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The Stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: variations in paleoenvironments and implications for paleoclimatic interpretations

Patricia M. Anderson[a,]*, Anatoly V. Lozhkin[b]

[a]*Quaternary Research Center, Box 351360, University of Washington, Seattle, WA 98195-1360, USA*
[b]*Northeast Interdisciplinary Research Institute, Far East Branch, Russian Academy of Science, 16 Portovaya St., Magadan, 685000, Russia*

Abstract

Questionable chronologies have limited detailed reconstructions of past vegetation and climate trends for the Karginskii/Middle Wisconsinan interstade (abbreviated here as MW). However, recent results from continuous lake records, while not resolving all the dating issues, do provide a new framework within which to examine this intriguing period. Paleobotanical data suggest significant regional variations in the interstadial vegetation of Beringia. *Larix* forests were relatively common in central and western areas of western Beringia through much of the middle and late MW, whereas tundra dominated most eastern Beringian landscapes. During relatively warm intervals within the interstade, western Beringia was more extensively forested (at times achieving modern forest distribution) than was eastern Beringia, where *Picea* forests were limited to lowlands of interior Alaska and the Yukon Territory. A period of maximum tree-cover occurred between ca. 35 and 33 ka BP, but forests were also present in western Beringia and the Yukon Territory between ca. 39 and 33 ka BP. Although a period of maximum warmth probably occurred throughout Beringia between ca. 35 and 33 ka BP, significant regional variability characterized other intervals, with differences not only in the timing of climatic changes but also in their trends (e.g., warming in the upper Kolyma region, cooling in interior Alaska). The paleovegetational data suggest that the greatest interstadial warming occurred in far eastern and far western regions of Beringia, with areas that are now closest to Bering Strait showing more moderate climatic fluctuations. The causes for either the long-term or the rapid climatic variations within interstadial Beringia do not seem to relate simply to either Milankovitch or sub-Milankovitch scale forcings.

1. Introduction

Beringia encompasses the arcto-boreal region extending from the Lena River eastward to northwestern Canada. This vast subcontinent spans ca. 98° of longitude and 20° of latitude and, if placed over North America, would stretch from the Bering Strait to eastern Hudson Bay. Unlike most areas at high latitudes, much of Beringia remained ice-free during the late Pleistocene (here defined as marine isotope stages 2, 3, and 4 age equivalents; Shilo et al., 1987; Clague, 1991). Consequently, it is one of the few northern regions where it is possible to explore long-term regional responses to various combinations of global climatic controls. This paper summarizes paleoenvironmental data from the last major interstade (marine isotope stage 3 age equivalent), one of the more poorly understood climatic intervals in Beringia. Global climatic controls during this period were characterized by: (1) a Laurentide ice sheet that was of significant size but below its 14 ka BP extent; (2) atmospheric concentrations of CO_2, sea surface temperatures, and sea levels that were lower than present; and (3) summer insolation that at times was greater than present but never exceeded values for the early Holocene (Bartlein et al., 1991). Although an absolute chronology remains problematic for the paleorecords, interesting patterns do emerge and suggest that this interstade represents a unique late Quaternary climatic scenario.

Much of the late Pleistocene paleoenvironmental data from Beringia, especially information gathered during the pioneering stages of research, was obtained in connection with gold mining activities and exploration for gold-bearing deposits exposed in river and beach terraces. Stratigraphy and chronology dictated the scope of

* Corresponding author. Tel.: 1-206-685-7682; fax: 1-206-543-3836.
E-mail address: pata@u.washington.edu (P.M. Anderson).

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Quaternary research during this early period. As more data were collected and analyzed, new questions arose that guided Beringian research for the next two decades. These questions centered on glaciations and changing sea levels, extinct herbivorous megafauna (e.g., *Mammuthus, Bison, Equus*), types of vegetation that supported these animal populations, and timing and routes of human migration from the Old to the New World (e.g., Hopkins et al., 1982). During this second phase of research, the focus typically was on stadial events, particularly on the paleoenvironments of the latest Pleistocene glaciation (equivalent to marine isotope stage 2), because data were relatively abundant and radiocarbon age control was possible. In the past 15 yr, a third phase of Beringian research has begun. This phase has concentrated more specifically on describing past climates and examining possible mechanisms responsible for the observed climatic changes and associated regional variations (e.g., Bartlein et al., 1991; Anderson and Brubaker, 1994; Mock and Anderson, 1997). With this paleoclimatic emphasis, attention has turned increasingly to the transition from glacial to interglacial conditions in Beringia from 14 to 8 ka radiocarbon years ago and during the climatic optimum of the last interglaciation (isotope stage 5e equivalent).

In all phases of research, the stage 3 interstadial complex has been given little attention, even though these deposits provide a rich source of paleoenvironmental data and early studies suggested Beringia might have been extensively reforested at this time. This apparent lack of interest (with the notable exception of Schweger and Matthews, 1985) was due to the limited distribution of sites, questionable chronologies, and the uneven quality of the paleoenvironmental data. The most problematic evidence is from exposures that provide limited "snapshots" of some poorly defined, albeit interstadial time. Sections that contain longer radiocarbon-dated records are more useful, but the uncertain temporal continuity of these deposits, variations in depositional environments (which can affect the types and preservation of paleoenvironmental indicators, such as pollen), and the predominant representation of floodplain settings somewhat limit the interpretations possible with these data alone. Lacustrine sediments are preferable, because they provide continuous information about paleovegetational and inferred paleoclimatic changes integrated over a broader landscape. Ancient lakes of appropriate age are rare, and like the exposures, they also suffer from dating problems. A final factor in examining the last interstade is the recent reassignment of many "classic" interstadial sites to the last interglaciation (Hamilton and Brigham-Grette, 1991; Schweger and Matthews, 1991; Sher, 1991). Nonetheless, sufficient information of certain and probable interstadial age is available to offer some preliminary reconstructions of its landscapes and climates. Furthermore, the recent publication of several continuous lacustrine records that span all or most of this interstadial period provides a new framework within which to re-examine the earlier studies.

The following is a brief summary of published data preserved in alluvium, buried peats and other paleosols, yedoma, loess, and lake sediments. Paleobotanical evidence is used as the primary source for paleoenvironmental interpretations, but sedimentological and faunal analyses provide additional key information. In western Beringia (northeastern Siberia; here abbreviated as WB), interpretations are based on a generalized summary from an official stratigraphic scheme (Shilo et al., 1987) with additional detail provided for key sites. In eastern Beringia (Alaska and northwestern Canada; here abbreviated as EB), we have concentrated on summaries of key sites only.

2. Modern Beringia

EB and WB differ significantly in their physiography, vegetation, and climate. The topography of EB is dominated by the Brooks and Cordilleran ranges to the north and the Alaska and coastal ranges to the south (elevations exceeding 2000 and 3500 m, respectively). A series of alternating uplands and lowlands lies between the east–west trending mountain systems, giving extensive areas of low rolling terrain throughout much of the central region. A broad coastal plain extends from the northern mountains to the arctic seas. Three large river systems, the Yukon, Kuskokwim, and Colville, crosscut EB, with their main valleys paralleling the mountains.

WB is topographically more complex than EB. The highest mountains of WB are found in the upper Kolyma–upper Indigirka region (elevations 600–2500 m). The Chukchi (1000–1800 m) and Koryak (1000–1600 m) uplands dominate the central area. Two large lowlands occupy much of the northern coast (Indigirka–Yana–Kolyma lowland) and southeastern areas of Chukchi Peninsula (Anadyr–Penzhina lowland). Tectonic depressions are a dominant feature both along the coast and interior of WB, giving a basin and range appearance in some areas. The WB river systems, with the exception of the Kolyma and Indigirka drainages, tend to be shorter, with valleys that more typically trend north–south.

Boreal forest and shrub tundra are the primary vegetation types in both EB and WB. These two major ecotones are demarcated by both latitudinal (determined by summer temperatures) and longitudinal treelines (influenced by the cool marginal seas bordering the area of Bering Strait). However, the composition of the forests and high shrub tundra varies markedly between the two continents. The forests of EB are dominated by *Picea glauca* and *Picea mariana*, the former occupying sites with warm, well-drained soils and the latter occurring in areas with cool, moist substrates. *Larix laricina* is the third EB conifer species. Its distribution is limited to central Alaska, and the tree never forms a major component of the regional vegetation. Hardwoods include *Betula papyrifera*, *Populus balsamifera*, and *Populus tremuloides*. All these species are associated with disturbed sites and are part of the successional series culminating in well-developed *Picea* forests. Today, interior EB supports extensive closed forest, often characterized by large areas of *Picea mariana* muskeg.

Larix dahurica is the dominant tree species in WB, forming a more open-canopied forest than is found in EB. *Picea obovata* is restricted to a small, disjunct population near Magadan and grows in association with the tree *Alnus hirsuta*. Broadleaf deciduous tree species include *Betula platyphylla*, *Populus suaveolens*, and *Chosenia arbutifolia*. These species are restricted to disturbed or poorer quality microhabitats and tend to form dense gallery forests (often in association with *Alnus fruticosa*) along the main rivers. *Pinus pumila* is the main component of the forest understory. Lichens (in particular *Cladonia rangifera*) are also more typical of forest ground cover in WB than in EB.

Tundra composition differs less between EB and WB compared to the differences found in the boreal forests. *Betula*-Ericales–*Salix* tundra commonly occurs beyond altitudinal and latitudinal treeline. On favorable mountain slopes, dense shrub thickets of *Alnus* (EB) or *Pinus pumila* (WB) often form altitudinal belts just beyond treeline. Alpine tundra occupies higher elevations, although many of the mountaintops support little or no vegetation. High (up to 3–4 m) *Pinus pumila*–*Alnus* shrub tundra is unique to southern and central Chukotka. Cyperaceae–Poaceae–*Salix* tundra is typical of mesic to wet coastal areas in both EB and WB.

Twenty-six atmospheric patterns determine modern Beringian conditions, with most of these patterns resulting in heterogeneous climatic responses at the surface (Mock et al., 1998). This heterogeneity is most strongly expressed as variations between EB and WB. The major circulation controls of winter climates are the spatial variations associated with the Siberian high, the Pacific subtropical high, and the Aleutian low. Summer conditions are influenced by the long wave patterns over the Northern Hemisphere and specifically the configuration of the East Asian trough and adjacent ridges. Topographic variations also play an important role in determining local conditions. For example, steep precipitation gradients characterize southern coastal areas of Beringia due to rainshadow effects associated with the higher mountain ranges. Although topographic and maritime influences provide variations within and between EB and WB, both temperature and precipitation tend to decrease as latitude increases.

3. Chronology and paleogeography

The stage 3 interstadial complex of Beringia is classified by a variety of names that are applied from local to continental spatial scales (Table 1). Although Hopkins (1982) proposed the term Boutellier Interval (to include both the Siberian Karginskii interval and local stratigraphic names in Alaska and Canada), it has not gained wide usage. Therefore, we have chosen to use a term that is generally more familiar. Because Beringia spans two continents, we have arbitrarily decided to use the North American nomenclature: late (LW), middle (MW), and early Wisconsinan (EW).

The classical age assignment of the MW in North America is 70–28 ka BP. More recently the MW has been equated to marine isotope stage 3 (ca. 65–25 ka BP). Kind (1974) originally placed the MW of Siberia between ca. 50 and 22 ka BP, whereas data from WB suggest the MW–LW boundary more appropriately dates to ca. 27–26 ka BP (Kotov et al., 1989; Lozhkin and Anderson, 1996). The latter age is in better agreement with glacial and palynological studies that indicate the LW–MW boundary in EB dates to ca. 25 ka BP (Hamilton, 1982; Ten Brink and Waythomas, 1985; Hamilton et al., 1988a) to ca. 26 ka BP (Anderson, 1988). In EB, early workers followed the classical North American glacial sequence. Subsequently, Hopkins (1982) assigned the MW of Beringia to ca. 65–30 ka BP. More recently, MW studies have referred to the marine isotope scale as a temporal framework (e.g., Schweger and Matthews, 1985; Begét, 1990). Defining the age of the MW–EW boundary is difficult, because absolute age control for early MW sites is absent. However, glaciological evidence from western Canada indicates that a period of reduced glaciation began some time prior to 59 ka BP and ended between ca. 30 and 25 ka BP (Clague, 1991; Klassen, 1987). The oldest finite radiocarbon dates by region are Kolyma–Yana–Indigirka lowlands (33 ka BP), Chukotka (33 ka BP), Priokhot'ye (37 ka BP), upper Kolyma–upper Indigirka (45 ka BP), the Yukon Territory (41 ka BP), southern Alaska (32 ka BP), interior Alaska (43 ka BP), and northwestern Alaska (37 ka BP). Many of these ages are near the finite range of radiocarbon dating and thus must be interpreted with caution. Furthermore, many of the older dates have large standard deviations and in most instances should be viewed as nonfinite or minimum age limits. Because of these dating concerns, the focus on this paper will be the middle to late portions of the MW with some liberties taken by us in our chronological interpretations of radiocarbon dates older than 35 ka BP.

One of the most distinctive features of Beringia is the land bridge that periodically connected North America and Asia. Such massive changes in paleogeography strongly influenced the regional paleoclimates within Beringia (e.g., turning coastal areas into landlocked

Table 1
Stratigraphic and climatostratigraphic units of Beringia

Late Pleistocene units:			
Siberia[a]	*Marine isotope stage approximate age equivalent*	*North America*	
Kazanskii	5	Sangamon	
Zyranskii	4	Early Wisconsinan	
Karginskii	3	Middle Wisconsinan	
Sartanskii	2	Late Wisconsinan	
Siberia — Middle Wisconsinan units:			
Karginskii Interval[b]	50–45 ka BP	Unnamed warm interval	
	45–43 ka BP	Unnamed cool interval	
	43–33 ka BP	Malokhetskii warm interval (climatic optimum)	
	33–30 ka BP	Konotzel' skii cool interval	
	30–22 ka BP	Lipovskoy–Novoselovskii warm interval	
Northeast Siberia — Middle Wisconsinan units:			
Name	*Area*	*Age*	*Climate*
Khudzhakhskii Horizon[c]	NE Siberia	ca. 40 ka BP	Slightly cooler than present
Kirgilyakhskii Interval[d]	Upper Kolyma	45–43 ka BP	Cool
Kubalakhskii Horizon[e]	Upper Kolyma	MW	Modern climate
Nemkinskiye Layers[f]	Priokhot'ye	24–37 ka BP	Modern or near modern
Molotkovskii Horizon[f]	Yana-Indigirka-Kolyma lowland	24–48 ka BP	Varying climates
Longovskii Horizon[f]	Chukotka	MW	Varying climates

[a] Saks (1953).
[b] Kind (1974).
[c] Goldfarb and Lozhkin (1975).
[d] Lozhkin (1991).
[e] Vaskovskii and Terekhova (1970).
[f] Shilo et al. (1987).

interior; Fig. 1). Unfortunately, the sea level history of Beringia is poorly understood (Hopkins, 1982), even for the most recent glacial–interglacial cycle (Elias et al., 1992). A tentative and incomplete sea-level curve proposed by Hopkins (1982) suggests that seas bordering Beringia dropped to between ca. -25 and -50 m during the late to middle MW. Global estimates of late MW sea levels vary from near modern (Finkelstein and Kearney, 1988) to -75 m (Shackleton, 1987). Values between -35 and -40 m are generally accepted as a worldwide maxima, although regional sea-level changes may differ (Chappell and Shackleton, 1986; Colman et al., 1989).

4. Middle wisconsinan records from Beringia

In this section, we provide a brief historical perspective on MW research, followed by descriptions of key sites from both WB and EB. We subdivide WB into four regions (Fig. 1): Yana–Indigirka–Kolyma lowlands (region A), Chukotka (region B), Priokhot'ye (region C), and the upper Kolyma–upper Indigirka (region D). EB is also subdivided into: the Yukon Territory, Canada (region E), southern Alaska (region F; note: material from the Alaskan panhandle is not included), interior Alaska (region G), and northwestern Alaska (region H).

4.1. Western Beringian sites

Saks (1953), Kartashov (1963, 1966), and Kind (1974) pioneered research into late Pleistocene stratigraphy, paleoclimates, and paleoenvironments of Siberia. Saks (1953) formulated the first stratigraphic scheme for the late Quaternary and defined the basic terminology of Kazanskii, Zyryanskii, Karginskii, and Sartanskii intervals (equivalent to marine isotope stages 5–2, respectively; Table 1). This scheme, defined mostly from work in the Yenisei valley, was interpolated to deposits across Siberia and is still in use today.

Kartashov (1966) provided the initial paleovegetational and paleoclimatic interpretations of the Karginskii interval, which he correlated with the MW of North America. Reinterpreting pollen data from the upper Kolyma region, originally analyzed by Baskovich (1959), Kartashov inferred the vegetation to be a light (i.e., *Larix*-dominated) coniferous taiga. The distribution of vegetation zones during this interval was similar to that of the Holocene warm period. He further postulated that

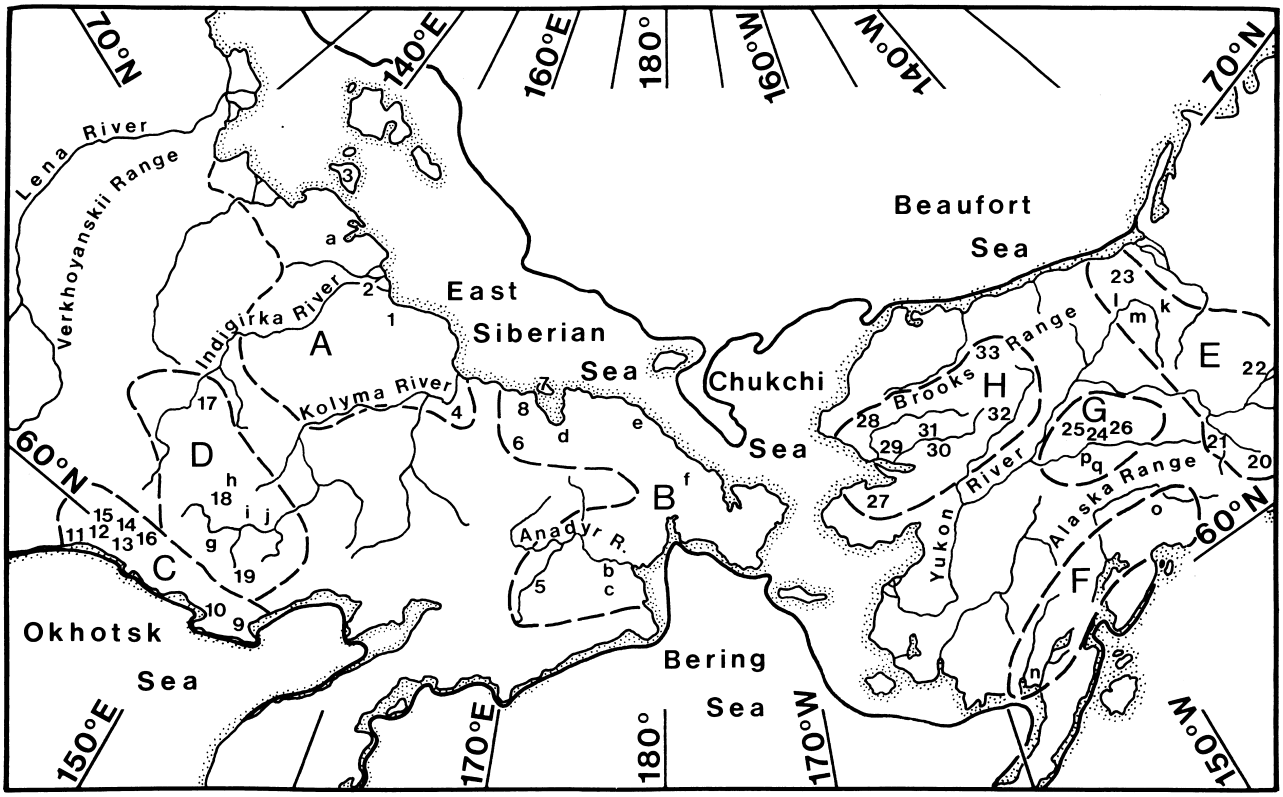

Fig. 1. Map showing location of MW sites in Beringia. The -50 m shoreline (heavy solid line) illustrates the likely maximum sea level depression for the MW; the modern shoreline is stippled. The regions discussed in the text are outlined by a dashed line and indicated by capital letters. Site locations are indicated by numbers; lower-case letters are used to show general localities or in areas where sites were too closely spaced to be shown individually. The key to areas and sites is:

REGION A (YANA–INDIGIRKA–KOLYMA LOWLAND)
Sites:
1. Bolshoii Khomus-Yuryakh River
2. Shandrin mammoth, Shandrin River
3. Bolshoii Lyakhovskii mammoth, Bolshoii Lyakhovskii Island
4. Malyii Anyuii River

Area:
a. Khroma-Keremesit

REGION B (CHUKOTKA)
Sites:
5. Lyedovyi Obry, Main River
6. Enmynveem mammoth, Enmynveem River
7. Aion Island
8. Rauchua River

Areas:
b. Anadyr–Tnekveem Rivers
c. Velikaya River
d. Chaun-El'khkakvun–Omrel'kai Rivers
e. Val'karaiskaya Depression–Ryveem River
f. Vankaremskaya Depression–Amguema River

REGION C (PRIOKHOT'YE)
Site:
9. Yama River
10. Tanon River
11. American River
12. Urak River
13. Bolshoi Marekan River
14. Kukhtuii River
15. Selemdzha River
16. Asibergan River

REGION D (UPPER KOLYMA–UPPER INDIGIRKA)
Sites:
17. Kuobakh-Baga River
18. Kirgilyakh mammoth (Dima), Kirgilyakh River
19. Elikchan 4 Lake

Areas:
g. Detrin River
h. Berelekh-Khatakchan River
i. Debin River
j. Kolyma River

REGION E (YUKON TERRITORY)
Sites:
20. Silver Creek
21. Antifreeze Pond
22. Mayo Indian Village
23. Hanging Lake

Areas:
k. Bell Basin
l. Old Crow Basin
m. Bluefish Basin

REGION F (SOUTHERN ALASKA)
Areas:
n. Nushagak Lowland
o. Copper River Basin

REGION G (INTERIOR ALASKA)
Sites:
24. Isabella Creek
25. Fox Permafrost Tunnel
26. Harding Lake

Areas:
p. Fairbanks loess
q. Tanana-Kuskokwim Lowland

REGION H (NORTHWESTERN ALASKA)
Sites:
27. Imuruk Lake
28. Kaiyak Lake
29. Squirrel Lake
30. Joe Lake
31. Epiguruk
32. Site EIV, John River
33. Ahaliorak Lake

the MW climate of the northeastern USSR was: (1) a uniformly warm period with no climatic fluctuations; (2) most similar to the Holocene optimum; and (3) more like an interglacial rather than a typical interstadial period. These interpretations are in contrast to the reconstructions of Baskovich (1959), who described the interval as being cooler than present, based on the presence of abundant shrub *Betula* pollen.

Radiocarbon analysis was first used by Kind (1974) to define an absolute chronology for parts of the Wisconsinan. Like Saks, much of the original data were gathered in the Yenisei region, but inferred to be applicable for all of Siberia. She also discussed in more detail the climate changes for Siberia during the late Pleistocene. The age of the Karginskii interval was proposed as ca. 50–22 ka BP and was classified as an interglacial, as opposed to interstadial, period. The Karginskii interval was subdivided into five periods: an early warm period between 50 and 45 ka BP; an early cool time at ca. 45 ka BP whose length is problematic; the climatic optimum (Malokhetskii warm interval) between 43 and 33 ka BP; a late cool period from 33 to 30 ka BP (Konotzel'skii cool interval); and a late warm period (the Lipovskoy–Novoselovskii warm interval) from 30 to 22 ka BP.

As fieldwork increased in WB proper, Goldfarb and Lozhkin (1975) proposed the name Khudzhakhskii Horizon for MW sediments of northeastern Siberia. This definition was based on sandy and loamy alluvial sediments exposed along the 20 m terrace of the Khudzhakh and the 35–45 m terrace of the Bolshaya Kuobakh–Baga (site 17) Rivers, upper Indigirka basin. Climate was interpreted to be similar to or slightly cooler than modern based on pollen analyses which contained high percentages of tree and shrub *Betula*, minor amounts of *Larix*, and occasionally significant amounts of *Alnus* and *Pinus pumila* pollen.

As the WB database grew over the next decade, an official Quaternary stratigraphic scheme for the eastern USSR was compiled (Shilo et al., 1987). This volume summarizes data from known sites, provides descriptions of type sections, and describes the regional stratigraphic terminology. Much of this scheme is still valid today, and some of its details are described below. A major exception is for the upper Kolyma region where Lozhkin (1991) described fluctuations in the MW based on research at the Kirgirlakh mammoth site. Here he defined the Kirgilyakhskii cool interval based on paleobotanical data associated with the remains of the baby mammoth called Dima. Further details of this analysis are provided below, but briefly Lozhkin stated that dramatic changes in vegetation and climate were characteristic of the MW across all of northeastern Siberia. These fluctuations suggest that at times climate was as warm as during the Holocene, whereas other cool periods approximated stadial conditions.

The following summary is divided into four geographic regions: the Yana–Indigirka–Kolyma lowland (region A), Chukotka (region B), Priokhot'ye (region C), and the upper Kolyma–upper Indigirka (region D; Fig. 1). Thus we will trace the temporal and spatial patterns of the MW environments along what is today the northern coastal region of western Beringia, moving to the southern coastal region, and finishing in the mountainous interior. See Fig. 1 for individual site locations.

4.1.1. *Yana–Indigirka–Kolyma lowland* (*region A*)

This region encompasses the lower courses of the Yana, Indigirka, and Kolyma drainages and comprises the largest lowland in WB. The landscape is characterized by "yedoma relief", that is, a series of small hills, steep-sided plateaus, and small meadowy depressions. This topography is caused by thermokarst processes acting on sediments resulting in "thick syngenetic ice wedges, separated by polygonal blocks of perennially frozen loess" (Tomirdiaro, 1982, p. 30). The modern vegetation is a mosaic of open *Larix* forest, *Betula–Salix*–Ericales tundra, and moist graminoid (i.e., Cyperaceae and Poaceae) tundra. MW sediments are typically well-sorted silts. However, sand lenses containing numerous peat layers occasionally occur. The origin of the nonorganic members of yedoma has been debated for decades, with some researchers favoring an alluvial history (e.g., Sher, 1971; see also Sher, 1997) and others claiming an aeolian origin (Tomirdiaro, 1980). The reliability of radiocarbon dates and associated chronological interpretations of the yedoma suite have been questioned (Lozhkin, 1987; Sher and Plakht, 1988; Sher, 1991). Sher (1991) suggested that many of these deposits more appropriately belong to the last interglaciation (marine isotope stage 5e equivalent). Lozhkin (1987) has argued that many radiocarbon dates of <28 ka BP are younger than would be expected by stratigraphies developed in bordering regions. However, materials with >28 ka BP ages are correctly assigned to the MW, although the exact age within the interstade can be suspect.

A 25 m high exposure (site 4) on the right bank of the Malii Anyuii River, a tributary to the Kolyma River, contains the type section for the MW, regionally referred to as the Molotkovskii Horizon. The MW sediments are composed of three distinct sediment facies that occur between 12 and 19 m above the river level (Shilo et al., 1987). This horizon has been radiocarbon dated from ca. 48–24 ka BP, although, as mentioned above, the exact chronology remains questionable (Sher, 1991, pers. comm.).

The lowermost MW layer (ca. 4 m thick) is dominated by lacustrine silts and includes a 0.5 m thick peat, which is radiocarbon dated from ca. 48–34 ka BP. The dominant pollen taxa in the peat include *Betula* sect. Albae and Poaceae with pollen of *Alnus* and *Betula* sect. Nanae of secondary importance. Minor types include Bryales, *Salix*, and *Larix*. These spectra are interpreted as a *Larix–Betula* forest with an understory of shrub *Betula* and *Alnus*. Tree-cover is interrupted by open boggy areas. Climate is inferred to be as warm as present.

The middle MW layer (10 m thick) consists of silt crosscut by ice-wedges and pseudomorphs. Radiometric dates are absent, but presumably these silts were deposited sometime between 34 and 28 ka BP. The main taxa are Poaceae and *Artemisia* pollen and Bryales spores with a subdominance of Chenopodiaceae, Caryophyllaceae, and Compositae pollen. The vegetation is interpreted as steppe tundra. Faunal remains include *Mammuthus primigenius, Equus cabalus, Alces alces, Ovibos pallantis, Bison priscus*, and *Coelodonta antiquitatis*. Based on the paleobotanical data, the climate is thought to be cooler than today. Although poor radiocarbon control precludes a definitive age, these middle deposits are perhaps correlative to the Konotzel'skii cool interval.

The upper level (3 m thick) includes a lens of lacustrine silt overlain by peat. This association is characteristic of an ancient *alas* complex (i.e., meadows within thermokarst depressions). Radiocarbon dates bracket this part of the section between ca. 28 and 24.5 ka BP. The lake muds are dominated by pollen of tree and shrub *Betula*. The peat contains abundant *Sphagnum* spores and *Alnus* pollen. Subdominant taxa in both the peat and lake sediments include Ericales, *Pinus pumila*, and Cyperaceae pollen and spores of bryophytes. *Larix* and *Salix* pollen occur sporadically and in small amounts. Like in the lower MW deposits, the vegetation is inferred to be a mosaic formed by *Larix–Betula* forests and open boggy areas. Climatic conditions were similar to those of today.

No other site from the lowland encompasses the complete Malii Anyuii record. However, dated deposits from other localities are consistent with the interpretation from the type site. For example, radiocarbon dates of 32,030 $\pm$ 1170 BP (MAG-316A) and 32,100 $\pm$ 900 BP (MAG-316) obtained from the skin and muscle of a mammoth preserved in yedoma on Bolshoii Lyakhovskii Island (site 3) also fall within the Konotzel'skii interval (Lozhkin, 1990). The palynological spectra are typical of this cool period (i.e., abundant Cyperaceae, Poaceae, *Artemisia*, and Caryophyllaceae pollen and Bryales spores) and indicate a dry herb tundra. The ancient and modern pollen spectra from this site are both dominated by herbaceous taxa, but unlike today, the MW assemblage lacks significant input of *Alnus*, *Betula*, or *Pinus* pollen. These shrubs are currently absent on the island, but their pollen is transported from more southerly areas by continental off-shore winds. The absence of these exotic taxa in the MW deposits suggests that shrub tundra lay at a greater distance to the south and that climate was cooler than today.

The occurrence of another cool period some time prior to 28 ka BP is supported by work at the Khroma Keremisit site (area a) in the Yana–Kolyma lowland. A silty exposure in a river terrace yielded radiocarbon dates of 31,000 $\pm$ 1000 BP (MAG-425) and 30,000 $\pm$ 600 BP (MAG-660) which date the initiation of ice-wedge growth. Paleobotanical analysis indicates the presence of a herb–moss association typical of a more northerly arctic tundra (Ovander et al., 1987).

Periods of relative warmth during the MW are indicated by the Shandrin River mammoth site (Indigirka lowland; site 2). Here a stomach lodged within the animal's skeletal remains was preserved in the permafrost (Arslanov et al., 1980). The stomach was dated at 41,750 $\pm$ 1290 BP (LU-505) and 40,350 $\pm$ 880 BP (LU-595). Analysis of the stomach contents indicated the presence of *Larix* trees. Pieces of *Larix* wood, dated to 41,200 $\pm$ 2000 BP (MAG-108), were also found in a river terrace on the Bolshoii Khomus Yuriyakh River (Indigirka lowland; site 1). Horizontal layers of silt and fine sand with occasional paleosols characterize this exposure. Shrub tundra dominates both areas today, indicating a northward extension of treeline at ca. 40 ka BP. Arslanov et al. (1980) postulated that *Larix* grew as far north as the modern land–sea boundary during the MW, which at this time probably experienced a more continental climate than today because of lowered sea levels. Lozhkin (1991) argued that mean annual temperatures at this time were probably relatively cool across WB, but summers remained sufficiently warm to permit trees to survive in the lower Kolyma basin.

4.1.2. Chukotka (*region B*)

Chukotka occupies the far northeastern tip of Asia, separated from the Alaskan mainland by only 100 km. High rolling uplands and tectonic depressions give the region a variable topography, and with the exception of the Anadyr drainage, Chukotka lacks the large river systems of interior WB. *Larix–Pinus pumila* forest is limited to westernmost Chukotka. Other areas are dominated by either a high *Pinus pumila–Alnus* shrub tundra (to the south and center) or a mid-arctic *Betula*–Ericales–*Salix* shrub tundra (to the north and east).

Late Quaternary sites are few and dating control is relatively poor in Chukotka. Much of the primary data are unpublished or in reports that are difficult to obtain. Consequently, the following description is based largely on the Shilo et al. (1987) summary.

MW deposits (termed the Longovskii Horizon) of northwestern Chukotka are found on the first terrace of the Chaun, El'khkakvun, and Omrel'kai Rivers (area d); the 25 m high terraces of rivers on Aion Island (site 7); and terraces of various heights along the Rauchua River (site 8), the Val'karaiskaya (area e) and Vankaremskaya (area f) depressions; and the Malyii Anyuii basin (site 6). A variety of depositional environments are represented, characterized by alluvial gravel and sand, lacustrine sand intercalated with peat, yedoma silt with ice wedges, and lagoonal sand and gravel. Significant variations in the interstadial vegetation and climate are inferred from analyses of these deposits.

The Enmyveemskii mammoth site (site 6) is one of the more informative from northwestern Chukotka. The right leg of a mammoth, dated to 32,890 $\pm$ 1200 BP (MAG-1001B) and 32,810 $\pm$ 720 BP (MAG-1001A), was excavated from frozen silts exposed along the Enmynveem River (Malyii Anyuii basin). Palynological analysis of the fur and the attached silt indicates high percentages of Poaceae and *Artemisia* pollen and *Selaginella rupestris* and Bryales spores. Taxa of secondary importance include Ranunculaceae, Cyperaceae, and Caryophyllaceae. Pollen of shrub species does not exceed 10% and is dominated by *Salix*. Pollen concentrations are 4–10 times lower than today. The vegetation is inferred to be a *Salix*–herb tundra, perhaps similar to the modern arctic tundra near the East Siberian Sea (Lozhkin et al., 1988). Today the Enmynveem River flows through an open *Larix* forest. Paleobotanical data indicate cooler than present conditions that probably correspond to the Konotzel'skii interval.

Other sites found in northern Chukotka provide only tantalizing glimpses of MW conditions. For example, in the Chaun depression pollen analysis associated with an Upper Paleolithic *Mammuthus–Rangifer tarandus* site suggests the region supported a shrub and/or shrub–herb tundra with *Betula, Salix*, and perhaps *Pinus pumila* being present. Herb-dominated paleobotanical data preserved in lagoonal sand and pebbles near Cape Shmitt and along the Ryveem River (area e) were dated to ca. 33.7 ka BP (MGU-338) and 33.2 ka BP (Ri-191). This evidence suggests that areas of the Val'karaiskaya depression were tundra steppe and conditions were significantly more severe than today. In the Vankaremskaya depression (area f), the 40–55 m high terrace of the lower valley of the Amguema River includes lake, peat, and mixed clay–peat deposits that have been dated to ca. 30.6 ka BP (MAG-578) and 28.9 ka BP (MAG-574). Associated faunal remains are typical of the Upper Paleolithic complex of *Mammuthus-Rangifer tarandus*. The paleovegetation is reconstructed as tundra–steppe, although shrub *Betula* and *Salix* communities perhaps were present at times. These data, too, suggest climates that were greatly to slightly cooler than present.

In southern Chukotka, MW sites are limited to the Koryak–Anadyr area. As in the north, the sites are generally poorly dated. Paleobotanical data suggest the widespread presence of a *Betula* forest tundra and a high shrub tundra. For example, 5–7 m thick sections of lagoonal and deltaic sand, gravel, clay, and peat found near Kresta Gulf are characterized by high percentages of tree and shrub *Betula*, shrub *Alnus*, and *Salix* pollen and *Lycopodium*, *Sphagnum*, and *Selaginella rupestris* spores. A single radiocarbon date places this assemblage at ca. 27 ka BP. The regional vegetation is inferred to be a high shrub tundra with *Betula* forest–tundra limited to the most favorable sites. In the nearby lower reaches of the Anadyr, Tnekveem, and Velikaya Rivers (areas b and c), MW sediments are preserved in 5–15 m thick deposits exposed along the second terraces. These terraces, generally 10–15 m high, consist of alluvial sand and coarse gravel and occasionally marine clays. Radiocarbon dates are absent. Pollen analysis suggests the presence of shrub tundra and *Betula* forest–tundra, and climate is inferred to be warmer than present.

Perhaps the richest MW sites in all of Chukotka are located in river exposures along the Main River in the southern Anadyr Basin (Kotov et al., 1989; Lozhkin, 1991). The most detailed work has been done at the Lyedovyi Obryv site (site 5), a 33–35 m high exposure rich in plant remains. Radiocarbon dates encompass the interval from 42,000 $\pm$ 1300 BP (MAG-804) to 19,500 $\pm$ 500 BP (MAG-815). The MW–LW boundary was dated at 27,400 $\pm$ 500 BP (MAG-810) and 27,000 $\pm$ 500 BP (MAG-811). Consequently, Kotov et al. (1989) proposed that the MW–LW transition occurred at ca. 27 ka BP rather than 22 ka BP as proposed by earlier researchers (e.g., Kind, 1974). The section contains alluvial sand, lacustrine muds, and yedoma of MW and LW ages. Pseudomorphs and syngenetic ice-wedges up to 2.5 m wide and more than 6 m long occur in the yedoma. Stadial and interstadial pollen spectra are dominated by Poaceae, Cyperaceae, and *Artemisia* (L.P. Zharikova, unpublished data). The LW deposits are characterized by a herbaceous macroflora. Paleobotanical remains lying below the LW–MW boundary (i.e., river level to ca. 23.5 m) are characterized by large shrubs with thick branches. An intense study of wood types was not done, but the large shrubs likely are *Salix*, perhaps with a minor component of shrub *Alnus* and *Betula* (Lozhkin, unpublished data). *Pinus pumila* is probably absent. Although there is little change in the palynological spectra at the site, the macrofossil materials suggest that the MW vegetation was a high shrub tundra, whereas a herb–moss tundra characterized the stadial period. The MW climate is inferred to be cooler than present but warmer than Sartan times.

4.1.3. *Priokhot'ye* (*region C*)

Priokhot'ye is that region which borders the Okhotsk Sea. Its coastal section is characterized by a series of tectonic depressions, bordered to the north by mountain ranges. River systems are typically short and braided. The bottoms of the depressions support Cyperaceae meadows, with nearby areas of moderate elevation supporting *Larix–Betula platyphylla* forests. A *Pinus pumila* shrub tundra occurs just beyond altitudinal tree line. A disjunct forest of *Picea obovata* and *Alnus hirsuta*, more typical of areas 1000 km distant, grows in the Yama depression (site 9). MW deposits are informally classified as Nemkinskiye layers based on the work in the Okhoto–Kykhtuii depression (Shilo et al., 1987). TL dates from Moscow State University indicate that these layers date from 37,000 ± 500 BP to 24,000 ± 400 BP. Nemkinskiye layers have been described at various sites: 3 m thick beds of well-rounded fluvial pebbles on the 4–6 m high terrace on the American River (site 11); 2.5 m thick layers of alluvial pebbles, sand, and gravel from the upper part of the 8 m terrace of the Urak River (site 12); beach pebbles and sand from the lower 16 m terrace of the Bol'shoi Marekan (site 13) and Kukhtuii (site 14) Rivers; alluvium from the 8 m terrace of the Selemdzha River (site 15); and alluvial gravel, clay, and sand from the 3 m terrace of the Asibergan River (site 16). The palynological assemblages from these sites are similar and characterized by *Larix, Picea,* and *Betula* sect. Albae, *Alnus* (tree and shrub), *Pinus pumila*, Ericales, and Poaceae pollen and *Sphagnum* spores. These spectra are interpreted as a mosaic of pure *Larix* forest and a mixed *Larix–Betula–Picea–Alnus* forest with an understory of *Betula*, *Alnus*, and *Pinus pumila* shrubs. Such vegetation suggests modern climatic conditions.

The lower Tanon River basin contains several sites that provide some of the best data about late Quaternary environments of Priokhot'ye (Lozhkin, 1989, 1991). Two different sections exposed at a quarry site (site 10) in the Tanon depression, dated to the Holocene (local site number 5) and late Pleistocene (local site number1), are particularly informative. Holocene sediments include clay, sand, and paleosols with shrub macrofossils and tree stumps (in growth position). Radiocarbon dates indicate that these sediments were deposited between ca. 6.6 and 5.9 ka BP. The palynological assemblage from local site number 5 is dominated by *Betula, Pinus pumila*[1] and *Alnus* pollen and *Sphagnum* spores (zone TA3; Fig. 2). Ericales, *Rubus chamaemorus*, and *Lycopodium* species are minor but ecologically significant components of the diagram. The vegetation is interpreted to be similar to modern and comparable to the pollen assemblage described below for zone TA-1.

A 10–12 m thick sand, exposed at local site number 1, is assigned to the late Pleistocene. The upper horizon is a brown-gray color and contains pseudomorphs with vertical depths of ca. 2 m. A date of > 20,760 BP (MAG-1180) obtained from a piece of wood suggests this upper sand is LW in age. An age of 21,600 ± 200 BP (GIN-6309) was obtained from the tusk of *Mammuthus primigenius*. Although not in situ, the displaced tusk was found lying near the wood. Palynological analysis of the upper sand (zone TA2) yields spectra composed of up to 90% herb pollen. Poaceae is the most abundant (up to 45%) herb taxa, but Cyperaceae, *Artemisia*, and Caryophyllaceae also occur in significant amounts. Percentages of shrub pollen vary from 5 to 15%. The minor herb pollen flora are more diverse than in zone TA3. *Selaginella rupestris*, Hepaticae, and Bryales are the dominant spore types. This assemblage is similar to other LW spectra from WB and comparable to the pollen spectrum from sediments coating the mammoth tusk. The vegetation is interpreted to be a herb-dominated tundra, perhaps with local *Salix* thickets. Climate was more severe than present.

Thin horizontal layers of silty sand and medium-grained sand of fluvial and lacustrine origin characterize the basal horizon at the quarry. This sand also includes paleosols composed of fine organic matter and roots of herbs in growth position. Radiocarbon dates of > 26,580 BP (MAG-1197), > 30,130 BP (MAG-1198), > 31,900 BP (MAG-1200), and > 33,400 BP (MAG-1201) suggest the lower sand is of indeterminant but likely MW age. During the MW, this flat-bottomed depression was probably occupied by the floodplain of slow-moving rivers. The floodplain was dotted by numerous small lakes. Pollen from trees and shrubs comprises 25–45% of the spectra from the MW sand (zone TA1). Of note are the higher percentages of *Pinus pumila, Betula*, and *Alnus* pollen in comparison to zone TA2. Poaceae, Cyperaceae, and *Selaginella rupestris*, taxa that were dominant in the LW zone, are of less importance, whereas *Artemisia* and Caryophyllaceae percentages remain relatively high. The vegetation is reconstructed as a *Pinus–Betula–Alnus* high shrub tundra. *Pinus pumila* perhaps formed a broad belt along the slopes of the mountains that border the depression. *Larix* appears in four levels, suggesting that a forest-tundra was established at least briefly. Because *Larix* is an under-represented and easily degraded pollen type, forest may have occupied the area longer than implied by the pollen diagram. The presence of herb taxa and *Selaginella rupestris* suggests the occurrence of disturbed and/or sparsely vegetated areas (e.g., higher elevations in the mountains), although the increase in Polypodiaceae spores implies locally moist habitats. Because climate is inferred to be cooler than present, but more hospitable than in LW times, altitudinal tree limit might have been lower than today.

4.1.4. *Upper Kolyma–Upper Indigirka* (*region D*)

The upper Kolyma–upper Indigirka region includes the upper basins and tributaries of interior WB. It is a mountainous area that today supports *Larix* forests at lower and mid-elevations, and a belt of *Pinus pumila* shrub tundra, at times in association with shrub *Betula* and *Alnus*, beyond altitudinal treeline. Most of the higher mountain tops are devoid of plants or support a discontinuous alpine tundra. This region includes the richest and most informative MW sites within WB. Paleoenvironmental data were obtained from alluvial, lacustrine, and mammoth sites discovered in river exposures (e.g.,

[1] *Pinus pumila* is shown in the figures as *Pinus* Haploxylon. We feel confident that this curve represents *P. pumila*, because it is the only Haploxylon type present in northeast Siberia and its size is larger than other *Pinus* grains.

the Kolyma (area j), Detrin (area g), Berelokh (area h), Khatakchan (area h), and Debin (area i) Rivers).

The first descriptions of MW sediments were done by Vaskovskii and Terekhova (1970) based on work at the lower portion of an alluvial exposure on the 15–17 m terrace of the upper Kolyma River. They named these sediments the Kubalakhskii Horizon. This horizon consists of coarse-grained alluvial sediments and included skulls of *Bison priscus deminutus*. No paleobotanical analysis was done at the time, but Grichuk et al. (1975) later described the palynological characteristics from undated alluvial sediments from the 10–15 m terrace of the Khatakchan River. The dominant pollen taxa are from arboreal taxa (ca. 52%) of which 15–20% are from tree

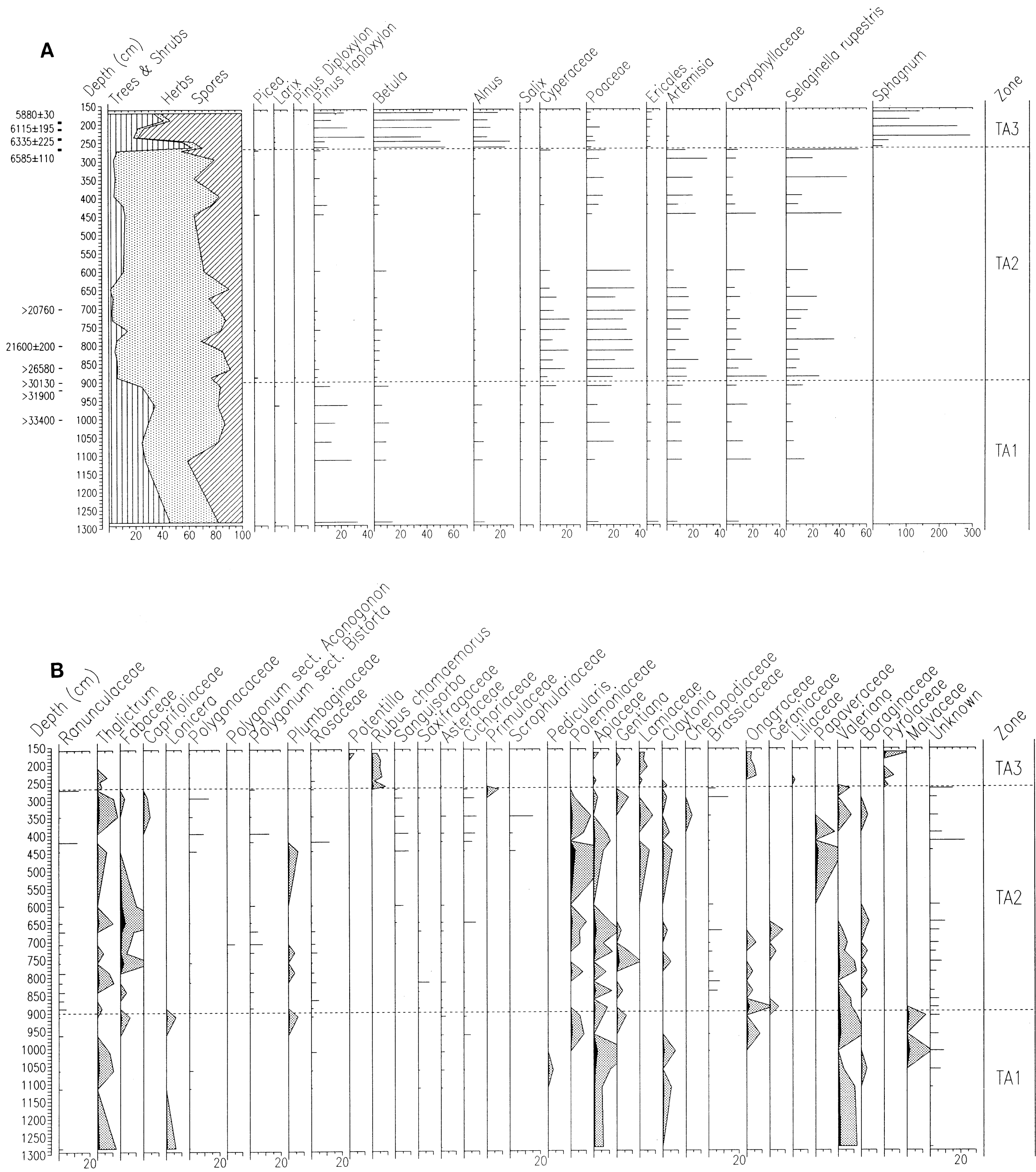

Fig. 2. Percentage diagrams from the Tanon quarry site. These diagrams are a composite of two exposures. Zone TA3 is from local site number 5; zones TA2 and TA1 are from local site number 1. The summary groupings to the left of the diagram are based on a sum of all terrestrial pollen and spores. Percentages of individual taxa are calculated on a sum of all terrestrial pollen. The main pollen and spore taxa are shown in Fig. 2a; Figs. 2b and c illustrate minor herb and spore taxa, respectively. Stippled curves indicate a 10X exaggeration.

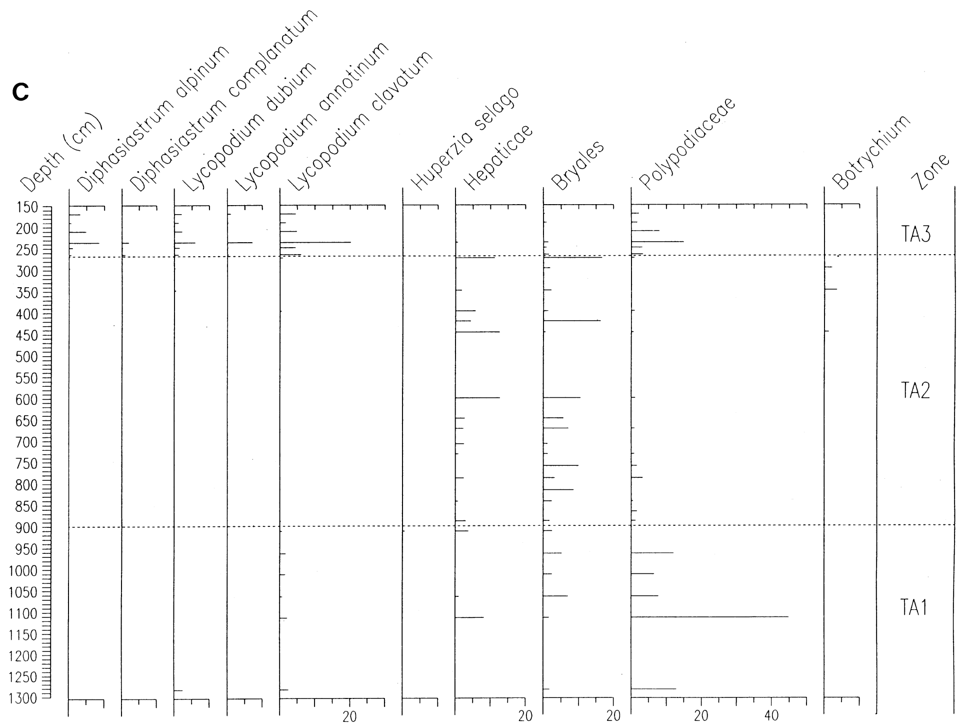

Fig. 2. Continued.

species. Total herb pollen is at its maximum of ca. 30%, but this occurs only in the lower layers of the section. Herb pollen in the upper sediments averages 8–15%. Exotic species in this assemblage vary from 26 to 36% including four to five species of trees (e.g., *Betula platyphylla, Betula* sect. Fruticosae). These spectra are inferred to reflect an altitudinal gradient characteristic of mountainous regions (i.e., low to mid-elevation forest ending in alpine tundra in the highest areas). These MW forests included species typical of the area today (e.g., *Betula exilis, Betula middendorffii, Pinus pumila, Chosenia, Alnus, Larix*). Boggy areas were also present in the valley bottoms. Climate is inferred to be like today.

Macrofossil, paleontological, radiocarbon, and sedimentological analyses of the Kirgirlakh mammoth (site 18) and surrounding deposits added significantly more detail for interpreting MW paleoenvironments (Shilo et al., 1983; Lozhkin, 1991). This famous site is now considered the MW stratotype, and the palynological data (Fig. 3) associated with the baby mammoth (informally called Dima) form the basis for defining the Kirgilyakhskii cool interval (45–43 ka BP). Exposures of three buried river terraces (Shilo et al., 1983), originally draped by colluvium, were uncovered along with the frozen mammoth during mining excavations along the Kirgilyakh River in 1976. The base of the third terrace overlies bedrock (measured at 6 m above river level). The terrace itself is a ca. 10 m thick deposit of paleosols and alluvial gravels with wood remains. The base of the second terrace lies no more than 1 m above river level and the sediments (8 m thick) consist of alluvial and nonalluvial cobbles and gravels, silts, and paleosols. The second terrace includes numerous ice wedges. The first terrace overlies bedrock measured to be 3 m below the current river level. Sediments comprised a 7 m thickness of alternating alluvial sands and gravels. Four paleosols are preserved in this sequence and are associated with fossils of *Larix* found in growth positions.

A suite of radiocarbon dates were obtained from buried wood, organic-rich alluvium, and the mammoth remains in order to define the original sedimentary association of Dima. The mammoth was discovered near the surface of the second terrace, but pieces of fur were found in the third terrace, suggesting that this was its initial resting place. Radiocarbon dates of 44,600 $\pm$ 2000 BP (MAG-378) and 43,500 $\pm$ 1500 BP (MAG-495) were obtained from wood found in coarse alluvial gravels near the bottom of the third terrace. Similar dates were obtained from muscles, meat, and intestines of the mammoth: 40,600 $\pm$ 700 BP (MAG-366A; pieces of skin, muscles, and intestine); 41,000 $\pm$ 900 BP (MAG-576; pieces of skin and large intestine); 41,000 $\pm$ 1100 BP (MAG-366B; pieces of skin and intestines). Similarity in dates and the location of the fur remnants suggest that the mammoth, which had been encapsulated in an ice envelope, had moved from the third to second terrace as the result of solufluction.

Pollen analysis of materials from the stomach and intestines of the mammoth and sediments that included the mammoth body indicates spectra that are dominated by herb pollen (ca. 60–80%), consisting primarily of Cyperaceae and Poaceae pollen (Fig. 3a). Other herb pollen taxa include 28 species that are similar to the modern flora of the upper Kolyma region. The most abundant of these minor types are *Artemisia*, Brassicaceae, and Ranunculaceae. Of the shrub taxa, *Betula* pollen (10–25%) has the highest percentages, with *Alnus* and *Salix* pollen having values of <10 and $<5\%$, respectively. *Larix*, *Pinus*, Ericales, Alismataceae, *Potamogeton*, and *Myriophyllum* pollen occur in trace amounts in all samples. Macrofossil analysis of the digestive tract revealed *Pleurozium* and other Bryales spp., *Selaginella rupestris*, *Carex* sp., *Ranunculus* sp., *Potentilla*, *Mnium affine*, *Aulacomnium*, *Rumex acetosella*, and numerous fungi. These plants all grow today in subarctic and arctic settings.

Pollen analysis of the sediments from the third alluvial terrace delineates two distinctive assemblages (Fig. 3b). Similarity of radiocarbon dates (ca. 45–43 ka BP) and the zone D3a pollen assemblage argue for approximate contemporary deposition of the mammoth and the basal sediments in the third terrace. Herb pollen is dominant, with Cyperaceae being most abundant. Other important herbaceous taxa include Poaceae, *Artemisia*, and Caryophyllaceae. *Salix* and *Betula* are the most common shrub pollen. *Larix* pollen occurs consistently, although *Pinus* pollen remains $<5\%$. *Sphagnum* and *Selaginella rupestris* alternate in dominance of the spore taxa.

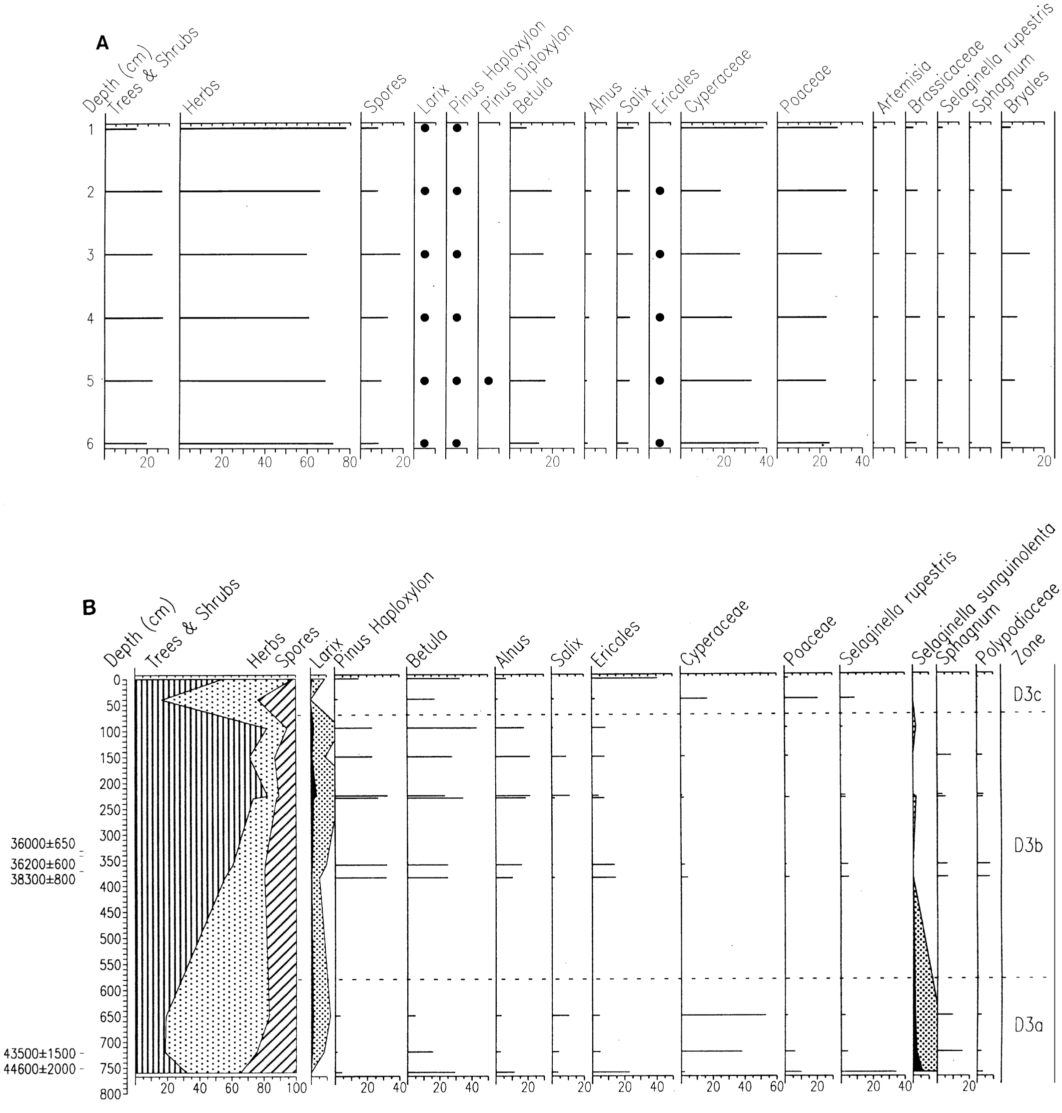

Fig. 3. Percentage diagrams from the Kirgirlakh mammoth site. Percentages are calculated in the same way as described for Fig. 2. Dots indicate trace ($<2\%$) amounts. Stippled curves represent a 10X exaggeration. Diagrams are from: (A) the materials preserved in the mammoth's digestive tract and sediments in the ice deposit that encapsulated the mammoth's body; (B) the exposure from the third alluvial terrace (MW); (C) the exposure from the first alluvial terrace (mid-Holocene to recent); and (D) the exposure from the second alluvial terrace (late MW and Holocene). The sources of spectra illustrated in Fig. 3a are: (1) fur and organic silt which coated the upper side of the mammoth; (2) silt with fur found under the mammoth's body; (3) mammoth fur; (4) mammoth's colon; (5) mammoth's large intestines; and (6) mammoth's stomach.

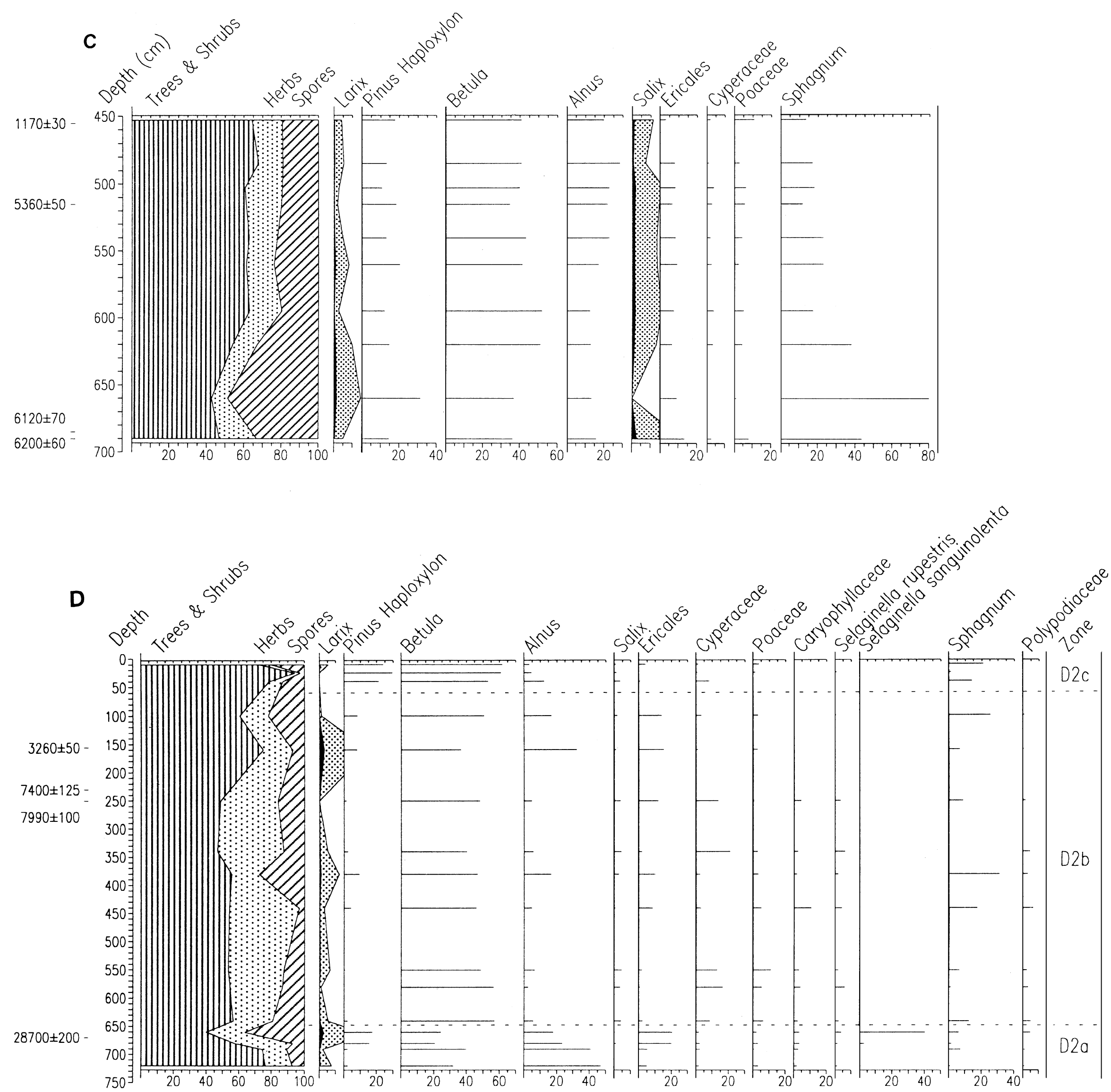

Fig. 3. Continued.

A forest-tundra mosaic probably characterized the vegetation between ca. 45 and 43 ka BP. *Larix* was present on lower mountain slopes and valley bottoms. Dominant shrubs included *Betula* and *Salix* growing in streamside thickets and in the forest understory. *Alnus* shrubs and Ericales may have been present, but if so, they were restricted to only the most favorable sites. Mid-elevation slopes likely supported a *Betula* shrub and herb tundra, depending on exposure and altitude. An altitudinal belt of *Pinus pumila*, so characteristic of the area today, was probably absent. The highest elevations were almost certainly unvegetated or sparsely vegetated. Lozhkin (1991) defined this period as the Kirgilyakhskii cool interval. Summers are interpreted to be cooler than present, although the presence of *Larix* suggests that average July temperatures in the valley bottoms were not less than 10°C. Winters were likely cool and dry, with insufficient snow cover to protect the *Pinus* shrubs from the frigid winter temperatures. Summer conditions may have been relatively moist as indicated by the presence of many Bryales, water plants, and *Salix*.

The pollen spectra of zone D3b (ca. 36–38 ka BP) are comparable to modern and mid-Holocene assemblages from the area (Fig. 3c and 3d). The major differences between zones D3b and D3a are the greater abundances of *Pinus* and *Alnus* shrub pollen and lesser amounts of Cyperaceae pollen in the former zone. *Larix* continues to occur in minor but consistent amounts in zone D3b, and Ericales pollen increases slightly from zone D3a.

Pollen evidence from zone D3b indicates that the vegetation was similar to present. Relatively dense *Larix* forests occupied the valley bottoms and lower slopes.

Pinus pumila, at times in local association with *Betula* and *Alnus* shrubs, probably established at mid-elevations. The relative abundance of Ericales pollen approaches recent values, perhaps indicating an abundance of heaths which are now an important component of the vegetation. These data suggest more moderate conditions as compared to zone D3a, and climate is inferred to be similar to present. Although zone D3b is dated to 36–38 ka, this climatic amelioration probably continued to ca. 33 ka BP with the onset of the Konotzel'skii cool interval (33–30 ka BP).

The second terrace consists primarily of Holocene deposits, but the lower meter has a radiocarbon date of 28,700 $\pm$ 200 BP (wood; MAG-503; Fig. 3d). The associated pollen spectra (zone D2a) show percentages of *Pinus, Betula*, and *Alnus* that are generally similar to the late Holocene (Fig. 3c). These data suggest that the later portion of the MW experienced warm, moist conditions that perhaps were modern or near modern.

As rich as the Kirgirlakh exposures are, they provide only a discontinuous view of MW environmental change. In contrast, the Elikchan 4 Lake (site 19) core encompasses a continuous record of vegetation change since the EW (Lozhkin et al., 1995; Lozhkin and Anderson, 1996). The Elikchan Lakes, a series of four interconnected basins, occupy the base of a narrow tectonic valley near the divide between the Kolyma and Okhotsk Sea drainages. Elikchan 4, the largest of the lakes (ca. 3.9 km long $\times$ 0.9–1.3 km wide), drains southward to the Ola River and is thus part of the Okhotsk drainage. However, we include this site within the upper Kolyma region, because its mountainous character and modern vegetation are more similar to sites within this region than sites from Priokhot'ye.

Silts and organic-rich silts comprise the 9.4 m long core. The pollen record indicates two times of severe conditions (LW and EW stades), characterized by a herb-*Salix* tundra, and two times of warmer climates (Holocene interglacial and MW interstade), indicated by the presence of light coniferous forest (Fig. 4a). Poaceae, Cyperaceae, and *Artemisia* pollen often with high percentages of *Selaginella rupestris* spores and a variety of minor herb taxa dominate the tundra assemblages (zones EL1, EL3). The warm periods (zones EL2, EL5) are indicated by spectra where pollen of *Pinus*, *Betula*, and *Alnus* are dominant, and *Larix* pollen occurs in consistent, albeit minor, amounts. A brief transitional period (zone EL 4) is dominated by *Betula* and *Alnus* pollen.

Zone EL2 is assigned to the MW (Fig. 4b). The most noticeable characteristic of this zone is the marked fluctuation in *Pinus* pollen percentages which at times are similar to recent pollen spectra (e.g., zone EL2c) and at other times occur in modest amounts (e.g., zone EL2a3). Four intervals have *Pinus* pollen of less than 20% (zones EL2a2/a3, EL2b, EL2d, EL2f) and three intervals have greater percentages (zones EL2a1, EL2c, EL2e). The periods with low *Pinus* pollen also tend to have higher percentages of Cyperaceae, Poaceae, and *Artemisia* pollen. However, of the proposed warm intervals only zone EL2c has an assemblage that is similar to the late Holocene (i.e., modern) spectra. Variations in the percentages of *Pinus* pollen and in the sum of pollen from trees and shrubs suggest important changes in the vegetation during the MW. Although there are likenesses to the modern pollen spectra, the moderate amounts of Cyperaceae, Poaceae, and *Artemisia* suggest that the MW forests differed somewhat from present. Except for zone EL2f and perhaps zone EL2b, where *Larix* pollen is absent, the MW vegetation can be generally characterized as a *Larix* forest with an understory of *Pinus pumila, Betula*, and *Alnus*. The greater abundance of herb pollen, as compared to the Holocene, may suggest a more open forest or woodland and/or greater numbers of herbaceous species occupying the bordering mountain slopes. The absence of *Larix* and increase in Poaceae (and *Artemisia* in zone EL2b) pollen in zones EL2f and EL2b may indicate the presence of a shrub tundra or a very sparsely forested valley bottom. Zone EL2b especially suggests a near-stadial type of environment with the reduction in shrub pollen and increase in herb pollen. The earliest portion of the MW (zone EL2a) seems to be characterized by a changing environment, although *Larix* trees are likely to be present throughout this zone. *Pinus* percentages are sufficiently low in zone EL2a2/a3 as to suggest that the shrubs were absent. If so, the vegetation was likely a *Betula–Alnus* shrub tundra at mid-elevations and cooler valley sites, with *Larix* growing in the more favorable areas of the lowlands. This *Larix–Betula–Alnus* association is also typical of the late glacial-Holocene transition in the upper Kolyma region.

These fluctuations in *Pinus* and sum of trees and shrub pollen and the presence/absence of *Larix* pollen contrast to the more uniform nature of the Holocene pollen assemblage and suggest that the MW was characterized by several cool (lower *Pinus*/sum trees and shrubs; absence of *Larix*) and warm (higher *Pinus*/sum trees and shrubs; presence of *Larix*) events. During the warm periods, *Pinus pumila* was able to grow to higher elevations on the mountain slopes and probably formed denser thickets due to the warmer summer temperatures. Snow depths may have been greater during these times thereby providing the needed protection against winter desiccation. During cooler periods, summer temperatures may have decreased and/or winters become drier, resulting in a die-back of the shrubs at higher elevations. Although dating control is poor for this part of the Elikchan core, the sequence of these climate fluctuations is similar to that proposed by both Kind (1974) and Lozhkin (1991). That is, zones EL2f and EL2e probably represent the terminal MW warm period, suggested by Lozhkin, with near-modern conditions. Zone EL2f is likely to be cooler than zone EL2e because *Larix* pollen is absent and *Pinus*

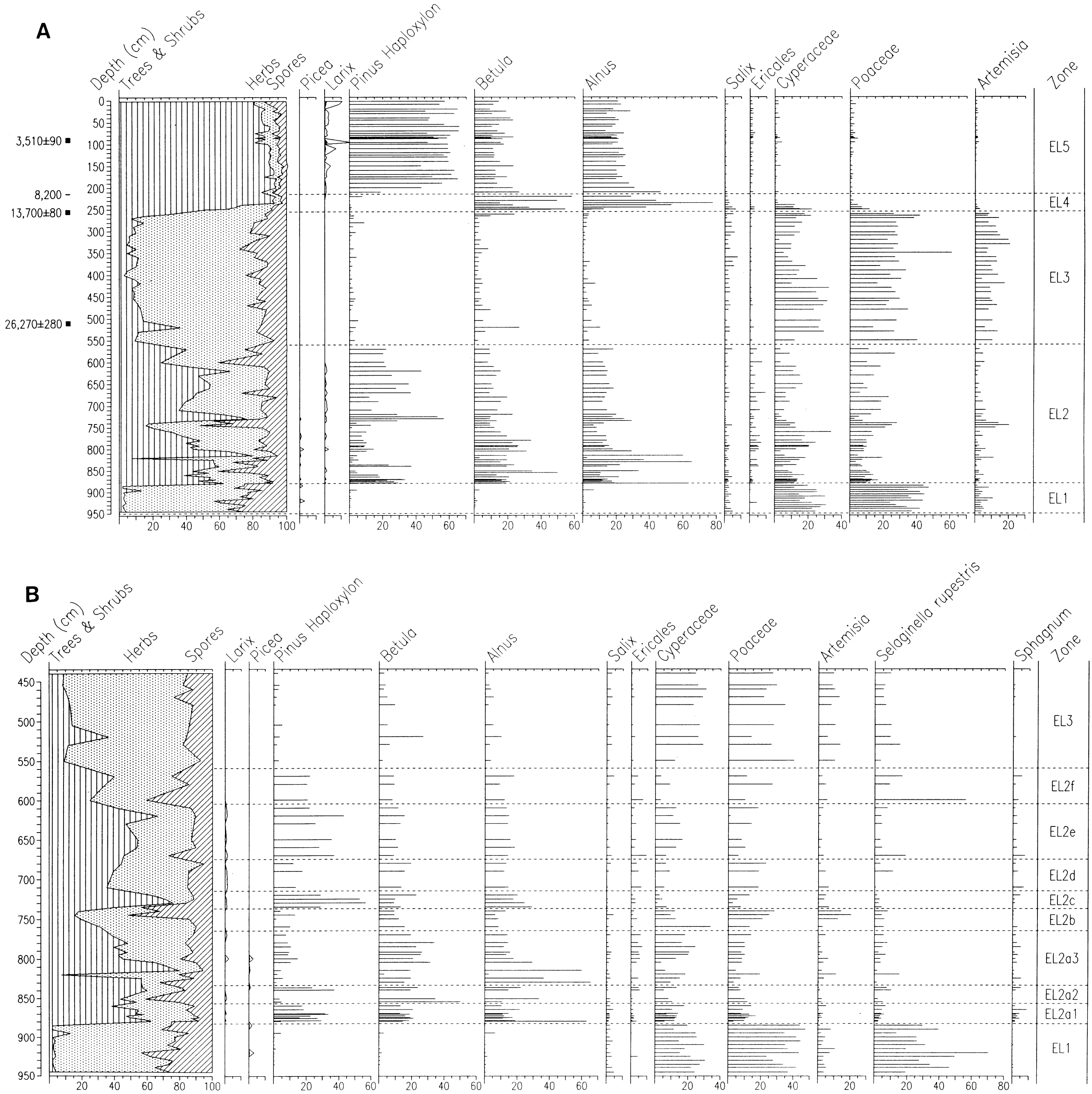

Fig. 4. Percentage diagrams from Elikchan 4 Lake. Percentages are calculated in the same way as described for Fig. 2. Summary diagrams are: (a) main taxa for the entire core; and (b) main taxa for the MW portion of the core.

pollen is reduced. However, the nearly 20% *Pinus* pollen suggests that climates were still warmer and moister than during the LW. We tentatively assign these two zones an age of 30–26 ka BP, based on comparison to other sites from central WB. A cooling is indicated in zone EL2d (ca. 33–30 ka BP) preceded by a period that probably represents the climatic optimum for the MW (zone EL2c; 39–33 ka BP). Zone EL2b (45–39 ka BP) suggests another cooling event. The earliest times of the MW seemingly are characterized by climatic fluctuations that generally have a warm signal (zone EL2a).

4.1.5. Summary — Western Beringia

All WB sites of possible MW age indicate that the interstadial environments differed markedly from the preceding and subsequent stadial conditions. Unlike glacial times, when a herb-*Salix* or herb tundra was the dominant vegetation, the MW is characterized by periods of widespread occurrence of *Larix* forests which approximated the modern distribution (Table 2). Even with questions about the chronologies, sites with continuous or semi-continuous records document major fluctuations in the MW vegetation and climate. The core from Elikchan 4 Lake, perhaps the strongest MW record from WB, indicates that the MW was characterized by one period of near modern conditions; two times when climate was greatly cooler than present though warmer than stadial times; a moderately warm period that approaches modern conditions; and two intervals characterized by varying but generally moderate conditions (although still cooler and drier than present). Although temporal correlations are uncertain, the Elikchan data are consistent with the previous paleoclimatic schemes presented by Kind (1974) for Siberia and by Lozhkin (1991) for the upper Kolyma and Indigirka–Yana–Kolyma lowlands. In contrast to records from western and central WB, available data from Chukotka suggest lesser vegetational and climatic variations within the MW. However, the spatial distribution of sites is limited and temporal control is poor for this region. The available evidence from Priokhot'ye also indicates little change in the MW vegetation, at least between ca. 37 and 24 ka BP. However, like Chukotka, the number and quality of sites are limited.

Despite concerns about the paleoenvironmental data, some general patterns are indicated. From ca. 45–39 ka BP, open *Larix* forests occupied protected valley bottoms in the upper Kolyma region, with an altitudinal treeline that was lower than present. Climates that were cooler than present (though warmer than during the LW or EW) prevailed in this mountainous region. Between ca. 39 and 33 ka BP, *Larix* forests occupied higher elevations and seem to have a modern regional distribution and composition. Kind (1974) thought this may be the period of maximum warmth for the MW, an interpretation that is also consistent with the Elikchan 4 record. This warm period is followed by a time of cool, dry climates, as evidenced by the appearance of herb- or *Betula*-dominated tundras in areas that previously were forested and by the occurrence of active ice-wedge formation throughout much of WB. Conditions perhaps were quite severe in northern WB, as suggested by the presence of tundra steppe. The age of this period varies from area to area, but the Kirgirlakh mammoth site provides the best dating control, indicating a climatic deterioration corresponding to ca. 33 to 30 ka BP. The latest portion of the MW seems to be characterized by the widespread establishment of *Larix*, with forested areas in the lowlands of the Yana, Indigirka, and Kolyma rivers and in the coastal areas of Priokhot'ye. *Larix* distribution was probably more restricted in the mountainous interior, where it likely survived on only the most protected low elevation sites. These paleovegetational data suggest that the MW ended under relatively warm conditions. Lozhkin (1991) suggested that this interval represented the MW climatic optimum, based on data from the upper Kolyma and Yana–Indigirka–Kolyma regions. The Elikchan data, however, do not support this hypothesis, at least for the latest part of the MW. Rather, the core suggests that conditions were slightly cooler than present but still warmer than stadial times.

4.2. Eastern Beringian sites

Prior to the 1980s, interpretations of MW (Boutellier interval) paleoenvironments were based on a handful of sites located in the St. Elias Range, Old Crow Basin, the Fairbanks area, Seward Peninsula, and the middle Kobuk River (Colinvaux, 1964; Denton and Stuiver, 1967; Matthews, 1974a, b; Péwé, 1975a, b; Morlan, 1980; Schweger and Janssens, 1980; Hopkins, 1982; Schweger, 1982; Shackleton, 1982). All sites, with the exception of Imuruk Lake on Seward Peninsula, provided only discontinuous MW records, but together they indicated that landscapes differed significantly from those of stadial times. Widespread peat growth, ice-wedge collapse, and relatively warm soils generally characterized the interstade. Distributions of dated plant macrofossils (Hopkins

Table 2
Summary of Mid-Wisconsinan vegetation patterns, Western Beringia

Yana–Indigirka–Kolyma Lowland	Chukotka	Priokhot'ye	Upper Kolyma–Upper Indigirka
Larix-tree *Betula* forest-tundra mosaic (28–24.5 ka BP)	South: *Salix* (*Betula*, *Alnus*?) high shrub tundra (42–27 ka BP; perhaps limited stands of *Betula* trees	*Larix* forest with *Betula, Alnus*, and *Picea* trees (37–24 ka BP?)	Mosaic of *Larix* forest and high shrub tundra (30–25 ka BP)
	North: tundra steppe (30–29 ka BP)		
tundra steppe (28–24.5 ka BP)	North: *Salix*-herb tundra (33–32 ka BP)		Herb or shrub *Betual* tundra (33–30 ka BP)
Larix-tree *Betula* forest (48–34 ka BP)			*Larix* forest (39–33 ka BP)
			Shrub tundra with limited *Larix* growth (45–39 ka BP)

et al., 1981) suggested that *Picea, Larix*, tree *Betula*, and *Populus* "may have extended to near their present geographic limits as late as 35,000 years ago, [but] they were certainly much sparser than at present" (Hopkins, 1982, p. 8). Thus, the sedimentological and paleobotanical evidence suggested that the interstade was a period of relatively warm, moist climate, but conditions still remained more severe than during the Holocene.

Although some records had independent chronological control, many of these first sites were assigned to the MW based on their stratigraphic relationship to the Old Crow tephra and the presence of "warm" proxy data. This tephra, which initially was believed to be of early MW to EW age (60–80 ka BP; Westgate et al., 1983), is now thought to date to 140,000 $\pm$ 10,000 BP (Westgate et al., 1990). Consequently, former MW sites (e.g., zone i at Imuruk Lake, the Eva formation) have now been assigned to the previous interglaciation (i.e., isotope stage 5 and often stage 5e; Hamilton and Brigham-Grette, 1991; Schweger and Matthews, 1991). More recently, research has focused on lacustrine sediments that span the Holocene, LW, and all or part of the MW (Cwynar, 1982; Anderson, 1985, 1988; Nakao and Ager, 1985; Eisner and Colinvaux, 1990; Anderson et al., 1994). These sequences also suffer from dating problems. However, they do provide continuous records of change and better chronological control for intersite comparisons. Unfortunately, most of these sites are located in the northern sectors of EB and thus do not provide complete geographic coverage. However, continued research on non-lacustrine sediments also offers new information from other areas of EB (e.g., Ager, 1989; Elias et al., 1996; Lea et al., 1991; Begét, 1990).

An equivalent stratigraphic scheme described for WB (Shilo et al., 1987) is absent in EB, because details of MW paleoenvironmental changes are more poorly defined, partially due to of the dating issues described above. Thus, the following summary draws only from key alluvial, lacustrine, paleosol, and organic-rich/peat deposits that are radiometrically dated to the MW. For purposes of discussion, EB has been divided into the following four areas (Fig. 1): Yukon Territory (Region E), southern Alaska (Region F), interior Alaska (Region G), and northwestern Alaska (Region H). See Fig. 1 for individual site locations.

4.2.1. Yukon territory (region E)

The Yukon Territory occupies the northwestern sector of Canada, with MW sites located within two sub-regions. The southern Yukon Territory includes the mountainous areas of the upper reaches of the Yukon drainage. The northern Yukon Territory is dominated by the Old Crow, Bluefish, and Bell structural basins which are crosscut by the upper Porcupine drainage, and the Keefe, Olgivie, Richardson, British, and Barn Mountains. The entire Yukon Territory is a mosaic of *Picea* forest and tundra, reflecting variations in elevation and soil conditions. *Picea glauca* and *Picea mariana* are the primary conifer species, whereas hardwoods include *Betula papyrifera, Populus balsamifera*, and *Populus tremuloides*. Shrub tundra that includes *Betula glandulosa, Alnus crispa, Salix* ssp. and/or Ericales species occupies areas at mid- to high elevations or lowland sites unsuitable for tree growth. Highest elevations support alpine tundra or fell fields.

The MW interstade in EB originally was defined from bluff exposures along Silver Creek (site 20), St. Elias Range, southwestern Yukon Territory (Denton and Stuiver, 1967; Denton, 1974). Coarse-grained alluvial sediments with organic-rich silts lie between till of the LW Kluane glaciation and till and outwash of the EW Icefield glaciation. Radiocarbon dates ranging from $>49{,}000$ BP (Y-148) to 29,600 $\pm$ 460 (GSC-769) were obtained from organic detritus and wood fragments. Paleobotanical analyses of fine-grained pond deposits revealed pollen spectra dominated by Cyperaceae (up to 85%), Poaceae (up to 20%), and *Artemisia* (up to 20%) and macrofossils of the mosses *Calliergon gigateum, Drepanocladus brevifolias*, and *Scorpidium scorpioides* (Schweger and Janssens, 1980). These data indicate tundra with locally abundant *Salix* shrubs and scattered areas of immature minerotrophic soils, the latter reflecting the recently deglaciated terrain. Because *Picea* stands dot the modern landscape, climate is inferred to be colder and probably drier than present.

Antifreeze Pond (Rampton, 1971, site 21), currently located within the *Picea* forest, provides a continuous record of vegetation change over the last ca. 32 ka BP. Unfortunately, the lower portion of the core suffers from several dating reversals, but the data suggest that the area near the Alaska–Yukon border was tundra between ca. 27 and 31 ka BP. The basal pollen spectra contain *Picea, Populus, Alnus*, Cyperaceae, Poaceae, and *Artemisia*, perhaps indicating the presence of a *Picea* woodland sometime before ca. 31 ka BP.

At the Mayo Indian Village site (site 22), located along the Stewart River, ca. 5 m of an organic-rich alluvial silt with lenses of organic detritus is covered by tills of the LW McConnell glaciation (Matthews et al., 1990). These sediments were deposited as overbank deposits or infill of an abandoned stream channel. A date of 29,640 $\pm$ 260 BP (TO-292) was obtained from *Corispermum* seeds found in the alluvium. Pollen spectra are dominated by graminoids (ca. 20–25%) and *Artemisia* (30–45%). In the upper 4 m, *Picea* pollen appears consistently, with values of ca. 5–10%. Plant remains in the alluvial detritus include: seeds of aquatic taxa of Potamogetonaceae, Haloragaceae and Najadaceae, *Carex* seeds, bud scale and capsule from *Salix*, seeds of *Betula glandulosa*, and seeds from various forbs (*Oxyria digyna*, *Rumex*, Chenopodiaceae, Caryophyllaceae, *Ranunculus* spp., *Papaver*, *Draba*, and *Potentilla*). Insect remains contain some species that typically occur beyond either altitudinal or latitudinal treeline (e.g., *Elaphrus parviceps*, *Amara glacialis*, and *Pterostichus caribou*). Other types are limited to low arctic settings (e.g., *Notaris*, *Lepidophorus lineaticollis*, and *Ceratomegilla ulkei*). Many additional beetle species are restricted to dry, poorly vegetated riverbanks.

The combined paleoenvironmental data suggest that a treeless environment existed in the Mayo lowlands ca. 30 ka BP. *Betula* and *Alnus* shrubs may have been present, but if so, they were extremely rare. The lack of *Picea* macrofossils indicates that trees were absent along river courses or existed as small pockets in only the most protected areas. If the latter, treeline was lowered by ca. 850 m. Climate was probably no colder than that of modern low arctic tundra. Preliminary analyses of organic materials from a nearby exposure, dated to 38,100 $\pm$ 1330 BP (GSC-4554), suggest that *Picea* may have been more common ca. 38 ka BP as indicated by a horizon containing erect *Salix* stumps, *Picea* needles, and relatively high percentages of *Picea* pollen.

The northern portion of the Yukon Territory contains four areas with MW records: the Old Crow (area l), Bluefish (area m), and Bell (area k) Basins and Hanging Lake (site 23). The Hungry Creek site (Hughes et al., 1981) in the Bonnet Plume Basin was originally thought to be of interstadial age. However, Schweger and Matthews (1991) argue on the basis of the paleobotanical data that this site is more likely of the last interglaciation and that the radiocarbon date is in error. Thus, it is not included in our discussion.

The Old Crow Basin contains some of the most important stratigraphic sections for inferring late Quaternary paleoenvironments in EB (reviewed in Schweger, 1989; Matthews et al., 1987). Bluefish Basin also contains MW sites; their records are similar to those from Old Crow and thus will not be described here. The upper portions of the Old Crow sections are characterized by thick lacustrine clays deposited between ca. 30 and 12.5 ka BP, when the northern Yukon River drainage and glacial meltwaters were funneled into the structural basins of the northern Yukon Territory. Beneath this unit is an organic-rich alluvium that has yielded only infinite radiocarbon dates. However, unpublished sections from localities 12 and 15 (reported in summary form in Schweger and Matthews, 1991) have provided the first dated materials from the MW: 31,400 $\pm$ 660 BP (GSC-2739) and 35,500 $\pm$ 1050 BP (GSC-2507) from allochonous peats at locality 12 and 38,880 $\pm$ 2000 BP (GSC-2756) and 41,100 $\pm$ 1650 BP (GSC-2574) from wood fragments at locality 15.

The pollen analysis is preliminary, but samples from locality 12 suggest local variations in vegetation and climate during the last ca. 41 ka BP. The pollen assemblage for 41.1 ka BP contains ca. 10% *Betula*, 10% *Salix*, 5% *Picea*, 3% *Alnus*, and 40% Cyperaceae and is inferred to represent either an open *Picea* woodland or a *Betula*-dominated shrub tundra with isolated populations of *Picea*. By ca. 38.8 ka BP, *Betula* percentages have increased to ca. 70% with 15% Cyperaceae and $< 5\%$ *Picea* and *Alnus* pollen. Although Schweger suggests these spectra represent a vegetation like that at 41.1 ka BP, the extremely high percentages of *Betula* pollen would argue more strongly in favor of a *Betula* shrub tundra. The 35.5 ka BP pollen assemblage is similar to that at 41.1 ka BP, documenting either the continued presence of *Picea* or the expansion of *Picea* following a period of *Betula* shrub tundra. The ca. 31.4 ka BP spectrum, with ca. 8% *Salix*, 65% Cyperaceae, 20% Poaceae, and 7% *Artemisia* pollen, indicate the presence of a *Salix*-graminoid tundra. These palynological data suggest fluctuations in MW climates with a relatively warm interval (ca. 38.8–35.5 ka BP) marked by the presence of *Picea*, with more moderate conditions associated with *Betula* shrub tundra (ca. 41 and 39 ka BP) and harshest conditions with herb-dominated tundra (ca. 31.4 ka BP).

An unpublished site, reported in Schweger and Matthews (1991) from the Rock River, Bell Basin, contains ca. 5 m thick horizon of organic silts and in situ peats. These sediments are overlain by a sequence of alternating ripple-marked sands and lacustrine clays indicative of the LW flooding of the Bell, Bluefish, and Old Crow Basins. Radiocarbon dates of 34,220 $\pm$ 170 BP (TO-124) at 5.1 m above river level and $> 43{,}000$ BP (GSC-2585) at ca. 3.75 m were obtained from fecal pellets and a *Salix* wood fragment, respectively, preserved in the organic-rich deposits. The entire section shows abundant, but fluctuating, pollen percentages of *Betula* (ca. 10–55%), Cyperaceae (ca. 5–55%), and Ericales (ca. 2–30%). The lower meter contains only trace amounts of Poaceae and *Artemisia*, but their values increase up core to ca. 25 and 15%, respectively (although between ca. 2 and 3.5 m, measured from base, *Artemisia* is absent). *Alnus* pollen has maximum percentages of ca. 25% in the basal meter, with up-section values between ca. 3 and 10%. *Picea* pollen also varies with three peaks of $> 10\%$ at 0, 3, and 5.1 m. Although unclear from the description, these variations presumably do not reflect changes in sediment type. If true, this pollen diagram suggests vegetation and climatic fluctuations through the MW. An open *Picea* woodland (or perhaps a shrub tundra with isolated *Picea* groves), associated with relatively mild interstadial climates, perhaps was present two times prior to and at ca. 34 ka BP. Shrub tundra dominated by either *Betula*, *Betula*-*Alnus*, *Salix*, or *Betula*–Ericales seems to have been present throughout this interval at least on some areas of the landscape, suggesting conditions that were cooler than present, but not as severe as during the LW.

Hanging Lake (site 23) is the best dated continuous record of vegetation change in the Yukon Territory (Cwynar, 1982; Ritchie and Cwynar, 1982). The site is located in the rolling uplands to the northeast of Old Crow Basin. The ca. 4 m long core has 21 radiocarbon dates, giving a calculated basal age of 30 ka BP. The basal pollen zone, which is assigned an age of 18.5–30 ka BP, is dominated by Poaceae (20–40%) and *Artemisia* (20–50%). Cyperaceae, *Betula*, and *Salix* pollen generally are less than 10% in this zone. Total pollen accumulation rates are low, varying between 5 and 50 grains $cm^{-2}\,yr^{-1}$. The LW and MW are not differentiated based on the pollen spectra. The vegetation is inferred to be a depauperate tundra perhaps similar to modern fell field. Such vegetation indicates climates that were more severe than present.

4.2.2. Southern Alaska (region F)

Southern Alaska includes two separate regions: an eastern area centered on the Copper River basin (area o) and a western area in the Nushagak lowlands (area n). The Copper River basin is surrounded by mountains of the Alaska Range and the Talkeetna, Wrangell, and Chugach Mountains. A mosaic of *Picea mariana* and *Picea glauca* forests and treeless bogs characterizes the vegetation in the lowland. Shrub tundra is more restricted than in other regions of Alaska, occupying only the higher elevations or sites with poor soils. The Nushagak lowland is bordered to the north and west by the Kilbuck and Kuskokwim Mountains and to the east and south by the Aleutian Range. The lowland lies at the southwestern border of the *Picea* forest, but isolated stands of *Picea glauca, Populus balsamifera*, and *Betula papyrifera* occur in favorable sites. The predominant vegetation is mesic to wet shrub tundra with large areas of Cyperaceae–Poaceae meadows. Shrubs, which can occur in both high and low growth form, include *Betula glandulosa, Alnus crispa, Salix* spp. , and Ericales.

Data from this region, although sparse, are beginning to accumulate, especially in the Nushagak and Holitna Lowlands. Deposits in the latter area remain rather ambiguous in their dating (Waythomas, 1996). Therefore, we summarize results only from the Nushagak area.

Some of the earliest work in the Nushagak Lowlands was done at a 12 m high outcrop on Kvichak Peninsula bordering Nushagak Bay (Ager, 1982). Here glacial tills are overlain by glaciofluvial sands, aeolian silts, and peats. Ice-wedge pseudomorphs and cryoturbation structures are sometimes present in the nonorganic deposits. Two radiocarbon dates of $>33{,}000$ BP (AU-83 and AU-84) were obtained between 7 and 8 m height within the section. Peats at ca. 10 and 11 m heights were dated to $12{,}760 \pm 300$ BP (AU-82) and 7600 ± 100 BP (AU-81), respectively. Pollen analysis of samples pre-dating 12.7 ka BP are dominated by Cyperaceae (ca. 30–60%) and Poaceae (ca. 10–40%) pollen. Percentages of *Salix* and *Betula* are generally $< 10\%$, but spectra associated with samples taken from levels with infinite radiocarbon dates show slightly higher percentages. Ager suggested that these levels are likely MW in age and perhaps indicate slight increases in the shrub component in the tundra vegetation. Although climate may have ameliorated slightly from stadial times, the pollen evidence suggests that the MW conditions were still colder and drier than present.

More recent work in the bluffs bordering Kvichak and Nushagak Bays revealed two nonglacial units deposited under a LW aeolian sand and silt (Lea et al., 1991). The upper unit, named the Etolin Complex, has nine radiocarbon dates of which only one is finite ($35{,}600 \pm 1500$ BP (W-3576)), although its large standard deviation suggests it be treated as a nonfinite or minimum age limit. Preliminary conclusions are that this complex is MW (probably early MW) in age. The Etolin Complex itself is characterized by two organic units. The lower unit (ca. 0.1–0.35 m thick) is a detrital peat with intercalated sand lenses and a discontinuous layer of white tephra (not identified at time of publication but possibly may be the Old Crow tephra). The upper organic unit, separated from the lower unit by ca. 0.2–2.5 m of moderately sorted sandy silt, is a ca. 0.2–1.5 m thick sandy silt that includes thin peat lenses and humified organic material. These sediments are inferred to be from small, shallow ponds and colluvium deposited over swales and lower slopes.

Poaceae (25–64%) and Cyperaceae (27–55%) dominate the pollen assemblage with more moderate abundances of *Artemisia* (up to 18%) and *Salix* (up to 17%). The beetle assemblage includes high percentages of the *Cryobius* group with the Omaliinae, Hygrophilous, and *Stenus* groups varying from moderate to low representation. No boreal species were present. Faunal and floral remains suggest a cold, mesic to wet graminoid tundra where local *Salix* thickets were present. Conditions were cooler than present. The tentative correlation of the Holitna and Etolin complexes is mentioned here but the dating remains somewhat unclear for the Holitna lowland deposits (Waythomas, 1996). The nonglacial sediments from Holitna indicate mesic-to-wet tundra and climatic conditions intermediate between those of full-glacial and interglacial intervals.

A number of deposits possibly of MW age have been reported from the Copper River Basin, an area that today supports a *Picea*-dominated boreal forest (see Ager, 1989). However, as in other areas, radiometric dating does not permit unambiguous assignment of many sites to interstadial as opposed to interglacial times. The strongest data come from a peaty fine sand and silt unit exposed along the Sanford River. Radiocarbon dates of $28{,}300 \pm 1000$ BP (W-1343) and $31{,}300 \pm 1000$ BP (W-843) bracket the deposit (Ferrians et al., 1983; Williams and Galloway, 1986). The pollen assemblage is dominated by Cyperaceae, Poaceae, Caryophyllaceae, and *Artemisia* which are inferred to represent alpine tundra. This interpretation is consistent with plant macrofossil data from Tyrone Bluff, ca. 125 km to the west where an alluvium that separates glacial till has been dated between 31,070 + 860/-960 (DIC-1862) and 21,730 ± 390 BP (DIC-1861; Thorson et al., 1981). Another horizon, lying between till and gravel units along the Nelchina River, has yielded an herb-dominated pollen assemblage that Ager (1989) interprets as a moist Cyperaceae meadow occupying the Copper River Basin lowland, possibly during early parts of the MW. A > 38,000 BP (W-842) date was obtained from these deposits. At a nearby outcrop, another > 38,000 BP (W-295) date was obtained from peaty material that lay stratigraphically below the tundra horizon. Pollen analysis of this material indicates the presence of boreal forest similar to modern. Other assemblages from the Basin are also dominated by *Picea* (e.g., Tyrone Creek, Dadina River), but it is uncertain whether the forest was associated with an early MW warm period or dates to the last interglaciation. Ager (1989) argued that a treeless vegetation most likely characterized all or most of the MW interstade, whereas a boreal association more likely occurred during interglacial times.

4.2.3. *Interior Alaska* (*region G*)

Interior Alaska encompasses the central portions of the Tanana–Kuskokwim lowland and the Yukon–Tanana upland, located to the south and north of Fairbanks, respectively. Closed *Picea mariana* forests or muskegs occupy much of the lowland. *Larix laricina* can be locally abundant on poorly drained soils. *Betula papyrifera, Populus tremuloides*, and *Populus balsamifera* are typical on sites disturbed by fire or floods. Above 760 m, the forests are discontinuous and occupy only the most favorable microhabitats. Shrub heath or high *Betula* shrub tundra characterize areas beyond treeline.

This region was one of the first to yield ancient records and thus provided the data to form some of the first reconstructions of Wisconsinan environments. Originally, some of the organic-rich deposits (e.g., the Isabella basin and adjacent areas; Matthews, 1974b) were assigned to the MW, leading to the belief that interior Alaska was extensively forested during the interstade. However, subsequent dating and identification of the OCT have caused the MW age assignments of many of these sites to be questioned (Hamilton and Brigham-Grette, 1991).

Matthews' (1974b) detailed analysis of the 27 m long core near Isabella Creek (site 24) suffers from dating problems common to much of the earliest research into interstadial environments. Of most interest here is Zone A, which is an organic-rich silt with peaty inclusions. Matthews originally assigned this zone a ca. 35–32 ka BP age, based on radiocarbon dates of > 31,900 (I-4775) and 34,900 ± 2950 (I-3083) obtained from near the top and bottom of this zone, respectively. However, Hamilton et al. (1988a) subsequently proposed that 34.9 ka BP should be considered a minimum age because of the large counting error. The > 31.9 ka BP date which is located ca. 7 m up-section from the 34.9 ka BP date also lends support to this interpretation.

Zone A is divided into two subzones. Subzone Aa, the lower of the two, is characterized by spectra with generally less than 10% *Picea, Betula*, and *Alnus*, 10–40% Cyperaceae, 10–20% Poaceae, and 5–20% *Artemisia* pollen. *Larix* pollen is present in only one sample. Although the pollen assemblage is similar to that at Antifreeze Pond (interpreted by Rampton (1971) as graminoid-moss tundra), plant macrofossils include *Picea, Rubus idaeus*, and *Discus cronkhitei*, which are not indicative of tundra. Matthews (1974b, p. 836) interpreted the vegetation as an "open sedge dominated, environment with scattered spruces and practically no alders." *Picea* was probably restricted to only the most favorable sites with south-facing exposures. Matthews postulated that this zone is equivalent to the period of ice-wedge growth (ca. 33/34–30 ka BP) at the Fox permafrost tunnel (see below).

Subzone Ab has higher *Picea* pollen percentages (up to 25%) than in subzone Aa and includes ca. 20–30% *Alnus* and 15–30% *Betula* pollen. Graminoid pollen is generally less than 10%. Traces of *Larix* pollen occur throughout the zone. Matthews viewed these spectra as similar to modern, suggesting the presence of boreal vegetation. Unlike subzone Aa, *Alnus* shrubs were probably abundant, and both shrub and tree *Betula* (*B. papyrifera*) grew near the site. Altitudinal treeline was undoubtedly higher than during Zone Aa, but perhaps not as high as present. Macrofossil evidence indicates the nearby presence of a pond or lake and a general increase in moisture from the bottom to the top of the subzone. Matthews suggested this is a general time of warming accompanied by thawing of the permafrost.

The Fox Permafrost tunnel (site 25) has been excavated through late Quaternary sediments of the Goldstream valley. It is an exceptionally rich site that has provided detailed information on changing biota and geocryological conditions over more than 40 ka years (Hamilton et al., 1988a). Fluvial gravels are exposed at the base of the tunnel. This deposit is overlain by two silt units separated by an unconformity marked by lenses of buried organic material and ground-ice features. The lower unit is a ca. 4.5 m thick well-sorted coarse silt. A basal layer ca. 1–1.5 m thick consists of organic-rich sediments with peat lenses and wood fragments (up to 4 cm in diameter). The lower 1 m thick silt also contains large ice-wedges whose tops are truncated by thaw processes. The upper silt unit is 8–11 m thick and consists of well-sorted medium to coarse silt, grading into more organic-rich silts. The upper silt unit includes epigenetic ice-wedges that probably formed during the LW.

The basal gravels exposed in the tunnel were part of a broader drainage system (e.g., a braided stream) that eventually shifted course leaving behind finer grained overbank deposits. The streamside was populated by *Salix* thickets. The floodplain was eventually covered by aeolian loess, perhaps beginning ca. 43 BP. Initially, this deposition was slow, as indicated by inclusions of autochthonous peats. After 43 ka BP until ca. 39 or 38 ka BP, vegetation was probably a shrub or herb–shrub tundra that included *Betula* and *Alnus*. Sedimentation rates then increased under seemingly sparsely vegetated conditions dominated by xeric habitats. This rapid deposition continued until ca. 34 or 33 ka BP. Permafrost aggradation accompanied the loess accretion. A period of slow loess deposition followed (ca. 33-30 ka BP) associated with dramatic ice-wedge growth. The active layer likely deepened and thermokarst activity was more pronounced. Macrofossil and pollen data imply a moister environment than previously and one that may be somewhat analogous to that found in modern polygon systems. Loess began to accumulate again after 30 ka BP, eventually replacing the *Salix*–graminoid tundra with more xeric communities of the LW.

A 6.5 m long core from Harding Lake in the Tanana valley (site 26; Ager, 1983; Nakao and Ager, 1985; Ager and Brubaker, 1985) is estimated to be ca. 30 ka years old based on six radiocarbon dates. However, the lower three dates contain a reversal (13,690 $\pm$ 500 BP from 4.03 to 4.16 m; 26,500 $\pm$ 400 BP from 4.5 to 4.7 m; 15,900 $\pm$ 1300 BP from 5.34 to 5.44 m). The paleoenvironmental interpretations are based on preliminary sedimentological and palynological findings. The basal pollen zone is assigned an age of $>$ 26.5 ka BP and is presumed to be MW in age. Although no pollen diagram is presented, the zone is described as a *Picea–Betula*–Ericales–*Sphagnum*–Cyperaceae zone. The local vegetation is inferred to be a mosaic of muskeg and bogs. Sedimentological analysis implies that lake levels were lower than present. Nakao and Ager (1985) made no climatic conclusions, but the data suggest relatively warm summer conditions, perhaps with sufficient summer thaw of the permafrost to yield cool moist soils characteristic of muskeg.

The late Pleistocene loess exposures near Fairbanks (area p) have been the focus of numerous paleoenvironmental studies, beginning with the pioneering research of Péwé (1975a, b). More recent work based on magnetic susceptibility readings from six sites in the 20 m high deposits of the Goldstream valley suggest that the MW was a time of rapid climatic and environmental changes (Begét, 1990; Fig. 5). Although independent dating does not exist at all sections, in some cases, correlation among sites was possible using a buried peat layer (3950 $\pm$ 60 BP; B-24229) and/or forest bed (formed between 6300 $\pm$ 60 BP (B-28543) and 7290 $\pm$ 60 BP (B-28544)). Begét argued that the general trends in the susceptibility curves are similar, and when combined with the radiocarbon-dated levels, provide adequate chronological control. Organic Holocene deposits (Zone A) are underlain by a massive, organic-poor loess (Zone B) and an oxidized layer (Zone C). Zones A and B are differentiated by their magnetic susceptibility, with Zone B having higher values than Zone A. Zone C, assigned to the MW, is characterized by fluctuating susceptibility, but with values that are generally lower than the upper LW zones. The magnetic susceptibility curves from the sites generally are characterized by peaks at the LW–MW

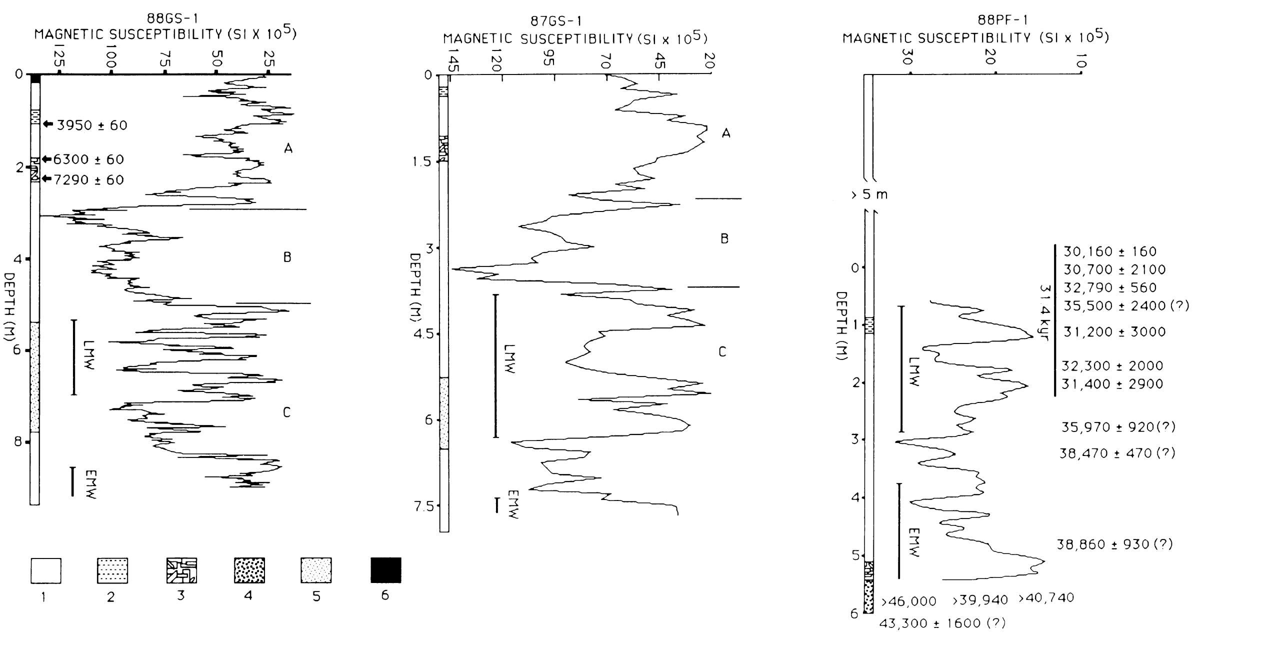

Fig. 5. Magnetic susceptibility profiles, radiocarbon dates, and lithostratigraphy of loess sections near Fairbanks, Alaska, as described by Begét (1990). The profiles are from the Goldstream valley (sites 88GS-1, 87GS-1) and the Fox permafrost tunnel (88PF-1). Capital letters indicate general ages: A (Holocene), B (LW), C (MW), LMW (late MW), and EMW (early MW). The radiocarbon sequence in site 88PF-1 is correlated with zone C, and suspect dates are indicated by a question mark. This figure has been reproduced with the permission of *Géographie Physique et Quaternaire*.

boundary and during the early MW, separated by sediments of varying susceptibilities. Low susceptibility is interpreted as indicating weak surface winds and warmer temperatures, whereas high susceptibility is associated with stronger winds and cooler conditions. By the proposed dating scheme, a period of warmth occurred between 30 and 32 ka BP, similar to the time of permafrost degradation in the Fox permafrost tunnel, which is located ca. 1 km to the north. A second warm interval is recorded at ca. 50–60 ka BP, whereas a cool interval is noted at ca. 42 ka BP.

Research in the Tanana–Kuskokwim lowland (area q) has provided sites of MW age, but dating problems limit conclusive interpretations (Elias et al., 1996). Of the four sites described, the Foraker Slump Site shows the greatest possibility of being from the MW. A ca. 15 m thick composite section from the upper reaches of the Foraker River (site 28) consists of alternating organic-rich and silty sediments. Radiocarbon dates of 6540 ± 230 BP (GX-16177) and 8075 ± 290 BP (GX-16178) from the upper organic horizons date Unit 2 to the Holocene. A massive frozen silt, presumed to be deposited during the LW, separates Holocene deposits from an organic-rich layer (Unit 1) that yielded a $> 42{,}000$ BP date (GX-16176). Underlying Unit 1 is another massive silt that is similar to loess deposits of the EW stade described from nearby areas. A third organic deposit (peaty silt) occurs below the loess and although unsampled, is believed to represent a prior interglacial interval. Given the > 42 ka BP radiocarbon date, Elias et al. (1996) suggested that Unit 2 accumulated during the early MW. A single pollen sample has percentages (24% *Picea*, 35% *Betula*, 22% *Alnus*) that are broadly similar to modern polster spectra from the region, although the amount of *Sphagnum* spores (145%) is significantly greater than the modern values ($< 15\%$). Remains of *Tachinus brevipennis* (an arctic rove beetle) and the ground beetle group *Pterostichus* (*Cryobius*) were also found in Unit 2, as were *Picea* twigs and needles. The paleovegetation is inferred to be a mosaic of open *Picea* forest and meadows, indicating a climate that was cooler than present.

A second site located ca. 100 km to the southeast on the Toklat River (site 27) has tentatively been assigned by Elias et al. (1996) to the MW, but infinite radiocarbon dates on wood and organic detritus ($> 42{,}000$ BP (GX-16164); $> 42{,}900$ BP (GX-16167); $> 35{,}800$ BP (GX-16168) and large standard errors in the finite dates ($40{,}400 \pm 4200$ BP (GX-16165); $35{,}300 \pm 2900$ BP (GX-16166)) make a definitive age assignment impossible. Like at the Foraker Slump site, the Toklat High Bluff section contains organic-rich sediments (Unit 2: felted and silty peat with wood detritus at the base) deposited between silts. A 1.8 m long pollen diagram from Unit 2 has been separated into two zones with Zone II (0.85–1.8 m) likely of early to mid-Pleistocene age. The upper zone (Zone I; 0–0.85 m) is dominated by pollen of *Alnus* (60–80%) with lesser frequencies of *Picea* (3–25%) and *Betula* ($< 10\%$). *Typha latifolia* pollen was identified at 20 cm indicating an altitudinal range extension for this aquatic plant. Twigs and a cone of *Picea glauca* were recovered at 1 m depth, whereas twigs and cones of *Picea mariana* were found in all samples. The insect data are dominated by boreal or arcto-boreal species associated with open ground habitats within forested environments. The presence of dytiscid and hydrophilid water beetles suggests a local open-water environment, which was well vegetated (probably with dense thickets of *Alnus*) at its margins. The regional vegetation is inferred to be an open *Picea* forest. Climatic conditions were likely near modern.

4.2.4. *Northwestern Alaska* (*region H*)

Northwestern Alaska includes the Kotzebue Sound and Koyukuk drainages and the Anaktuvuk River valley. Latitudinal *Picea* treeline, comprised of *Picea glauca*, is located along the southern flanks of the Brooks Range, whereas longitudinal treeline lies in easternmost Seward Peninsula and eastern shores of Kotzebue Sound. Forests are restricted to valley bottoms and favorable sites on nearby slopes. Dense *Alnus* thickets occur beyond altitudinal treeline (ca. 450 m), up mountain draws, and bordering bodies of water. To the south of the mountains, a *Betula*–Ericales–Cyperaceae shrub tussock tundra occupies mid- to upper elevations and areas with soils too cool or moist to support tree growth. This shrub tussock tundra is the predominant vegetation in the northern foothills. Here, streamsides often support thickets of *Salix* shrubs.

Like interior Alaska, this region is relatively data-rich, with five lake records that span at least the latest portion of the MW. The earliest recovered lake record was from Imuruk Lake (site 27; Colinvaux, 1964). As mentioned above, this 8 m long core has presented several interpretive problems. Early interpretations suggested that *Picea* occupied areas of central Seward Peninsula during the MW (correlated to pollen zone i; Colinvaux, 1964; Schweger and Matthews, 1985). This reconstruction contrasts sharply to better dated cores located to the west, which indicate the presence of tundra vegetation during the interstade. The presence of a MW forest refugium on Seward Peninsula is unlikely given: (1) new dates for the Old Crow tephra, which underlies zone i; (2) the subsequent repopulation of Alaska by *Picea* during the early Holocene, which indicates an expansion from sources to the east and not near Bering Strait (Anderson and Brubaker, 1994); (3) the high percentages of *Picea* at Imuruk Lake, which suggest a relatively dense boreal forest, whereas pollen spectra from the Kotzebue Sound basin show only trace amounts (Anderson, 1985; Anderson et al., 1994); and (4) the complex history of the Imuruk basin which has resulted in discontinuous sediment deposition (Ager, 1982). More recently Colinvaux

(1996) proposed a new chronology, assigning zone i to the Sangamon interglaciation and zone J to ca. 70–13 ka BP or the entire Wisconsinan glaciation. Vegetation for this latter period is "a form of tundra with abundant *Artemisia* not present on the contemporary earth" (Colinvaux, 1996, pp. 91–92) that showed no response to interstadial amelioration.

A suite of lake records from the eastern Kotzebue drainage span some or most of the MW. Kaiyak Lake (site 28; Anderson, 1985) lies just beyond latitudinal treeline in the Noatak River valley. Pre-Holocene sediments are dominated by silts and sandy silts, yielding bulk radiocarbon dates of 14,300 $\pm$ 140 BP (WIS-1216), 21,690 $\pm$ 330 BP (WIS-1219), and $>$ 37,000 BP (WIS-1222). The associated pollen spectra, assigned to a single zone, are characterized by high percentages of Poaceae (ca. 25–55%), Cyperaceae (ca. 10–45%), *Artemisia* (ca. 10–25%), *Thalictrum* (ca. 10–30%), and *Salix* (ca. 10–15%; Fig. 6a). Although the exact core depth associated with MW times is unclear, it is likely that at least the latest portion of the interval is present. Vegetation is inferred to be a graminoid tundra with locally abundant *Salix* shrubs, similar to the LW vegetation. Climate is colder and drier than present.

Squirrel Lake (site 29; Anderson, 1985; Berger and Anderson, 1994) lies ca. 100 km to the south of Kaiyak Lake near the confluence of the Squirrel and Kobuk Rivers within the forest-tundra ecotone. Both radiocarbon and TL dates (Berger and Anderson, 1994) indicate the entire MW is preserved in the lacustrine silts. The pollen spectra are generally similar to that described for Kaiyak Lake, with the main characteristics of the LW and MW assemblages being alike. These data indicate herb-dominated tundra occupied areas of the lower Kobuk drainage and that conditions were more severe than present.

Joe Lake (site 30; Anderson, 1988; Anderson et al., 1994) is located ca. 50 km to the southeast of Squirrel Lake, within the forest–tundra ecotone of the Selawik Uplands. A suite of 18 bulk and AMS radiocarbon dates, ranging from 5960 $\pm$ 140 BP (Beta-20075) to 36,970 $\pm$ 1250 BP (Beta-41047) indicates the ca. 9 m composite core includes MW sediments. Zone JO2 is assigned to the MW (Fig. 6b). It is typified by loss-on-ignition percentages that are $>$ 10%, whereas the underlying and overlying zones (tentatively assigned to the EW and LW, respectively) have organic contents that are generally $<$ 10%. The top of Zone JO2 is dated to 25,160 $\pm$ 270 BP (Beta-41045). Other within-zone dates include 26,000 $\pm$ 280 BP (Beta-18852), 28,310 $\pm$ 480 BP (Beta-11221), 31,620 $\pm$ 500 BP (Beta-41046), and 36,970 $\pm$ 1250 BP (Beta-41047).

Unlike Kaiyak and Squirrel Lakes, pollen assemblages from Joe Lake differ between samples assigned to the LW and MW. Zone JO2 is dominated by Cyperaceae (25–30%), Poaceae (20–35%), and *Betula* (ca. 20–50%) pollen percentages. Like the stadial periods preceding and succeeding this zone, the vegetation is inferred to be a graminoid- dominated tundra. However, *Betula* pollen percentages are sufficiently high to indicate its presence in better drained sites of the Selawik Upland. A more productive ecosystem is also implied by the higher organic content of sediments in zone JO2, which equal or are greater than Holocene values. Minor herb taxa indicate a mosaic of micro-habitats, as are present in the LW and EW, but slight increases in more mesic taxa suggest that moister sites may have been more common locally. Analog analysis implies a vegetation similar to that of northern coastal Alaska; i.e., a mesic low arctic tundra (Anderson et al., 1994). Analogs also indicate a climate that was warmer than previously, but still cooler and drier than present.

Epiguruk (site 31), a 12–36 m high exposure, lies to the north of Joe Lake in the *Picea* forests that border the middle reaches of the Kobuk River (Schweger, 1976, 1982; Ashley and Hamilton, 1993; Hamilton and Ashley, 1993; Hamilton et al., 1993; Hamilton et al., 1988b). This 3.5 km long cutbank includes four alluvial–eolian complexes and two paleosol units. The entire section spans more than the last 44 ka years. These deposits represent a complex history of alluviation, eolian activity, incision, and soil formation. The entire exposure is rich in vertebrate and paleobotanical fossils. The early work of Schweger concentrated on the pollen-bearing lacustrine, peat, and paleosol units dating to the LW and MW. Contemporaneous and subsequent work by Hamilton and Ashley focused on the sedimentological aspects of the paleoenvironmental reconstructions.

Schweger's MW work concentrated on an organic-rich horizon (referred to as the upper paleosol unit) that includes lacustrine and peat layers. It is overlain by bedded alluvial sediments with basal dates of 20,700 $\pm$ 440 BP and 24,290 $\pm$ 720 BP. The upper paleosol formed between 33 and 24 ka BP with radiocarbon dates of 23,560 $\pm$ 160 BP (USGS-1442; *Salix* roots); 23,620 $\pm$ 110 BP (USGS-1438; portion of mammoth jaw); 33,670 $\pm$ 280 BP (USGS-1443; detrital *Salix* twigs); 34,620 $\pm$ 560 BP (USGS-1440; mammoth rib); 36,850 $\pm$ 750 BP (USGS-1514; horse bone, probably redeposited). A lower paleosol accumulated at some time prior to 44 ka BP. Pollen diagrams from the upper paleosol are dominated by Cyperaceae (ca. 20–80%), Poaceae (ca. 5–20%), and, at times, *Artemisia* (ca. 5–50%). *Salix* (ca. 5–20%) is also an important component of the pollen spectra, and numerous macrofossils of *Salix* attest to its presence locally. *Betula* pollen percentages are variable, with maxima of ca. 20% and minima of ca. 3%, perhaps relating to changes in the types of organic sediment. The vegetation during the MW at Epiguruk has been interpreted as both a *Salix*-graminoid tundra (Hamilton et al., 1993) and a *Betula* shrub tundra with numerous floodplain microhabitats (Schweger,

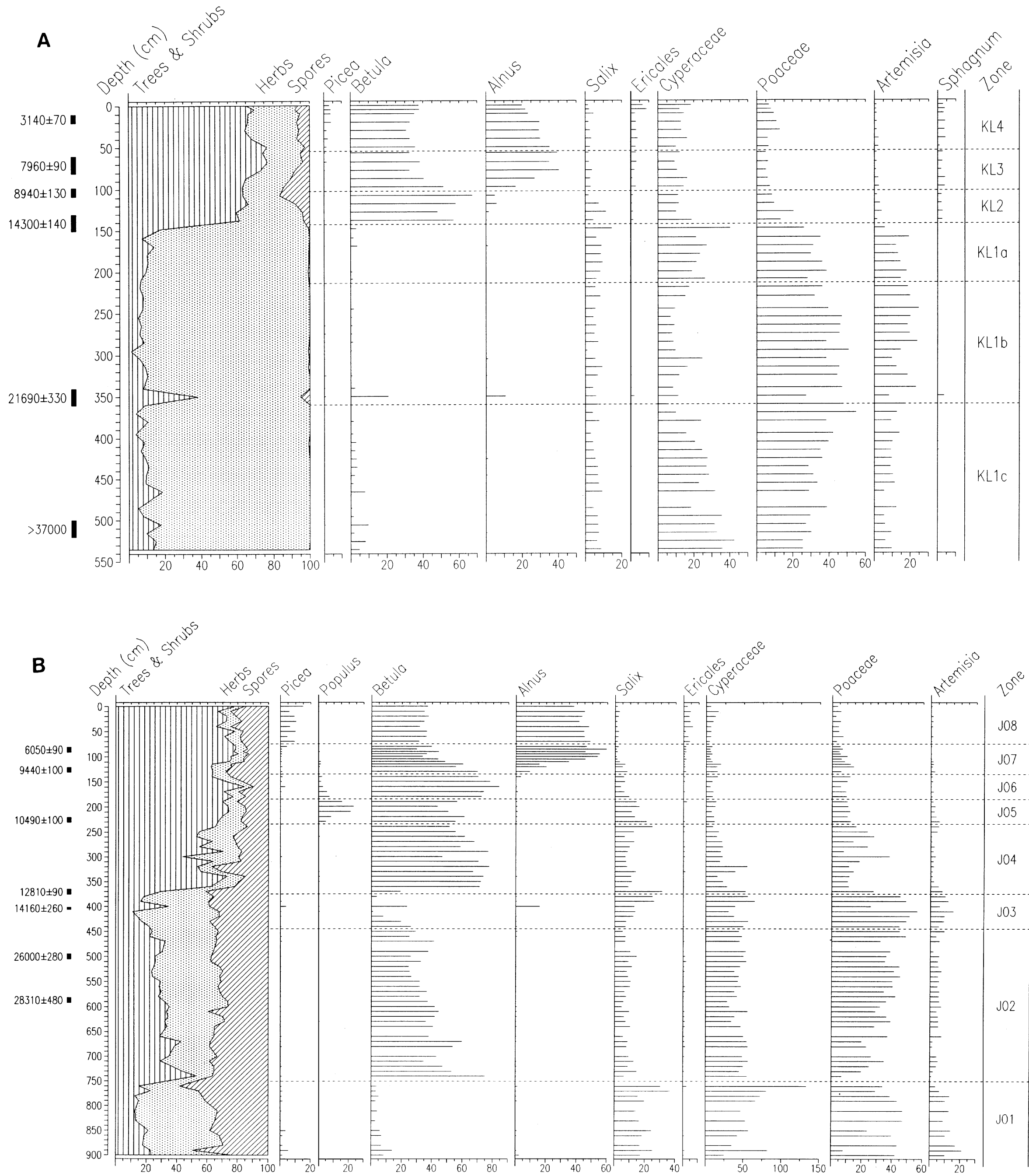

Fig. 6. Percentage diagrams of main taxa from: (a) Kaiyak Lake; and (b) Joe Lake. Percentages are calculated in the same way as described for Fig. 2.

1982). The vegetation was evidently sufficiently productive to support populations of Pleistocene megafauna (Hamilton et al., 1993).

Ashley and Hamilton (1993) examined the depositional environments preserved at Epiguruk as a means to infer late Quaternary climates (Fig. 7). Their interpretations are based on detailed field descriptions and particle size analysis of six major lithostratigraphic units that included river channel, floodplain, dune, sand sheet, loess, paleosol, and pond deposits. Periods of glaciation in the Brooks Range occurred during times of alluviation in the Kobuk valley, whereas interstadial intervals were characterized by downcutting of the river. This pattern is strongly influenced by eolian sedimentation (i.e., periods of great eolian influxes lead to excess sediment loads and alluviation, whereas times when eolian activity is low downcutting of the river bed occurs). River incision is probably associated with relatively warm, moist climates, which favor more extensive vegetative cover and dune stabilization. Periods of cold, dry climate and dune field

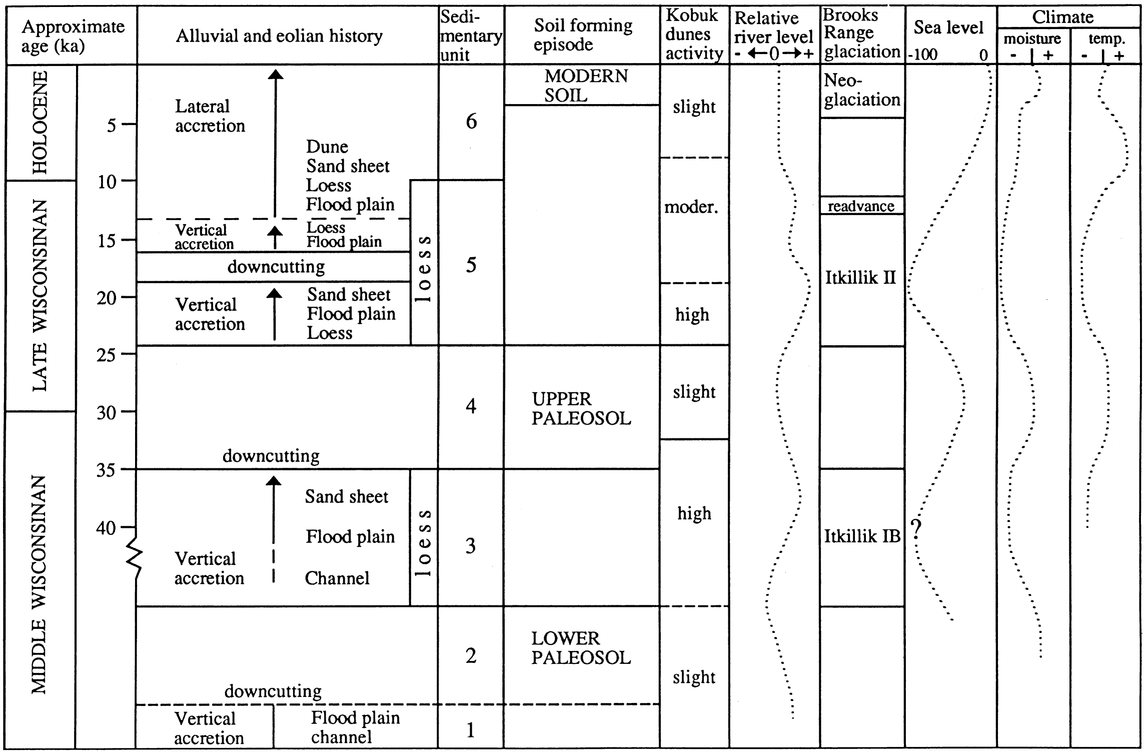

Fig. 7. Summary diagram from Ashley and Hamilton (1993) describing the late Quaternary depositional history at the Epiguruk site, northwestern Alaska, and comparing it with changes in regional glaciation, sea levels, and climate. Relative river level is based on data in Hamilton et al. (1993); glaciation from Hamilton (1986); and sea levels from Chappell and Shackleton (1986). Vertical accretion of lithofacies is shown by the arrows. This figure is reproduced with the permission of the *Journal of Sedimentary Petrology*.

activity are typical of an aggrading river bed. The dates from Epiguruk suggest that from ca. 35–25 ka BP dune activity was slight, river levels were downcutting to achieve near-modern levels, and surfaces were stabilized sufficiently to permit extensive vegetation and soil development. Climates are inferred to be slightly drier than present with temperatures that may have been near-modern. However, this interval was significantly wetter and warmer than the immediately preceding and succeeding stadial times.

Bluff exposures along the Noatak River include several sections with sediments dated between ca. 35 and 30 ka BP (Hamilton, 1996; Hamilton et al., 1996, 1987). These exposures are described in detail in Hamilton (2001). Hamilton interpreted these deposits as indicative of slightly milder conditions resulting in local glacial retreat. Glacial sediments both over- and underlie the interstadial peaty silts, suggesting glacial readvances during the LW and some time prior to 35 ka BP.

The Koyukuk River drainage is another area rich in late Quaternary exposures (e.g., Schweger, 1982; Hamilton and Brigham-Grette, 1991; Schweger and Matthews, 1985). Many of these deposits were once postulated to represent interstadial conditions but are now thought to be more appropriately assigned to the last interglaciation. A 27 m high exposure of fluvial and eolian sands and silts on the John River (Exposure EIV; site 32) has a basal date of 29,000 $\pm$ 700 BP (SI-1880; *Salix* wood) and includes the last stages of the MW. Like in the Epiguruk spectra, the palynological assemblage is dominated by graminoid pollen with lesser but important amounts of *Salix* and *Artemisia* (Schweger, 1982). *Betula* pollen, while present consistently, is typically less than 10%, suggesting the shrub was absent or exceedingly rare. These data indicate a herb-dominated vegetation where *Salix* shrubs were locally abundant in streamside thickets during late MW times.

Ahaliorak Lake (site 33) in the Anaktuvuk River valley provides the longest record of late Pleistocene vegetation change from north of the Brooks Range (Eisner and Colinvaux, 1990). The chronology of the ca. 3 m long core is poor, but available dates (22,740 $\pm$ 820 BP, 31,300 $\pm$ 850 BP, and $>$ 35,000 BP) and the pollen stratigraphy indicate a pre-Holocene age for the lower 2.8 m of the core. Eisner and Colinvaux (1990) argued that the basal pollen zone is likely of MW age, based on the radiocarbon dates and correlation to sites farther south. The pollen spectra are dominated by Cyperaceae (ca. 20–40%) and Poaceae (ca. 20–30%) with lesser but important amounts of *Artemisia* (ca. 5–20%), *Betula* (ca. 5–20%), *Salix* (ca. 5–10%), and *Juniperus* (trace to ca. 5%). The vegetation is inferred to be shrub tundra with fewer *Betula* than present. Increased areas of summer-dry habitat supported *Artemisia-Juniperus* communities, which are rare on the modern landscape. The MW climate is interpreted to be "perhaps as warm as some Holocene climates but was significantly drier" (Eisner and Colinvaux, 1990, p. 35).

4.2.5. Summary of Eastern Beringia

Despite chronological questions and reassignment to the last interglaciation of many sites once thought to be MW in age, certain temporal and spatial patterns are suggested by the EB data (Table 3; Fig. 8). Like WB, the data are strongest for middle to late portions of the interstade and will be the focus of this summary. The interstadial conditions in EB were characterized by fluctuations between relatively warm and cool conditions, as inferred from the paleovegetational, sedimentological, and paleomagnetic data. Unlike WB, neither climate nor vegetation was ever similar to today. For example, *Picea* established in areas of the Yukon Territory and interior Alaska, but these open forests did not achieve their modern latitudinal, longitudinal, nor altitudinal distributions. The greatest (although still limited) distribution of *Picea* in the Yukon Territory likely occurred between ca. 38 and 34 ka BP, indicating an interval of maximum regional warmth and effective moisture. The climatic optimum (Fox thermal event) in interior Alaska is marked, in part, by the presence of a *Picea* forest-tundra between ca. 35 and 30 ka BP. The Harding Lake record suggests *Picea* muskegs may have been present in interior Alaska as late as 26 ka BP, but given its problematic chronology and evidence from nearby sites, this interpretation seems unlikely. Age discrepancies for forest presence between interior Alaska and the Yukon Territory may well be an artifact of the radiocarbon chronologies and not reflect a true regional variation in vegetation or climatic conditions.

In far northwestern Alaska, a herb-*Salix* tundra was likely the dominant regional vegetation throughout the entire interstade. However, local populations of shrub *Betula* apparently flourished ca. 35–26/25 ka BP in protected areas of the major river valleys. This period of shrub *Betula* expansion corresponds in part to the warm periods of interior Alaska and the Yukon Territory, although the vegetation response was not nearly as strong as in the more easterly areas. Tundra also dominated the landscape in southern Alaska for much if not all of the MW, indicating regional climates that were much colder and drier than present. The occurrence of herb or herb-*Salix* tundra assemblages in interior Alaska (30–26 ka BP and 39–34/33 ka BP) and the Yukon Territory (26–31 ka BP) indicate intervals when temperatures approached glacial conditions. The harshest MW environments, as indicated by fell fields, seem to be restricted to the northern Yukon Territory and higher elevations. An herb-*Betula–Salix* low shrub tundra characterized the regional

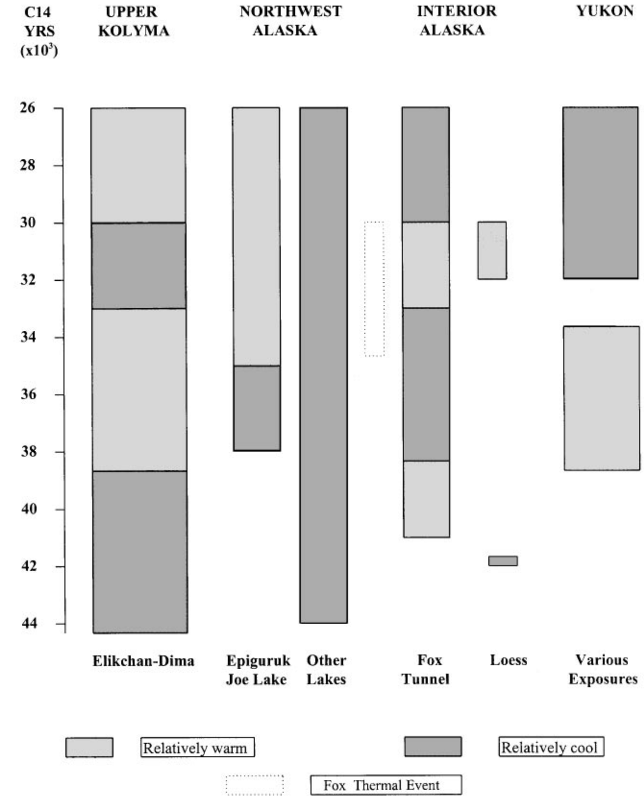

Fig. 8. Summary diagram of main climatic trends in Beringia for late and middle MW. The categories of relatively warm/relatively cool are used within the framework of the MW in which conditions are thought generally to be cooler and/or drier than present. The exception is the interval from 39 to 33 ka BP in WB (illustrated by the upper Kolyma) when climates are inferred to be modern.

vegetation of the Yukon Territory at ca. 39 ka BP and at two times prior to 43 ka BP, suggesting conditions that were warmer than during glacial times but still greatly cooler than present. Even during the warmest intervals when *Picea* forests established, tundra likely was still an important community on the landscape, occupying areas beyond altitudinal treeline and cooler low elevation sites.

Unlike WB, the EB data indicate a heterogeneous climatic pattern for the middle and late MW (Fig. 8). For example, the paleoclimatic scheme presented by Hamilton et al. (1988a), based on the Fox permafrost tunnel results, suggested that interior Alaska experienced relatively warm conditions (although still cooler than modern) from ca. 43 to 38/39 ka BP; a cool period from ca. 38/39 to 33/34 ka BP; a warmer interval from 33/34 to 30 ka BP; and cool conditions that differed little from those of the LW between 30 and 26 ka BP. Hamilton and Fulton (1990) proposed that the younger period of relative warmth, which they dated between ca. 35 and 30 ka BP, be called the Fox thermal event. The vegetational response to this amelioration seems greatest in interior Alaska and relatively moderate to the northwest (i.e., the

Kotzebue Sound drainage), although in the latter area the warming possibly lasted longer (until ca. 25/26 ka BP). Records from the Yukon Territory indicate a MW climatic optimum ca. 38 ka to at least 34 ka BP, followed by an extremely cool period between ca. 34 and 26 ka BP.

The paleoenvironmental data from northwestern Alaska present the most contradictory picture of MW environments. Several lake records from the Kotzebue Sound region imply no environmental changes during the LW, MW, and EW. In contrast, evidence from exposures along the Kobuk and Noatak valleys and from Ahaliorak Lake suggest that summer temperatures may have been as warm as present but effective moisture was less. If summer temperatures were like today, *Picea* forests would be expected to occupy areas south of the Brooks Range. The absence of the trees is not likely caused by delayed migration, considering the rapid movement of *Picea* across the late glacial and early Holocene landscapes (Ritchie and MacDonald, 1986; Anderson and Brubaker, 1994). However, *Picea* (in particular *Picea glauca*) also is limited by effective moisture. The presence of local populations of shrub *Betula* in northwestern Alaska further supports the idea that conditions were drier than present, presuming temperatures were near-modern. If this interpretation is correct, then an east–west gradient of effective moisture characterized EB (i.e., drier to the west, greater effective moisture in the eastern interior) for at least a portion of the MW.

5. Discussion

The MW records from Beringia provide an intriguing glimpse into landscapes and climates of ca. 45–26 ka BP, but fall well short of describing definitive patterns of paleoenvironmental change. The innate limitations due to discontinuous sequences and/or problematic radiometric dating may always prevent a complete understanding of interstadial conditions. Nonetheless, the available records, as troublesome as their chronologies are, suggest certain patterns that are consistent with our knowledge of the spatial variations, which characterized Beringia over the last 18 ka BP. The MW of Beringia clearly is a heterogeneous period. That is, MW paleoenvironments in a given region of Beringia varied through time, and at any given time, these environments often differed from one region to another. This pattern stands in marked contrast to the more homogeneous environments of the Wisconsinan stadial events. We conclude this paper with comments on the paleovegetational and paleoclimatic patterns suggested by the data and explore some possible mechanisms that might account for them. We focus our discussions on times younger than ca. 40 ka BP, because this period has better chronological control and greater site density. Our interpretations are influenced by the quality of the chronology at each site, with better dated records given more weight.

5.1. MW vegetation in Beringia

Comparisons of the regional summaries (Tables 2 and 3) suggest striking contrasts in the vegetation histories of EB and WB during the MW. WB was more extensively reforested during warm intervals, as compared to EB. *Larix* forests were present from Priokhot'ye across the mountainous interior to the northern Yana–Indigirka–Kolyma lowlands from ca. 39 ka to at least 33/34 ka and between 30 and 26 ka BP. During both intervals, the forests, although approximating their present-day range limits, were perhaps restricted to lower elevations in the river valleys and favorable sites, although modern forest composition and distribution is more likely to have occurred in the older interval. Evidence from Chukotka suggests that a high shrub tundra was present during at

Table 3
Summary of Mid-Wisconsinan vegetation patterns, Eastern Beringia

Yukon Territory	Southern Alaska	Interior Alaska	Northwest Alaska
North: herb-*Salix*, herb, or steppe tundra (32–25 ka BP)	West: poorly dated fluctuations in herb and herb–*Salix* shrub tundra	Herb or herb–*Salix* tundra (30–25 ka BP)	Herb, herb-*Salix* tundra throughout MW; local *Betula* shrub thickets (35–26 ka BP)
South: herb-*Salix* or herb tundra (35–25 ka BP?)	East: herb tundra (31–28 ka BP and probably all of MW)	*Picea* forest-tundra (35–30 ka BP)	
North: *Picea* forest–tundra (36–34 ka BP)		Herb tundra (38/39–34/35 ka BP)	
South: *Picea* forest-tundra (38 ka BP)		*Betula-Alnus* shrub tundra or herb tundra (43–38/39 ka BP)	
North: *Betula* shrub tundra (39 ka BP)			
North: *Picea* forest–tundra (41 ka BP)			

least portions of both periods of forest establishment. *Picea* populations likely expanded between ca. 38 and 34 ka BP in the Yukon Territory and ca. 35–30 ka BP in the Fairbanks area. These forests were probably open woodlands in valley bottoms or perhaps isolated stands occupying protected sites. In contrast to WB, the *Picea* forest does not approximate its modern distribution, evidently being absent from both southern and northwestern Alaska.

The number and timing of forest-tundra fluctuations during the MW are difficult to assess, because continuous records are so few and chronologies poor. However, the single lacustrine record from WB indicates that numerous changes from a forested to a shrub tundra environment occurred during the interstade. The trends described from the Elikchan core are also in general agreement with stratigraphic schemes pieced together from the nonlacustrine records in WB, suggesting the Elikchan record is not idiosyncratic. In contrast, pollen diagrams from lakes in EB tend to show no change, changes that are less dramatic (e.g., from herb-*Salix* to *Betula–Salix*-herb tundra), or ones that are more locally confined. Pollen data from all types of sites suggest that the period between ca. 35 and 33 ka BP was the time of maximum forest development for all of Beringia (although forest distribution remains limited in EB). The EB records show no equivalent expansion of forest, as occurred in WB ca. 30–26 ka BP. Differences in the forest histories of EB and WB might reflect the differences in the behaviors of *Larix* and *Picea*. However, the lateglacial and Holocene records indicate that both species can respond quickly to climatic amelioration, and when using fossil pollen to examine patterns of change on the scale of Beringia, edaphic factors are not significant (e.g., Ritchie and MacDonald, 1986; Anderson and Brubaker, 1994; Lozhkin et al., 1991).

Much of EB supported tundra through all or most of the late to middle MW, whereas long-term tundra dominance in WB was limited to areas of Chukotka. Reconstructions of tundra types are difficult (Ritchie and Cwynar, 1982; Anderson et al., 1994), and interpretations are further complicated for the MW data by having to combine records from different depositional environments (e.g., lacustrine vs. alluvial deposits). However, herb–*Salix* tundra was probably the most common MW tundra type in both EB and WB. Depauperate fell fields, *Betula* shrub, or *Betula–Pinus pumila–Alnus* tundra were also present at times during the MW. The establishment of a high shrub tundra in areas of Chukotka seems to correspond to the expansion of *Larix* forest in more western areas of WB. *Betula* shrub tundra occupied areas of the central Kobuk valley from ca. 35 to 26 ka BP, although the occurrence of a graminoid-*Salix* tundra at other nearby sites suggests that the presence of shrub *Betula* may relate more to edaphic than climatic conditions.

The paleobotanical data suggest that the WB vegetation had an almost interglacial character, with forests quickly re-establishing in low elevation areas during two distinct intervals of the middle to late MW. The widespread collapse of the forest also occurred twice, with replacement by tundra types more typical of glacial times. The temporal and spatial "flickering" of the WB forests does not have a parallel in EB. Forests were re-established in interior Alaska and the Yukon Territory some time between ca. 38 and 30 ka BP, but the EB vegetation did not approximate the distribution or composition of the modern forests (i.e., even at times when *Picea* was abundant, herb-*Salix* tundra was still widespread). Rather, the vegetation patterns in EB retained more of their stadial characteristics, with forest-tundra distributions unlike any seen during the Holocene.

5.2. MW climates in Beringia

Where there are sufficient numbers of WB sites with reasonable dating control, the paleobotanical data present a consistent picture of climatic change. Thus we use Lozhkin's proposed scheme for the upper Kolyma region as representative (Fig. 8). Middle to late MW climates were as warm or nearly as warm as present in WB from 30 to 26 ka BP and 39 to 33 ka BP. Cool, dry intervals occurred between 33 and 30 ka BP and 45 and 39 ka BP. In contrast, paleoclimatic interpretations for EB are more spatially complex. An interval when conditions were cooler than present but still warmer than stadial times occurred in the Yukon Territory from ca. 38 to 34 ka BP. Conditions in this region were extremely cold and dry between ca. 34 and 26 ka BP. A period of relative warming occurred in interior Alaska between ca. 35 and 30 ka BP, preceded by times of cool (ca. 38/39–35 ka BP) and milder conditions (ca. 43–38/39 ka BP). Little climatic change is registered for the entire Wisconsinan in several lacustrine records from northwestern Alaska, suggesting long-term cold, dry conditions. However, other records suggest warm and/or dry climates ca. 35–26 ka BP, with cooler conditions from ca. 40–35 ka BP.

Although chronological problems limit the reliability of region-to-region comparisons, certain trends seem evident. The MW–LW transition is clearly marked in WB with a shift from warm/moist to severely cool/dry climates at ca. 26/27 ka BP. Yet in much of EB this transition is essentially seamless, with cool conditions prevailing from at least ca. 30 ka BP (except for some areas of northwestern Alaska). In EB, the youngest MW climatic moderation is registered in interior Alaska ca. 33/35–30 ka BP. In contrast, this interval in WB is characterized, at least in part (33–30 ka BP), by a return to cool/dry conditions. The WB MW climatic optimum, when conditions probably were near modern, occurred between ca. 39 and 33 ka BP. This interval also seems to be a time of maximum warmth (though still cooler than present) in the Yukon Territory. The timing of the Fox thermal event (35–30 ka BP) does not correspond exactly to these warm intervals in WB or the Yukon Territory. If the dates are accurate, the MW optimum occurred at slightly different times in Beringia, with greatest similarity between areas of far eastern and far western Beringia. Even with inaccuracies in the dating, evidence indicates that the MW did experience a thermal maximum some time between ca. 39 and 30 ka BP with moderate to no responses in those areas that now lie closest to Bering Strait (i.e., eastern Chukotka and western Alaska). An east–west precipitation gradient may have characterized EB from ca. 35 to 25/26 ka BP, but evidence is absent for a similar difference in WB. As today, these patterns suggest east-west climatic gradients that are as or more important than latitudinal variations.

5.3. Implications for paleoclimatic interpretations

Modern and paleoenvironmental data indicate that Beringia cannot be considered a homogeneous climatic unit (Barnosky et al., 1987; Anderson and Brubaker, 1993; Lozhkin et al., 1993; Anderson and Brubaker, 1994; Mock and Anderson, 1996; Mock et al., 1998). The spatial variations in climatic patterns seem strongest during periods of warm or warming conditions, whereas differences are more muted during cooler intervals. The modern synoptic patterns are driven primarily by changes in circulation, especially as related to the Pacific subtropical high and the East Asian trough (Mock and Anderson, 1996; Mock et al., 1998). However, causes of late Quaternary climate change in Beringia must consider the effects of boundary conditions (e.g., insolation, atmospheric composition, extent and configuration of ice-sheets, sea-surface temperatures) that differed significantly from today. In cases with relatively rapid climatic changes, as seen in Beringia during the MW, sub-Milankovitch scale forcings may also play an important role.

To understand the broad spectrum of possible responses of regional climates to global forcings it is necessary to examine a wide range of boundary controls. Atmospheric general circulation models (GCMs) provide a means to examine the problem by simulating physically consistent responses (and thus feasible paleoclimatic scenarios) of atmospheric systems to specified sets of boundary controls (e.g., Wright et al., 1993). Unfortunately, as yet no MW experiments have been done. However, it is possible to develop conceptual models, based on insights from available data-model comparisons that can be used to hypothesize about times that lack GCM results. Such a conceptual model was developed for EB climates (Bartlein et al., 1991). In this model, the MW was proposed to be an extremely cold interval, dominated by continental conditions and with little climatic variability throughout the interstade. Although the limited re-establishment of *Picea* forests, local occurrence of *Betula* shrub tundra, downcutting of the central Kobuk valley, changes in ice-wedge and soil activity, and variations in mineralogy of the loess suggest some time(s) of amelioration, much of the MW of EB was characterized by conditions that were more severe than present. In the broadest sense, the conceptual model is correct, in that MW conditions in EB generally were harsher than today. A similar observation is not true for WB, in that climates seemingly were modern or near-modern during long periods of the MW. Although the Bartlein et al. model was never proposed to apply beyond EB, this comparison emphasizes the heterogeneous character of Beringian climates during at least the last 45 ka BP and argues against the naïve extrapolation of EB models to WB.

The conceptual model described above did not predict the non-uniformity of MW climates. The "blindness" of the conceptual model to the relatively rapid MW climatic fluctuations observed in the paleoenvironmental data suggests that causal explanations might be sought in sub-Milankovitch events or in moderating influences of regional feedbacks. The best documentation of interstadial sub-Milankovitch events is from the North Atlantic sector, where marine and ice-core data indicate an instability in MW conditions that is possibly related to major ice-discharge events from the Laurentide ice-sheet into the North Atlantic (Johnsen et al., 1992; Rasmussen et al., 1997; Broecker, 1994.). GCM-fossil data comparisons indicate that changes in the configuration of the North American ice sheet since the last glacial maximum did influence full- and late-glacial climates of EB (Barnosky et al., 1987). Possibly the amelioration–degradation "cycles" in Beringia may relate to changes in ice–sheet configuration, although the GCMs indicate this would require a major reduction in ice-sheet height or extent. Variations in the size or extent of the Laurentide ice–sheet during the MW is poorly understood (e.g., Dredge and Thorleifson, 1987; Andrews, 1987), but some researchers have suggested that the Cordilleran ice–sheet may have disappeared for at least part of the interstade (Fulton, 1976, 1984). GCM sensitivity tests which examine sub-Milankovitch events are few, but one experiment, looking at the Younger Dryas, indicates teleconnections between the North Atlantic and North Pacific Oceans (Mikolajewicz et al., 1997). These results suggest that cooler conditions over the North Atlantic, caused by a temporary shut-down in the North Atlantic deep water formation, resulted in cooling in the North Pacific sector. Fluctuations in the deep water formation may be a more probable forcing than changes in ice-sheet configurations, but the timing and numbers of North Atlantic ice-rafting events/climatic shifts do not correspond to the inferred Beringian climatic fluctuations. Thus, the mechanisms responsible for the observed paleoclimatic changes in Beringia remain unclear, even though we have improved documentation of the characteristics and regional variations of interstadial environments and inferred climates for this vast subcontinent. This uncertainty underscores the importance of examining the MW and its particular combination of global to regional scale controls as a means to improve understanding of Arctic climate system dynamics.

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References

Ager, T.A., 1982. Vegetational history of western Alaska during the Wisconsin glacial interval and Holocene. In: Hopkins, D.M., Matthews Jr., J.V., Schweger, C.E., Young, S.B. (Eds.), Paleoecology of Beringia. Academic Press, New York, pp. 75–93.

Ager, T.A., 1983. Holocene vegetational history of Alaska. In: Wright Jr., H.E. (Ed.), Late-Quaternary Environments of the United States, The Holocene, Vol. 2. University of Minnesota Press, Minneapolis, pp. 128–141.

Ager, T.A., 1989. History of Late Pleistocene and Holocene vegetation in the Copper River Basin, south-central Alaska. In: Carter, L.D., Hamilton, T.D., Galloway, J.P. (Eds.), Late Cenozoic History of the Interior Basins of Alaska and the Yukon, U.S. Geological Survey Circular No. 1026. U.S. Government Printing Office, Washington, DC, pp. 89–92.

Ager, T.A., Brubaker, L.B., 1985. Quaternary palynology and vegetational history of Alaska. In: Bryant Jr., V.M., Holloway, R.G. (Eds.), Pollen Records of Late-Quaternary North American Sediments. American Association of Stratigraphic Palynologists, Austin, TX, pp. 353–384.

Anderson, P.M., 1985. Late Quaternary vegetational change in the Kotzebue Sound area, northwestern Alaska. Quaternary Research 24, 307–321.

Anderson, P.M., 1988. Late Quaternary pollen records from the Kobuk and Noatak River drainages, northwestern Alaska. Quaternary Research 29, 263–276.

Anderson, P.M., Brubaker, L.B., 1993. Holocene vegetation and climate histories of Alaska. In: Wright Jr., H.E., Kutzbach, J.E., Webb III, T., Ruddiman, W.F., Street-Perrott, F.A., Bartlein, P.J. (Eds.), Global Climates Since the Last Glacial Maximum. University of Minnesota Press, Minneapolis, pp. 386–400.

Anderson, P.M., Brubaker, L.B., 1994. Vegetation history of north-central Alaska: a mapped summary of late-Quaternary pollen data. Quaternary Science Reviews 13, 71–92.

Anderson, P.M., Bartlein, P.J., Brubaker, L.B., 1994. Late Quaternary history of tundra vegetation in northwestern Alaska. Quaternary Research 41, 306–315.

Andrews, J.T., 1987. The late Wisconsin glaciation and deglaciation of the Laurentide ice sheet. In: Ruddiman, W.F., Wright, Jr., H.E. (Eds.), North America and Adjacent Oceans during the Last Deglaciation, Geological Society of America, The Geology of North America Series, K-3, Boulder, CO, pp. 13–37.

Arslanov, Kh.A., Vereschagin, N.K., Lyadov, V.V., Ukraintseva, V.V., 1980. About the chronology of the Karginskii Interstadial and the reconstruction of the landscape of Siberia: results from investigations of mammoths and mammoth fauna. In: Geochronology of the Quaternary Period. USSR Academy of Science, Moscow, pp. 208–213 (in Russian).

Ashley, G.M., Hamilton, T.D., 1993. Fluvial response to late Quaternary climatic fluctuations, central Kobuk valley, northwestern Alaska. Journal of Sedimentary Petrology 3, 814–827.

Barnosky, C,W., Anderson, P.M., Bartlein, P.J., 1987. The northwestern U.S. during deglaciation; Vegetational history and paleoclimatic implications. In: Ruddiman, W.F., Wright, Jr., H.E. (Eds.), North America and Adjacent Oceans during the Last Deglaciation, Geological Society of America, The Geology of North America Series, K-3, Boulder, CO, pp. 289–322.

Bartlein, P.J., Anderson, P.M., Edwards, M.E., McDowell, P.F., 1991. A framework for interpreting paleoclimatic variations in eastern Beringia. Quaternary International 11/12, 73–83.

Baskovich, R.A., 1959. Spore-pollen complex of Quaternary deposits of northeast USSR. In: Anikeev, N.P. (Ed.), Works of Joint Meeting about Stratigraphic Schemes of Northeast USSR. Magadan Book Publishing Office, Magadan, pp. 439–450 (in Russian).

Begét, J., 1990. Middle Wisconsinan climate fluctuations recorded in central Alaskan loess. Geographie physique et Quaternaire 44, 3–13.

Berger, G.W., Anderson, P.M., 1994. Thermoluminescence dating of an arctic lake core from Alaska. Quaternary Sciences Reviews 13, 479–501.

Broecker, W.S., 1994. Massive iceberg discharges as triggers for global climate change. Nature 372, 421–424.

Chappell, J.M.R., Shackleton, N.J., 1986. Oxygen isotopes and sea level. Nature 324, 137–140.

Clague, J.J., 1991. Quaternary glaciation and sedimentation. In: Gabrielse, H., Yorath, C.J. (Eds.), Geology of the Cordilleran Orogen in Canada: The Geology of North America v. G-2. Geological Society of America, Canada Communications Group Publishing, Ottawa, pp. 419–434.

Colinvaux, P.A., 1964. The environment of the Bering Land Bridge. Ecological Monographs 34, 297–329.

Colinvaux, P.A., 1996. Pollen records: Point Barrow, Pribilof Archipelago, and Imuruk Lake. In: West, F.H. (Ed.), American Beginnings: The Prehistory and Palaeoecology of Beringia. The University of Chicago Press, Chicago, pp. 83–96.

Colman, S.M., Mixon, R.B., Meyer, R., Bloom, A.L., Johnson, G.H., 1989. Comment on Late Pleistocene barrier-island sequence along the southern Delmarva Peninsula — implications for middle Wisconsin sea levels. Geology 17, 84–85.

Cwynar, L.C., 1982. A late Quaternary vegetation history from Hanging Lake, northern Yukon. Ecological Monographs 52, 1–24.

Denton, G.H., 1974. Quaternary glaciations of the White River Valley, Alaska, with a regional synthesis for the northern St. Elias Mountains, Alaska, and Yukon Territory. Geological Society of America Bulletin 85, 871–892.

Denton, G.H., Stuiver, M., 1967. Late Pleistocene glacial stratigraphy and chronology, northeastern St. Elias Mountains, Yukon Territory, Canada. Geological Society of America Bulletin 78, 485–510.

Dredge, L.A., Thorleifson, L.H., 1987. The middle Wisconsinan history of the Laurentide ice sheet. Geographie Physique et Quaternaire 41, 215–235.

Eisner, W.R., Colinvaux, P.A., 1990. A long pollen record from Ahaliorak Lake, arctic Alaska. Review of Palaeobotany and Palynology 63, 35–52.

Elias, S.A., Short, S.K., Phillips, R.L., 1992. Paleoecology of late-glacial peats from the Bering land bridge, Chukchi Sea shelf region, northwestern Alaska. Quaternary Research 38, 371–378.

Elias, S.A., Short, S.K., Waythomas, C.F., 1996. Late Quaternary environments, Denali National Park and Preserve. Alaska. Arctic 49, 292–305.

Ferrians, Jr., O.J., Nichols, D.R., Williams, J.R., 1983. Copper River basin. In: Pewe, T.L., Reger, R.D. (Eds.), Richardson and Glenn Highways, Alaska: Guidebook to Permafrost and Quaternary Geology: Guidebook 1, International Conference on Permafrost, 4th, University of Alaska, Fairbanks, State of Alaska Department of Natural Resources, Division of Geological and Geophysical Surveys, Anchorage, July 18–22, pp. 137–175.

Finkelstein, K., Kearney, M.S., 1988. Late Pleistocene barrier-island sequence along the southern Delmarva Peninsula — implications for middle Wisconsin sea levels. Geology 16, 41–45.

Fulton, R.J., 1976. Quaternary history, south-central British Columbia and correlations with adjacent areas. In: Easterbrook, D.J., Sibrava, V. (Eds.), Quaternary Glaciations in the Northern Hemisphere — Report of International Correlation Program Project 73-1-24. Geological Survey, Prague, pp. 1–24.

Fulton, R.J., 1984. Quaternary stratigraphy of Canada — a Canadian contribution to IGCP Project 24. Geological Survey of Canada Paper 84–10.

Goldfarb, Yu.I., Lozhkin, A.V., 1975. The history of vegetation of the northeastern USSR during the late Pleistocene and Holocene. Bulletin of the Commission for the Study of the Quaternary Period No. 43, 78–89 (in Russian).

Grichuk, M.P., Karevskaya, I.V., Polosukhina, Z.M., Ter-Grigoryan, E.V., 1975. Paleobotanical Data for Age Correlation Late Cenozoic Deposits in Indigirka-Kolyma Mountain Region. VINITI, Moscow (in Russian).

Hamilton, T.D., 1982. A late Pleistocene glacial chronology for the southern Brooks Range — stratigraphic record and regional significance. Geological Society of America Bulletin 93, 700–716.

Hamilton, T.D., 1986. Late Cenozoic glaciation of the central Brooks Range. In: Hamilton, T.D., Reed, K.M., Thorson, T.M. (Eds.), Glaciation in Alaska - The Geologic Record. Alaska Geological Society, Anchorage, pp. 9–49.

Hamilton, T.D., 1996. Late Pleistocene stratigraphic sections from northern Alaska. In: West, F.H. (Ed.), American Beginnings: The Prehistory and Palaeoecology of Beringia. The University of Chicago Press, Chicago, pp. 21–34.

Hamilton, T.D., 2001. Quaternary glacial, lacustrine, and fluvial interactions in the western Noatak basin, northwest Alaska. Quaternary Science Reviews 20, 371–391.

Hamilton, T.D., Ashley, G.M., 1993. Epiguruk: a later Quaternary environmental record from northwestern Alaska. Geological Society of America Bulletin 105, 583–602.

Hamilton, T.D., Brigham-Grette, J., 1991. The last interglaciation in Alaska: stratigraphy and paleoecology of potential sites. Quaternary International 11/12, 49–71.

Hamilton, T.D., Fulton, R.J., 1990. Middle and late Wisconsin environments of eastern Beringia. Unpublished manuscripts, 44 pp.

Hamilton, T.D., Lancaster, G.A., Trimble, D.A., 1987. Glacial advance of late Wisconsin (Itkillik II) age in the upper Noatak River valley - a radiocarbon-dated stratigraphic record. In: Hamilton, T.D., Galloway, J.P. (Eds.), Geological Studies in Alaska by the U.S. Geological Survey During 1986, Geological Survey Circular No. 998, Washington, DC, pp. 35–39.

Hamilton, T.D., Craig, J.L., Sellmann, P.V., 1988a. The Fox permafrost tunnel: a late Quaternary geologic record in central Alaska. Geological Society of America Bulletin 100, 948–969.

Hamilton, T.D., Galloway, J.P., Koster, E.A., 1988b. Late Wisconsin eolian activity and related alluviation, central Kobuk River valley. In: Galloway, J.P., Hamilton, T.D. (Eds.), Geologic Studies in Alaska by the U.S. Geological Survey During 1987, Geological Survey Circular No. 1016, Washington, DC, pp. 39–43.

Hamilton, T.D., Ashley, G.M., Reed, K.M., Schweger, C.E., 1993. Late Pleistocene vertebrates and other fossils from Epiguruk, northwestern Alaska. Quaternary Research 39, 381–389.

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

Hopkins, D.M., Smith, P.A., Matthews Jr., J.V., 1981. Dated wood from Alaska and the Yukon: implications for forest refugia in Beringia. Quaternary Research 15, 217–249.

Hopkins, D.M., Matthews, Jr., J.V., Schweger, C.E., Young, S.B. (Eds.), 1982. Paleoecology of Beringia. Academic Press, New York.

Hughes, O.L., Harington, C.R., Janssens, J.A., Matthews Jr., J.V., Morlan, R.E., Rutter, N.W., Schweger, C.E., 1981. Upper Pleistocene stratigraphy, paleoecology, and archaeology of the northern Yukon interior, eastern Beringia 1. Bonnet Plume Basin. Arctic 4, 329–365.

Johnsen, S.J., Clausen, H.B., Dansgaard, W., Fuhrer, K., Gundestrup, N., Hammer, C.U., Iversen, P., Jouzel, J., Stauffer, B., Steffensen, J.P., 1992. Irregular glacial interstadials recorded in a new Greenland ice core. Nature 359, 311–313.

Kartashov, I.P., 1963. New data about the stratigraphy of Anthropogene deposits and the age of glaciations of north east USSR. Kolyma 2, 24–28 (in Russian).

Kartashov, I.P., 1966. Numbers and ages of glaciations in north east USSR. In: Quaternary Period of Siberia. Nauka, Moscow, pp. 289–296 (in Russian).

Kind, N.V., 1974. Late Quaternary Geochronology According to Isotopes Data, Transactions, Vol. 257. Nauka, Moscow (in Russian).

Klassen, R.W., 1987. The Tertiary-Pleistocene stratigraphy of the Liard Plain, southeastern Yukon Territory. Geological Survey of Canada Paper 86-17. Geological Survey of Canada, Ottawa.

Kotov, A.N., Lozhkin, A.V., Ryabchune, V.K., 1989. The condition and formation of frozen facies of the late Quaternary of the Main River valley (Chukotka). In: Relief Formation, Correlation of Deposits and Placer Deposits of North East USSR, pp. 117–131. North East Interdisciplinary Research Institute, USSR Academy of Science, Far East Branch, Magadan, pp. 117–131 (in Russian).

Lea, P.A., Elias, S.A., Short, S.K., 1991. Stratigraphy and paleoenvironments of Pleistocene nonglacial deposits in the southern Nushagak Lowland, southwestern Alaska, U.S.A.. Arctic and Alpine Research 23, 375–391.

Lozhkin, A.V., 1987. Paleogeography and formation of yedoma sediments of the Indigirka lowland. In: Quaternary Period of Northeast Asia, North East Interdisciplinary Research Institute, USSR Academy of Science, Far East Branch, Magadan, pp. 135–144 (in Russian).

Lozhkin, A.V., 1989. About the late Quaternary deposits of Northern Priokhot'ye. Doklady Akademii Nauk 308 (5), 1194–1196 (in Russian).

Lozhkin, A.V., 1990. Two questions about the interdisciplinary study of mammoth sites in northeast Siberia. Bulletin of the Commission for Study of the Quaternary 59, 156–159.

Lozhkin, A.V., 1991. The questions of geochronology and paleogeography of the Quaternary period, north east USSR. In: The Geology of the Zone between the Continent and the Ocean of North East Asia, North East Interdisciplinary Research Institute, USSR Academy of Science, Far East Branch, Magadan, pp. 47–59 (in Russian).

Lozhkin, A.V., Anderson, P.M., 1996. A late Quaternary pollen record from Elikchan 4 Lake, northeast Siberia. Geology of the Pacific Ocean 14, 18–22.

Lozhkin, A.V., Pavlov, G.F., Ryabchune, V.K., Gorbachev, A.L., Zadal'skii, S.V., Shubert, E.E., 1988. A new mammoth discovery in Chukotka. Doklady Akademii Nauk 302 (6), 1440–1444 (in Russian).

Lozhkin, A.V., Anderson, P.M., Belaya, B.V., 1995. Radiocarbon dates and pollen zones from lacustrine sediments from the region of the Kolyma-Okhotsk drainage divide. Doklady Akademii Nauk 342, 396–399 (in Russian).

Matthews Jr., J.V., 1974a. Quaternary environments of Cape Deceit (Seward Peninsula, Alaska): Evolution of a tundra ecosystem. Geological Society of America Bulletin 85, 1353–1385.

Matthews Jr., J.V., 1974b. Wisconsin environment of interior Alaska: pollen and macrofossil analysis of a 27 m core from Isabella Basin (Fairbanks, Alaska). Canadian Journal of Earth Sciences 11, 812–841.

Matthews Jr., J.V., Schweger, C.E., Hughes, O.L., 1990. Plant and insect fossils from the Mayo Indian Village section (central Yukon): new data on middle Wisconsinan environments and glaciation. Geographie physique et Quaternaire 44, 15–26.

Matthews Jr., J.V., Harington, C.R., Hughes, O.L., Morlan, R.E., Rutter, N.W., Schweger, C.E., Tarnocai, C., 1987. Schaffer Mountain Lookout and Old Crow Basin stratigraphy/paleontology. In: Morison, S.R., Smith, C.A.S. (Eds.), XIIth INQUA Congress Field Excursions A20a and A20b - Research in Yukon. National Research Council, Ottawa, pp. 75–83.

Mikolajewicz, U., Crowley, T.J., Schiller, A., Voss, R., 1997. Modeling teleconnections between the North Atlantic and North Pacific during the Younger Dryas. Nature 387, 384–387.

Mock, C.J., Anderson, P.M., 1997. Some perspectives on the late Quaternary paleoclimate of Beringia. In: Issacs, C.M., Tharp, I.V. (Eds.), Proceedings of the Thirteenth Annual Pacific Climate (PACLIM) Workshop, April 15–18, 1996, Interagency Ecological Program, Technical Report 53, California Department of Water Resources, pp. 193–200.

Mock, C.J., Bartlein, P.J., Anderson, P.M., 1998. Atmospheric circulation patterns and spatial climatic variations in Beringia. International Journal of Climatology, 18, 1085–1104.

Morlan, R.E., 1980. Taphonomy and Archaeology in the Upper Pleistocene of the Northern Yukon Territory: A Glimpse of the Peopling of the New World. National Museum of Man, Mercury Series, Archaeological Survey of Canada Paper 94, National Museums of Canada, Ottawa.

Nakao, K., Ager, T.A., 1985. Climatic changes and vegetational history in the interior Alaska. In: Nakao, K. (Ed.), Hydrological Regime and Climatic Changes in the Arctic Circle, Report of the Alaskan Paleohydrology Research Project - 1982/83, Laboratory of Hydrology Department of Geophysics Hokkaido University, Hokkaido, pp. 43–52.

Ovander, M.G., Lozhkin, A.V., Bashlavin, D.K., Zhigulevtseva, S.H., 1987. Paleogeography situation during the formation of yedoma suite in the Yana-Indigirka lowland. In: Quaternary Period of Northeast Asia, North East Interdisciplinary Research Institute, USSR Academy of Science, Far East Branch, Magadan, pp. 119–134.

Péwé, T.L., 1975a. Quaternary Geology of Alaska. Geological Survey Professional Paper 835, U.S. Government Printing Office, Washington, DC.

Péwé, T.L., 1975b. Quaternary stratigraphic nomenclature in unglaciated central Alaska. U.S. Geological Survey Paper No. 862, U.S. Geological Survey, Washington, DC.

Rampton, V., 1971. Late Quaternary vegetational and climatic history of the Snag-Klutlan area, southwest Yukon Territory, Canada. Geological Society of America Bulletin 82, 959–978.

Rasmussen, T.L., Van Weering, T.C.E., Labeyrie, L., 1997. Climatic instability, ice sheets and ocean dynamics at high northern latitudes during the last glacial period (58-10 ka BP). Quaternary Science Reviews 16, 71–80.

Ritchie, J.C., Cwynar, L.C., 1982. The late Quaternary vegetation of the north Yukon. In: Hopkins, D.M., Matthews Jr., J.V., Schweger, C.E., Young, S.B. (Eds.), Paleoecology of Beringia. Academic Press, New York, pp. 113–126.

Ritchie, J.C., MacDonald, G.M., 1986. The patterns of post-glacial spread of white spruce. Journal of Biogeography 13, 527–540.

Saks, V.N., 1953. The Quaternary period in the Soviet Arctic. Vodtransizdat, 402–430 (in Russian).

Schweger, C.E., 1976. Late Quaternary Paleoecology of the Onion Portage region, Northwestern Alaska. Ph.D. Thesis, University of Alberta.

Schweger, C.E., 1982. Late Pleistocene vegetation of eastern Beringia: pollen analysis of dated alluvium. In: Hopkins, D.M., Matthews Jr., J.V., Schweger, C.E., Young, S.B. (Eds.), Paleoecology of Beringia. Academic Press, New York, pp. 95–112.

Schweger, C.E., 1989. The Old Crow and Bluefish Basins, northern Yukon: development of the Quaternary history. In: Carter, L.D., Hamilton, T.D., Galloway, J.P. (Eds.), Late Cenozoic History of the Interior Basins of Alaska and the Yukon, U.S. Geological Survey Circular No. 1026, U.S. Government Printing Office, Washington, DC, pp. 30–33.

Schweger, C.E., Janssens, J.A.P., 1980. Paleoecology of the Boutellier nonglacial interval, St. Elias Mountains, Yukon Territory, Canada. Arctic and Alpine Research 12, 309–317.

Schweger, C.E., Matthews Jr., J.V., 1985. Early and middle Wisconsinan environments of eastern Beringia: stratigraphic and paleoecological implications of the Old Crow tephra. Geographie physique et Quaternaire 39, 275–290.

Schweger, C.E., Matthews Jr., J.V., 1991. The last (Koy-Yukon) interglaciation in the Yukon: comparisons with Holocene and interstadial pollen records. Quaternary International 10–12, 85–94.

Shackleton, J., 1982. Environmental Histories from Whitefish and Imuruk Lakes, Seward Peninsula, Alaska. Ohio State University Institute of Polar Studies Report No. 76, Columbus, OH.

Shackleton, N.J., 1987. Oxygen isotopes, ice volume and sea level. Quaternary Science Reviews 6, 183–190.

Sher, A.V., 1971. Mammals and Stratigraphy of Pleistocene from Far North East USSR and Northern America. Commission for Study of the Quaternary Period, Nauka, Moscow (in Russian).

Sher, A.V., 1991. Problems of the last interglacial in arctic Siberia. Quaternary International 11/12, 215–222.

Sher, A.V., 1997. Yedoma as a store of paleoenvironmental records in Beringida. Beringian Paleoenvironments Workshop, Program and Abstracts, pp. 140-144.

Sher, A.V., Plakht, I.P., 1988. Radiocarbon dating and problems of Pleistocene stratigraphy, northeast U.S.S.R. News of the USSR Academy of Science, Geological Series 8, 17–31 (in Russian).

Shilo, N.A., Lozhkin, A.V., Titov, E.E., Schumilov, Yu.V., 1983. Kirgirlakh Mammoth: Paleogeography Aspect. Nauka, Moscow (in Russian).

Shilo, N.A., Siderov, A.A., Baranova, Yu. P., Lozhkin, A.V. et al., 1987. Resolution: Interdisciplinary Stratigraphic Congress for Quaternary Systems of Eastern USSR. Russian Ministry of Geology, USSR Academy of Science, Far East Branch, Magadan (in Russian).

Ten Brink, N.W., Waythomas, C.F., 1985. Late Wisconsin glacial chronology of the north-central Alaska Range - a regional synthesis and its implications for early human settlements. In: Powers, W.R. (Ed.), North Alaska Range Early Man Project, National Geographic Society Research Reports, Vol. 19, pp. 15–32.

Thorson, R.M., Dixon Jr., E.J., Smith, G.S., Batten, A.R., 1981. Interstadial proboscidean from south-central Alaska: implications for biogeography, geology, and archaeology. Quaternary Research 16, 404–417.

Tomirdiaro, S.V., 1980. Loess-Ice Formation of Eastern Siberia in the Late Pleistocene and Holocene. Nauka, Moscow (in Russian).

Tomirdiaro, S.V., 1982. Evolution of lowland landscapes in northeastern Asia during late Quaternary time. In: Hopkins, D.M., Matthews Jr., J.V., Schweger, C.E., Young, S.B. (Eds.), Paleoecology of Beringia. Academic Press, New York, pp. 29–37.

Vaskovskii, A.P., Terekhova, V.E., 1970. Anthropogene (Quaternary System). In: Geology of the USSR Book 1, Nedra, Moscow, pp. 501–547 (in Russian).

Waythomas, C.F., 1996. Late Quaternary aeolian deposits of the Holitna lowland, interior southwestern Alaska. In: West, F.H. (Ed.), American Beginnings: The Prehistory and Palaeoecology of Beringia. The University of Chicago Press, Chicago, pp. 35–52.

Westgate, J.A., Hamilton, T.D., Gorton, M.P., 1983. Old Crow tephra: A new late Pleistocene stratigraphic marker across north-central Alaska and western Yukon Territory. Quaternary Research 19, 38–54.

Westgate, J.A., Stemper, B., Pewe, T.L., 1990. A three million year record of Pliocene-Pleistocene loess in interior Alaska. Geology 18, 858–861.

Williams, J.R., Galloway, J.P., 1986. Map of western Copper River Basin, Alaska, showing lake sediments and shorelines, glacial moraines, and location of stratigraphic sections and radiocarbon-dated samples. U.S. Geological Survey Open-File Report 86-390, scale 1:250,000.

Wright Jr., H.E., Kutzbach, J.E., Webb III, T., Ruddiman, W.F., Street-Perrott, F.A., Bartlein, P.J. (Eds.), 1993. Global Climates Since the Last Glacial Maximum. University of Minnesota Press, Minneapolis.