Espizúa, L. (2004). Pleistocene glaciations in the Mendoza Andes, Argentina. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 69–73. Elsevier.

Espizúa, L., and Bigazzi, G. (1998). Fission-track dating of the Punta de Vacas glaciation in the Río Mendoza valley, Argentina. *Quaternary Science Reviews* **17**, 755–760.

Hall, K. (2004). Quaternary glaciation of the sub-Antarctic Islands. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 339–345. Elsevier.

Heine, J. (1995). Late Quaternary glacier advances in the Ecuadorian Andes: a preliminary report. *Quaternary of South America and Antarctic Peninsula* **9**, 1–22.

Heine, J. (2004). Late Quaternary glaciations of Ecuador. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 165–169. Elsevier.

Ingólfsson, O. (2004). Quaternary glacial and climate history of Antarctica. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 3–44. Elsevier.

Kaplan, M., Douglass, D., Singer, B., Ackert, R., and Caffe, M. (2005). Cosmogenic nuclide chronology of Pre-last glacial maximum moraines at Lago Buenos Aires, 46 °S, Argentina. *Quaternary Research* **63**, 301–315.

Mahaney, W. (2004). Quaternary glacial chronology of Mount Kenya massif. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 227–231. Elsevier.

Mark, B., Seltzer, G., and Rodbell, D. (2004). Late Quaternary glaciations of Ecuador, Peru and Bolivia. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 151–163. Elsevier.

Meglioli, A. (1992). Glacial geology of Southernmost Patagonia, the Strait of Magellan and Northern Tierra del Fuego. PhD Dissertation, Lehigh University, Bethlehem, USA. Unpublished.

Mercer, J. H. (1976). Glacial history of Southernmost South America. *Quaternary Research* **6**, 125–166.

Osmaston, H. (2004). Quaternary glaciations in the east African mountains. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 139–149. Elsevier.

Porter, S. (1981). Pleistocene glaciations in the Southern Lake District of Chile. *Quaternary Research* **16**, 263–292.

Rabassa, J., and Clapperton, C. (1990). Quaternary glaciations of the Southern Andes. *Quaternary Science Reviews* **9**, 153–174.

Rabassa, J., Coronato, A., and Salemme, M. (2005). Chronology of the Late Cenozoic Patagonian glaciations and their correlation with biostratigraphic units of the Pampean region (Argentina). *Journal of South American Earth Sciences* **20**, 81–103.

Rodbell, D. (1993). Subdivision of Late Pleistocene moraines in Cordillera Blanca, Peru, based on rock-weathering features, soils and radiocarbon dates. *Quaternary Research* **39**, 133–143.

Schubert, C., and Clapperton, C. (1990). Quaternary glaciations in the Northern Andes (Venezuela, Colombia and Ecuador). *Quaternary Science Reviews* **9**, 123–135.

Shanahan, T., and Zreda, M. (2000). Chronology of Quaternary glaciations in East Africa. *Earth and Planetary Science Letters* **177**, 23–42.

Singer, B., Ackert, R., and Guillou, H. (2004). $^{40}Ar/^{39}Ar$ and K-Ar chronology of Pleistocene glaciations in Patagonia. *Geological Society of America Bulletin* **116**, 434–450.

Suggate, R. (2004). South Island, New Zealand; ice advances and marine shorelines. In *Quaternary Glaciations – Extent and Chronology*, Part III (J. Ehlers and P. Gibbard, Eds.), pp. 285–291. Elsevier.

Ton-That, T., Singer, B., Mörner, N., and Rabassa, J. (1999). Datación de lavas basálticas por $^{40}Ar/^{39}Ar$ y geología glacial de la región del Lago Buenos Aires. *Revista de la Asociación Geológica Argentina* **54**(4), 333–352.

Late Pleistocene Events in Beringia

S A Elias, Royal Holloway, University of London, Surrey, UK
J Brigham-Grette, University of Massachusetts, MA, USA

Introduction

The region of the western Arctic stretching from the Lena River in northeast Russia to the Mackenzie River in Canada, is geographically know as Beringia. From a glaciological standpoint, Beringia, including the vast continental shelves of the Bering and Chukchi Sea, has presented a great anomaly throughout the Cenozoic. While nearly all other high latitude regions of the Northern Hemisphere were covered by ice sheets during each glaciation, the lowlands of Beringia remained ice-free, thus providing a refuge for high-latitude flora and fauna. Due to the build up of continental-sized ice sheets over North American, Greenland, parts of Eurasia, and other regional mountain ranges during Pleistocene glaciations, global sea level repeatedly dropped significantly exposing large portions of the continental shelves. For instance, during the last glaciation, global sea level dropped by 120–130 m (Lambeck *et al*., 2002) exposing the continental shelf regions of the Bering and Chukchi seas between Siberia and Alaska. These vast shelves lie only about 30–110 m below modern sea level. Therefore, as continental ice sheets grew, sea level dropped, and the shelf regions between northeast Russia and Alaska became dry land forming the Bering land bridge. By mapping the 120 m contour of the sea floors for the Bering and Chukchi seas, it is possible to roughly trace the outline of the former land bridge region (**Fig. 1**). This land bridge was no narrow isthmus between two continents. Rather, from north to south, the land bridge extended more than 1,000 km and may have existed for nearly 10,000 years (from 30,000 to 20,000 years ago) during the last glacial advance. From east to west, the land bridge was narrow at the center (in the Bering Strait region) but broad at

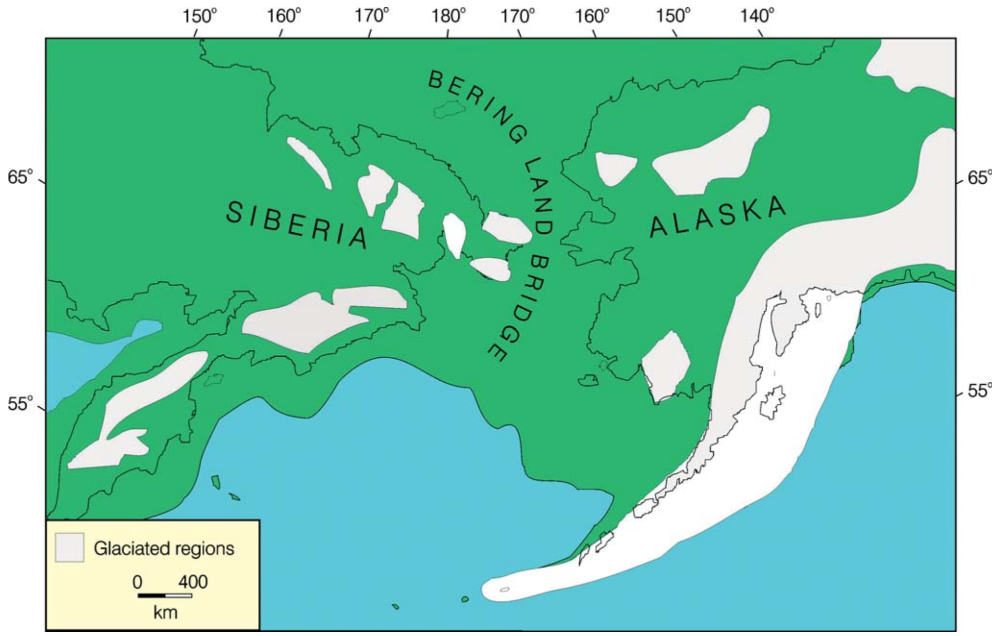

Figure 1 Map of northeast Asia and northwestern North America, showing the regions of western and eastern Beringia, and the Bering Land Bridge.

both the north and south ends. The area of the land bridge was larger than the state of Texas.

When it existed, the Bering Land Bridge cut off circulation between the waters of the North Pacific and Arctic oceans. This in turn greatly increased continentality and diminished the in-land flow of relatively warm, moist air masses from the North Pacific. Beringia was also influenced by being downwind of large ice sheets in Scandinavia and the Eurasian north, themselves creating widespread aridity during full glacial conditions (Siegert *et al.*, 2001). Moisture stripped from the westerlies by these ice sheets left little but strong dry winds to sweep the landscapes of Beringia. Consequently, despite being cold enough to accumulate glacial ice, the desiccating environment of the land bridge restricted glaciation to isolated mountain ranges and kept glacial ice out of the lowlands of Alaska, the Yukon Territory, and eastern Siberia, plus the land bridge itself. The largest of the Alpine glacier complexes was, in fact, the westernmost extension of the Cordilleran Ice Sheet that once occupied western North America. This ice sheet covered the Alaska Range, and extended south and west to the Alaskan Peninsula and the continental shelf regions of Prince William Sound and the Gulf of Alaska (Mann and Hamilton, 1995). The Beringian lowlands remained ice-free throughout the Pleistocene.

Beringia is most conveniently partitioned for discussion into three geographic sectors: western Beringia, central Beringia, and eastern Beringia. Western Beringia is comprised of the regions of northeast Russia, from the Verkhoyansk Mountains in the west, to the Bering and Chukchi Sea coasts in the east, and the Sea of Okhotsk in the south. The Verkhoyansk Mountains lie along the eastern margin of the lower Lena River Basin, and form the western limit of some characteristic Beringian plant taxa. The central sector of Beringia is that area of the Bering Land Bridge that is now submerged beneath the Bering and Chukchi seas, forming what was once a broad, relatively flat landscape with only a few highlands (the modern islands of the northern Bering Sea region). Eastern Beringia is defined as the unglaciated regions of Alaska and the Yukon Territory. The eastern boundary of Beringia has been considered to be the lower Mackenzie River in the Canadian Northwest Territories (Hopkins, 1982). This places the eastern edge of Beringia along the margin of the Laurentide ice sheet at its maximum extent.

The vast regions of Beringia contained enormous topographic diversity, spanning over more than 20° of latitude and 80° of longitude (**Fig. 1**). One of the principal differences between western and eastern Beringia is that the former region is topographically more complex and rugged than the latter (Brigham-Grette *et al.*, 2004). The northward-draining river basins of the Lena, Indigirka, and Kolyma are broken by steep mountain ranges, as well as broad uplands. The geography of eastern Beringia is marked by three sets of mostly east-west trending mountain ranges. From south to north these comprise (1) the Pacific coastal ranges that continue southwest along the Alaska Peninsula, (2) the Alaska Range, about 100 km inland from the Gulf of Alaska, and (3) the Brooks Range, separating the interior lowlands from the North Slope region. The interior regions of Alaska and the Yukon Territory lie in the rain shadow of two high mountain ranges, which block Pacific moisture.

Contrasts in Paleoclimate Across Eastern and Western Beringia

During the past 20 years, cooperation between Russian and North American scientists has added significantly to our understanding of Beringian paleoenvironments, especially concerning the last glaciation. For instance, we have learned there were significant differences between the climate and vegetation of western versus eastern Beringia during the marine isotopic stage 3 interstadial (MIS 3, ca. 48–28 ka). In western Beringia, temperatures apparently rose to nearly-modern levels, allowing the treeline to migrate northwards and into mountain ranges to virtually its modern limit (Anderson and Lozhkin, 2001). In contrast to this, however, very little evidence exists for the expansion of coniferous forest in eastern Beringia during MIS 3. Based on fossil beetle evidence, there appears to have been different degrees of warming in the northern and southern sectors of eastern Beringia during this interstadial (Elias, 2001). The Arctic regions experienced mean summer temperatures as much as 1.5° C warmer than modern levels at about 35 ka, whereas the synchronous warming in sub-Arctic regions remained about 2° C cooler than modern levels.

As in other high-latitude regions, the Last Glacial Maximum (LGM) was an interval of climatic cooling. Paleotemperature estimates based on fossil beetle assemblages (Elias, 2001) suggest that average summer temperatures (TMAX) were as much as 7.5° C cooler at 22 ka than they are today in the Yukon Territory. However, it appears that this cooling of summer temperatures was far from uniform across Beringia. On the Seward Peninsula, for instance, TMAX was depressed by a maximum of 2.9° C. Most eastern Beringian paleotemperature estimates based on LGM beetle data fall within a range of 3–5°C colder than modern TMAX. Mean January temperatures (TMIN) in eastern Beringia were probably depressed by very little, if at all (Elias 2001). The fossil beetle evidence (Alfimov and Berman, 2001) indicates that LGM summer temperatures in western Beringia were essentially as high as they are today in many regions, and TMIN values remained close to modern levels as well. However, not all researchers agree on these temperature reconstructions. For instance, periglacial features that developed during the LGM, indicate that mean annual temperatures dropped significantly (Hopkins, 1982). Most paleoclimatologists would agree that on a very broadscale, Beringia was relatively dry and cold with cooler summers during the LGM. More mesoscale patterns indicate east to west trends in temperature and moisture gradients with colder and drier conditions dominant over eastern Beringia (Brigham-Grette *et al*., 2004). Guthrie (2001) postulated that during the LGM, western Beringia was kept arid by a huge, stable, high-pressure system north of the Tibetan Plateau. North Atlantic sea ice cover and the Scandinavian/Eurasian ice sheets would have cut off the main source of westerly moisture, enhancing aridity in Asiatic steppe regions. Lowered sea level and the Cordilleran Ice Sheet would have reduced the moisture available to interior regions of eastern Beringia.

Sea-Level History

Six marine raised beaches have been documented along the more stable parts of the Alaskan coastline, providing evidence that sea level was higher than present during some earlier interglacial warm periods. The oldest four high sea-level events occurred during the late Pliocene to Middle Pleistocene and are significant with respect to paleoclimate (Kaufman and Brigham-Grette, 1993). The two Late Pleistocene highstands are known as the Pelukian transgression, dated between 125,000 and 115,000 years BP (MIS 5e), and the Simpsonian transgression, dated between about 88,000 and 70,000 years BP (MIS 5a; **Table 1**) (Brigham-Grette and Hopkins, 1995).

The last interglacial, or MIS 5, broadly includes the period from about 130,000 years to 75,000 years BP. The time period is normally characterized as including the peak of the last interglacial (MIS 5e) followed by two mild and two cooler intervals (MIS 5 d through a) which eventually terminated in the first of the two major glacial intervals, MIS 4 (**Fig. 2**). The peak of the last interglacial, MIS 5a, is remarkable across Beringia because eustatic sea level stood roughly 6–7 meters higher than it is today and terrestrial climates were generally warmer. In the Bering and Chukchi seas, it has been estimated that for much of this interval, winter maximum sea ice limits were as much as 800 km further north, well north of the narrow Bering Strait; summer sea ice was also less extensive and marine mollusks from the Bering Sea extended their range to Barrow, Alaska (Brigham-Grette and Hopkins, 1995). Similar range extensions

Table 1 Timing of major late Pleistocene cold events and marine transgressions in Beringia

Cold Event	*W. Beringian terminology*	*E. Beringian terminology*	*Age (calibrated yr BP × 1000)*	*Marine transgression event*
			125 to >100	Pelukian
Middle MIS 5			100–88	
			<88 to >70	Simpsonian
MIS 4	Zyryan glaciation	Early Wisconsinan glaciation	70–55	
MIS 2	Sartan glaciation	Late Wisconsinan glaciation	25–15	

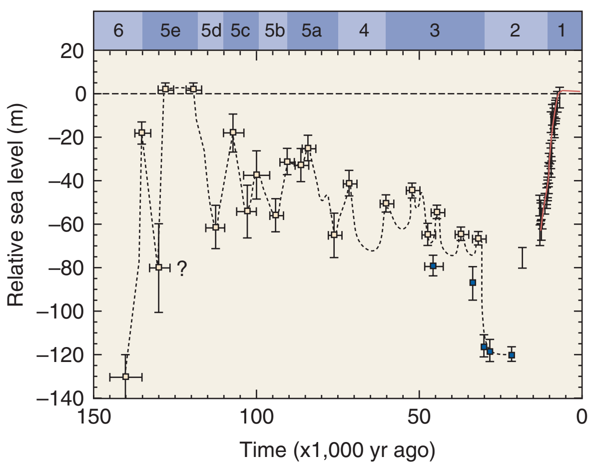

Figure 2 Relative sea-level curve for the last glacial cycle for Huon Peninsula, Papua New Guinea, supplemented with observations from Bonaparte Gulf, Australia. Error bars define the upper and lower limits. Time scale is based on U-series ages of corals older than about 30 ka and on calibrated ^{14}C ages of the younger corals and sediments (after Lambeck *et al.*, 2002).

have been documented in marine sediments on the Russian coast and confirmed with the expansion of species of foraminifer from northern Japan as far north as eastern Chukotka (Brigham-Grette *et al.*, 2001). The treeline spread to the Russian coast across most of western Beringia with range extensions among some tree species of 600 to 1,000 km (Lozhkin and Anderson, 1995). At the same time in Alaska the northward shift of treeline was more modest but extended at least 50 km north of the Brook Range (Brubaker *et al.*, 2005). Temperatures estimates based on pollen and fossil beetle assemblages across the interior of Alaska suggest that summer temperatures were similar to today, but conditions were considerably wetter (Muhs *et al.*, 2001). In general across Beringia, summer temperatures were 0–2° C warmer, winters a few degrees cooler and conditions were generally wetter.

Rapid Onset of Glaciation During MIS 5

Numerous workers have now documented that parts of the high Arctic can become rapidly glaciated at the end of interglacials while eustatic sea level is still high. The Beringian region is no exception. Various researchers including Pushkar *et al.* (1999), demonstrated that the mountains to the north and west of Kotzebue Sound had to have been rapidly glaciated near the end of MIS 11 (possibly stage 9). Such an explanation was necessary in order to explain the continuous deposition of extensive pro-deltaic interglacial marine sediments overlain by glacially deformed glaciomarine muds. Because these deposits lie some 400 km from the edge of the shallow continental shelf, eustatic sea level had to have remained high as a consequence of widespread glaciomarine sedimentation in Kotzebue Sound in front of thin but extensive valley glaciers emanating from the western Brooks Range.

Recent research in Chukotka (Brigham-Grette *et al.*, 2001) and parts of southwest Alaska (Kaufman *et al.*, 1996; 2001) has lead to the realization that these regions also experienced rapid valley glacier expansion at times directly after the peak of MIS 5e. On outer Chukotka near the narrow Bering Strait, Brigham-Grette *et al.* (2001) reassigned the Pinakul Formation, once thought to be of Early Pleistocene age, to MIS 5a–e. The depositional sequence consists of an upward shallowing interglacial marine sequence characterized by warm faunas (the lower Pinakul Formation of MIS 5) overlain by a thin coarse gravel buried beneath a thick glaciomarine sequence containing classic dropstones (upper Pinakul Formation). The entire sequence is capped by glacial till, dated to MIS 4. Brigham-Grette *et al.* (2001) interpreted the section to represent 5e marine deposits overlain by an unconformity and then the rapid deposition of glaciomarine sediments in MIS 5d. Deposition of such a thick section of glaciomarine mud in the shallowest coastal reaches of the Bering Strait requires that eustatic sea level remained near modern levels as rapidly expanding valley glaciers in coastal mountains reached the sea. Dating of the deposits is based on amino acid age estimates, radiocarbon, U/Th ages on concretions, and ESR ages.

Valley glaciers responded in a similar fashion at roughly the same time in the Bristol Bay region of southwestern Alaska. Kaufman *et al.* (2001) have demonstrated that mountain glaciers extended as much as 100 km from their cirques and intersected the sea before eustatic sea level had dropped. Like the Pinakul sequence, this advance probably occurred either at the MIS 5e/5d transition or the MIS 5a/4 transition based on stratigraphic relationships with the underlying Old Crow Tephra (ca. 140 ka) and a TL age estimate of 70 ka $\pm$ 10 on basalt overrun by advancing glacier ice. However, calibrated amino acid age estimates and infrared-luminescence ages offer the best geochronological control.

The rapid glacierization in Beringia at the end of MIS 5 was presumably driven by a precipitous drop in insolation while oceanic temperatures remained warm and eustatic sea level remained relatively high. However, changes in oceanic circulation also contributed to the onset of widespread Arctic glaciation in sensitive regions. Marine evidence suggests that major climate shifts such as the MIS 5e/5d transition also were driven by changes in deep water flux (Adkins *et al.*, 1997). There is evidence for a rapid reorganization of deepwater production at the 5e/5d transition over a period of less than 400 years (Adkins *et al.*, 1997). Hemispheric changes in the ocean and atmospheric system likely contributed to the rapid build up of the Eurasian Ice Sheet at the 5e/5d transition. Reconstructions of the Barents and Kara Ice sheets at this time suggest that the glaciers came on to shore and dammed large proglacial lakes in the Ural Mountains as early as 85–90 ka cal yr BP (Mangerud *et al.*, 2001). Ice sheets then retreated northward and re-advanced in Stage 4 to the Markhida Moraine by about 60 ka cal yr BP.

Changes in the height and extent of the Scandinavian and Barents/Kara sea ice sheets probably had a significant influence on the temporal and spatial response of the eastern Siberia and western Beringia (northeast Siberia) to hemispheric scale climate change. Ice-sheet and GCM modeling of the Scandinavian and Eurasian ice sheets by Siegert and Marsait (2000) suggests the extent to which changes in the size of these ice sheets diminished the temperature and precipitation influence of the North Atlantic eastward across the Russian Arctic. Fine-tuned models of ice sheet size for parts of the Late Pleistocene (Siegert *et al.*, 2001) allow for more realistic assessments of how the physical stratigraphy of western Beringia may have been influenced by 'downwind' effects while being upwind of the maritime influences of the Bering Strait and the sea ice conditions in the Bering Sea.

The timing of ice-sheet growth in Eurasia and Beringia fits well with the notion for early ice build up over parts of the Canadian Arctic at this same time (Brigham-Grette and Hopkins, 1995). In Northern Alaska along the Beaufort Sea coast, the Flaxman glaciomarine deposits are dated to substage 5a (Carter *et al.*, 1991). It is also well-known and mapped below 7 m altitude on the Alaskan Arctic Coastal Plain (Brigham-Grette and Hopkins, 1995) recording both a high sea-level event that flooded the Bering Straits and the Heinrich-like collapse into the Beaufort Sea via Amundsen Gulf of a large ice sheet building over the Canadian Arctic islands presumably starting in stage 5d. Erratics in the Flaxman beds are exclusively of Canadian provenience reaching all the way to the Chukchi Sea coast of Alaska. TL ages on sediments, several ESR ages on shells, and a uranium-series age on whale bone, all place the age of the sediments at about 75 ka BP. The presence of large whale bones in these deposits east of Barrow is especially significant since it requires that the Bering Strait was submerged enough (sea level above -55 m) to allow for the seasonal migration of bowhead and graywhales from the open Pacific, as they do today.

Flooding of the Bering Strait in MIS 5a, coupled with an insolation high, may have contributed to the collapse of these Canadian and Eurasian ice sheets. The collapse of these ice sheets may have then allowed the penetration of moisture across the continent, permitting the expansion of valley glaciers in the Bering Straits region and elsewhere in the high latitudes. Afterwards, increasing continentality, sea ice cover, and the expansion of the Scandinavian/Eurasian ice sheet began to limit the available moisture supply across most of Beringia.

Early Wisconsin/Zyryan Glaciation (Marine Isotope Stage 4)

Glaciation during MIS 4 is referred to as the early Wisconsin in North America and the Zyryan in northeastern Russia, using traditional time stratigraphic terms. With the exception of the Alaska Range and Chugach/St Elias Mountains facing the Gulf of Alaska, the early Wisconsin/Zyryan glacial ice caps and mountain glacier complexes throughout much of Beringia were considerably more extensive than during the subsequent late Wisconsin/Sartan glaciation in MIS 2 (**Fig. 3**). Satellite and aerial photo mapping followed up by field studies of moraine limits across much of Chukotka indicates that Zyryan glaciers advanced 50–200 km from their source areas during MIS 4 (Heiser and Roush, 2001). Glushkova (1995) suggested that nearly 40% of Chukotka was ice covered during this interval and

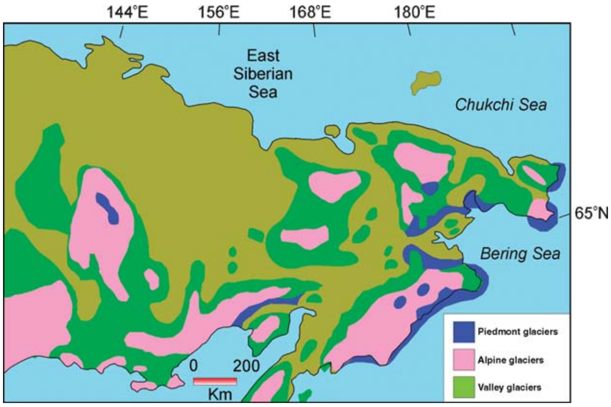

Figure 3 Map of western Beringia, showing extent of glaciers during the Zyryan glaciation (after Brigham-Grette *et al.*, 2004).

in general, glacial advances were up to three times more extensive than during the Sartan glaciation. On the southeast corner of Chukotka, mountain glacier complexes coalesced into a piedmont lobe of sufficient size that the glacier terminated on St Lawrence Island leaving a moraine belt mappable on the sea floor (Brigham-Grette *et al.*, 2001).

Correlation between Zyryan glacial events in western Beringia and early Wisconsinan glacial events in eastern Beringia can be difficult (Heiser and Roush, 2001) because of dating problems. The ages of MIS 4 glacial advances in Beringia are most commonly based on morphostratigraphic relationships coupled with a variety of relative age estimate criteria, stream terrace stratigraphy, and most recently, cosmogenic isotopic dating.

Several glacial advances have been documented in Alaska between the middle and Late Pleistocene. Among the most distinctive features of these glaciations are the Salmon Lake moraines of the Kigluaik Mountains of the Seward Peninsula, Alaska (Kaufman and Calkin, 1988) (**Fig. 4**). The moraines of the Itkilik I stage in the Brooks Range are also thought to represent MIS 4 glaciation (Hamilton, 1986). Ice advanced on to St Lawrence Island, north of the Bering Strait, sometime in the late Pleistocene. Stratigraphic study of sedimentary sequences on the island indicate that this glaciation occurred after the last interglacial, and before the late Wisconsin glaciation, so it seems most likely that it occurred in MIS 4 (Heiser and Roush, 2001).

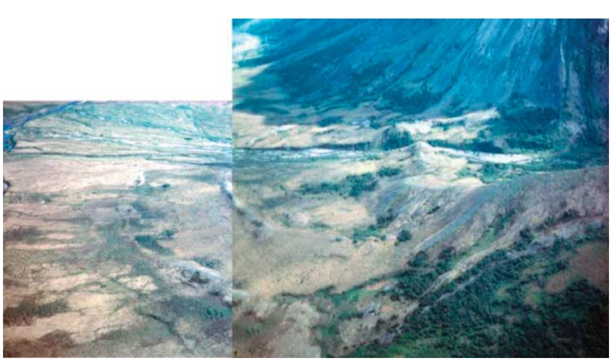

Figure 4 Ridge of the early Wisconsinan Salmon Lake moraine exiting Cobblestone Creek, Seward Peninsula. Photo by Brigham-Grette.

Paleotemperature estimates for MIS 4 in eastern Beringia are indicative of cold summers. TMAX cooling of 6°C below modern levels has been estimated from beetle assemblages in Alaska and the Yukon (Elias, 2001). However, TMIN values were estimated to be near modern levels. This seasonal temperature profile would aid the growth of mountain glaciers, as colder summer temperatures would inhibit summer melting of the snowpack, and winter temperatures at these high latitudes have been sufficiently low to allow glacial ice to form throughout the Pleistocene.

Middle Wisconsin/Karaginsky Interstadial

The Beringian interstadial warming associated with MIS 3 is known as the Karaginsky interstadial in western Beringia, and as the middle Wisconsinan interstadial in eastern Beringia. During this interval glacial margins retreated throughout Beringia. As discussed above, the paleobotanical evidence from western Beringia indicates warmer-than-modern conditions there during this interval, especially in the Arctic. Fossil beetle evidence (Sher *et al.*, 2002) indicates that TMAX was 1–3°C warmer than modern in the Arctic, and that TMIN was lower than modern, pointing towards increased continentality during the Karaginsky interval. This level of interstadial summer warming apparently did not occur in eastern Beringia. The fossil beetle evidence suggests that early MIS 3 environments had TMAX values about 3°C colder than modern and TMIN values about 3°C warmer than modern. This reconstruction fits well with both the summer and winter insolation pattern at the beginning of Stage 3, about 60,000 years BP. Younger (stratigraphically higher) Stage 3 samples from the Yukon yielded TMAX closer to modern values, with TMIN values remaining warmer than present. Eastern Beringian samples dated between 32,400 and 29,600 years BP are indicative of mean summer temperatures that were 1–4°C colder than today.

Paleobotanical evidence from eastern Beringia shows some expansion of coniferous forests in interior regions, but not approaching modern northern treeline in Alaska. The warmest interval of the interstadial documented in the vegetation comes from interior Alaska. The ‘Fox Thermal Event’, dated from 35,000–30,000 ^{14}C yr BP, is associated with expansion of spruce-forest tundra (Anderson and Lozhkin, 2001).

Late Wisconsin/Sartan Glaciation

The last Pleistocene glaciation in western Beringia is called the Sartan glaciation. The lack of significant moisture across Beringia during the LGM prevented the growth of large ice masses. As shown in Figures 5 and 6, the Sartan and late Wisconsinan ice limits were well inside those of some previous glaciations. LGM glaciation in western Beringia was essentially limited to valley and cirque glaciation in some mountain ranges (Brigham-Grette *et al*., 2003). Glacial extent decreased from west to east in western Beringia, probably because of increasing aridity towards the Bering Strait region. The Verkhoyansk Mountain region developed glacial ice that extended downslope almost to the Lena River (Fig. 6) (Zamoruyev, 2004). Radiocarbon dating of wood and organic matter, combined with cosmogenic isotope surface exposure ages, indicate that LGM ice in western Beringia reached its maximum extent between 27,000 and 20,000 cal yr BP (Brigham-Grette *et al*., 2003). A study of fossil diatom assemblages from Bering Sea sediment cores (Brigham-Grette *et al*., 2003) suggests that the Bering Sea experienced sea ice cover for as much as nine months per year during the LGM. The combination of Bering sea ice cover and the 1000-km-long land bridge in the center of Beringia meant that Pacific moisture was far removed from the interior regions of Beringia during the LGM.

The transition from MIS 3 to MIS 2 is imprecisely dated in eastern Beringia, but by about 32,000 cal yr BP, the climate changed from interstadial to glacial, as evidenced by changes in regional floras and faunas. Apparently the glacial advances in eastern Beringian mountain ranges were not synchronous. The late Wisconsin advance began in the mountains of southwestern Yukon by 31,000 cal yr BP (Hamilton and Fulton, 1994), and mountain glaciers expanded down the north flanks of the Alaska Range about 29,000 cal yr BP. The Cordilleran Ice Sheet advanced throughout the coastal regions of the Pacific Northwest about

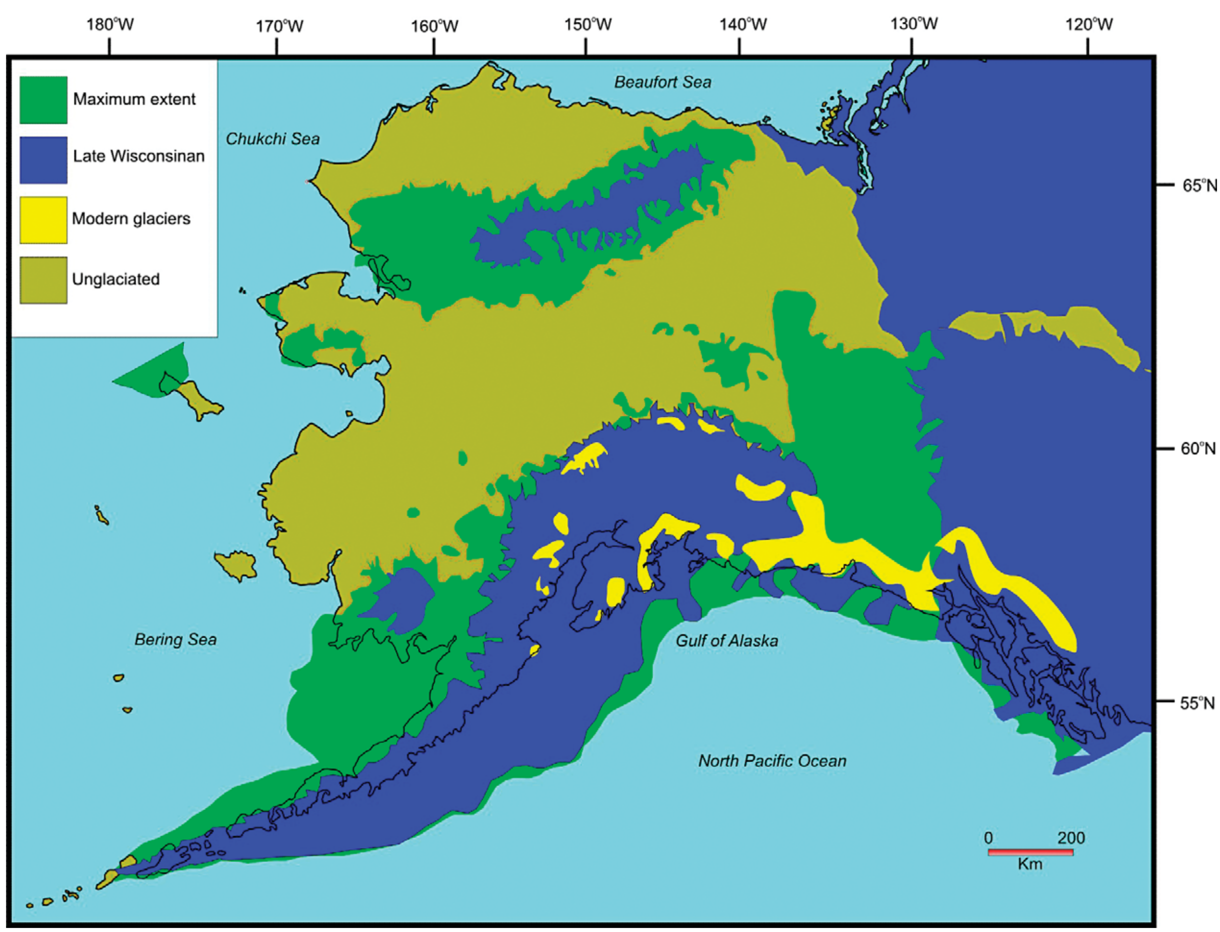

Figure 5 Map of eastern Beringia, showing extent of glaciers during the early and late Wisconsinan glaciations (after Manley WF and Kaufman DS, 2002, Alaska PaleoGlacier Atlas, and Duk-Rodkin, 1999.

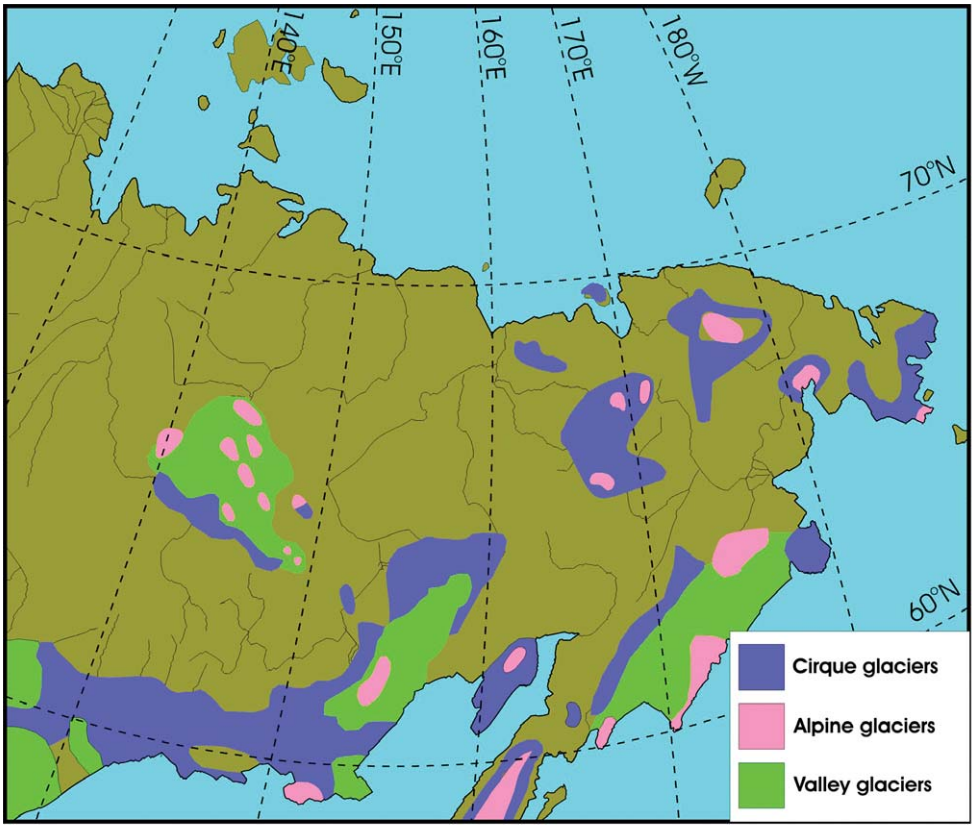

Figure 6 Map of western Beringia, showing extent of glaciers during the Sartan glaciation (after Brigham-Grette *et al.*, 2004).

28,000 cal yr BP. At that time, ice advanced in the Brooks Range, signaling the beginning the Itkilik II glaciation (Hamilton and Ashley, 1993). Ice advances in the highlands of the Seward Peninsula are termed the Mount Osborne glaciation (Fig. 7). In southwestern Alaska, glaciers advanced toward the outer continental shelf after 27,000 cal yr BP. Evidence for stabilization of sand sheets and dunes in the Kobuk Dunes region of northwestern Alaska after 22,000 cal yr BP may correlate with the Port Moody Interstade documented from southwestern British Columbia (Clague *et al.*, 2004).

During the LGM, the Cordilleran Ice Sheet covered southern and central Yukon and parts of southern Alaska (Fig. 5), where it formed ice fields and large valley and piedmont glaciers, flowing out of the St Elias Mountains, the Chugach Range, and the Alaska Range. The ice extended out on to the continental shelf to the south, and into the broad, low country drained by the Yukon River and its tributaries to the north (Mann and Hamilton, 1995). Ice expanded from the coastal mountains of British Columbia into the Alexander Archipelago of southeastern Alaska, reaching the outer continental shelf in some regions (Carrara *et al.*, 2002).

Figure 7 Late Wisconsinan Mt Osborn moraines exiting a small cirque into in the valley of Cobblestone Creek, Seward Peninsula. Photo by Brigham-Grette.

One of the earliest dates for deglaciation following the LGM in eastern Beringia comes from the Hungry Creek site, Yukon Territory. Hughes (1983) suggested that the western limit of Cordilleran ice began retreating here by about 19,000 yr BP. Mann and Hamilton (1995) suggest that ice retreat was time-transgressive in southern Alaska, with northern shelf areas becoming ice free by 16,000 yr BP. Much of southeastern Alaska was deglaciated by about 15,000 cal yr BP (Clague *et al*, 2004). Deglaciation in the Brooks Range began by at least 13,400 cal yr BP (Hamilton, 1983). Postglacial environmental change throughout eastern Beringia brought about wholesale changes in vegetation, the regional extinction of much of the magefauna, and the entrance of *Homo sapiens*.

See also: **Glaciations**: Overview. **Plant Macrofossil Records**: Arctic Eurasia.

References

Adkins, J. F., Boyle, E. A., Keigwin, L., and Cortijo, E. (1997). Variability of the North Atlantic thermohaline circulation during the last interglacial period. *Nature* **390**, 154–156.

Alfimov, A. V., and Berman, D. I. (2001). Beringian Climate during the Late Pleistocene and Holocene. *Quaternary Science Reviews* **20**, 127–134.

Anderson, P. M., and Lozhkin, A. V. (2001). The Stage 3 interstadial complex (Karginskii/Middle Wisconsinan interval) of Beringia: variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* **20**, 93–125.

Brigham-Grette, J., Hopkins, D. M., Ivanov, V. F., Basilyan, A. E., Benson, S. L., Heiser, P. A., and Pushkar, V. S. (2001). Last Interglacial (Isotope Stage 5) glacial and sea-level history of coastal Chukotka Peninsula and St. Lawrence Island, Western Beringia. *Quaternary Science Reviews* **20**, 419–436.

Brigham-Grette, J., Gualtieri, L. M., Glushkova, O. Y., Hamilton, T. D., Mostoller, D., and Kotov, A. (2003). Chlorine-36 and ^{14}C chronology support a limited last glacial maximum across central Chukotka, northeastern Siberia, and no Beringian ice sheet. *Quaternary Research* **59**, 386–398.

Brigham-Grette, J., Lozhkin, A. V., Anderson, P. M., and Glushkova, O. Y. (2004). Paleoenvironmental conditions in Western Beringia before and during the Last Glacial Maximum. In *Entering America. Northeast Asia and Beringia Before the Last Glacial Maximum* (D. B. Madsen, Ed.), pp. 29–61. University of Utah Press, Salt Lake City.

Brigham-Grette, J., and Hopkins, D. M. (1995). Emergent marine record and paleoclimate of the Last Interglaciation along the northwest Alaskan coast. *Quaternary Research* **43**, 159–173.

Brigham-Grette J and Hayden T (2002) *Land/sea correlations: separating climate events at the 5e/d and 5a/4 transitions in Chukotka and across the western Arctic*. Geological Society of America Annual Meeting, Program and Abstracts, 145–5.

Brubaker, L. B., Anderson, P. M., Edwards, M. E., and Lozhkin, A. V. (2005). Beringia as a glacial refugium for boreal trees and shrubs: new perspectives from mapped pollen data. *Journal of Biogeography* **35**, 833–848.

Carrara, P. E., Ager, T. D., Baichtal, J. F., and Van Sistine, D. (2002). *Late Wisconsin glacial limits in southern Southeastern Alaska, as indicated by a new bathymetric map*. Institute of Arctic and Alpine Research, University of Colorado, Boulder. 32nd International Arctic Workshop, Program and Abstracts, 41–43.

Carter L. D., Hamilton T. D., Galloway J. P. (1991) Late Cenozoic History of the Interior Basins of Alaska and the Yukon. *U.S. Geological Survey Circular* 1026.

Clague, J. J., Mathewes, R. W., and Ager, T. D. (2004). Environments of northwestern North America before the Last Glacial Maximum. In *Entering America. Northeast Asia and Beringia Before the Last Glacial Maximum* (D. B. Madsen, Ed.), pp. 63–94. University of Utah Press, Salt Lake City.

Duk-Rodkin A (1999) *Glacial limits map if Yukon*. Geological Survey of Canada, Open File Report 3694.

Elias, S. A. (2001). Mutual Climatic Range reconstructions of seasonal temperatures based on Late Pleistocene fossil beetle assemblages in Eastern Beringia. *Quaternary Science Reviews* **20**, 77–91.

Glushkova, O. Y. (1995). Paleogeography of Late Pleistocene Glaciation of North-Eastern Asia. In *Proceedings of the International Conference of Arctic Margins* (K. V. Simakov and D. K. Thurston, Eds.), p. 350. Russian Academy of Sciences, Far East Branch, Northeast Science Center, Magadan.

Guthrie, R. D. (2001). Origins and causes of the mammoth steppe: a story of cloud cover, woolly mammoth tooth pits, buckles, and inside-out Beringia. *Quaternary Science Reviews* **20**, 549–574.

Hamilton, T. D. (1983). Glaciation of the Brooks Range. In *Glaciation in Alaska* (R. M. Thorson and T. D. Hamilton, Eds.), pp. 35–41. University of Alaska, Fairbanks, Alaska Quaternary Center.

Hamilton, T. D. (1986). Late Cenozoic glaciation of the central Brooks Range. In *Glaciation in Alaska* (T. F. Hamilton, K. M. Reed and R. M. Thorson, Eds.), pp. 9–49. Alaska Geologic Society, Anchorage.

Hamilton, T. D., and Ashley, G. M. (1993). Epiguruk: A late Quaternary environmental record from northwestern Alaska. *Geol. Soc. Am. Bull* **105**, 583–602.

Hamilton, T. D., and Fulton, R. J. (1994). Middle and Late Wisconsin environments of Eastern Beringia. In *Ice-Age Peoples of North America* (R. Bonnichsen, G. C. Frison and K. Turnmire, Eds.). Texas A & M University Press, College Station, Texas.

Harris, S. A. (2005). Thermal history of the Arctic Ocean environs adjacent to North America during the last 3.5 Ma and a possible mechanism for the cause of cold events (major glaciations and permafrost events). *Prog. Phys. Geog.* **29**, 1–19.

Heiser, P. A., and Roush, J. J. (2001). Pleistocene glaciations in Chukotka, Russia: moraine mapping using satellite synthetic aperture radar (SAR) imagery. *Quaternary Science Reviews* **20**, 393–404.

Hopkins, D. M. (1982). Aspects of the paleogeography of Beringia during the Late Pleistocene. In *Paleoecology of Beringia* (D. M. Hopkins, J. V. Matthews, C. R. Schweger and S. B. Young, Eds.), pp. 3–28. Academic Press, New York.

Hughes, O. (1983). Quaternary geology, Yukon Territory and western District of Mackenzie. In *Glaciation in* (R. M. Thorson and T. D. Hamilton, Eds.), pp. 51–56. University of Alaska, Fairbanks, Alaska Quaternary Center.

Kaufman, D. S., and Calkin, P. E. (1988). Morphometric analysis of Pleistocene glacial deposits in the Kigluaik Mountains, northwestern Alaska, USA. *Arctic Alpine Research* **20**, 272–284.

Kaufman, D. S., and Brigham-Grette, J. (1993). Aminostratigraphic correlations and paleotemperature implications, Pliocene-Pleistocene high-sea-level deposits, northwestern Alaska. *Quaternary Science Reviews* **12**, 21–33.

Kaufman, D. S., Forman, S. L., Lea, P. D., and Wobus, C. W. (1996). Age of pre-late-Wisconsin glacial-estuarine sedimentation, Bristol Bay, Alaska. *Quaternary Research* **45**, 59–72.

Lambeck, K., Esat, T. M., and Potter, E.-K. (2002). Links between climate and sea levels for the past three million years. *Nature* **419**, 199–206.

Lozhkin, A. V., and Anderson, P. M. (1995). The last interglaciation in northeast Siberia. *Quaternary Research* **43**, 147–158.

Mangerud, J., Astakhov, V., Jakobsson, M., and Svendsen, J. I. (2001). Huge ice-age lakes in Russia. J. *Quaternary Science* **16**, 773–777.

Manley W. F. and Kaufman D. S. (2002) *Alaska PaleoGlacier Atlas: Institute of Arctic and Alpine Research (INSTAAR)*, University of Colorado, http://instaar.colorado.edu/QGISL/ak_paleoglacier_atlas, v. 1.
Mann, D. H., and Hamilton, T. D. (1995). Late Pleistocene and Holocene paleoenvironments of the north Pacific coast. *Quaternary Science Reviews* **14,** 449–471.
Muhs, D. R., Ager, T. A., and Begét, J. E. (2001). Vegetation and paleoclimate of the Last Interglacial period, central Alaska. *Quaternary Science Reviews* **20,** 41–62.
Pushkar, V. S., Roof, S. R., Cherepanova, M. V., Hopkins, D. M., Ivanov, V. F., and Brigham-Grette, J. (1999). Paleogeographic and paleoclimatic significance of diatoms from Middle Pleistocene marine and glaciomarine deposits on Baldwin Peninsula, northwestern Alaska. *Paleogeography, Paleoclimatology, Paleoecology* **152,** 67–85.
Sher A, Alfimov A, Berman D, and S Kuzmina (2002) *Attempted reconstruction of seasonal temperature for the Late Pleistocene arctic lowlands of north-eastern Siberia based on fossil insects.* Sixth Quaternary Environment of the Eurasian North (QUEEN) Workshop, Programme and Abstracts, 50.
Siegert, M. J., Dowdeswell, J. A., Hald, M., and Svendsen, J. (2001). Modelling the Eurasian Ice Sheet through a Full (Weichselian) Glacial Cycle. *Global Planetary Change* **31,** 367–385.
Zamoruyev, V. (2004). Quaternary glaciation of north-eastern Asia. In *Quaternary Glaciations – Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica* (J. Ehlers and P. L. Gibbard, Eds.), pp. 321–323. Elsevier.

Late Pleistocene of the SW Pacific Region

E Colhoun and J Shulmeister, University of Newcastle, Newcastle, Australia

Australia and New Zealand on either side of the Tasman Sea form the Southwest Pacific region. The region lies in the latitudes of prevailing westerly winds. These moisture-bearing winds are uplifted as they cross highland regions giving abundant rain and snow as the low-pressure systems pass eastwards. During the Quaternary Epoch of the last 2.6 million years, ice accumulated at various times in the mountainous areas. Four areas, each with distinctive characteristics of glaciation, can be identified. The most northerly was the Snowy Mountains of southeast Australia, where at 36° 45'S cirque and short valley glaciers were formed only late in the Quaternary. Further south between 41° 30'S and 43° 30'S, ice formed extensively over the central and western parts of Tasmania as cirque, valley glaciers, icecaps, and small ice sheets at various times during the past million years. Several large volcanoes dominate the highest peaks at 39°S in North Island New Zealand that both today and during the Quaternary sustained cirque, valley, and small icecap glaciers. On South Island, an ice sheet was developed numerous times on the Southern Alps from 42°S in Nelson to 47°S in Southland. Outlet and piedmont glaciers terminated on the plains east and west of the mountains, and in Fjordland the ice sheet extended offshore. This article summarizes current knowledge on the glaciations of the regions (**Fig. 1**).

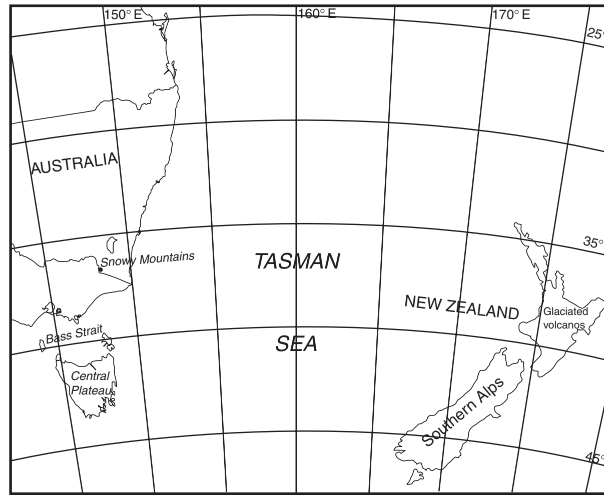

Figure 1 Location map.

Early Quaternary

Evidence for glaciation between 2.6 million years ago and the last reversal of the earth's magnetic field at the Brunhes–Matuyama boundary dating to 780,000 years ago is here regarded as early Quaternary. The first evidence for terrestrial glaciation in Tasmania occurred approximately 1 million years ago. The onset of cold conditions is recorded much earlier in New Zealand, where marine sediments adjacent to North Island and glaciers in Westland and Nelson indicate cold glacial conditions by approximately 2.6 million years ago.

Early Quaternary of Tasmania

The oldest glaciations in Tasmania were the most extensive. They are identified as a maximum limit of Quaternary ice in **Figure 2**. Individual ice advances have not been distinguished and correlated between areas, but at least two occur in the Forth, upper Pieman, and middle Franklin valleys.

An ice sheet occupied the western Central Plateau and Highlands, and an ice cap formed on the West Coast Range. These ice masses were probably connected at times in west central Tasmania. Both contributed to outlet glaciers descending into adjacent northern, western, and southern valleys.