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Review

The relationship of sea level changes to climatic change in northeast Asia and northern North America during the last 75 ka B.P.

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Abstract: The Arctic air mass is the cold, dry body of air slowly moving eastwards around the North Pole in the northern hemisphere. Its southern boundary consists of four planetary waves known as the Rossby waves that mark the interface with subtropical air bringing heat polewards. The Arctic air mass is constantly being modified by the addition of heat and moisture over the oceans, as well as by winter cooling over the land masses due to limited incoming solar radiation and constant reradiation of heat into the atmosphere. The coldest air in winter is located over northeastern Siberia and moves east, cooling Canada. Warm ocean currents add large quantities of heat to the air mass moving over them, but without this addition of heat, the Arctic air mass becomes significantly colder. Research in Tibet and Northeast Asia on depression of sea level shows that during the Late Wisconsin cold event (65–10 ka B.P.), vast quantities of sea water were sequestered on land primarily as ice sheets, exposing the sea bed in the Bering Strait from 50–10 ka B.P. together with the bottom of the South China Sea between 30–20 ka B.P.. The East China Monsoon failed to reach Tibet and much of Northeast China, resulting in severe cooling of northeast Siberia and northern Tibet. This, in turn, caused severe cooling in eastern Canada together with the development of a vast, predominantly cold-based ice sheet. As the sea levels started to rise (about 19 ka B.P.), the East China Monsoon slowly redeveloped and a gradual warming took place on both continents. However, along the west side of the North American Cordillera, the Late Wisconsin glaciation only began in 29 ka B.P. but continued along the west coast until about 10 ka B.P. This paper explores the relationship of the Late Wisconsin history on the two continents, together with the mechanisms causing the landforms and climatic differences. Finally, the probable effects of these climatic changes on the early peopling of North America are discussed.

Keywords: Late Wisconsin Glaciation; Tibet Plateau; Northeast Asian climate; North American climate; Arctic air mass; sea level changes and climate

1. Introduction

For decades, scientists have debated the causes of extreme periodic cold climatic changes in the Northern Hemisphere during the last 3.5 Ma without really conclusively determining either their causes or their mechanisms. We do know that there are many potential factors that influence the mean annual air temperature system of the Earth to varying extents, and that most of these are changing simultaneously to produce the considerable variability at all time scales [1]. The Milankovich cycles are closely related to the frequency of cold events during the last few cold events resulting in glaciations and/or the development of permafrost at high elevations and at high latitudes [2]. The common practice has been to suggest possible causes for which there is some credible evidence, but little attention has been paid to the details of the actual operation of the climate system. What happens if one of the changes in the controls such as sea level triggers a chain reaction that causes a drastic alteration in the way the system works? One way to find out is to examine the operation of the system during a recent period of time, examining the effects of failure of one of the key controls on the resultant climate.

Recently, considerable progress has been made in determining the sequence of climatic events during the last 75 ka in northeastern Asia [3,4] and it is now possible to compare and correlate this sequence and associated mechanisms with those for northern North America. In this paper, first the climatic mechanisms currently operating in the Northern Hemisphere will be examined, and then the sequences of climatic changes during the last 75 ka on the two sides of the Pacific Ocean will be summarized. It will be shown that the Arctic Air mass is constantly being modified as it moves around the northern hemisphere, and that there is a strong correlation of these modifications to the climatic history of the areas over which it moves. Failure of the movement of heat from the Tropics to the north by ocean currents due to sea level changes can result in spectacular changes in the climate of very large areas downwind to the east.

2. The climate system

The climate system that operates today is discussed in many textbooks [5,6]. It includes an Arctic Polar air mass (A) circulating around the North Pole and crossing alternately areas of land and sea. This air mass is often split into Polar (P) and Arctic (A) air masses in the older literature, but this distinction is difficult to justify because the air mass is constantly being modified as it moves eastward at varying rates over different surfaces. Likewise the modifiers, maritime (m) and continental (c), are of limited value in this case, though they are important in indicating the properties of the tropical (T) and equatorial (E) air masses located nearer the equator. The Arctic air mass is separated from the Tropical air mass by a sharp, wavy boundary, usually accompanied by high speed winds aloft known as jet streams. Figure 1 shows a typical north-south cross-section of the air masses over western North America in 1980. These planetary waves are usually referred to as the Rossby waves [7].

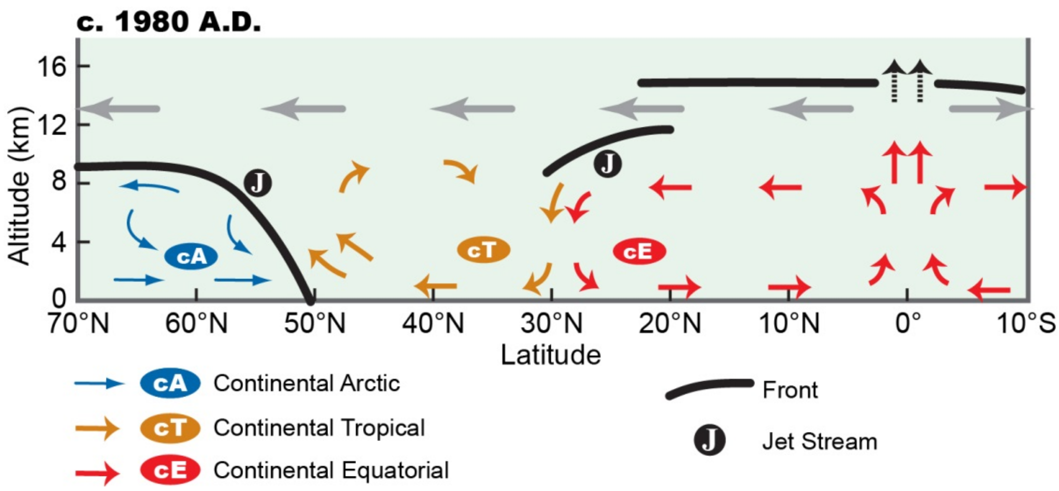

Figure 1. North-south section through the lower atmosphere in western North America showing the circulation of the air masses under the climatic conditions occurring in 1980 A.D. [8]. Note that there are three basic air masses, viz., Arctic, Tropical and Equatorial, with jet streams along their margins. The more northerly black line marks the polar front, and the southern one is the subtropical front separating the tropical air (in orange) from the equatorial air mass (in red). The grey arrows indicate the air movement in the upper atmosphere. Each air mass operates as a distinct band around the globe, interacting with the others primarily along the boundaries between air masses.

2.1. Rossby waves and the arctic polar air mass

The air in the lower atmosphere surrounds the Earth with no obvious impediments to air mass movement other than the effects of the higher land areas generally above 4000 m. The rotation of the Earth results in a slow eastward movement of the Arctic air masses with time as indicated by arrows in Figure 2. There are usually four individual Rossby waves circling the North Pole that are constantly changing shape to spawn discrete islands of cold air south of the main cold polar air mass, straightening out the southern margin of the main air mass in the process. The Arctic air masses in winter are generally cold and dry except for those that have traversed large bodies of open water, e.g., the north Pacific or Atlantic Oceans.

The north-south elongation of the waves varies with time. This variation is usually greater in winter, apparently due to the greater contrast in temperature between the polar and subtropical air masses. At that time, the southern margin of the lobes of Arctic air may extend south to 40°N, while the northern extension of the cT air can reach beyond 60°N. During the process of movement of the fronts, cyclones and anticyclones develop, the cyclones resulting in precipitation of a portion of the water mass available in the adjacent air masses. The greater the contrast between the two air masses, the greater the probability of violent storms, hence the “tornado alley” along the west side of the tropical, moisture-laden air moving north from the Caribbean Sea in the eastern parts of North America. The discrete air masses are usually high pressure zones, and low pressure cyclones tend to develop between the contrasting zones, particularly on the east side of the lobes forming the waves.

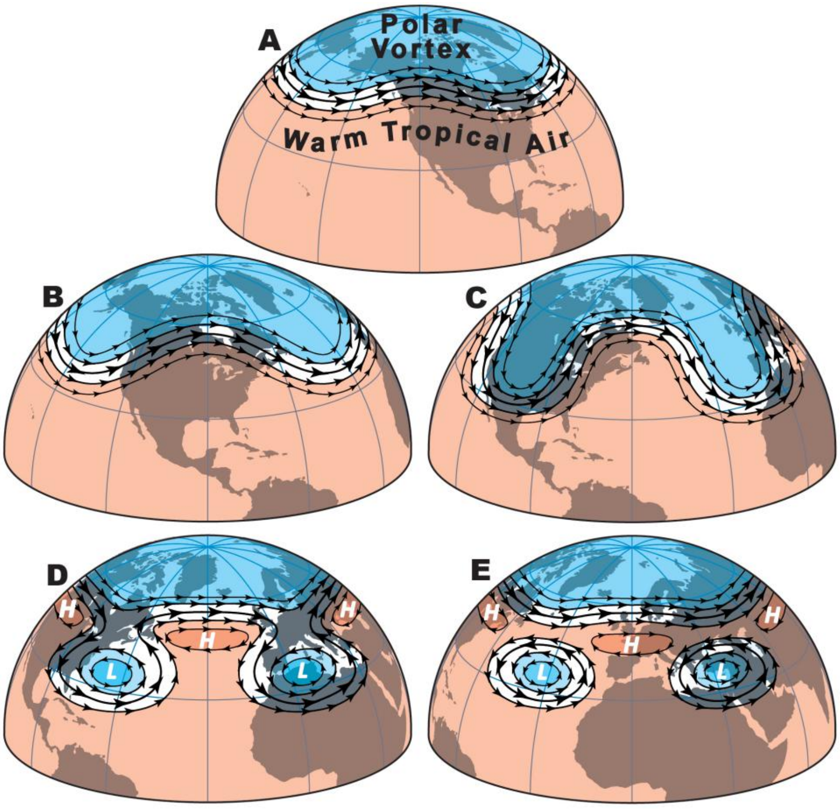

Figure 2. Evolution of the southern edge of the Rossby waves as the cold air mass (in blue) moves eastwards [9]. H represents a high air pressure cell (an anticyclone) while L represents one of low pressure (a cyclone).

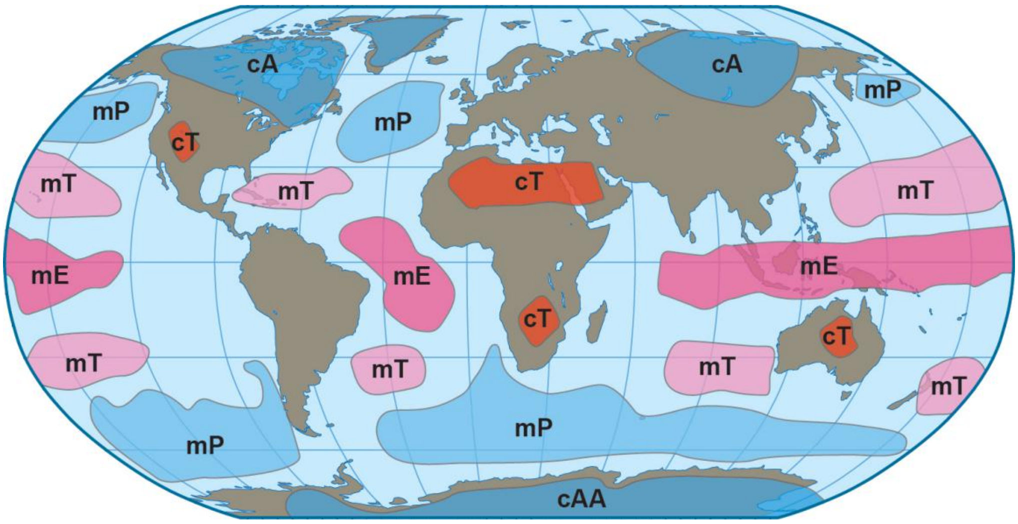

Figure 3. Distribution of the source areas of the main air masses today (modified from [6]). cA is continental Arctic air, cAA is continental Antarctic air, mP is maritime Polar air, mT is maritime Tropical air, cT is continental Tropical air and mE is maritime Equatorial air.

2.2. The tropical air masses

Figure 3 shows the position of the major air masses around the world. In North America, the continental (cT) air mass comes north from the deserts of California and Arizona, whereas the maritime Tropical (mT) air mass comes from either Hawaii or the Caribbean Sea. In Europe and western Asia, the cT air comes from the Sahara Desert, the Arabian Peninsula, Iran and Afghanistan. The cA air mass in Siberia is more widespread and colder than that in northern Canada, and is partly fed from the south in winter by the cold, dry cT air mass draining north from the very high altitude ranges of the Tibetan Plateau, the Tien Shan and the Pamir Knot. The rest of the air represents the cA air moving east from Europe to eastern Siberia, cooling as it moves over the land that is reradiating heat into space during the long winter nights at the higher latitudes. The mT air mass from Hawaii brings moisture to the western part North American continent particularly in summer, while the Caribbean air interacts with the Arctic air masses along the south-central and eastern seaboard of the continent throughout the year.

2.3. Redistribution of heat and moisture

Because of the curvature of the Earth, considerably greater radiation is received in the tropics on a given surface than in polar areas. To produce the present-day distribution of mean annual air temperatures (MAAT), 30% of the heat absorbed in the tropics and subtropics south of 40° latitude must be moved north to compensate for the disparity [5]. This is primarily achieved by warm ocean currents along the eastern seaboards of continents and by high speed, deep water, saline currents. Tropical air masses also move northwards but may carry far less heat. If any of these fail, then a cold event will occur in the areas at higher latitudes that lose part of their heat supply. The cold, dry Arctic air passing over the warm ocean currents absorbs heat from them and transports it over the adjacent land to the east [10–12]. It also absorbs substantial quantities of moisture from the ocean surface as it warms in transit, though the amount of warming and increase in moisture depends on whether or not the sea over which it passes is covered in ice, as well as the distance of contact with the warmer water surface. The cT air masses are hot and dry, resulting in drying out of the land areas over which they pass, and can result in the development of deserts and semi-desert conditions unless there is additional precipitation from another source, e.g., the wet, westerly winds in the mountains of southern British Columbia.

2.4. Effect of Topography

Where very high mountain ranges occur in the path of the surface winds, these air masses can be deflected to areas of lower relief as in Northern British Columbia and the central Alaska-Yukon area (Figure 4). The key is the relative thickness of the air mass compared to the height of the crest of the mountain range. The Wrangell-St. Elias mountain crests are over 5500 m high so that the Arctic air tends to go either north (paths II and III) or south of these coastal mountains (path I). Although the boundary between the Arctic and subtropical air masses is generally moving east in a fairly random fashion, there is a tendency for it to use the two separate paths that are often followed past the high Wrangell-St. Elias ranges (>5000 m high) in southern Alaska, while the third is the route (path III) in northern Canada directly across the Arctic Ocean to the vicinity of Hudson Bay where the elevation of the land is closer to sea level. In Europe, a preferred route for the Arctic air masses is between the mountains of northern Scandinavia and the Alps. The mountains of Spain and Italy currently act as a barrier to the westerly winds, while the east-west trending Alps and Carpathian ranges protect northern Europe from the full effect of the hot, tropical air coming north from the Sahara desert in summer.

Figure 4. Map showing the distribution of permafrost in the Arctic [9] together with the mean surface air January isotherms (°C) and the adjacent warm and cold ocean currents. Also shown are the three main paths (I to III) taken by the Arctic air as it moves from Asia to northern Canada, and the positions of the main warm ocean currents currently bringing heat from the Tropics.

As the Arctic air mass moves further east from Europe, its southern boundary moves north, due to the effects of both the Gulf Stream and the warm tropical air of the eastern part of the Mediterranean in summer. This causes the Mediterranean climate to extend north into the southern Ukraine, and eastwards into northern Iran, and Afghanistan. There is a sharp climatic divide between the Arctic and Subtropical air masses at the southern boundary of the Pamir knot and the western boundary of the Tibetan Plateau and Himalayas due to the great altitude of these mountains (c.5500–6000 m). This divide has persisted for a very long time, since the flora of Afghanistan is quite different from that found to the north and east [13–16].

In cold climates, the presence or absence of adequate supplies of moisture also control whether glaciers are present or the ground develops permafrost [9]. Development of glaciers requires persistent supplies of moisture that exceed the annual losses by melting and sublimation of the ice. In northeast China, there is very little precipitation apart from that brought inland by the East Asian monsoon. In eastern Siberia, permafrost is ubiquitous north of the Hexi Corridor which is marked by the Great Wall of China in the east (Figure 4). The corridor has an altitude of about 1000 m elevation and lies north of the Tibet-Qinghai Plateau and Pamir Knot. This area currently lacks permafrost except for patches of relict frozen ground at depth but exhibits wedge structures in its sediments indicating past colder climates [3]. The peaks of the mountain ranges to the south rise to over 6000 m, so that permafrost is widespread across the northern slopes of these mountains above 4500–5000m. In winter, cold, dense air drains down their northern slopes to cross the Hexi Corridor and form a southern extension to the vast Siberian Arctic air mass (Figure 4). The coldest winter temperatures recorded in the northern hemisphere were obtained in the interior valleys of the mountains in northern Siberia at Verkhoyansk located at 67°33′N, 133°23′E. (−67.8 ℃ on January 15th, 1885), and equally cold temperatures were recorded at Oymyakon (located at 63°15′N, 143°09′E). These are sometimes referred to as the northern poles of cold. Similar cold temperatures (−65 ℃) have been recorded in a cold air drainage event in a mountain valley 100 km west of Fort Nelson, British Columbia [17]. These are still considerably warmer than the higher altitude stations inland in Antarctica. In 2018, many stations in Antarctica experienced colder temperatures, the current record low being −98 ℃ at 81°S, 63.5°E measured on the East Antarctic Plateau [18]. This is at least partly due to there being no ocean currents transporting heat into the approximately circular Antarctic land mass from the north. Cold air can descend down the outer slopes of mountain massifs and linear mountain ranges such as the Alps and drain away across the adjacent lowlands, so these mountain margins experience warmer average winter temperatures (Figure 4).

2.5. Anthropological effects

The thermal modifications to the landscape by Mankind are generally small, e.g., the replacement of prairie by wheat fields in Manitoba only increased the surface ground temperature by about 1 ℃. However, the loss of forests to be replaced by cultivation is another matter. In summer, the difference in summer afternoon air temperature between the inside of a closed canopy of western Red Cedar forest and the adjacent secondary forest may be 15–20 ℃, e.g., at the Giant Cedars site in Glacier National Park in British Columbia. This is accompanied by a marked contrast in relative humidity, hydrology and the soil properties. When agriculture replaced 85% of the forest in Costa Rica, the mean annual precipitation decreased by 30%, the soil density changed from 800–1100 kg/m^3 to about 1500 kg/m^3, and the soil structure changed from fine crumb to coarse angular blocky structure. Surface runoff increased spectacularly, soil permeability was greatly reduced, and the soil moisture content was greatly decreased. The mean relative humidity of the air dropped from over 90% to below 70%.

Heat island effects in urban areas can result in 2–4 ℃ increases in mean annual air temperature in the urban core, resulting in lower air pressure which can suck in passing thunderstorms as occurs in northwest Calgary. This results in more frequent extreme precipitation events as the city increases in size [19]. The local changes in smog in urban areas in Europe and China resulting from changes in heating from burning coal to using other cleaner forms of heating have altered both the quantity of insolation received at the surface of the ground as well as reducing the quantity of heat lost by re-radiation back into space. This has been found to be a contributor to the warming of the mean annual temperature of the surrounding air.

3. Modifications of the Arctic air mass as it circles around the Northern Hemisphere today

The year is traditionally divided into spring, summer, autumn and winter. This division was developed and is based on humid temperate climates such as Europe, but it does not work too well in areas with continental climates such as North America and Siberia. Instead, these regions experience a long winter and a shorter summer, with these alternating in short bursts during the equinoxes. Accordingly, the following discussion is divided into typical winter and summer regimes.

3.1. Typical winter weather patterns in North America

The coldest Arctic air leaves northeastern Siberia in winter with temperatures in the –50 °C to –60 °C range following three paths (Figure 4). Path I is long and passes eastwards across the North Pacific Ocean. The air is warmed on the way by the open water and picks up great quantities of moisture, arriving at Haida Gwai and the west coast of Vancouver Island with typical mean January air temperatures of –5 °C to –15 °C. It produces extreme orographic precipitation (snow) as it passes over successive ranges of mountains, descending on the lee side of the Rocky Mountains as a dry, warmer air mass. The resulting dry Chinook wind causes the snow pack to thaw and/or sublime. This air then joins the cP air circulation and may represent an escape route for excess Arctic air when there are excessive volumes trapped by the Subtropical air masses to the south. Lakes in the deep montane valleys help replenish some of the moisture in the air [20]. Rising over the highest mountains by the continental divide, they produce enough precipitation to produce glaciers, and support a temperate rain forest on the western slopes of Mount Robson.

In winter, Arctic air following path II moves across the Bering Strait, coming ashore in Alaska and the Yukon Territory. It is warmed on the relatively short journey, but only to –30 °C to –45 °C. It is therefore not dense enough on arrival to be able to push its way south. It therefore stays in place, cooling by re-radiation into space, due to there being negligible solar radiation at these latitudes in winter. Where there is tundra on the mountain tops but boreal forest at lower elevations, the lack of vegetation on the tundra produces enhanced cooling by re-radiation. When the air on the mountain tops becomes sufficiently colder and denser than that in the valleys, cold air drainage takes place, producing MAATs as low as –65 °C in the valley bottoms [17]. Meanwhile more Arctic air piles up behind the cooling air mass, waiting to continue moving on. Once the stationary air mass has cooled sufficiently thus becoming relatively dense, it commences moving south with the main mass of the Arctic air following behind it. However the cold air trapped in the deep valleys remains, behind as can be seen by examining infra-red imagery. This is the reason for the southern boundary of continuous permafrost swinging south over the Yukon part of the mountains (Figure 4). In western Alaska and on the eastern slopes of the Asian side of the Bering Strait, the cold air drains away down-slope to the Bering Sea.

The third path (III) carries cold, dry air from north central Siberia over the Arctic Ocean to the northern Canadian Archipelago and the exit from Hudson Bay (Figure 4). As the air mass passes over the Arctic Ocean, it warms somewhat to arrive in northern Canada with temperatures around –25 to –40 °C. Once again, the air usually cools before moving south into eastern Canada and

New England. Occasionally during break-outs, a cold cell breaks off and comes west across the Prairie Provinces, but most of the cold air moves south across Labrador towards Florida. These three paths are all in use periodically during the current weather patterns, but paths II and III vary in relative frequency from year to year.

3.2. Summer weather patterns in North America

Two main subtropical air masses dominate the south and central portions of the continent (Figure 3). The cT air from the deserts of Arizona and Nevada brings hot, dry air north in western North America up to about latitude 57°N where it interacts with the Arctic air. The hot, humid mT air mass from the Caribbean Sea moves north-northeast from the Gulf of Mexico across the central and east coast, bringing rain to these areas. When they interact with much colder air, tornados are common, while hurricanes can move across the southern United States or Mexico from either the Atlantic or Pacific Oceans. The warm Gulf Stream carries large quantities of heat along the eastern seaboard before turning east towards Iceland and Europe. Periodically, hot, humid mT air moves northeast from the vicinity of Hawaii to bring precipitation to western Canada.

4. Evolution of the climate over the past 75 Ma

Since the Arctic air mass is generally moving eastwards, what happened in eastern Asia has a direct effect on what happened in North America during this period. Accordingly this is best discussed by describing the climatic changes experienced in north-east China and Siberia, then those in North America, followed by a discussion of how they appear to be related.

4.1. Terminology for the ages of the cold Late Wisconsin cold events

There are several ways of subdividing the Wisconsin cold events, i.e., those occurring after 120 ka B.P.. The earlier subdivision into Early, Middle and Late Wisconsin in North America does not fit the newer data, so that the division into the Early Glaciation, a relatively warmer Interstadial, and Late Wisconsin Glaciation will be used here, as proposed by [21]. As indicated by the previous discussion, the North American cold climatic events are very dependent on those that take place in China, whose ages are relatively well established. Accordingly the terminology for the events will be based on the sequence that has been observed in Northeast China. It follows that used elsewhere as far as possible, but using the term Late Wisconsin for the glacial/periglacial cold events occurring after about 65 ka B.P.

An alternative is the use of the Marine Isotope Stage numbers (MIS) [22], but these are based on the sequence of oxygen isotope values obtained from shells of foraminifera in ocean cores obtained from the sea bed. They are dependent on the changes in temperature that have taken place at that particular location within the water column and on the sea floor rather than on the land areas of the world. While they have been very useful as a general reference for the comparison of climatic changes in many locations, there are over 100 of these numbers but only about 13 major cold events on land in the last 3.5 ka [23]. As we obtain more data from land areas, it is becoming clear that the climate changes vary from place to place on land at any one time, as will be seen in the following review. Accordingly, the MIS numbers are omitted here, and the results of dating of sediments representing the results of changes in climate in the relevant area are used instead.

4.2. Eastern Asia and North-east China

Prior to 75 ka B.P., this region was experiencing warm conditions, but shortly before 65 Ma, glaciers of the last glaciation on the higher mountains started to advance [24]. The available evidence suggests that in spite of minor climatic fluctuations, this was the start of the Late Wisconsin cold event in this region [4]. Between 65 ka and 30 ka B.P., the climate fluctuated between glacial advances and warmer intervals. The site where the most complete mammoth fossil skeleton was found yielded dates between 60 ka and 27 ka B.P. for the associated deposits followed by a major discontinuity [25]. The glacial outwash gravels found in the Mengyuan section on the lower slopes of the Qilian Mountains are overlain by sediments dated at about 19–30 ka B.P. [26,27]. The underlying gravels contain the casts of large blocks of ice blocks that have been infilled by the overlying sediments when the climate ameliorated sometime after about 20 ka B.P. [3,27]. Thus these blocks of ice in the gravels must have remained at sub-zero temperatures from the time they were deposited until at least 19 ka B.P..

About 30 ka B.P., the East China Monsoon failed either due to low sea levels or to reduced insolation due to the effect of the Milankovitch cycles, or in all probability a combination of both. This exposed much of the South China Sea floor as dry land [28]. Subsequently, north-east China experienced a period of at least 10 ka when exceptionally dry and cold conditions produced rock tessellons over a wide area at the higher elevations on the Qinghai-Tibet Plateau [29], as well as the development of both loess and sand tessellons. The evidence from the Menyuan section located on the lower slopes of the Qilian Mountains dates this extreme cold period at c.30 ka to about 19 ka [27]. There is also abundant evidence for the development of other primary tessellons across the adjacent Hexi Corridor at altitudes between 1100–4400 m a.m.s.l., as well as the presence of mirabilite at the same period of time in these deposits [4].

Thus the area was being subjected to intense cold for over 10 millennia, followed by an amelioration of temperature accompanied by wetter conditions as the East Monsoon rains reappeared, and the floor of the East China Sea gradually became inundated by the rising sea levels as glaciers in other parts of the world started to melt. The summer rain penetrated the ground forming ice-wedges and gradually warmed the region until about 10,500 years B.P., resulting in the gradual thawing of the permafrost on the higher ground. Aeolian sand and loess were deposited in the Eastern Qinghai-Tibet Plateau from 12.4–11.5 ka B.P. and 10.0–8.5 ka B.P. [30], indicating short, colder and drier events. The climate became warmer and wetter in the Gonghe Basin (2800–3300 a.m.s.l.) until about 8,500 years B.P., although more cold periods were recorded at 9.8–9.5 ka and 9.2–8.7 ka B.P. [31].

The available evidence suggests that the warm Megathermal/Altithermal event only lasted about 2.2 ka in Northeast China, starting about 8.5 ka B.P. and ending around 6.3 ka B.P. at Yituli'he. There, ice-wedges started growing on floodplain sediments that have survived subsequent climatic events until the present day [32,33]. These conclusions are supported by stable isotope data which show a gradient in deuterium excess increasing eastwards and northwards from the environs of Lake Baikal to northeast China and Japan. Around Lake Baikal, boreal forest started to develop around 8.1 ka, reflecting a change from colder and drier conditions [34,35].

Northwards in the coldest part of Siberia, Murton et al. [36] have examined the upper 50–80 m of sediment exposed in a mega slump at Batagnika in the Yana Uplands near Verkhoyansk. There, two ice complexes are separated by a fine sand unit with syngenetic ice and multiple palaeosols. The sands accumulated by wind deposition on the slopes of hills while ice bodies developed within them. The upper ice body overlies a lens containing woody remains of *Larix* on the top of the upper sand unit. The wood yielded a radiocarbon date of 49, 320 +/− 3150 radiocarbon years B.P. This suggests that the upper ice body corresponds to the equivalent of the 20–30 ka B.P. cold event on the Tibetan Plateau, except that the cold period with sufficient moisture to produce ice bodies started earlier here. The woody layer was interpreted as corresponding to an interstadial period, whereas the underlying sands contain a herbaceous pollen flora suggesting the presence of open tundra. This appears to indicate that most of the Late Wisconsin climate at this location was intensely cold. The main evidence for Holocene amelioration in the climate comes from a more coastal site at Berelyokh where mammoths dated at 8431 radiocarbon years B.P. appear to have sunk into thawing sediments on the floodplains, resulting in their burial. Lozhkin and Anderson [37] suggest they were buried by sediments slumping down the adjacent hill slopes or by the results of their struggles in the thawing sediments.

4.3. North America

The last major Glaciation in North America (Wisconsin–Central North America; Fraser–British Columbia; Pinedale–Central US Rocky Mountains; Devensian–Europe) took place between 120 ka and 10 ka B.P. It followed the high sea levels and higher air temperatures of the Sangamon Interglacial (Eemian–Europe) with variable but lower mean sea levels (Figure 5), indicating the accumulation of ice in glaciers elsewhere on land. At first during the Early Wisconsin glaciation, the reduction in sea level was generally about 30 m although there were significant fluctuations. Around 80 ka B.P., there was a period when the Bering land bridge may have existed (>50 m sea level lowering), signalling significant glacial advances. This was followed by a short, warmer period with higher sea levels.

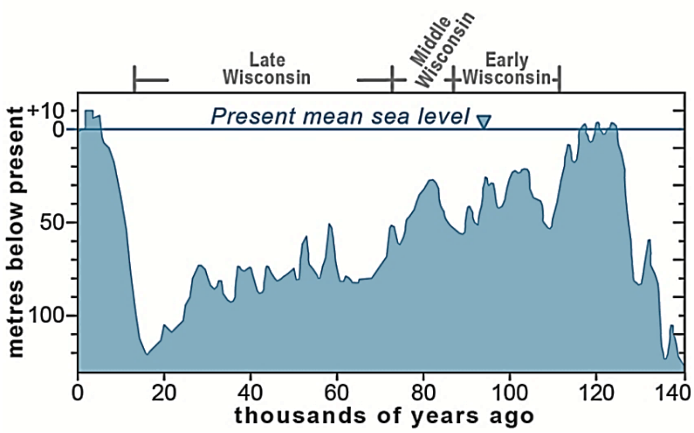

Figure 5. Mean sea levels during the last 140 ka B.P. (http://rst.gsfc.nasa.gov/Sect16/Sect16_2html).

Currently, the Wisconsinan glaciation is usually divided into early and late events, separated by a relatively brief, warm period [21]. The Early Wisconsin glaciations mainly affected the north and northeast parts of Canada, and partial deglaciation took place around 90 ka B.P. although it was both diachronous and incomplete. The Early Wisconsin glaciation also affected the extreme south of the Canadian Cordillera, with that ice finally disappearing about 59 ka B.P. [38], so that all three paths of transport of cold Arctic air from northeast Siberia were in use. The core ice centres that survived the Middle Wisconsin warming included the Keewatin, Baffin Island, Inuitian, Labradorian and Cordilleran centres, but these centres did not always produce glacial advances at the same time, and their exact locations are still being debated [39].

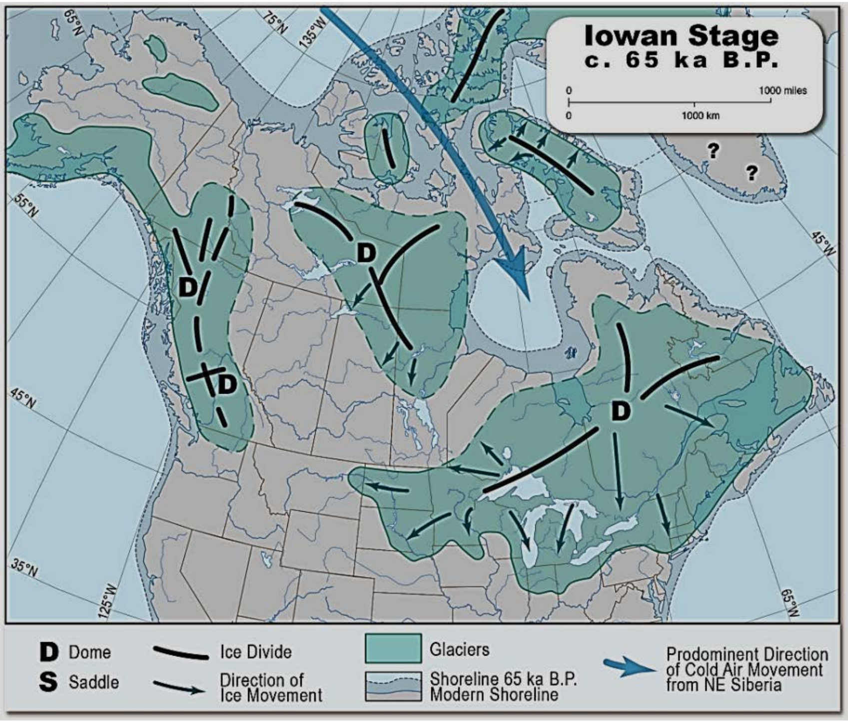

Figure 6. The apparent distribution of glaciers in North America during the Iowan Stade (65.5 ka) [40].

Figure 6 shows the locations of the main core ice centres together with the known area covered by glaciers during the first (Iowan) major ice advance across the northern tier of the American States. They are dated at about 65 ka, and are presumably the result from the interaction over that area of the cold Arctic air of the Labradorian Ice Core and the mT air coming up from the Caribbean Sea. To produce this, the path III of the Arctic air must have been dominant, although some Arctic air crossed from northeast China via paths I and II. Shortly afterwards, the cold air supply using path I ceased and the ice sheet in southern British Columbia melted.

The available evidence suggests that the subsequent early Late Wisconsin ice advances consisted of numerous localized lobes that advanced and retreated/ablated at different times. In the earliest and last stages, they alternated with non-glacial events in southern Canada. In northeast Canada, the ice sheets are thought to have persisted through the preceding warm period [41], and expanded southwards periodically during the Late Wisconsin events. In the central United States of America, the Wadena lobe entered western Minnesota from the west and is regarded as being older than 40 ka B.P., while another lobe entered Wisconsin and Michigan from the north, advancing back and forth three times across northeast Illinois and twice across Michigan and much of Indiana between about 45 ka and 29 ka B.P. [42–44]. The resultant tills are separated by interstadial deposits indicating warmer conditions. A single till containing wood dated at older than 44.2 ka B.P. divides the long, cool interstadial interval (Port Talbot) into two parts, representing the early part of the Late Wisconsin deposits along the north shore of Lake Erie in Southern Ontario [45]. Thus the period dated at 65–30 ka B.P. involved periodic advances and retreats of ice into this region, as was the case in China.

During this time in Canada, a glacial lobe moved westwards across southern Manitoba towards Saskatchewan from the Labrador Ice Centre with a large glacial lake ahead of it about 33 ka B.P. [46]. Christiansen [47] described a till older than 28 ka B.P. in the Floral Formation. Their actual ages are unknown, but the tills in Manitoba and Saskatchewan appear to represent the last part of the first cooling phase in eastern North America of the Late Wisconsin glaciation (65–30 ka B.P.). The till in the Floral Formation appears to be the result of a short cold event when an ice stream advanced from the Keetwatin Ice Centre lying to the north. These advances were punctuated by warmer intervals represented by nonglacial sediments containing cold temperate faunas, e.g., the Riddell Local Fauna [48]. However, similar glacial deposits dated at about 65 ka B.P. have not been reported from the western Canadian Prairies, indicating that the central Canadian Prairies were ice-free.

These and similar ice advances on other continents resulted in a further drop in sea level of over 50 m at 30 ka B.P. (Figure 5) that lasted until about 19 ka. As a result, the Bering land bridge opened at about 58 ka B.P. and persisted until about 11 ka B.P. when the sea level rose above 50 m a.m.s.l once again [49]. The marked drop in mean world sea level coincided with the extreme cold and dry climate on the northeast Qinghai-Tibet Plateau. This corresponds to some $4x10^9$ km^3 of water having been removed from the ocean basins and presumably sequestered as glacial ice somewhere on the land areas. It amounts to a volume of water that would be more than 100 m deep covering an area of the entire land surface of the Earth. Obviously, the ice in the centres of glaciation had to expand enormously to account for the resulting ice mass.

The Bering Land Bridge prevented Pacific water from entering the Arctic Basin and would have helped in the development of an ice cover over most of the Arctic Ocean, and cold, dry air leaving Siberia would not be modified too much when crossing along paths II and III (see Figure 4). After 59 ka until about 29 ka, path I south of the Wrangell-St. Elias Range operated as today, and path II only increased in frequency of use after 40 ka B.P. at about the same time that the Inuitian ice centre became an important ice centre [50].

Modelling the deglaciation of the Eurasian ice-sheet complex, Patton et al. [51] concluded that only 17 m of this sea level change could be explained by snow accumulation on that ice sheet by 29 ka B.P. The maximum ice volume added to that centre coincided with its maximum extent at about 22.7 ka B.P. and amounted to about 20 m decrease in sea level. There is insufficient evidence from the Alps to determine the ice volumes there, but it is possible that the present estimates of ice distribution in the Pre-Alps are too low [52]. Even so, the total volume was probably rather small. This leaves up to 30 m of sea level reduction to be explained by increases in ice accumulation elsewhere. Figure 7 shows the apparent distribution of glaciers at about 25 ka, assuming that 30 m of the reduction in sea level was accommodated in the Laurentide-Greenland ice sheet, and assuming that these ice sheets fused [49].

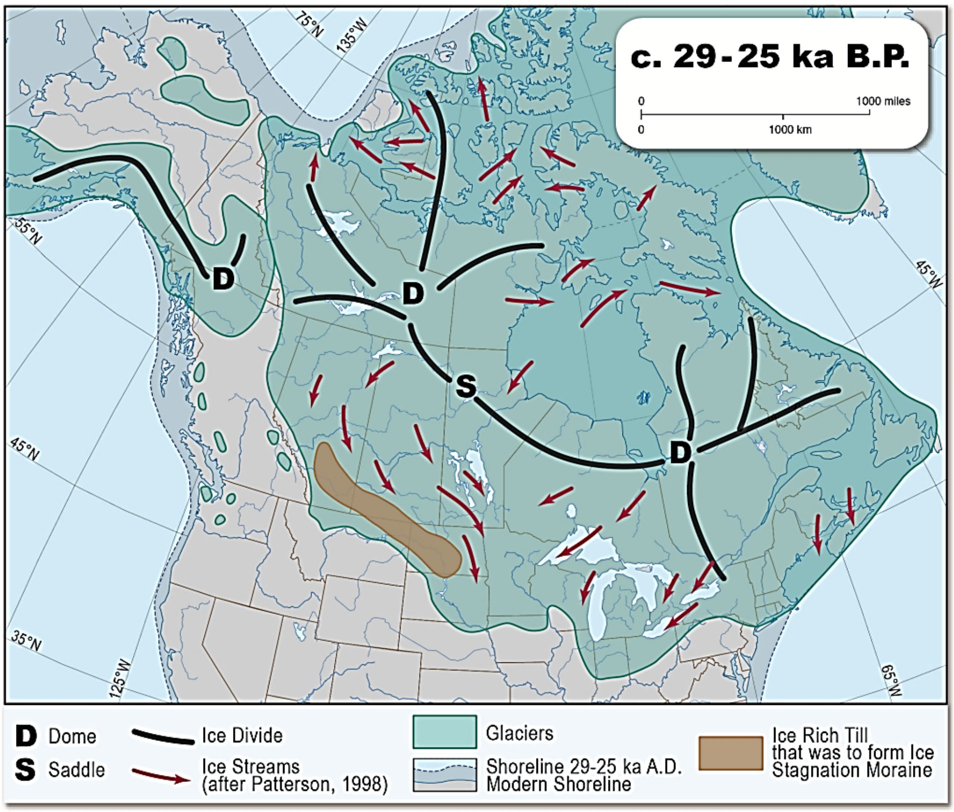

Figure 7. Probable distribution of the Laurentide ice sheet and Cordilleran glaciers during the period of 29 ka–25 ka B.P. Also shown are the ice streams from Patterson [53], together with the position of the ice-rich till that was to produce the arcuate ice stagnation moraine.

Accommodating this ice as part of the Laurentide Ice Sheet indicates that the concept of limited glaciation during the period prior to 30 ka [54–57] is impossible. Recently, Andriashek and Barendregt [58] have encountered multiple tills in boreholes in North-central Alberta. Recently, it has been realized that the movement of ice in polythermal ice sheets often moves fast in valleys and low-lying areas where there is a substantial body of water at its base, but more slowly over ridges due to greater friction [59,53]. The absence of tills from much of the prairies at this time can be readily explained by the ice sheets being primarily cold-based except where bodies of water drained along elongate depressions underneath the ice. The probable ice streams underlying the Laurentide ice sheet about 25 ka (Figure 7) are described in detail by Margold et al. [60] and were similar to those currently described from Antarctica. While tills are normally absent from the ridges where the ice was certainly cold-based and slow moving, tills are sometimes found along the beds of the former ice streams. The ice streams formed a dendritic pattern with their speed of movement increasing with the width of the channel. The dendritic pattern of the ice streams changed with time resulting in criss-crossing of channels, especially during deglaciation.

Close to the Rocky Mountains and along the southern margin of the glaciated areas on the Prairies, other authors [46,47,61–69] have described multiple tills in particular sections or transects, usually separated by interglacial deposits. These areas were probably warmer due to the occurrence of chinooks descending the eastern slopes of the Cordillera, and were the location of warm-based glaciers during deglaciation.

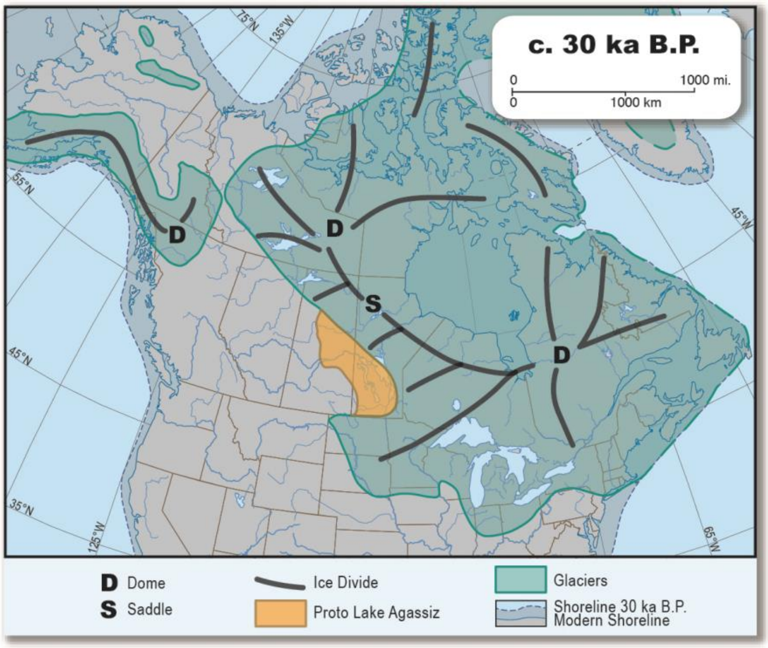

Figure 8. About 30 ka B.P., coalescence of the Keewatin and Labrador lobes resulted in the development of a large Proto Lake Agassiz accumulating against the front of the ice, preventing the margin of the ice freezing to underlying sediments. Part of the unfrozen lacustrine sediments were entrained in the ice as it advanced west southwestwards up the Saskatchewan River valleys, aided by heavy lake effect snowfall. Elsewhere, the advancing ice lobes were generally cold-based due to the extremely cold Arctic air.

In the beginning of this glacial expansion at about 30 ka, the eastern ice centres must have blocked the exit of river drainage from the west to the North Atlantic via the Hudson Bay. This would have produced a sequence of large lakes around the vicinity of the present-day Great Lakes, ultimately resulting in the development of a major Proto-Lake Agassiz glacial lake in front of the eastern ice masses (Figure 8). Subsequently, the western margins of these glaciers appear to have raced upslope across the southern Prairies to arrive at the gates of Waterton National Park in Alberta by about 12 ka B.P. They were probably fed in part by increased up-slope precipitation on the western slopes of the ice cap, fueled by the lake effect causing vastly increased precipitation on east side of the adjacent proglacial lakes. The unfrozen sediments in the lake were incorporated into the westward-moving glaciers, resulting in the development of very sediment-rich ice. When the ice began to melt about 12 ka B.P., a large area of stagnant ice remained in an arc south from Edmonton Alberta to North Dakota [70,71]. Ice in this moraine lasted for up to 4 ka until about 8 ka B.P. after the ice elsewhere on the prairies had thawed [9]. Unfortunately, the exact age of this advance is unknown, but the ice streams go around these moraines (see Figure 7).

Ryder et al. [72] concluded that there were no Late Wisconsin glacial advances in southern British Columbia until 29 ka B.P.. The glaciers along the Western Cordillera only expanded a limited distance out of the cirques until about 25 ka B.P. after which they started advancing and coalescing in the valleys [73,74]. The Last Glacial Maximum (LGM) is usually regarded as occurring at 21 ka B.P., but this was probably not the case in North America in southern British Columbia. At about 15 ka B.P., the Cordilleran glaciers in the southern Rocky Mountains of Canada had formed a dome centered at about 54°N, infilling all the valleys [73,74]. Jackson and Clague [75] provide a summary of the ice-flow directions. This resulted in a persistent chinook wind that descended down the east slope of the Cordillera [76], thawing that margin of the Cordilleran ice sheet before the arrival of the Laurentide ice along the Oldman River valley [64,66,77,78]. This substantial variation in timing of the onset of deglaciation across the continent makes estimation of the time and volume of maximum the ice cover over North America very difficult to calculate.

Along the southern margin of the ice sheets in Southern Ontario and the coastal areas of New England, warm-based lobes and sheets of glacial tills reappeared after about 24 ka B.P. [45,79]. About 20 ka B.P., a major lobe of warm-based ice advanced from the north across Southern Ontario [80], followed in the subsequent millennia by several more ice advances marking interuptions in the retreat of the parts of the active ice sheets to a morainal position around Hudson Bay dated at about 9 k radiocarbon years B.P. [9]. On the adjacent Prairies, similar sequences of tills appeared beginning about 17 ka B.P. marking dwindling ice advances and subsequent major retreats to this 9 ka B.P. moraine [46,81]. Thus the period from about 20 ka to 9 ka B.P. was marked by gradual warming, resulting in the redevelopment of a succession of warm-based glaciers around the ice sheets in eastern Canada, but which occupied steadily decreasing areas of land until 9 ka B.P. This indicates fluctuating but decreasing snowfalls accompanied by warming mean annual air temperatures.

This warming trend resulted in the development of Glacial Lake Agassiz along the ice front in Manitoba, Saskatchewan, North Dakota and Minnesota [82–85]. The highest level was the Lockhart Phase (12,875–12,560 a B.P.) when the water overflowed westwards to Alberta, then turning north to join the Clearwater-Athabasca spillway. Fisher et al. [85] attributed the Pre-boreal climate oscillation to the discharge of these waters into the Arctic Ocean at 11,335 cal. years B.P.. They consider this flood as increasing the albedo and generating a low salinity anomaly as the ice pack flushed through the Fram Strait and melted in the North Atlantic. This, in turn, decreased the formation of North Atlantic Deep Water. They suggested that this was the major part of the cause of the world-wide climate fluctuations between then and the Younger Dryas about 11.1 ka B.P.. Subsequently, the outlet of Lake Agassiz changed to the Mississippi drainage and channels along the Great Lakes area as the advances and retreats of the ice lobes permitted.

Glacial Lake Agassiz finally drained into the Tyrell Sea that was occupying the Hudson Bay basin in 7490 a B. P. [86]. The Keewatin ice centre became the first separate ice centre to disappear around 8 ka B.P., well before the more eastern ice centres [49]. The remnants of the Laurentide ice centre disappeared between 6 and 7 ka B.P., while the Baffin ice cap stabilized about 5 ka, remaining as the Penny Ice Cap.

There was a general change in vegetation in the Prairies around 10–11 ka. Although a steppe-tundra fauna was obtained from Bow River outwash gravels at Cochrane and Calgary dated at 11 ka B.P. [87–89], the main remaining active Laurentide ice had retreated back to the stand-still position around Hudson Bay by 9 ka B.P., except for the belt of stagnant ice [9] and smaller patches of stagnant ice scattered across the Prairies [71]. Ritchie [90] reported that as the ice melted across the western interior of Canada, forest moved in. In the Rocky Mountain foothills, MacDoinald [91] reported a transition from tundra pollen, with a cold ostracod and molluscan fauna to a Boreal Forest assemblage beginning at c. 11,150 a B.P. in a core from the bottom of Wedge Lake, lower Kananaskis valley, after which the climate ameliorated. By about 11 ka B.P., Gryba [92] found evidence for Boreal forest in the Foothills at Sibbald Flats.

A cold, higher precipitation period about 9.6 ka B.P. resulted in a readvance of ice from a stagnant ice in the foothills west of Innisfail, Alberta, producing a short-lived proglacial lake along the Bow River near Calgary, together with a substantial flight of terraces [76]. However this soon ended as the Altithermal/Megathermal event warmed the region. Evidence of climatic fluctuations prior to or early in the Megathermal event include fossil rock glaciers that extend down to 200 m below the present limit of permafrost in the mountains of Banff and Jasper National Parks, and 8 ka-old tree trunks that have thawed out of the Sunwapta glacier ice as it recedes today [93]. Additional undated fluctuations include the Crowfoot moraine [94] occurring between Jasper National Park and northern Montana, and the double Chateau Lake Louise moraines which are common in front of glaciers and glacial cirques from Summit Lake on the Alaska Highway west of Fort Nelson, British Columbia, south to the Kicking Horse Pass [95]. Some of these fluctuations may be the same age as the climatic fluctuations dated between 14 ka and 8 ka in China, although the dating control is too imprecise in Canada for a direct correlation at this time.

There was a significant change in vegetation identified by most palynologists at 6.5–6.0 ka B.P. indicating the end of the Altithermal/Megathermal warm period in the Canadian part of North America. Zoltai et al. [96] obtained ^{14}C dates of this age from the organic matter buried in oscillating earth hummocks from both the Low and High Arctic sites indicating the presence of permafrost throughout that area during this time period. The dates from these earth hummocks also provide evidence of the presence of permafrost in both the Arctic and Subarctic between 5.0–4.5 ka, 3.5–2.0 ka, and 1.5–0.5 ka B.P.. Cold conditions returned in approximately 1500 A.D to 1900 A.D. Glacial advances and growth of rock glaciers, peat plateaus and palsas, together with the development of patterned ground also occurred. These provide the best evidence for the Neoglacial events, with an estimated cooling of 2 °C below 1980 temperatures [97].

5. Discussion and conclusions

5.1. The Arctic air mass and its controls

It is clear that the Arctic air mass consists of a large discrete body of air circling around the North Pole in an eastward direction. As it moves, it is either warmed and picks up moisture over the oceans, or becomes greatly cooled in winter over land due to negligible solar insolation and intense re-radiation of energy back into space. Today, the coldest temperatures occur inland in eastern Siberia around Verkhoyansk (−67.8 °C) and this is in spite of the import of heat from the Tropics primarily by ocean currents, as well as the Eastern Summer Monsoon coming from the South China Sea. If part of these heat sources fail, then the winter cold temperatures in Siberia would be closer to the < −80 °C temperatures of Antarctica. Thus the sea levels are critical in keeping the ocean gateways such as the Bering Strait open and in determining the degree of cold in the Arctic air emanating from Siberia in winter. As shown in Figure 4, this air moves east over North America, determining the winter temperatures there. Meanwhile, the heat that is presently transported north in the South China Sea would tend to remain in the Tropics when that sea no longer existed. When the warmer, moist, maritime Subtropical air moved north to meet the extremely cold Arctic air, they would interact more violently than at present and would have produced greater snowfalls. The belts occupied by these major air masses definitely changed so that aeolian sand dunes were active in Guyana in what is now Equatorial Rain Forest, and the amplitude of the Rossby waves may have been much greater. In North America, there was a north-south temperature gradient resulting in heavy cold snowfall in the north, grading southwards into heavy cold rain in the south. Using pollen analysis, Ritchie [90] reported the presence of Boreal forest type vegetation in the unglaciated areas between the glaciers that extended south to Kansas. Cold-blooded invertebrates such as snakes only survived the coldest period in the Southern United States south of the latitude of northern Texas, but are now spreading north to northern Michigan and the St Lawrence River [98,99].

5.2. The relationship of sea level to the Late Wisconsin climate

Sea level indirectly affects climate by altering the position of the ocean currents or eliminating them entirely, e.g., that along the China coast. They also are involved in opening or closing ocean gateways, e.g., the Bering Strait. These cause changes in the heat supply to areas over which the Arctic air passes resulting in temperature changes downwind. Most sudden changes in world sea level are due to sequestering of liquid water on land in the form of ice and snow. Obviously, there should be a close relationship between sea level fluctuations during cold climatic events with the growth or demise of the major ice sheets on land.

The winter temperature of the Arctic air leaving Siberia during the early part of the period after 30 ka B.P. would be in the −60 °C to −70 °C range and would have ensured that the Arctic Ocean would be covered by an ice pack that could not greatly warm the air passing over it. On arrival over the Canadian ice centres, it would lower the temperatures on the ground, thus ensuring that they consisted of cold-based ice. Meanwhile the maritime Subtropical air coming from the Gulf of Mexico would be warmer and carry more moisture. When the two air masses interacted, there would be enhanced snowfall on to very cold land surfaces which would tend to produce larger cold-based glaciers and ice streams that would have advanced westwards from the existing ice centres. These glaciers would leave no obvious till deposits and would eventually reach an equilibrium situation after major ice advances. These would also provide the ice for the numerous ice streams that have been postulated by various authors in different parts of the North American landscape (see the summary by Patterson [53]. These ice streams would be part of the major cold-based ice advances across the landscape from the contemporary ice centres headed towards the mountains of the Western Cordillera, in contrast to the concept of Dredge and Thorleifson [54] who suggested that the 30–19 ka B.P. period was one of warmer, minimal glacial conditions.

5.3. The origin of the vast area of stagnation moraine stretching from SW Alberta to North Dakota

The vast ice-stagnation moraine stretching in a crescent-shaped arc from just east of Edmonton to North Dakota is clearly of a different origin. The till in this stagnation moraine contains c. 10–20% granite clasts as opposed to c. 45% granite clasts in the Laurentide ground moraine to the east. The Laurentide tills closer to the mountains only contain 1–5% granite clasts. Thus the glacial ice that left the hummocky moraine in an arc between Edmonton and North Dakota had picked up a considerable quantity of both Cordilleran and Laurentide clasts, and this landform is most easily explained as being formed by a glacier pushing into an adjacent glacial lake such as a proto-Lake Agassiz. The latter would have developed due to the drainage flowing east-north-eastwards downhill slope from the mountains of the western Cordillera to the Hudson Bay area. When cold air masses pass over water bodies, they pick up large amounts of moisture. This is then deposited as thick, heavy snowfalls downwind. This lake effect would have piled great quantities of snow on to the front of the ice sheets holding in the proto-Glacial Lake Agassiz, resulting in the ice advancing westwards over the unfrozen lake sediments, much of which would have been incorporated into the advancing ice mass. The formation of the stagnation moraine along the southwest margin of the prairies most probably predated the main development of the major Laurentide Ice Sheet since the pattern of the ice streams suggests that they were largely deflected east of the stagnation deposits before heading on southwards along the Mississippi valley. Some spillways come from the stagnation deposits and drained away along the low areas left by the ice streams, but the general pattern strongly suggests that the stagnant ice arrived at its final destination first.

5.4. The Late Wisconsin glaciation west of the continental divide along the Rocky Mountains

Other parts of the climate system besides sea level can change and result in a serious climatic change in a localized area. An example from the present study is found west of the continental divide in British Columbia where the extreme cold conditions of the Late Wisconsin period in northeast Asia and eastern Canada had less effect in the early part of the Late Wisconsin cold period. Apparently, warm conditions with merely cirque glaciations on the higher mountain peaks prevailed after 52 ka B.P. until the beginning of the cirque glacier advances in about 29 ka B.P. These started slowly but were not apparently related to changes in sea level air coming from Siberia. Since glaciers respond to both changes in precipitation as well as temperature, it seems probable that the climate of southern British Columbia between 59 ka and 25 ka B.P. was similar to today. The North Pacific would have had to be ice-free. Substantial growth took place after 25 ka B.P., probably signalling a gradual switch in the dominant cold Arctic II and III pathways to move to pathway I (Figure 4). By 20 ka B.P., path I became a major route resulting in the rapid growth of the Cordilleran Ice Sheet at the time of commencement of deglaciation in the central part of Canada. Downwasting of the ice sheet was probably aided by an increase in frequency of the pineapple express. As in the Eurasian ice sheet complex, the timing of the glacial maximum and deglaciation in the two parts of North America on either side of the continental divide were asynchronous, with each sector becoming deglaciated under the influence of its local microclimatic environment and geographical controls rather than sea level change.

5.5. Post 19 ka B.P.

On both sides of the Pacific Ocean, the slow recovery from the extreme cold was similar except in British Columbia. The warming was slow and punctuated by set-backs during the Pre-Boreal event, the Younger Dryas and the Holocene. On the whole, the two areas experienced the same general pattern of change. The regional variations in Late Wisconsin climatic history of subpolar and polar regions of both continents are very closely tied to the sea level changes along the coasts of both continents and the sudden discharges of large volumes of glacial meltwater into the Arctic and North Atlantic Oceans from both the Laurentide and Scandinavian ice caps about 11–12 ka B.P. The timing and degree of warming during the Altithermal/Megathermal events depended on local conditions since the residual ice centres disappeared at significantly different times. However the Neoglacial events started about the same time, but once again, the number and degree of cold depended on the geographic and local environmental conditions. Thus in detail, it is not possible to safely infer intercontinental climatic changes from those at a single location.

In British Columbia, glacial advances continued into coastal areas such as the Fraser River valley until about 10 ka B.P. Inland, the ice in the major valleys down-wasted *in situ* after about 14 ka B.P. resulting in numerous glacial lakes occupying the main valleys, each of which had its own history depending on the local geography and environmental factors. During the Holocene, at least one glacier or group of glaciers was advancing in each millennium of the Holocene in some area of northern British Columbia or Alaska, making identification of regional climatic changes difficult, particularly in the north. On both sides of the Pacific Ocean, the megathermal/hypsithermal warm event decreased in length when traced northwards until it was difficult to discern in the sedimentary record.

5.6. Effects on the biota and human migrations

The extreme climatic changes greatly affected the biota. The first human crossing of the Bering Land Bridge may have occurred about 33–40 ka B.P. [100–103] before the extreme cold set in when a few people, probably related to the ancestors of the Sakha, entered North America. No further successful crossings appear to have occurred until the advent of the Clovis people about 12–14 ka B.P. [104], almost certainly because of the extreme cold of the intervening times. The third wave consisted of the Late-glacial Siberians [105], followed by the Inuit around 9 ka B.P.], when Beringia had warmed up.

The Land Bridge extended for about 9° latitude at its maximum (Figure 9), so that there was almost certainly quite a variation in the cold climate with latitude and proximity to the seas. The assumption by many authors that there was a single, cold-environment vegetation on one or both sides of the present Bering Strait is unlikely [106]. The topography would consist of low, rolling hills and valleys with a maximum local relief of about 120 m, and the present areas that are currently land would have looked like much higher uplands. The streams would have had to adjust their thalwegs to their new vastly increased lengths, down cutting new deep river beds inland from the new and changing shorelines so that the river valleys would be quite incised into the older landscape. Wind erosion of the sediments on the uplands probably supplied the silts for the loess deposits found around the lower slopes of the adjacent mountains.

The literature suggests that the first settlers stayed put through the ensuing cold period [102]. The cold air drainage coming from the mountain valleys on both sides of the Strait would accumulate in the depressions and valley bottoms since the gradients along the valleys would be very low and their extent very considerable. This cold air would gradually drain into the adjacent ocean basins. The climate along the new Bering Sea coast would be unhospitable [107] because the Arctic Ocean would be frozen over. In contrast, the southern coast of Beringia would probably be warmer, thanks to the proximity to the Pacific Ocean. Before 25 ka B.P., there would be the possibility for any of the c. 3000 generations of any families living along that coast during the period from 30 ka to 19 ka B.P. to move south along the exposed sea bed in British Columbia where there was undoubtedly adequate wildlife both on the unglaciated land and in the sea to help them survive. After about 25 ka B.P., they would tend to move on south to warmer Pacific coast shores, but the remains of their activities and any settlements may have been largely destroyed when these were inundated as the sea level rose.

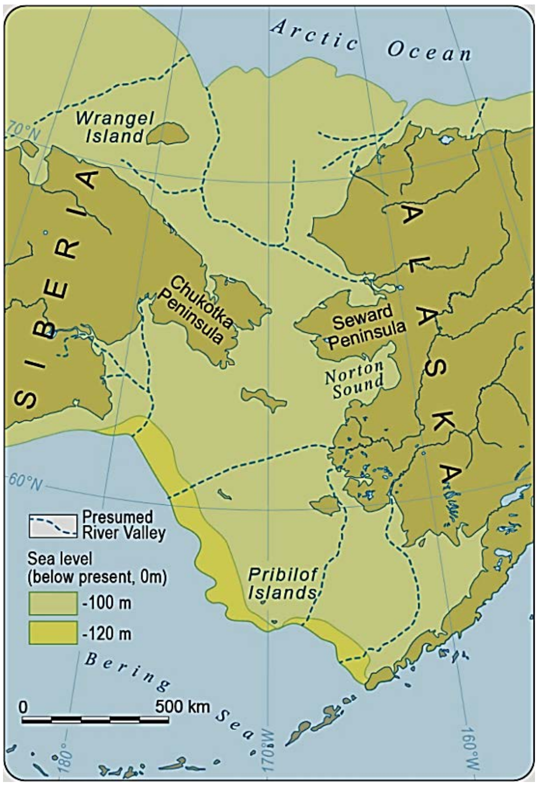

Figure 9. The Bering Strait showing the size of the land bridge and the associated inferred rivers.

By about 14 ka B.P., the climate had ameliorated inland in both Tibet and on the Bering Strait, and the glaciers had largely retreated into the mountain valleys along the eastern slopes of the Rocky Mountains which were receiving frequent Chinooks. In Tibet, the first foraging settlements were established by the Tibetans [108] about the same time that the second wave of settlers crossed the Bering Land Bridge. By about 14 ka, remains of the Clovis people had been left at the Whitefish Caves in the Yukon Territory and they had travelled as far south as El Fin del Mundo in northwestern Sonora (Mexico) by 13,390 calibrated years B. P. [109]. In 2001, remains of the Aubrey site in Denton County, Texas, produced an almost identical radiocarbon date [110]. At that time, the glaciers in coastal British Columbia were still advancing to Haida Gwai and along the Fraser River valley so that the coastal route south would have be a poor choice for the later Clovis people to reach warmer weather and more hospitable environments [111]. The third wave of settlers consisted of the Inuit-Aleuts who arrived about 9 ka B.P.. By then, conditions were suitable along the north coast of Beringia and parts of the Arctic Islands for the newcomers to use their hunting and fishing skills to make use of the relatively inhospitable coastal environment. On the whole, the role of climatic change has not been adequately considered in discussing migration patterns of biota, though this is partly due to the limited previous studies based on good palaeoclimatic data.

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Conflict of interest

The author declares no conflict of interest.

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