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Geologic Framework and Glaciation of the Western Area

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Geologic Framework and Glaciation of the Western Area

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The geologic framework for western North America consists of physical landscapes (geomorphic features) and stratigraphic sequences that can be used to provide a basis for understanding the chronologic and environmental context for Late Pleistocene human populations. The Western Area includes the region of North America from the Pacific coast to the Rocky Mountains and parts of the Great Basin and Colorado Plateau (figs. 1-2).

The Late Pleistocene geology of western North America has been significantly influenced by global-scale changes in climate. For instance, much of the northern part of the region was covered by glaciers of the Cordilleran Ice Sheet, and glaciers were also present in the southern part of the Western Area, from the Sierra Nevada bordering the west side of the Great Basin to the southern Rocky Mountains (fig. 2). Besides the consequences of the extent and timing of glacial advance and melting for the presence of landscape-habitats available for biotic communities, changes in climate and local geologic processes influenced the relative sea level along the Pacific coast and the availability of terrains. The geologic framework of glaciated western North America provides a means for evaluating the temporal and physical patterns associated with Late Pleistocene human populations, both in terms of the kinds and distribution of habitable landscapes that were present and in terms of geologic formation process that could have influenced the preservation and visibility of paleontological and archeological contexts.

History of Research

The extent, distribution, and timing of glaciation and its effects on sea level and terrestrial landscape contexts is of potential importance for evaluating Late Pleistocene human migration in North America (cf. W.A. Johnston 1933; K. Bryan 1941; H.E. Wright 1991; E.J. Dixon 2001). Two main routes of migration have been proposed: a route along the Pacific coast, and an inland route along the Mackenzie River Valley and the eastern front of the Rocky Mountains. The presence of an ice-free region between the eastern margin of the Cordilleran ice sheet and the Laurentide ice sheet has been considered since the late 1800s (Chamberlin 1894). The possibility that this inland route was a migration route was suggested (W.A. Johnston 1933:22, 42). Years before the advent of radiocarbon dating, Antevs (1935a:302) suggested that people first "spread along on the eastern foot of the Rockies where an ice-free corridor had formed some 20,000 to 15,000

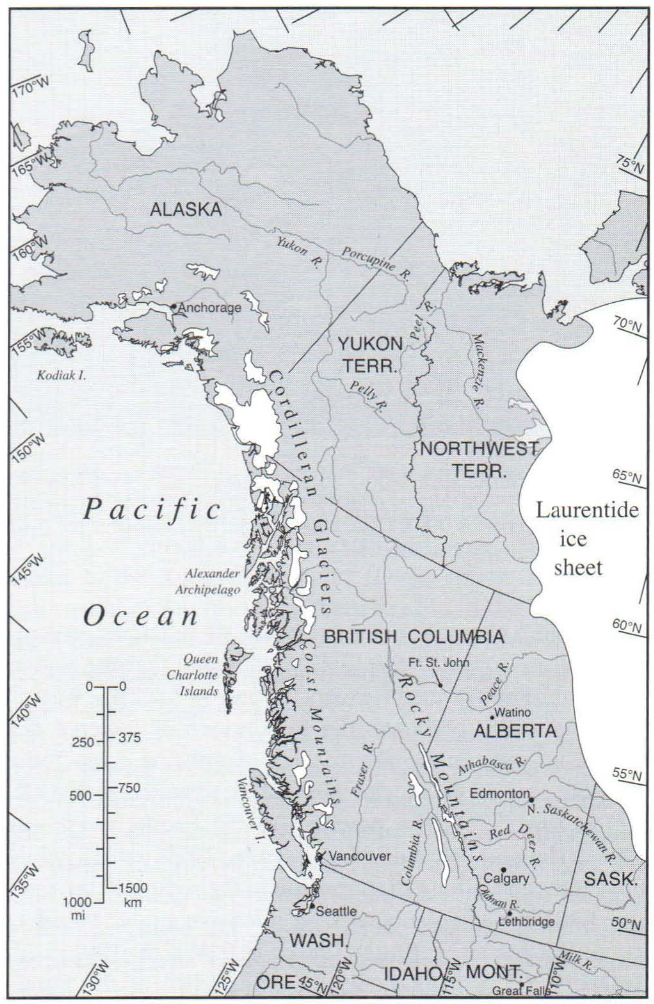

Fig. 1. Potential location of glaciers during the late Middle Wisconsin interstadial, isotope stage 3, around 33,000-30,000 B.C. (corresponding to the end of the Altonian or during the Farmdalian).

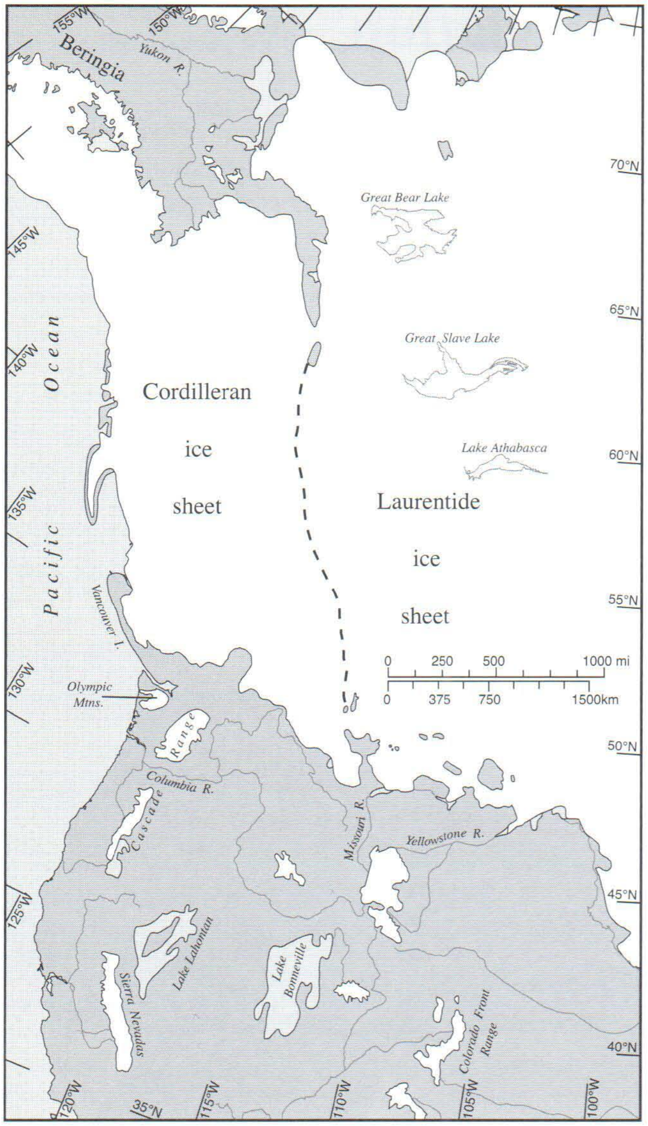

Fig. 2. Glaciated western North America around 19,700 B.C. during part of the Last Glacial Maximum (corresponding to the Woodfordian) (Dyke, Moore, and Roberson 2003; Dyke 2004). Under the Laurentide ice sheet are shown locations of present-day lakes.

years ago." Regarding the viability of a Pacific coastal route, W.A. Johnston (1933:42) concluded that the region could not "be traversed except in part at least by water." The possibility of boats, the importance of lowered sea levels, as well as the alternatives of either a coast route or route east of the Rocky Mountains were suggested (K. Bryan 1941:506). The potential of the Pacific coast was noted (Heusser 1960; Fladmark 1978, 1979, 1983), while the extent and timing of glaciation in the region of the interior route has been subjected to intensive study (Wendorf 1966; Reeves 1973; Rutter 1980a; H.E. Wright 1991; L.E. Jackson and A. Duk-Rodkin 1996). South of the continental ice sheets, alpine-mountain glaciation can be linked to climate change and landscape evolution, and especially the presence of pluvial lakes.

Temporal Framework

The development of a temporal framework that can be applied to western North America is based on the relationships between stratigraphic deposits, geomorphic features, artifactual and paleontological time markers, and chronometric age determinations. Regardless of whether the focus is on the geologic evidence of environmental conditions along the North American Pacific coast, the eastern front of the Rocky Mountains, or the glaciated mountains and lake-filled basin south of the continental ice sheets, it is useful to relate the evidence to the global-scale Pleistocene stages and events, as well as to the ages of Late Pleistocene artifact assemblages. For broad-scale correlations and comparisons relating the regional geologic evidence to the marine isotope record, the Greenland ice cores are of critical value (Gibbard and West 2000; Blunier and Brook 2001; Southon 2002), as is the calibration to calendar years of the chronometric age measurements (Hughen et al. 2004).

Age relationships between the geologic events and artifact assemblages in western North America follow Stanford (1999a) and V.T. Holliday (2000a), while the presence of mammal fossils is used as a first approximation of the potential viability of landscapes for Late Pleistocene human adaptations. The calibration of radiocarbon ages to calendar years was undertaken by applying the University of Cologne radiocarbon calibration program package (based on Hughen et al. 2004); the assignment expressed as years B.C. should be considered as an approximation of the average within the range likely for the respective ages. Using this procedure, the averages from Clovis sites (Aubrey, Anzick, Blackwater Draw, Colby, Dent, Domebo, and Lange-Ferguson) have a range from about 11,400 to 10,800 B.C., and Folsom sites (Hanson, Carter–Kerr-McGee, Hell Gap, Lindenmeier, Folsom, Lipscomb, Blackwater Draw, Lubbock Lake, Bonfire Shelter) range from about 10,900 to 9700 B.C. (V.T. Holliday 2000a).

The last part of the Late Pleistocene is divided into the Middle and Late Wisconsin. Correlations of the Late Wisconsin with oxygen isotope 2 and the Middle Wisconsin with isotope stages 3 and 4 have led to a proposed boundary at about 30,600-28,000 B.C. (R.B. Morrison 1991; Hoffecker and Elias 2003; Clague, Mathewes, and Ager 2004). The Last Glacial Maximum, centered about 19,000-18,500 B.C., correlates with global-scale conditions associated with lowered sea levels, while the Younger Dryas cooling event, dating to about 11,000-9500 B.C., marks the end of the Pleistocene. In the Greenland ice cores and in stratigraphic sequences from northern Europe, the Younger Dryas chronozone marks a short interval between the warmer climates of the Allerod (ca. 12,100-11,000 B.C.) and the beginning of the Holocene with the Pre-Boreal chronozone starting around 9500 B.C. The terms Bull Lake and Pinedale (Blackwelder 1915) are often used to designate the mountain glacial deposits in the Rocky Mountains. Originally applied to the Wind River Range, in Wyoming, Pinedale is now generally correlated with Late Wisconsin and marine isotope stage 2, while Bull Lake deposits are usually considered to be pre-Wisconsin, perhaps correlated with marine isotope stages 6 and 5d (K. Pierce 2004).

Time stratigraphic substages of the Wisconsin stage were related (C.V. Haynes 1964, 1969) to archeological periods with the following correlations: Altonian correlated with the "Early Period"; Farmdalian correlated with early "Middle Period"; Woodfordian correlated with later "Middle Period"; Two Creekan, end of the "Middle Period"; and Valderan, correlated with the "Late Period." The estimated age of the geochronologic and chronstratigraphic units as proposed by W.H. Johnson et al. (1997), based on a conversion to estimate the date expressed as years B.C. (cf. van Andel et al. 2003; Hughen et al. 2004), are: Altonian, prior to about 30,600 B.C.; Farmdalian, about 30,600-28,100 B.C.; and Woodfordian, about 28,100-11,660 B.C. The Two Creeks forest bed is radiocarbon dated to about 11,660 B.C. (Karrow, Dreimanis, and Barnett 2000) and is therefore probably slightly older than landscapes associated with people using Clovis artifacts. The Allerod and Younger Dryas chronozones may correspond to the time range associated with the presence of people using Clovis and Folsom artifacts.

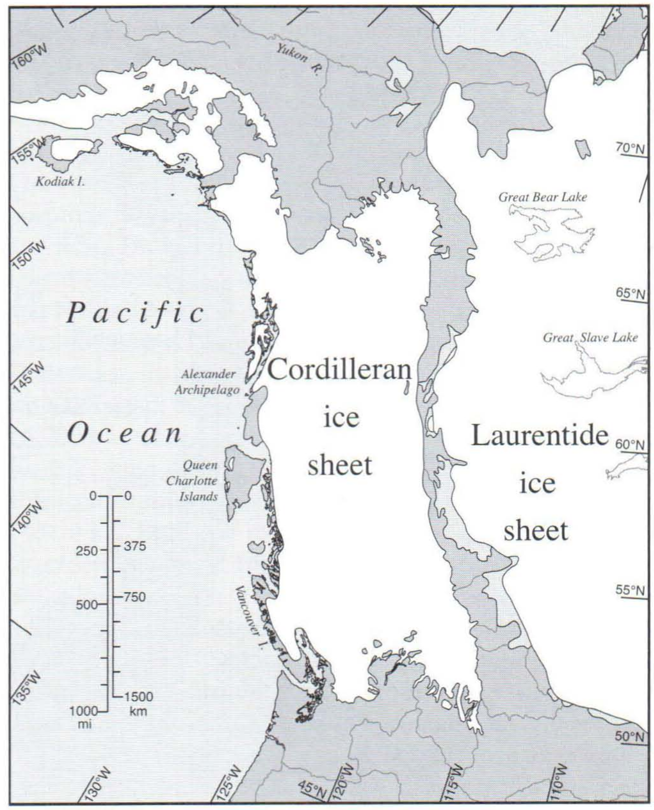

Fig. 3. Approximate extent of the Cordilleran and Laurentide ice sheets at about 12,700 B.C. (Dyke, Moore, and Roberson 2003; Dyke 2004).

Spatial Framework: Distribution of Ice Sheets

Continental and mountain glaciation in western North America was present in areas extending from the western Great Plains and eastern Rocky Mountains westward to the Pacific coast and Alaska. Continental glaciation is associated with the Cordilleran and Laurentide ice sheets (fig. 2). The dynamics of the western margin of the Cordilleran ice are important for understanding the environments associated with the southwestern region of western coastal Beringia southward to Vancouver Island and the Puget Lowland. The eastern margin of the Cordilleran ice and the northwestern margin of the Laurentide ice are important for developing an understanding of the environments associated with the southern part of eastern Beringia and potential migration routes east of the Rocky Mountains.

The geologic framework for the Late Pleistocene of western North America is summarized by reviewing data relevant to the environmental conditions from before the Last Glacial Maximum (Middle Wisconsin) to the end of the Pleistocene (the Younger Dryas–Pre-Boreal boundary). Generalized paleogeographic reconstructions for the extent of glaciation are presented in figures 1-4 (based primarily on Dyke, Moore, and Robertson 2003; Dyke et al. 2002; Dyke 2004).

There appear to have been two ice caps associated with the Cordilleran ice sheet. In the northern region, the ice sheet covered southern Alaska and Yukon Territory, leaving the remaining parts of the Yukon Territory and interior Alaska unglaciated. Ice initially formed in the Coast Range and in the Rocky Mountains. Along the Pacific coast (fig. 5) the Cordilleran ice sheet extended onto the continental shelf (Clague 1989). Along the southern margin, ice extended southward into the Columbia basin, forming a lobe that advanced into the Puget Lowland in the area of present-day Seattle. To the east, ice advanced onto the interior northern Great Plains, at certain places and times interacting with the western margin of the Laurentide ice sheet.

The Laurentide ice sheet expanded across the interior Plains to the Rocky Mountains (Dyke 2004). The timing and spatial extent of the southern margin of the Laurentide ice sheet from the eastern front of the Rocky Mountains to the Great Lakes and its implications for Late Pleistocene human populations are reviewed in "Geological Framework and Glaciation of the Central Area," this volume. Laurentide ice influenced the landscapes associated with the southeast region of Beringia (Yukon Territory and British Columbia) and the Saskatchewan and Missouri plateaus (Alberta and Montana). In terms of potential inhabitable landscapes and migration routes, the evidence for interaction or coalescence of the Laurentide and Cordilleran ice sheets is of particular significance. The character of the Wisconsin Laurentide western margin is considered here in relation to the eastern margin of the Cordilleran ice sheet and the availability of landscapes along the eastern front of the Rocky Mountains for migration and habitation.

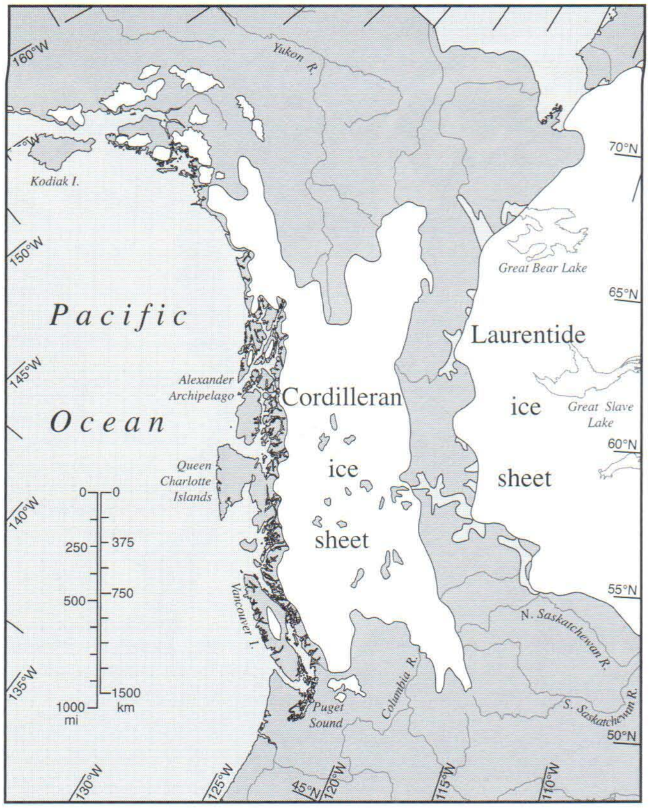

Fig. 4. General location of Cordilleran and Laurentide ice sheets at around 11,300 B.C., contemporaneous with the initial presence of Clovis artifacts (Dyke, Moore, and Roberson 2003; Dyke 2004).

Glaciers were present in the Rocky Mountain (Gillespie and Zehfuss 2004; Locke and Smith 2004; Dahms 2004; K. Pierce 2004) and Great Basin (Osborn and Bevis 2001; Osborn 2004) regions during the Late Pleistocene (fig. 2). There were several advances and retreats that can be related to a variety of sedimentary environments in these regions. The relationship between glacial conditions and evidence for changes in lake levels has implications for the availability of habitable landscapes in western North America.

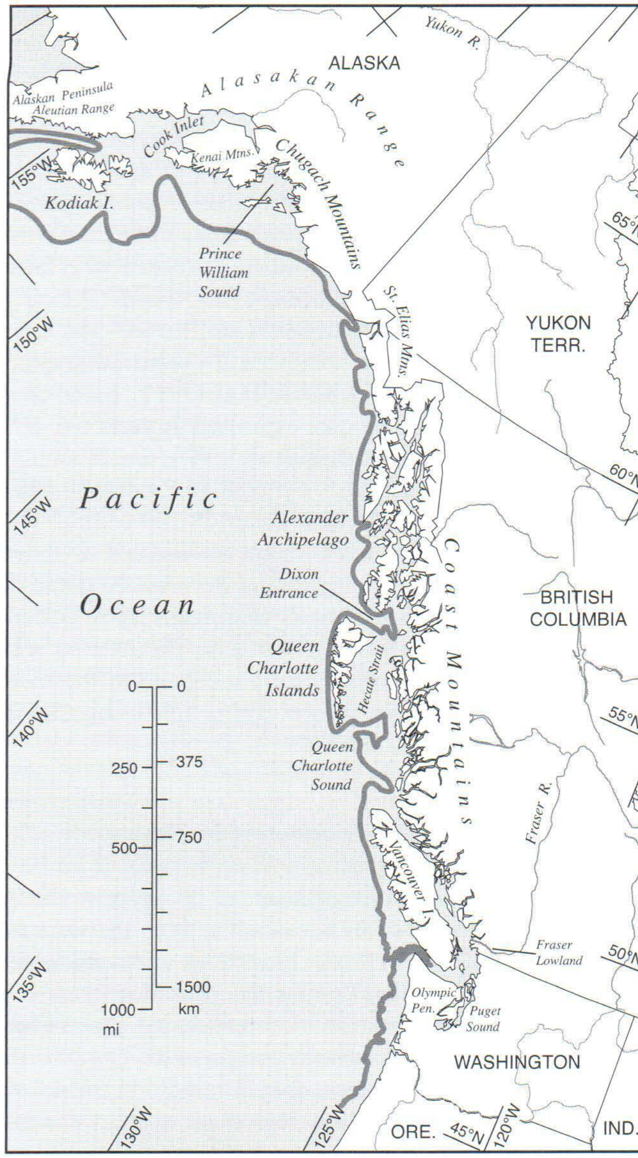

Fig. 5. Present-day Pacific coastline from the Alaska Peninsula to Oregon, showing the approximate extent of the continental shelf (200 m depth indicated by the heavy gray line) potentially exposed during maximum late-glacial sea-level lowering (Fedje and Josenhans 2000).

Physical Environments

Pacific Coast: Coastal Alaska, British Columbia, Puget Lowland

The Late Pleistocene geologic setting along the Pacific coast is important for understanding landscapes associated with a coastal route migration (W.A. Johnston 1933; K. Bryan 1941; Heusser 1960; Fladmark 1978, 1979, 1983; Mandryk et al. 2001; E.J. Dixon 2001; Barrie and Conway 2002, 2002a; Hetherington and Barrie 2004). Environments in the Arctic and Northwest Coast are reviewed by Stager and McSkimming (1984) and Suttles (1990a). Even in the unglaciated regions, Late Pleistocene geologic contexts were influenced by climate change, in particular changing sea levels linked to global-scale fluctuations between glacial and interglacial periods. Environmental conditions in this region of North America were dramatically influenced by changing sea levels (Peltier 2002). Large areas of land would have been available during lowered sea levels (fig. 5).

Glaciers developed on the island mountains and the mainland of southeastern Alaska and British Columbia. Expansion of local ice influenced local relative sea levels and coastal morphology. In some regions the advance of the ice led to depression of the local terrain and a consequent rise in local sea level. In contrast, uplift of the local terrain caused by the forebulge effect along the periphery of advancing glaciers could add to any conditions resulting from eustatic sea level lowering, providing potentially inhabitable landscapes.

The Pacific coast region extends from the Aleutian Range and Kodiak Archipelago southward to the Vancouver Range and the Olympic Peninsula (fig. 5). During the Late Wisconsin glaciation, ice advanced westward onto the Pacific coastal and island areas. Many of the landscapes were inundated by rising sea levels. Documentation of the sea level changes along the Pacific coast relies on studies of specific local geologic contexts and glacial chronologies. The sea level curves can be used to develop paleogeographic reconstructions of the continental shelf and paleo-shorelines (Muhs et al. 2004). Potential stratigraphic and geomorphic settings that might contain Late Pleistocene archeological and paleontological localities can be determined using sea level data to evaluate coastal paleogeographic contexts (Mathewes 1979). Besides geologic evidence for the advance and retreat of Cordilleran ice, landscape conditions were constrained by local tectonics (depression and uplift) and sea level response. Paleontological and archeological evidence has also been used to infer the presence and character of Late Pleistocene physical environments along the Pacific coast.

The southern region of Alaska was covered by the Cordilleran ice sheet (figs. 1-4). Cordilleran ice formed in two regions separated by the Cook Inlet. In the southwest, glaciers developed from the Aleutian Range to the Alaska Range, while in the southeast, glaciers extended from Kodiak Island to the Kenai, Chugach, and Saint Elias mountains and the Alexander Archipelago. The timing of the youngest major glacial events in Alaska can be correlated broadly with oxygen isotope stage 2 (corresponding generally to the Woodfordian); they occurred generally between about 27,000 and 11,000 B.C. (Kaufman, Porter, and Gillespie 2004; Kaufman and Manley 2004). The margins of the Cordilleran ice sheet in southern Alaska are linked to areas of the continental shelf that are presently below sea level and proglacial lakes. The limits and flow patterns in these areas are sometimes poorly defined due both to this submergence and to a complicated history linked to eustatic sea level changes, isostatic crustal depression, and late Quaternary tectonics.

Geologic processes have influenced the landscape evolution and coastal ecology along the Aleutian Islands and Alaska Peninsula (J.W. Jordan and H. Maschner 2000). Ice extended onto the continental shelf from the Aleutian Range, forming the western end of the Cordilleran ice sheet. During the Last Glacial Maximum (within the Woodfordian), glacial ice extended from the Aleutian Islands to the Gulf of Alaska. The maximum limits of the Late Wisconsin glaciers along the Alaskan Peninsula (the area from Uniak Island to the Cook Inlet) were probably reached after 26,000 B.C., perhaps around 24,600-19,700 B.C. (Mann and Peteet 1994; Stilwell and Kaufman 1996; Kaufman and Manley 2004). There were localized glacial readvances around 11,800 and 9500 B.C., some of which may correspond to the Younger Dryas. Based on radiocarbon dating of plant macrofossils the coastal lowlands were ice-free prior to about 11,300 B.C. (J.W. Jordan 2001).

The glaciated region extended beyond Kodiak Island to the outer continental shelf (Mann and Hamilton 1995). The Late Wisconsin advance in the Kodiak Island region occurred between 28,000 and 14,000 B.C. (T.D. Hamilton and R.M. Thorson 1983). Some areas of southwest Kodiak Island were not glaciated (Mann and Peteet 1994; Kaufman and Manley 2004). Even during the glacial maximum there were ice-free regions that contained proglacial lakes and landscapes that may have been biotic refugia. West of Kodiak Island, the continental shelf was deglaciated before 16,000 B.C. There were several smaller ice advances of the Kodiak Island glaciers at 14,500 and 11,700 B.C., but inhabitable landscapes with terrestrial vegetation were present after 14,000 B.C. Botanical evidence indicates a possibly cooler, dryer setting around 10,800-9500 B.C. (the local equivalent of the Younger Dryas) (Mann and Hamilton 1995).

Coastal south-central Alaska extends westward from the Cook Inlet to the Kenai and Chugach ranges (fig. 5). In south-central Alaska, glacial ice expanded from the Aleutian and Alaskan ranges and the Kenai Mountains, covering much of Cook Inlet (in the vicinity of Anchorage) and parts of the Gulf of Alaska. The Middle Wisconsin Eklutna advance, which ended around 49,000 B.C., was followed by a nonglacial interval and then the less pronounced Late Wisconsin advances (T.D. Hamilton and R.M. Thorson 1983). Glaciers were present on the Kenai Peninsula and Cook Inlet during the Last Glacial Maximum although some areas in the Cook Inlet and Kenai region remained ice free during the Last Glacial Maximum (Kaufman and Manley 2004). The dating for the maximum advance and beginning of deglaciation is somewhat problematic but may have been from 27,000/24,600-18,500 B.C. (Mann and Hamilton 1995). By about 22,000-19,700 B.C. isostatic depression had resulted in a marine transgression, even at a time of lower eustatic (global) seal levels (Schmoll, Yehle, and Updike 1999). Flooding continued as glaciers retreated. For example, in the Cook Inlet and lower drainage of the Susitna River areas, deglaciation had begun around 17,300-16,200 B.C. By around 17,900-17,300 B.C. deglaciation resulted in marine waters within the inlet, and waters reached present-day Anchorage by 16,100 B.C. Ice-free, vegetated landscapes were present on the Kenai Peninsula by 15,400 B.C. (Mann and Hamilton 1995). There was a readvance prior to about 14,700 B.C. and smaller glacial advances in the region continuing until about 11,500 B.C., but by 11,800 B.C. glaciers had retreated from the Anchorage lowland (T.D. Hamilton and R.M. Thorson 1983; Schmoll, Yehle, and Updike 1999). Ice had melted out of Prince William Sound by 15,400 B.C., although some coastal glaciers may have persisted until about 9500 B.C. (Mann and Hamilton 1995). The present coastline of the Gulf of Alaska was emergent and essentially ice free sometime after 10,400 B.C. (T.D. Hamilton and R.M. Thorson 1983).

The southeastern Alaska coastal region includes the Saint Elias Mountains and the Alexander Archipelago (fig. 5). In the Saint Elias Mountains, the Boutellier nonglacial interval dates to about 48,000 B.C. to 32,300 B.C. (Denton 1974). The Alexander Archipelago includes Heceta Island and Prince of Wales Island and is separated from the Queen Charlotte Islands by the Dixon Entrance. In the Alexander Archipelago, coastlines appear to have been unglaciated during much of the Wisconsin (E.J. Dixon et al. 1997; E.J. Dixon 2001). Glaciation was limited to alpine glaciers and a few valley glaciers along the coast that may have reached the outer continental shelf, and large areas were ice-free (Clague, Mathewes, and Ager 2004). Vertebrate remains indicate that the region was inhabitable before, during, and after the Last Glacial Maximum. Middle Wisconsin fossils of terrestrial fauna with ages ranging from about 43,000 to 26,000 B.C. include bear, mastodon, and caribou (Heaton and Grady 2003; Fedje et al. 2004). Ages on seal bones ranging from about 27,100 to 14,800 B.C. indicate ice-free conditions around On Your Knees Cave during and after the Last Glacial Maximum (Heaton and Grady 2003). Other proxy indicators of inhabitable landscapes are radiocarbon ages on bears and caribou ranging from about 12,420 -10,500 B.C. Prince of Wales Island also contains evidence of human presence slightly after 9550 B.C.; radiocarbon measurements corrected for the local marine and atmospheric carbon reservoirs have ages of about 9300 B.C. (Heaton, Talbot, and Shield 1996; E.J. Dixon et al. 1997; E.J. Dixon 2001; Fedje et al. 2004). Parts of the continental shelf off southeastern Alaska were deglaciated by 11,800 B.C. (T.D. Hamilton 1986).

A glacier extended from the British Columbia mainland separating the Prince of Wales Island area to the north from the Queen Charlotte Islands in the south (Barrie and Conway 1999, 2002, 2002a). Melting of this lobe of the Cordilleran ice sheet, after the maximum advance around 23,200 B.C., began by about 17,300-16,200 B.C., and glacial ice had disappeared from the area by 14,700-14,000 B.C. (Barrie and Conway 1999). Isostatic rebound after deglaciation led to a fall in relative sea level (a local sea level regression) that lasted until about 12,500 B.C. This resulted in the presence of an extensive coastal plain in the central Dixon Entrance until flooding began during a transgression starting around 12,500 B.C.

The Queen Charlotte Island region (fig. 5) has been studied in terms of its glacial chronology, sea level dynamics, and paleoecology. Forested conditions may have prevailed on the Queen Charlotte Islands during the Middle Wisconsin nonglacial interval (Warner, Mathewes, and Clague 1982; Clague, Mathewes, and Ager 2004). Large terrestrial mammals such as caribou have been documented for the Middle Wisconsin (Fedje et al. 2004). The maximum Late Wisconsin Cordilleran advance covered the north and east sides of the Queen Charlotte Islands around 25,900-23,200 B.C. (Lacourse, Mathewes, and Fedje 2003). It extended across the northern Hecate Strait and the adjacent Dixon Entrance (Hetherington et al. 2004a). Ice also extended southward into Queen Charlotte Sound and continued to the edge of the continental shelf. While the mainland subsided beneath ice, the areas surrounding the glaciers were raised. This forebulge uplift exposed large areas of the continental shelf. Because of the forebulge effects of isostatic uplift from the Cordilleran ice sheet, there appears to have been an ice-free coastal plain land connection between the islands and the mainland (Barrie et al. 1993; Josenhans et al. 1997; Fedje and Josenhans 2000).

The glacial retreat started around 17,600 B.C., and by 14,000 B.C. coastal landscapes were ice-free to the British Columbia mainland (Mann and Hamilton 1995; Barrie and Conway 2002a; Lacours, Mathewes, and Fedje 2003). During deglaciation, lowered relative sea levels exposed parts of the continental shelf, leaving a broad, low-relief plain. Inhabitable, ice-free landscapes were possibly present east of the Queen Charlotte Islands by about 15,200 B.C., although ice existed on the British Columbia mainland until about 9500 B.C. (Clague and James 2002; Hetherington et al. 2003; Hetherington et al. 2004, 2004a). By 11,100 B.C. local isostatic uplift exceeded rising eustatic sea level. This led to the development of a land bridge between the British Columbia mainland and the Queen Charlotte Islands. The width of the exposed coastal plain, around 150 kilometers, would have been available for biotic populations.

Relative sea level reached a maximum lowering of about 150 meters (Barrie and Conway 2002, 2002a). The continental shelf adjacent to the Queen Charlotte Islands contains evidence for vegetated terrestrial landscapes between 12,240 and 10,300 B.C. before rising sea levels and flooding (Fedje and Josenhans 2000; Lacourse, Mathewes, and Fedje 2003). Bear remains have been dated to about 15,700 and 8650 B.C. (Fedje et al. 2004; C. Ramsey et al. 2004). The presence of bear suggests a local Last Glacial Maximum biological refugium, or area of inhabitable conditions, soon after deglaciation, which has implications for the availability of environments that may have been viable for Late Pleistocene human populations.

The coastlines of the Queen Charlotte Island region changed in response to eustatic sea level and local uplift and depression. Even with rising eustatic sea levels after the Last Glacial Maximum, uplift resulted in a northern coastal plain in the Hecate Strait from about 16,100 to 8300 B.C., while a southern coastal plain persisted to after 6700 B.C. (Hetherington et al. 2004a). Despite rising eustatic sea levels, isostatic uplift resulted in a land bridge between the British Columbia mainland and the Queen Charlotte Islands. Local tectonic movements were a significant influence on sea-level change in the vicinity of the Queen Charlotte Islands at the end of the Pleistocene (Heatherington and Barrie 2004). Along the mainland, for example, relative sea level rose until about 9000 B.C. and then fell to about present-day levels by 7900 B.C. In contrast, other localities in the region experienced low sea levels prior to rises ranging from about 11,000 to 6800 B.C. (Hetherington and Barrie 2004).

The lowering of sea level in the Queen Charlotte Islands region has implications for landscape availability during the Late Pleistocene. Lowered sea levels before 12,240 B.C. might supply a context for interpreting artifacts found on the continental shelf southeast of the Queen Charlotte Islands at 153 meters below present sea level (Josenhans et al. 1995, 1997; Fedje and Christiansen 1999; Fedje and Josenhans 2000). The shelf was exposed and available for potential habitation during the lowered sea levels between 14,700 and 8900 B.C. (Josenhans et al. 1997). Archeological evidence has also been found on the southeastern side of the Queen Charlotte Islands in a drowned delta flood plain presently 53 meters below sea level (Fedje and Josenhans 2000). Between the Queen Charlotte Islands and Vancouver Island during the Last Glacial Maximum, glaciers from the Coast Mountains advanced to the outer edge of the continental shelf (Mann and Hamilton 1995).

There were several advances of the Cordilleran ice sheet in the Vancouver Island region (Clague and James 2002; Barrie and Conway 2002, 2002a; Heatherington and Barrie 2004). The Middle Wisconsin (designated the Olympia nonglacial interval) dates from before 42,300 B.C. to around 28,600 B.C. (J.E. Armstrong and J.J. Clague 1977; Stumpf, Broster, and Levson 2000). The early part of the Late Wisconsin Fraser Glaciation is represented by till as well as outwash deposited along the front of the glaciers. The Late Wisconsin glaciation had been initiated by 22,000 B.C. and had reached its maximum by 16,000 B.C. (Clague and James 2002; Lacourse 2005). Ice from the British Columbia mainland expanded across northern Vancouver Island and coalesced with ice in the Queen Charloote Strait. Deglaciation had begun before 14,900 B.C. Local sea level was lowered caused by isostatic rebound, exposing large areas of the continental shelf despite rising eustatic sea levels from about 11,500 to 11,000 B.C. (Lacourse, Mathewes, and Fedje 2003; Lacourse 2005).

Paleontological records can be used as proxy indicators for settings that could have been inhabitable on Vancouver Island during the Late Pleistocene. Fossil pollen records from southern Vancouver Island suggest coniferous forests were present during the Middle Wisconsin (Clague, Mathewes, and Ager 2004). During the Wisconsin, mammoths, mastodon, musk-oxen, and bison were present (Harington 1975). For example, mammoth remains dated to about 18,500 B.C. in outwash of the Fraser ice advance indicate viable habitats on southeastern Vancouver Island, while outwash in the Fraser Lowland region to the east of Vancouver Island contains fossils with ages ranging from 25,500 to 23,700 B.C. (Hicock, Hobson, and Armstrong 1982). Faunal remains from Port Eliza Cave on Vancouver Island with ages of about 19,723 to 17,600 B.C. may reflect an open parkland setting before glaciation (B.C. Ward et al. 2003).

Outwash from the Fraser Glaciation contains pine and spruce dated from about 31,600 to 27,200 B.C. (J.E. Armstrong and J.J. Clague 1977; Clague, Mathewes, and Ager 2004). By about 12,000 B.C. open pine woodlands characterized the landscape on northern Vancouver Island. The maximum for spruce in the woodlands dates to around 9800 B.C. The Younger Dryas chronozone appears to be recorded in the vegetational record indicating cool and moist terrestrial conditions (K.J. Brown and R.J. Hebda 2002; Lacourse 2005). Faunal materials include mountain goat with an age of about 11,800 B.C. and bison with an age of around 11,600 B.C., present prior to Younger Dryas when there may have been a transition from open parkland to coniferous forest (Nagorsen and Keddie 2000). These data imply that in this region of the Pacific coast, the glacial advance was relatively brief, starting sometime after 17,654 B.C. (B.C. Ward et al. 2003).

In the Vancouver Island region the early Fraser-age till and outwash are overlain by till of the Vashon Stade. The Vashon Stade was a more extensive advance in the Puget Sound area and reached its maximum after 15,800 B.C. (Porter and Swanson 1998; Porter 2004). Glaciers reached to the Pacific along western Vancouver Island during the Vashon Stade. The subsequent Sumas advance around 11,400 to 11,300 B.C. was followed by deglaciation around 9500 B.C. (Klovanen 2002). Coastal settings between Vancouver Island and the mainland were ice-free by 11,000 B.C. On Vancouver Island, for example, sea levels were lowered to about present-day levels slightly more than a thousand years after deglaciation, around 11,300 B.C.

Near Vancouver, at Lynn Creek, Middle Wisconsin plant communities indicate both tundralike and coniferous forest landscapes (J.E. Armstrong, J.J. Clague, and R.J. Hebda 1985). Glacial ice advanced into the coastal areas around Vancouver about 27,000 B.C. and was followed by interstadial conditions from about 21,000 to 19,700 B.C. when subalpine forest conditions prevailed (Lian et al. 2001; Clague, Mathewes, and Ager 2004). The early part of the Fraser glaciation in this region of southern British Columbia dates to about 19,700 B.C., and the Puget lobe had advanced to the Seattle area by about 15,800 B.C. (Porter 2004; Clauge, Mathewes, and Ager 2004).

Evidence for the Olympia Interglaciation dating from about 30,600-24,600 B.C. consists of nonglacial fluvial (floodplain) and lacustrine deposits in the Puget Lowland (Easterbrook 1986; Easterbrook et al. 2003). The Vashon advance reached the Seattle area by about 15,800 B.C. (Porter and Swanson 1998; Porter 2004). In the Fraser area there were several advances around 11,700 B.C. and 11,100 B.C. prior to the Younger Dryas. These were followed by smaller Younger Dryas advances (Clague et al. 1997; Friele and Clague 2002, 2002a; Fulton, Ryder, and Tsang 2004).

In the Olympic Mountains of Washington, glacial retreat from the Last Glacial Maximum position had begun by about 25,600 B.C., with several minor readvances (Thackray 2001; Kovenanen and Easterbrook 2001, 2002). Alpine glaciers in the Cascade Range had begun to melt away by the time the Puget lobe of the Cordilleran ice sheet reached its maximum, during the Vashon Stade. Deglaciation of the Olympic Mountains and the Puget Lowlands led to inhabitable landscapes. For example, as the late Wisconsin Cordilleran ice receded from the Olympic Peninsula the regional landscape became inhabitable by large mammals. Mastodon, bison, and caribou bones recovered at the Manis site, Washington, are found in deposits above till and below the Mazama tephra (Gustafson, Daugherty, and Gilbow 1979). The oldest dates for the deposits overlying the till are 12,200-11,000 B.C. and are associated with fossil pollen and plant macrofossils indicating the presence of a herb-and-shrub dominated landscape adjacent to regional coniferous forests (Petersen, Mehringer, and Gustafson 1983). Some of the readvances associated with the Sumas Stade of the Cordilleran Ice Sheet in the Fraser Lowland of British Columbia and Washington can be correlated with the Younger Dryas (Kovenen and Easterbrook 2001, 2002; Kovenen 2002).

Cordilleran Southern Margin

The southern margins of the Cordilleran ice sheet extended from the Cascade Range, Washington, to Glacier National Park, Montana (Porter, Pierce, and Hamilton 1983; Richmond 1986; D.B. Booth et al. 2004; Fulton, Ryder, and Tsang 2004; Porter 2004; Easterbrook et al. 2003). The environmental context of this region is reviewed by Chatters (1998). There was a set of major lobes or sublobes along this southern margin (fig. 6). The interpretation of the stratigraphic record of some aspects of these lobes is complicated by the presence of outwash that laterally grades into proglacial lakes and by the repeated erosion caused by the Missoula floods (Waitt and Thorson 1983; Benito and O'Connor 2003).

In north-central Washington, the advance of the Okanagon Lobe blocked the Columbia River and formed glacial Lake Columbia about 16,700-14,000 B.C., during the same time as the Missoula floods (Atwater 1987a; Easterbrook 1986; Easterbrook et al. 2003). The timing of the melting of the ice in the Okanogan valley is constrained by overlying Missoula deposits and the presence of Glacier Peak. Glacial Lake Columbia expanded northward as the Cordilleran ice retreated and was succeeded by glacial Lake Brewster.

The extent of the Columbia River lobe is marked by an end moraine south of the Columbia and Spokane rivers (Waitt and Thorson 1983; D.B. Booth 1987). The ice advance of the Columbia or Colville lobes formed Glacial Lake Spokane. In Idaho, ice of the Purcell Trench-Lake Pend Orielle lobe dammed the Clark Fork River along the northern edge of the Bitteroot Mountains during the Fraser Glaciation. The damming of the upper Clark Fork valley by the Purcell Trench lobe formed glacial Lake Missoula. Melting of the Purcell Trench-Lake Pend Orielle lobe led to the formation of Lake Clark (D.B. Booth et al. 2004). Continued melting of this margin of the Cordilleran Ice Sheet led to the expansion northward of Lake Clark and its coalescence with Lake Kootenay (Waitt and Thorson 1987).

There are distinct moraines that may be associated with the Fraser-Flathead lobe in western Montana (D.B. Booth 1987). The maximum advance is marked by the outer or Mission moraine that appears to have been formed prior to

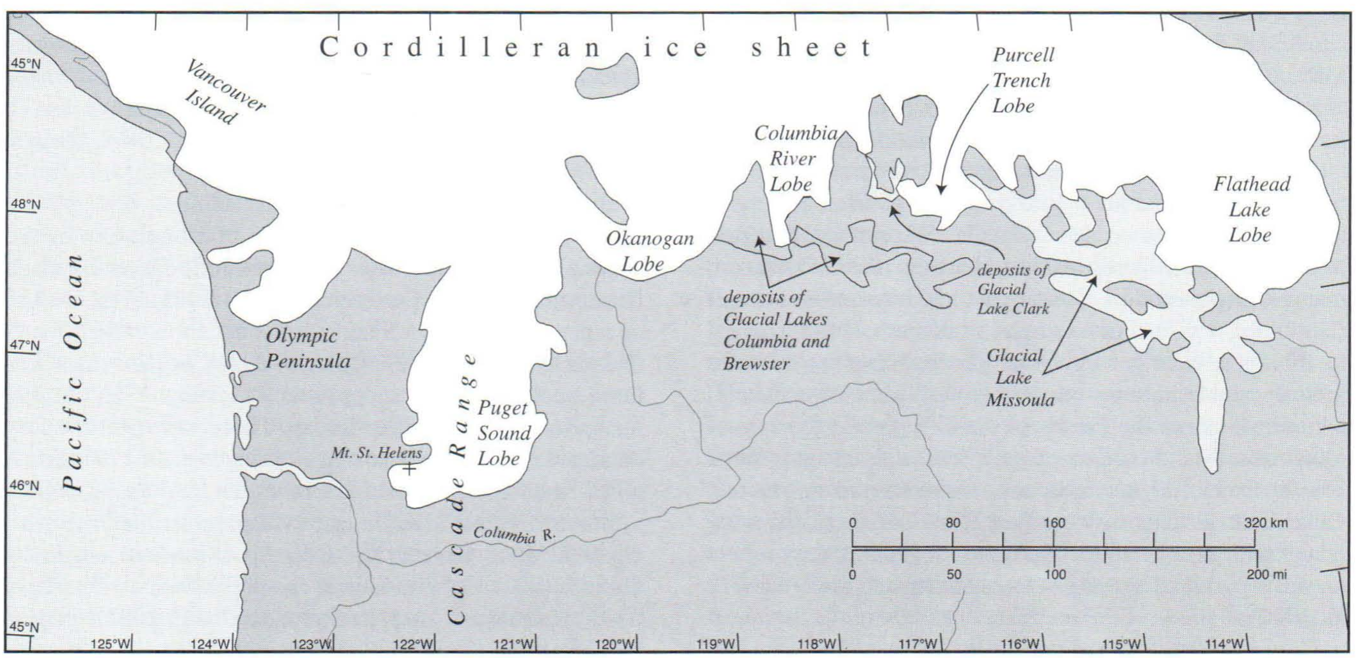

Fig. 6. Approximate extent of the southern margin of the Cordilleran ice sheets (Waitt and Thorson 1983; Dyke, Moore, and Robertson 2003; Pierce 2004).

glacial Lake Missoula. Outwash of the Polson and Kallispell moraines interfingers with or is overlain by lake sediments indicating they were contemporaneous with the later stages of Lake Missoula. Based on the presence of tephra attributed to Glacier Peak and Mount Saint Helens, the Flathead lobe had retreated north of Kalispell by 14,000-12,700 B.C. (Konizeski, Brietkrietz, and McMurtrey 1968; Waitt and Thorson 1983). The Mount Saint Helens set S tephra, dated at about 14,000 B.C., is used to constrain the age of Late Wisconsin Lake Missoula flood deposits (D.B. Booth et al. 2004). The Glacier Peak volcanic ashes B and G (circa 11,100 B.C.) appear to have been deposited after the last catastrophic floods (Carra and Trimble 1992).

Eastern Cordilleran and Western Laurentide Margins (Interior Route)

The eastern Cordilleran region extends from the Old Crow basin and Richardson Mountains in the north, to the eastern front of the Rocky Mountains and the Missouri and Saskatchewan river drainages in the south (cf. Gardner 1981). The eastern boundary is the set of geomorphic features and sedimentary deposits resulting from the melting of the west margin of the Laurentide Ice Sheet ("Geological Framework and Glaciation of the Central Area," this vol.). The location and age of the ice margins and the question of whether landscapes were present that could support large terrestrial vertebrates can be addressed by evaluating the timing and spatial patterns of mountain and continental ice advances (figs. 1-4). Geomorphic and stratigraphic evidence of nonglacial sedimentary contexts, proglacial lakes, glacial moraines, and associated paleontological and archeological materials provide the basis for interpretations.

Some areas of interior eastern Alaska and Yukon during the Middle Wisconsin may have had lowland boreal forests while uplands may have consisted of a shrub-herb tundra (Clague, Mathewes, and Ager 2004). Eastern Beringia, or the unglaciated Yukon refugium, was adjacent to the Laurentide ice sheet. Mountains surround the unglaciated interior basins of the northern Yukon. Glacial limits and the chronology for the northwest margin of the Laurentide ice sheet indicate a major advance before the global Last Glacial Maximum, possibly as early as 33,400 B.C. (Lemmen, Duk-Rodkin, and Bednarski 1994; Duk-Rodkin 1999). In the Mackenzie Mountains, alpine ice advanced down the major valleys, and Laurentide ice extended along the front (Duk-Rodkin and Hughes 1991; Duk-Rodkin et al. 1996).

Continental and montane glacial advances were not synchronous in the Richardson and Mackenzie mountains (Duk-Rodkin and Lemmen 2000; Duk-Rodkin et al. 2004; Zazula et al. 2004). By the time the mountain glaciers had reached their maximum limit the Mackenzie lobe of the Laurentide ice sheet was already receding. However, at a few localities the glaciers from the two sources coalesced. The mountain glaciers melted rapidly, while there was a relatively long succession of ice margins associated with the melting of the Laurentide ice eastward.

The Laurentide ice sheet covered the Mackenzie Plain and the eastern and northern slopes of the Richardson and Mackenzie mountains. Laurentide ice advanced to the eastern side of the mountains. The Wisconsin-age Laurentide glacier was the only continental ice to reach the Richardson and Mackenzie mountains. The maximum advance of the Laurentide ice sheet dammed the Peel River and Porcupine River drainages before 28,000 B.C., leading to a lake in the Old Crow basin (Lemmen, Duk-Rodkin, and Bednarski 1995; Zazula et al. 2004; Duk-Rodkin et al. 2004). The lake was blocked by the margin of a lobe of the Laurentide ice at McDougall Pass in the Richardson Mountains (Lemmen, Duk-Rodkin, and Bednarski 1994).

There was a retreat from of the Laurentide ice, and then a readvance designated the Katherine Creek phase, around 23,600 B.C. (Lemmen, Duk-Rodkin, and Bednarski 1995; Duk-Rodkin et al. 2004). The Katherine Creek phase advance of the Laurentide ice is linked to a second stage of glacial Lake Old Crow (Zazula et al. 2004). The Laurentide glacial deposits are truncated by the montane deposits of the Gayna River Glaciation. During the Gayna River Glaciation, glaciers from the Mackenzie Mountains merged with the Laurentide ice (Duk-Rodkin et al. 2004). In some areas montane glaciers overrode stagnant continental ice while in other places active ice from the mountain and continental sources merged.

The moraine formed during the Tutsieta Lake phase marks the Laurentide ice margin around 13,800 B.C. (O.L. Hughes 1987; Duk-Rodkin and Hughes 1995; Duk-Rodkin and Lemmen 2000). Glacial Lake Travaillant was formed as the glaciers receded around 12,700 B.C., and glacial Lakes Ontaratue and Mackenzie formed by 11,600 B.C. (D.G. Smith 1992; Duk-Rodkin et al. 2004). Glacial Lake McConnell, radiocarbon dated to about 9500 B.C. in the present Great Bear Lake Basin, contributed meltwater to Lake Mackenzie (Lemmen, Duk-Rodkin, and Bednarski 1994). There is evidence of a small readvance associated with the Younger Dryas of the MacKenzie lobe (Dyke 2004).

In terms of relating the glacial chronology to the potential availability of an ice-free migration route, the glacial advance dated to before 23,500 B.C., and the Katherine Creek phase dated to around 24,000 B.C. may have been a barrier to Late Pleistocene human presence and migration in this region. The northern region of the interior migration route appears to have been covered by the Laurentide Ice Sheet possibly before the global Last Glacial Maximum, based on studies in the Mackenzie and Richardson mountains. The end of the Katherine Creek phase of the Laurentide ice sheet coincides with a Cordilleran advance and coalescence of the ice sheets.

The paleogeographic maps based on Dyke, Moore, and Robertson (2003) and Dyke (2004) provide an initial basis for interpreting the ice margins and lake shorelines from the time of maximum glacial advance in the southern region (figs. 2-4). Paleontological remains and archeological sites provide some indication of when landscapes were inhabitable as glaciers melted from the area (Beaudoin and Oetelaar 2003). The Rocky Mountains of Alberta and Montana served as the source for mountain and Cordilleran ice in the southern region.

A Middle Wisconsin nonglacial (interstadial) on the Interior Plains of Canada dates from at least 44,000 to 24,600 years B.C. (Klassen 1989). Within the Peace River drainage at the Watino locality, northwest of Edmonton, spruce and willow dated to 44,500 to 30,400 B.C. imply that the Middle Wisconsin in this region of the Rocky Mountains may have been part of an open coniferous landscape (Catto et al. 1996; Clague, Mathewes, and Ager 2004). Coalesence of the mountain and continental ice in the Peace River region occurred after 28,700 B.C. (Levison and Rutter 1996). Mammoth remains north of Peace River at High Level and south of Peace River at Beaver Lodge suggest the presence of inhabitable landscapes from 25,300 to 24,600 B.C. In the Athabasca River valley the nonglacial, represented by an age of 32,100 B.C., was followed by the onset of glaciation (Levison and Rutter 1996). In the vicinity of the North Saskatchewan River, mammoth, horse, and spruce radiocarbon ages range from 43,500 to 23,600 B.C. (J.A. Burns 1996). Farther south, Middle Wisconsin fauna from January Cave in the front range of the Alberta Rocky Mountains suggest a cool and relatively dry environment during the nonglacial (J.A. Burns 1990), while the onset of glaciation around Eagle Cave happened after 26,000-25,500 B.C. Middle Wisconsin fauna including mammoth, bison, and horse have been documented for Alberta and the Missouri Plateau in Montana during the Middle Wisconsin and after the Last Glacial Maximum (J.A. Burns 1996; Hills and Harington 2003; L.V. Hill 2001). The Laurentide ice sheet buried fossil localities with ages of 37,300 and 24,900 B.C. (R. Young et al. 1999).

Deglaciation in the region of the Peace, Athabasca, and Red rivers had probably started by around 16,200 B.C. (Jackson, Phillips, and Little 1999; Dyke, Moore, and Robertson 2003; Dyke 2004). Based on a wood age from glacial Lake Peace of about 15,400 B.C. (Lemmen, Duk-Rodkin, and Bednarski 1995), Dyke suggests that "it is possible that the corridor was entirely opened by" 14,700 B.C., which would have been possibly several thousand years earlier than evidence for the presence of human groups using Clovis artifacts in western North America. This age marks the separation of the Cordilleran and Laurentide Ice Sheets. It does appear that large tracts of deglaciated land were present by 14,000 B.C. and that a thin space between the ice margins in some places containing lakes was present by 12,700-11,300 B.C. (figs. 3-4). Within the region between the Athabasca and Red rivers, geologic evidence indicates a potential overlap of the Cordilleran and Laurentide ice sheets between 23,300 and 19,700 B.C. (Catto et al. 1996; C. Cambell and I.A. Cambell 1997). North of the Peace River, deposits with artifacts at Charlie Lake Cave have an age range of at least 10,600-10,400 B.C. (J.C. Driver 1988; Fladmark, Driver, and Alexander 1988; J.C. Driver et al. 1996). The site is situated on deposits of glacial Lake Peace. Bison were present in the North Saskatchewan River region by 11,500 B.C. (J.A. Burns 1996).

Some evidence has been used to suggest that the southern iceposition for the Laurentide continental ice was the Lethbridge moraine in southwest Alberta, within the Saskatchewan River drainage (Stalker 1977; Rutter 1984), while other data seems to support a more southerly limit along the Missouri River in northern Montana (Fullerton and Colton 1986; Fullerton, Colton, and Bush 2004; Fullerton et al. 2004). The difference is potentially important in terms of the presence of available land for biotic communities and the timing of glaciation. A glacial margin to the south, near the present location of the Missouri River in northern Montana, appears to be supported by Late Wisconsin luminescence ages on lake deposits above and below Laurentide till within the Missouri drainage near Great Falls, Montana (C.L. Hill and J.K. Feathers 2002). The advance and melting of the Laurentide glacier resulted in changes in the Saskatchewan and Missouri river drainages; proglacial (ice-marginal) lakes that developed along the margin help to constrain the deglaciation chronology of the region (J.T. Teller 1987, 1995, 2004; Beaudoin and Oetelaar 2003).

The Missouri River drainage contains evidence for interaction between mountain and continental glaciers and for other areas where the space between them was always ice-free, although the rivers and lakes were affected by both mountain and continental glaciation (fig. 2). Vertebrate remains provide information on Middle and Late Wisconsin landscapes close to the glacial margins (C.L. Hill 2001, 2001a). Mammoth remains from the Sun River area provide evidence for deglaciated landcapes at the end of the Pleistocene around 11,500 B.C. (Marsters, Spiker, and Rubin 1969).

During the Last Glacial Maximum in southwestern Alberta, mountain valley glaciers blocked drainages, resulting in lakes and the deposition of the "Albertan till" (= M1, early Pinedale) that can be traced eastward from the Rocky Mountain front (L.E. Jackson and E.C. Little 2004). The mountain glacier complexes in southwestern Alberta appear not to have been obstructed by Laurentide ice during the maximum of the early Pinedale glaciation, dating to perhaps 25,100-24,600 B.C. Instead it appears the continental ice margin was northeast of the early Pinedale margin. The interstade between the early and middle Pinedale consisted of retreat of the active mountain ice margins and areas of stagnant and buried ice on the landscape (Fullerton, Colton, and Bush 2004).

Most of the early Pinedale montane till was overridden by Late Wisconsin Laurentide ice (Karlstrom

1987). Continental till overlies a montane till and grades westward into montane till; this relationship indicates a coalescence of the continental Laurentide and montane piedmont glaciers during the Last Glacial Maximum (L.E. Jackson and E.C. Little 2004). Laurentide ice obstructed the middle Pinedale lobes in southwestern Alberta. The continental ice advanced over early Pinedale deposits and geomorphic features. In the Belly River valley of Alberta, middle Pinedale mountain glacier ice and late Wisconsin Laurentide ice coalesced. In the Saint Mary River valley of Montana, lakes formed between the middle Pinedale valley glaciers and the Laurentide margin. Glacial lakes Twin River (in the Milk River valley) and Cutbank were also dammed by the maximum advance of the Laurentide ice. This margin of the Laurentide was also contemporaneous with an advance of the Two Medicine lobe mountain ice and the highest late Wisconsin phase of glacial Lake Cutbank (Fullerton, Colton, and Bush 2004; Fullerton et al. 2004). Thus, the maximum middle Pinedale valley glaciation appears to have been contemporaneous with the maximum late Wisconsin glaciation. The estimated age of this maximum is about 22,000 B.C., perhaps 2,000-3,000 years after the early Pinedale maximum.

The maximum of the late Pinedale valley glaciation is estimated to date about 16,200 B.C.; this advance does not appear to have been contemporaneous with a readvance of the Laurentide ice margin (Fullerton, Colton, and Bush 2004). Montane glaciers had begun to recede by about 12,200 B.C., before deposition of the Glacier Peak G tephra and Mount Saint Helens set Jy (Carra 1995; Locke and Smith 2004). By 11,300 B.C. the Cordilleran ice had melted back into the mountains, although there was a small advance that can be correlated with the Younger Dryas in the southern Alberta Rockies (Reasoner, Osborn, and Rutter 1994). The presence of Mount Saint Helens tephra within strata of glacial Lake Great Falls may serve to constrain the deglaciation chronology of the southern margin of the Laurentide ice sheet (M. Parker and D.G. Smith 2004).

The pattern of ice retreat along the southwest margin of the Laurentide ice sheet is useful for understanding the availability of inhabitable postglacial landscapes. One model proposes a series of phases of deglaciation starting after about 19,700 B.C. In this scenario glacial Lake Kincaid and other proglacial lakes drained into the Missouri basin until about 15,400 B.C. As the retreat continued, waters from glacial Lake Bigstick and Old Wives Lake continued to drain into the Missouri River until 14,000 B.C. The moraines and glacial lakes formed at about 11,800-8000 B.C. are correlated with a stage of Lake Agassiz (Klassen 1989). By about 9500 B.C. the Laurentide margin had receded to the Lake Athabaska region in northern Saskatchewan. Thus, by the end of the Pleistocene there were hundreds of square kilometers of deglaciated terrain available for habitation east of the northern Rocky Mountains.

Radiocarbon ages of the remains of extinct fauna or associated depositional contexts imply the presence of habitable, ice-free conditions within the Missouri and Saskatchewan river drainages of northern Montana and southern Alberta from about 11,800 to 9500 B.C. (J.A. Burns 1996; C.L. Hill 2001, 2001a; Hills and Harington 2003). Indications of a Late Pleistocene human presence include Clovis artifacts between the Milk and Oldman rivers (Kooyman et al. 2001) and artifacts within the stratified sequence at Vermilion Lakes on the Bow River (Fedje et al. 1995). The deposits containing artifacts at Vermilion Lakes dated as early as about 10,700 B.C. overlie sediments of glacial Lake Vermilion. Mammal fossils indicating potentially inhabitable landscapes in the region date about 11,200 and 11,000 B.C. (Kooyman et al. 2001).

Mountain Glaciations and Western Lakes

A relationship between the timing of glaciation and the presence of lakes in western North America has been proposed (Antevs 1935a, 1948). Changes in the Polar jet stream, for example, have been suggested as a control on paleolakes and glaciers in western North America (Liccardi 2001; Liccardi et al. 2004). A review of the geologic evidence for Pleistocene lakes in the Great Basin is provided by Mehringer (1986). The approximate extent of Late Wisconsin mountain glaciers and some of the larger pluvial lakes are depicted in figure 7.

At pluvial lake Chewaucan (present-day Summer Lake basin in south-central Oregon) there are indications for expansions and regressions during the Middle Wisconsin. Higher lake levels are contemporaneous with warm indicators (interstadial conditions) in the Greenland isotope record and low levels during colder conditions (stadials) (A.S. Cohen et al. 2000; Zic, Negrini, and Wigand 2002). A composite diagram for pluvial lakes Fort Rock and Chewaucan suggests a rise in lake levels after possibly fluctuating conditions 28,000-22,000 B.C., with a highstand around 17,300 B.C., a recession around 14,000 B.C., and a small transgression around 11,800 B.C. (Negrini 2002; Jenkins, Droz, and Connolly 2004). In Lake Chewaucan this pattern is reflected by radiocarbon ages indicating a rise in lake levels at 26,100 and 24,600 B.C., and a maximum highstand around 19,200 B.C. (Negrini 2002). Radiocarbon ages on gastropods have been used to argue for high lake levels around 11,800 B.C. at Chewaucan during the time that there were low lake levels in the Bonneville and Lahontan basins (Licciardi 2001).

During the Late Wisconsin, there appear to have been different timings for the glacial advances in the Rocky Mountains. Some localities imply glacial advance prior to the global Last Glacial Maximum (Licciardi et al. 2004). In the Sawtooth Mountains of Idaho, for example, there were two main ice advances several thousand years after the Last Glacial Maximum; they date to before 15,500 B.C. and 11,800 B.C. (Thackray 2004).

The age of the Pinedale advances has been compared in the Yellowstone and nearby Wind River region, in Wyoming.

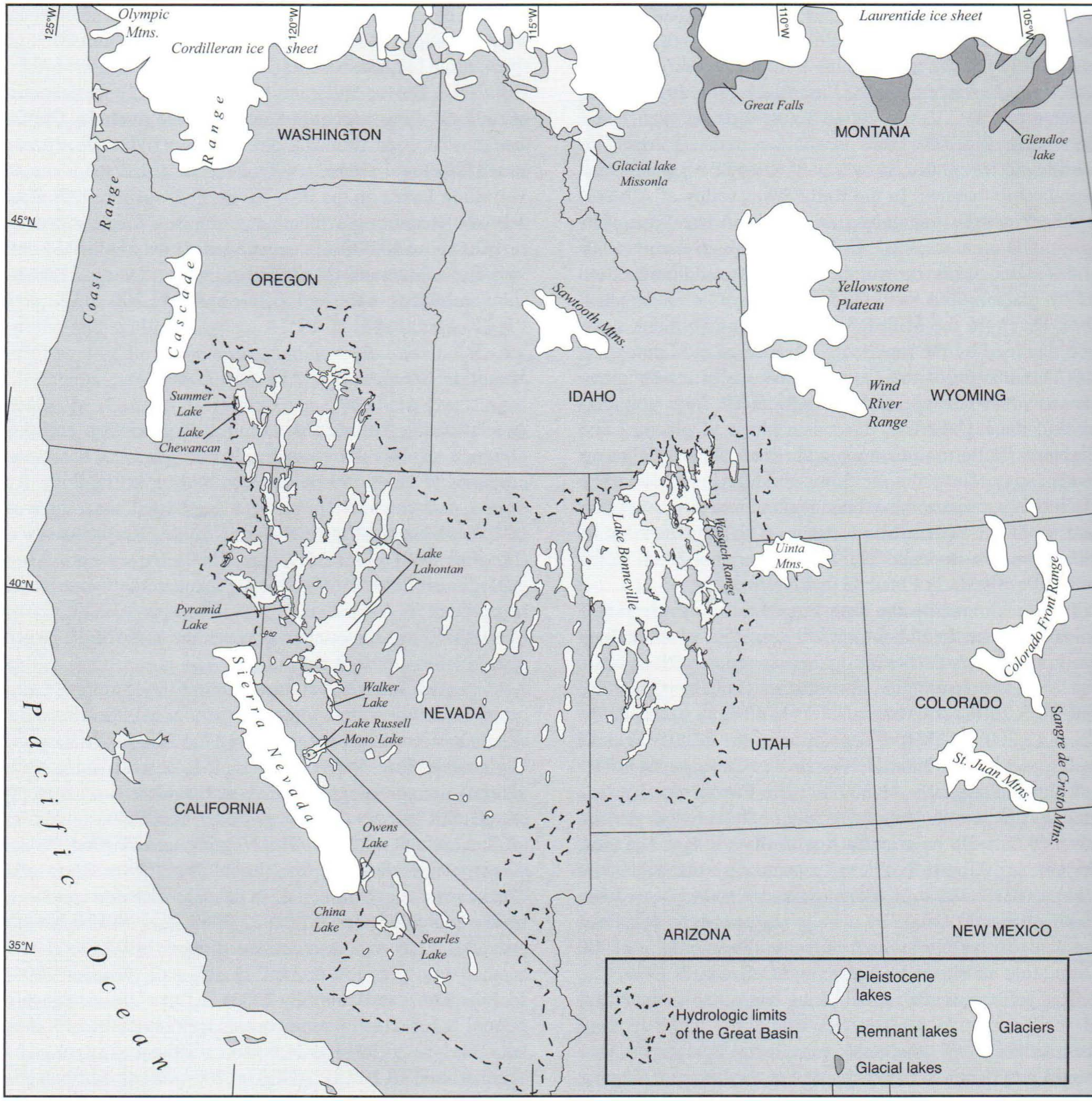

Fig. 7. Glaciated mountains and pluvial lakes south of the continental ice sheets in western North America (Dyke, Moore, and Robertson 2003).

The advance correlated with the Pinedale glacial maximum dates about 28,000-22,000 B.C., with deglaciation by 16,200 B.C. and a small readvance in cirque-valley glaciers around 11,800 B.C. (K. Pierce 2004). Sediments deposited during the melting of glaciers in the Yellowstone-Teton area have an age range of about 20,600-17,000 B.C. (Whitlock 1993). The Wind River Mountains contain Pinedale and earlier (Bull Lake) Late Pleistocene moraines that can be correlated with downstream terrace deposits (Chadwick, Hall, and Phillips 1997; Dahms 2004) as well as evidence for the Temple Lake advance prior to 12,100 B.C. (Easterbrook et al. 2003).

In the Colorado Front Range, Pinedale glaciation appears to have begun around 33,400 B.C. and ended before 11,800 B.C. with the glacial maximum likely in the range of 26,500 to 16,200 B.C. (Easterbrook et al. 2003). Maximum expansion of Pinedale ice is indicated by the presence of glacial Lake Devlin at roughly 25,000-23,200 B.C. while deglaciation had been initiated by about 15,100 B.C. (Dethier, Birkeland, and Schroba 2003). Along the Colorado Front

Range, the Pinedale glaciation has been correlated with the Broadway and Kersey alluvial terraces. These terraces contain mammoth, camel, and horse fossils at the Lamb Spring site, Colorado, and mammoth and Clovis artifacts at the Dent site, Colorado (cf. V.T. Holliday 1987; J.W. Fisher 1992; Madole 1986, 1995; Madole, van Sistine, and Michael 1999; C.V. Haynes et al. 1998).

Regionally, indications of the Younger Dryas are found in the Colorado Front Range (Menounos and Reasoner 1997), the San Juan Mountains of Colorado (Reasoner and Jodry 2000) and the southern Sangre de Christo Range. For example, in the Sangre de Cristo Mountains there is evidence of two Pinedale advances, with deglaciation occurring before 12,000 B.C., as well as a smaller advance around 9800 B.C. that could correspond to the Younger Dryas (Armour, Fawcett, and Geissmann 2002).

Both the Sierra Nevada and Wasatch Mountains were glaciated in the Pleistocene as were other mountains within the Great Basin (fig. 7). The glaciations in this region may have been generally contemporaneous with the formation of lakes in the basins, although the maximum expansions of some lakes appears to have occurred after the maximum of mountain glacial advances (Oviatt and McCoy 1992). It seems likely that the climatic factors that led to the formation of the lakes also played an important role in the dynamics of glaciation. For example, although Lake Bonneville was present during isotope stage 2, the highest level was reached around 17,300-16,200 B.C. (W.E. Scott et al. 1983; Madsen et al. 2001). By this time glaciers in the Wasatch Range had already reached their maximum extent and were melting.

The Middle Wisconsin (isotope stage 3) interstadial conditions in the Bonneville basin, northwest Utah, were associated with low lake levels. After about 30,600 B.C. lake levels began to rise with several major fluctuations until the Bonneville highstand at about 17,300 B.C. (Madsen et al. 2001; Oviatt 2002; Sack 2002). There are four major shorelines that mark lake levels in the Bonneville basin during the Wisconsin. The beach ridges of the Stansbury oscillation mark a rise around 26,000-22,000 B.C., while the Bonneville shorelines dated to about 17,300-15,800 B.C. mark the highest level of the lake. The lower Provo shoreline marks the lake level around 15,800-14,700 B.C., after overflow flooding through the Red Rock Pass into the Snake River drainage, while the Gilbert shoreline corresponds to the time of the Younger Dryas with an age range of 11,000-9500 B.C. (Easterbrook et al. 2003). It may reflect landscapes associated with people using Clovis and Folsom artifacts elsewhere in western North America.

Tills in the Wasatch Range have been related to Lake Bonneville. For instance, the age of the late Wisconsin Bells Canyon till is based on radiocarbon measurements of about 28,800 B.C. from a paleosol developed on the top of an underlying till (Madsen and Curry 1979) as well as on its relationship to sediments of Lake Bonneville. Glacial retreat occurred before the Bonneville highstand, since lake sediments overlie the glacial deposits and the Bonneville shoreline was formed on the moraines. Thus, lake levels were rising while the Wasatch Range glaciers were receding from their maximum position

High-water levels in several subbasins in northwest Nevada combined to form Lake Lahontan with a rise in lake levels after 33,400 B.C., reaching a maximum around 14,000 B.C., and then falling (K.D. Adams and S.G. Wesnousky 1998). Some of the water was derived from the Sierra Nevada and the mountains in Nevada. Within the Walker Lake subbasin there is evidence for a lake from 35,000 to 28,100 B.C. and no lake from 24,600-15,400 B.C. Camels dated to about 28,400 B.C. from the Wizards Beach level are present in the Pyramid Lake subbasin (Dansie, Davis, and Stafford 1988; Negrini 2002). By 27,000 B.C. several of the subbasins had been connected by rising water levels (L.V. Benson 1978; L.V. Benson et al. 1990). The highstand was reached between 15,800-14,000 B.C. (L.V. Benson 1993, 2004; T.C. Blair 1999). Camel bones dated to 13,700 B.C. mark the beginning of the recession (K.D. Adams and S.G. Wesnousky 1998, 1999). Lake levels had dropped to the Russell shoreline by 11,100 B.C. Transgression around 9500 B.C. has been linked to the Younger Dryas (L.V. Benson et al. 1992).

Glaciers in the Sierra Nevada fluctuated in response to Pleistocene climate change. This is reflected in glacial deposits in the mountains and paleolacustrine sediments in the basins; these have been used to estimate the timing of glaciations (L.V. Benson et al. 1998, 1998a; L.A. James et al. 2002; D. Clark et al. 2003). The drainage system included Lake Russell (present-day Mono Lake), and the Owens River system, including Owens Lake, China Lake, and Searles Lake. Within the Mono Lake basin, the deposits from pluvial Lake Russell can be related to glacial deposits from the Sierra Nevada. There appear to have been brief highstands around 38,200 and 28,800 B.C. followed by two high lake levels around 19,700 B.C. and 15,000-14,500 B.C., while the Tioga glaciation may range from about 29,000-15,400 B.C. (L.V. Benson et al. 1998). The Lake Russell highstand around 15,000 B.C. appears to have occurred after glaciers had begun to recede; the lake appears to have reached its maximum after glaciers in the Sierra Nevada had begun to melt from their maximum advance (L.V. Benson and Thompson 1987; L.V. Benson et al. 1998a).

Rock flour in Owens Lake in eastern California suggests glacial conditions in the adjacent mountains prior to the Middle Wisconsin and minor mountain glaciation during the Middle Wisconsin, oxygen isotope stage 3 (Bischoff and Cummings 2001; Bradbury and Forester 2002). The Tioga glacial advance, correlated with oxygen isotope stage 2, appears to have started around 27,600 B.C., with retreat beginning around 18,800-15,000 B.C. (L.V. Benson et al. 1998; Bradbury and Forester 2002; D. Clark et al. 2003). Alluvial fans along the front of the Sierra Nevada are primarily the result of sediment deposition during the maximum advance and then retreat of glaciers. The

Tioga 2 advance appears to have been the largest, dating to 24,500 B.C. (F.M. Phillips et al. 1996). Owens Lake was low around 15,000 B.C. and then began a several-thousand-year transgressive interval associated with wetter climates from about 11,000-9,000 B.C. (Mensing 2001; Easterbrook et al. 2003). Searles Lake may have had high water levels starting around 17,300-16,200 B.C. and lasting until after 15,400 B.C.; this highstand generally correlates with the lake levels in the Bonneville and Lahontan basins (cf. Lin, Broecker, and Hemming 1998). There is a rise in Searles Lake dated to about 14,000 B.C. and another prominent level around 11,000-10,400 B.C., within the range of the Younger Dryas. The last major late Pleistocene glaciation within the eastern Sierra Nevada, the Recess Peak advance, occurred before 11,800-10,900 B.C. (D. Clark and A. Gillespie 1997; D. Clark et al. 2003). There is some biotic evidence of a later cooling that may correspond to the Younger Dryas chronozone (Porinchu et al. 2003).

Within the Great Basin region the Younger Dryas may be linked to the presence of black mat spring deposits ("History of Research on the Paleo-Indian," fig. 10, this vol.) dated to about 9500 B.C., although the oldest black mats were present by about 11,600 B.C., slightly before the Younger Dryas interval (Quade et al. 1998). There seem to have been regional differences in western North America associated with the Younger Dryas. Some areas may have been cooler and wetter, while in others the episode may have been marked by dryer conditions C.V. Haynes 1991; V.T. Holliday 2000b, 2001). In general, late Wisconsin maximum glacial advances in the western Mountain ranges appear to have occurred before the highstands in Great Basin pluvial lakes.

Conclusion

The geologic framework for glaciated western North America appears to indicate the absence of widespread glacial conditions that can be generally associated with the Altonian and Farmdalian intervals, prior to about 28,000 B.C. Glacial conditions became more pervasive during the Woodfordian (about 28,000-11,600 B.C.), which includes the Last Glacial Maximum. The geologic conditions of the Late Pleistocene have had a significant effect on the presence of geomorphic features and sedimentological sequences that could contain paleontological and archeological evidence in western North America. Along the Pacific, global (eustatic) sea levels were about 125 meters or so lower during the Last Glacial Maximum. Large areas would have been exposed as a broad coastal plain. Also, as the glaciers advanced and then began to melt during the Late Wisconsin the geomorphic and ecologic character of the coastlines would have changed. Judging from the presence and age of mammal fossils, it seems very likely that there were habitable landscapes along the Pacific during the Middle and Late Wisconsin. Along the Mackenzie River valley and the eastern front of the Rocky Mountains some ice advance may have occurred prior to the Last Glacial Maximum. Several areas show evidence of the coalescence of Cordilleran and Laurentide ice. The deglaciation chronology indicates the possibility of ice-free conditions with glacial lakes starting possibly as early as 16,200-14,000 B.C. Within the Rocky Mountains, glaciation correlated with the Pinedale appears to have started in some places prior to the Last Glacial Maximum. The last major advance appears to fall within the interval of about 22,000-16,200 B.C. with some regional expressions of the Younger Dryas.

Late glacial conditions as reflected in the Greenland ice-core (Blunier and Brook 2001; Southon 2002) indicate fluctuating climate conditions with the climate warming of the Allerod chronozone (12,100-11,000 B.C.) disrupted by cooler conditions of the Younger Dryas (11,000-9500 B.C), prior to the beginning of the Holocene (Pre-Boreal). Changes associated with the Allerod and Younger Dryas intervals appear to be reflected in the geologic record of landscape evolution in some regions of western North America and may provide a contextual framework for understanding adaptations associated with the times when human groups were using Clovis and Folsom artifacts.