift exceeds 10 mm/year and surface uplift is most likely in the order of ~1 mm/year in the actively growing mountain ranges.

Most of the late Quaternary glacial geologic studies in Tibet and the bordering mountains are limited to the last 100 000 years. A range of surface uplift rates between 1 and 10 mm/year suggests that the mountains have not uplifted >100 m since ~100 ka. In the most rapidly rising regions, such as Nanga Parbat, uplift could be in the order of several hundred metres to as much as a kilometre. In most regions, however, uplift is not considered to have greatly influenced the style of late Quaternary glaciation. The Ladakh Range of northern India may be an exception. Here, glaciation may have become dramatically more restricted over the past few hundred thousand years. Taylor & Mitchell (2000) and Bovard (2001) suggest that the Greater Himalaya and Pir Panjal to the south may have uplifted enough to restrict the northward penetration of the south Asian summer monsoon, reducing precipitation and glacial cover.

Conclusions

An understanding of the nature of late Quaternary glaciation in Tibet and the bordering mountains is still in its infancy. However, the application of modern geomorphic and sedimentological techniques, and the development of new dating techniques, such as cosmogenic surface exposure and OSL dating, have opened up the possibility for regional and temporal correlations across this vast region. Although the extent of former glaciers is relatively well known, in most regions it is still not known when glaciers advanced to particular positions. Nevertheless, at this stage we are able to conclude the following:

- In most regions of Tibet and much of the Himalaya, glaciation was most extensive earlier in the last glacial cycle, possibly MIS-3 or MIS-5a to 5d. Furthermore, glaciation was generally most extensive prior to the last glacial cycle. It is possible, however, that the behaviour of glaciers influenced by the different climatic systems was different at different times, and more detailed geomorphic and geochronological studies are needed to fully explore regional variations.

- There is a strong topographic control on glaciation, particularly in regions that have rainshadow effects. This is most evident from the strong regional variability in present and past ELAs.
- Glaciation was more extensive in monsoon-influenced regions, and was strongly controlled by changes in insolation that control the geographic extent of the monsoon and consequently precipitation distribution. This strongly influences glacier mass balances, allowing glaciers to advance during times of increased precipitation in high altitude regions, and these are associated with insolation maxima during glacial times.
- Tectonic uplift has had little effect on glaciation during the last glacial cycle, but may have been important in controlling differences in glacial styles and extent over many hundreds of thousands of years.
- Drainage development, including lake level changes, has been strongly controlled by climatic oscillations, particularly on millennial time scales. However, centennial and decadal fluctuations are also discernible.
- Changes in glacial ice volume after the global LGM were relatively small in the Tibet–Himalayan region compared to the Northern Hemisphere ice sheets. It is therefore unlikely that meltwater draining from Tibet and the bordering mountains during the Lateglacial and early Holocene would have been sufficient to affect oceanic circulation upon flowing into the adjacent oceans. However, changes in surface albedo may have influenced the dynamics of monsoonal systems and had implications for global climate change.
- Despite the relatively small volume of glacial ice that existed in this region during glacial times, Tibet and the bordering mountains presently have the greatest concentration of glaciers outside the Polar Regions. Since the Little Ice Age, and particularly during this century, glaciers have been progressively retreating. This pattern is likely to continue throughout this century, exacerbated by human-induced global warming.

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