Varied climatic and topographic influences on Late Pleistocene mountain glaciation in the western United States

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ABSTRACT: The timing of Late Pleistocene mountain glacier fluctuations varied widely across the western United States. Glaciers in the maritime Olympic Mountains reached local maxima 125–60 ka and 34–31 ka, and were far less extensive during the global Last Glacial Maximum (LGM). Glaciers in the Cascade Mountains showed a similar general pattern but with contrasts in detailed timing. Most mountain glacier systems further inland exhibited maximum ice extent during Marine Isotope Stage (MIS) 2. In the Wallowa, Wind River, and possibly the Sawtooth Mountains, maximum ice extent occurred ca. 20–23 ka. Mountain glacier systems of the Cordilleran Ice Sheet (Puget Lobe), greater Yellowstone region and Uinta Mountains reached maximum extent ca. 19–16 ka, and those in the Wallowa, Sawtooth and Wind River Ranges were at near-maximum positions ca. 17 ka. The variable timing of maximum and near-maximum ice advances appears to be related to climatic setting and glacier mass balance characteristics. Mountain glacier systems under the strongest maritime influences reached maximum extent at times of strong westerly atmospheric flow. Olympic and Cascades glaciers appear to have been negatively influenced by weakened westerly flow at the time of the global LGM, perhaps associated with ice-sheet-induced anticyclonic flow. Glacier systems with maximum advances correlative with the global LGM appear to have responded more strongly to depressed temperature. The ca. 17 ka ice advances of several glacier systems are broadly correlative with Heinrich Event 1 in the North Atlantic region, and imply a cooling event and reinvigoration of moisture transport into the region. Post-LGM advances of Uinta Mountain glaciers appear to have arisen from moisture influences of pluvial Lake Bonneville. Copyright © 2008 John Wiley & Sons, Ltd.

KEYWORDS: glaciation; Late Pleistocene; United States; geochronology; palaeoclimatology.

Introduction

Glaciers occupied many mountain ranges in the northwestern United States during the last glaciation. Lying south of the Canadian ice sheets and spanning maritime to continental climatic zones across a broad cordillera, these glaciers ranged from isolated cirque glaciers in lower or more continental ranges to large valley glaciers and mountain ice caps in higher and/or maritime-influenced ranges. For many years, it was generally thought, albeit with insufficient chronological data, that mountain glaciers throughout the region fluctuated synchronously with the Canadian ice sheets and with each other (e.g. Chadwick et al., 1997). However, the ongoing revolution in the dating and analysis of glacial sequences has produced many new chronologies in the region, and the emerging patterns reveal regionally variable responses to climatic fluctuations. While those patterns might have once been viewed as betraying problems in dating and correlation, the volume of chronological data across several mountain uplands and the overall weight of evidence indicate that the chronological variation is real and important.

This synthesis focuses on glacier fluctuations across a broad swath of the northwestern United States and western interior. An exhaustive compilation of regional glacial chronologies has not been attempted, but focus is placed on several areas for which chronologies are particularly robust and revealing. The chronologies chosen range in setting from the maritime Olympic Mountains, Washington, to the fully continental Wind River Range, Wyoming, and the pluvial lake-influenced Uinta Mountains, Utah (Fig. 1). Glacial chronologies vary markedly along this transect, in some cases within the same upland area. Clearly, many episodes of snowline lowering occurred, with ice responses varying by climatic and topographic setting.

The current state of knowledge of regional glacial chronologies is growing, but is, of course, imperfect. Robust chronologies are limited to a handful of mountain ranges. By
their nature, terrestrial glacial chronologies are incomplete (e.g. Gibbons et al., 1984), with only the most extensive advances and subsequent less extensive phases typically recorded in landforms. Glacial chronologies typically feature ages that merely bracket the timing of the actual advance. The most widely applied dating methods, radiocarbon and cosmogenic nuclide exposure dating, have limited resolution, each for its own reasons. Luminescence dating, while having seen very limited application to glacial sequences in this region to date, also holds promise for dating glacial sequences here, albeit with its own limitations.

Nonetheless, the current state of knowledge is a marked improvement and reveals regional patterns of ice fluctuation, at least at a broad chronological scale. The goals of this synthesis are (1) to compile the most robust chronologies, (2) to elucidate regional patterns of glaciation, (3) to propose mechanisms of glacial responses to climatic fluctuations, and (4) to suggest specific research that could lead to a more thorough understanding of glaciation in this region.

Previous work

As a synthesis, this paper naturally draws from a long history of work in the region. Much of that previous work is referenced in the main text. Several previous syntheses have also addressed this region. Richmond (1965) summarised glacial fluctuations in the Rocky Mountains region, and Porter et al. (1983) provided a synthesis of Late Pleistocene mountain glaciation across the western United States. Colman and Pierce (1992) used calibrated weathering rind data to place age estimates on glacial landforms in several locations across the region, covering much of the area that is the focus of this review. They found evidence of glacier advances spanning Late Pleistocene time, as well as highly variable temporal patterns of glacier advance. Those efforts pre-dated the recent geochronological revolution and, while very revealing, included few firmly dated glacial sequences. More recently, Kaufman et al. (2004) included the maritime Pacific Northwest in a synthesis of glaciation that also included the Sierra Nevada, Alaska, and Hawaii. Similarly, Porter (2004) summarised glacial chronologies in the maritime Pacific Northwest, and Pierce (2004) reviewed glacial records in the Rocky Mountains region. In the context of these previous syntheses, this article is unique in its geographic breadth of coverage, from fully maritime to fully continental settings, and in taking advantage of recent geochronological advancements and recently published glacial chronologies.

Materials and methods

Geographic setting

The glacial systems described in this article span a 1300 km wide region of marked topographic and climatic complexity (Fig. 1). The western portion of the region lies within the Cascadia convergent margin and features largely north–south trending ranges and basins. Coastal ranges in Oregon and Washington originated through accretionary wedge processes and are mostly too low (peak elevations 1250 m above sea level (a.s.l.)) to have supported mountain glaciers. The principal exception is the Olympic Mountains, which rise to ~2700 m a.s.l. and supported extensive Pleistocene valley glacier
systems and support numerous modern glaciers. The coastal ranges lie in a fully maritime climatic regime, with modern precipitation as high as 5000 mm a⁻¹ in the Olympic Mountains. East of the coastal ranges lies the Puget–Willamette trough, a complex, low-elevation foreland basin that, in its northern portion, hosted the Puget Lobe of the Cordilleran Ice Sheet. The Puget–Willamette trough is bounded on the east by the Cascade Mountains, a complex range that has originated through a combination of terrane accretion, compressional tectonic forces and volcanic processes. The Washington Cascades, which are the focus of a portion of this review, are continuous for ~400 km and have an average width of 150 km. The range generally rises to 2800 m a.s.l. peak elevation, with local volcanic peaks as high as 4400 m a.s.l. The Cascade Mountains represent a prominent barrier to moist Pacific air masses, and precipitation consequently ranges from 3500 mm a⁻¹ on the western flank to 300 mm a⁻¹ on the eastern flank. The Columbia River Gorge transects the Cascade Range between the states of Washington and Oregon, and provides a conduit for moist maritime air to penetrate east of the Cascades.

Isolated mountain ranges mark a general low-elevation landscape across a region 500 km wide between the Washington Cascade Range and the numerous ranges of Idaho. Amongst those isolated ranges are the Wallowa Mountains, which lie in northeastern Oregon and intercept moist air masses traversing east through the Columbia River Gorge. Precipitation in the Wallowa Mountains region is as high as 2500 mm a⁻¹. In western and central Idaho, numerous ranges intercept moist Pacific air masses, engendering a pronounced east–west precipitation gradient. In the Sawtooth Mountains, which rise to 3300 m, annual precipitation is as high as 2000 mm a⁻¹, dropping below 500 mm a⁻¹ in adjoining eastern Idaho ranges.

As with the Columbia River Gorge, the Snake River Plain represents a regional lowland that funnels precipitation eastward, well into the continental interior (Fig. 1). The Snake River Plain is a relatively low volcanic province, 900–1800 m a.s.l. elevation, that transects the northern Basin and Range province. Moist air masses moving along the Snake River Plain rise against the Yellowstone Plateau and adjacent Teton Range. Portions of those uplifts thus receive anomalously high precipitation values (up to 2000 mm a⁻¹) for their location in the continental interior. The Yellowstone Plateau has a high average elevation (2400 m a.s.l.) and is surrounded by mountain ranges rising to 3400 m a.s.l., and was thus a major regional ice accumulation centre.

South-east of the Yellowstone–Teton uplifts, the Wind River Range lies firmly in the continental interior (Fig. 1). These mountains represent a Laramide-style uplift of Archaean cratonic rocks and Palaeozoic sedimentary cover, and rise to 4200 m a.s.l. The climate of the range is highly variable, being influenced by the remnants of moist Pacific storms and by moisture drawn west from the Gulf of Mexico and mid-continent. Annual precipitation averages 1700 mm a⁻¹. The Uinta Mountains lie 250 km south of the Wind River Range. This 300 km long, Laramide-style uplift rises to 4100 m a.s.l., exposing Neoproterozoic and Palaeozoic sedimentary rocks. While this range lies firmly in the continental interior, the presence of Pleistocene Lake Bonneville to the west appears to have influenced patterns of Uinta glaciation rather strongly.

Pleistocene regional atmospheric circulation patterns were probably broadly similar to modern patterns, with westerly flow being the dominant moisture transport mechanism at these latitudes. However, large Canadian ice sheets imparted strong influences on regional circulation. This was particularly so around the time of the ‘global’ Last Glacial Maximum (LGM, 21 ± 2 ka), and may also have been the case during other times when an extensive Canadian ice sheet system was present. In particular, the Canadian ice sheets appear to have created anticyclonic circulation and thus weakened westerly atmospheric flow and driven dry, easterly winds across the northern part of the region. The winter jet stream and storm tracks were diverted southward into the Great Basin region (e.g. Kutzbach et al., 1993; Thompson et al., 1993). The weakening of westerly atmospheric flow would have reduced precipitation delivery to mountain ranges across the region, and such a moisture reduction appears to be reflected in glacial chronologies in some mountain ranges.

**Methods**

This article uses only previously published chronological data. The chronologies presented are based upon radiocarbon and cosmogenic radionucleide (CRN) methods. While many glacial sedimentological sequences are amenable to luminescence dating, few, if any, published luminescence dates exist from the region.

In extracting chronological data from the literature, I have largely reported the numerical dates as published, with one exception. To render radiocarbon dates comparable with CRN dates, the radiocarbon ages have been calibrated (cal. ka BP) using CalPal2007 (quickcalc2007, version 1.5; Danzeglocke et al., 2007). The CRN ages have not been recalculated. While recalculating ages to a consistent production rate and scaling scheme would render the comparisons somewhat more rigorous, the broad geographical and temporal scale of this synthesis permits only broad correlations, and recalculation of ages would not influence the analysis strongly. Furthermore, calculation of CRN ages is still a topic of debate (e.g. Balco et al., 2008), especially with respect to production rates and their altitudinal and latitudinal scaling, and recalculation of dates used herein should await a consistent calculation scheme.

The global LGM is used formally to refer to the maximum Late Pleistocene global ice volume reflected in marine oxygen isotope records and, by correlation, to the inferred maximum extent of Northern Hemisphere ice sheets (ca. 21 ± 2 ka; Mix et al., 2001). In some regions the glacial maximum was not coincident with the global LGM and therefore this article distinguishes a local last glacier maximum.

**Results**

**Olympic Mountains**

The Olympic Mountains are the highest portion of the US coastal ranges, rising to ~2700 m in the northwestern part of Washington (Fig. 1). There, the accretionary wedge of the Cascadia subduction zone has been exposed beneath its basalt cap, and the interaction of rapid rock uplift and intensive erosion have produced a deeply incised mountain landscape (e.g. Batt et al., 2001). Annual modern precipitation ranges from ~1800 mm a⁻¹ at the Pacific coastline to ~5000 mm a⁻¹ at the crest of the range, and drops to ~460 mm a⁻¹ in the precipitation shadow at the northeastern flank of the range. That precipitation gradient engenders a steep, southwest–northeast gradient of ~12 m km⁻¹ in modern glacier equilibrium line altitude (Spicer, 1986) and a similar, but steeper, 20 m km⁻¹ gradient in the Pleistocene glaciation threshold.
(Porter, 1964). The Cordilleran Ice Sheet flanked the northern and eastern portions of the Olympic Mountains, and mountain glaciers may have been confluent with the ice sheet in those areas at the peak of the ice sheet advance.

Large valley glaciers in the Olympic Mountains descended repeatedly into the coastal lowlands during the last glaciation, leaving a rich geomorphic and stratigraphic record. Thackray (1996, 2001) constructed a detailed radiocarbon chronology for the Hoh and Queets valleys, identifying six major Late Pleistocene advances (Fig. 2). The Lyman Rapids advance, the most extensive, is constrained most closely by two minimum limiting radiocarbon dates, in valley-floor lacustrine sediments directly overlying glacial diamicton. Those two ages are reported as finite, at 54 and 51\(^{14}C\) ka BP; while these ages lie beyond radiocarbon calibration schemes, the ca. 6 ka calibration offset at 50 ka BP (CalPal2007; Danzeglocke et al., 2007) suggests these ages may correspond roughly to ca. 60 and 57 cal. ka BP, respectively. As these are very old radiocarbon ages, they should be considered \(\geq 60\) and \(\geq 57\) cal. ka BP. Several additional minimum limiting dates are reported as infinite ages (>42 \(^{14}C\) ka, >49 \(^{14}C\) ka BP), as are several maximum limiting ages (e.g. >42 \(^{14}C\) ka BP). Lyman Rapids outwash overlies a buried, last interglacial wave-cut surface MIS 5e or 5c and associated stratigraphic units (Thackray, 1998). Thus, the most conservative interpretation is that the advance occurred between ca. 60 and 125 ka (MIS 5e). An age interpretation of around 60 ka is feasible if the very old radiocarbon ages (ca. 60 and 57 cal. ka BP) are representative of original \(^{14}C\) content and not of even minor contamination.

The best dated advance, only slightly less extensive than the Lyman Rapids advance, spanned 34–31 cal. ka BP. The Hoh Oxbow 2 advance is constrained by numerous maximum limiting radiocarbon dates and a handful of minimum limiting dates, and is confirmed by independent dating of glacial sediments in the nearby Quillayute drainage by Florer (1972). Importantly, a much less extensive advance (Twin Creeks 1, 23–22 cal. ka BP) occurred during the global LGM. In fact, that LGM-correlative ice terminus position represents a retreat of 27 km from the maximum local Late Pleistocene positions.

Correlations of ice advances to cold periods revealed in local pollen data of Heusser (1972) indicate clear influences of precipitation delivery on mountain glaciation (Thackray, 2001). The Late Pleistocene pattern of generally decreasing stadial ice extent corresponds to generally decreasing palaeotemperature estimates, suggesting that the largest advances corresponded to cool periods of sustained moisture influx, but not to the coldest period. The global LGM-correlative (and far less extensive) Twin Creeks 1 advance corresponded to a cold but dry period represented in pollen records from throughout western Washington.

Puget Lobe of the Cordilleran Ice Sheet

The Puget Lobe of the Cordilleran Ice Sheet filled the lowland area between the Olympic and Cascade Mountains (Fig. 1). While the Puget Lobe was part of a subcontinental-scale ice sheet and of markedly larger scale than the mountain glacier systems at issue, the ice accumulated in the mountains of southwestern British Columbia under the same regional climatic influences. Thus, the chronology of the Puget Lobe (Fig. 2) bears relevance to the mountain glacier records.

![Figure 2](image-url)
The Puget Lobe advanced south into the Puget Lowland twice during Late Pleistocene time. During the Possession advance, which correlates broadly with MIS 4, the ice sheet terminated between Seattle and Tacoma (Porter, 2004; Borden and Troost, 2001). The Vashon advance is very well dated. The Cordilleran Ice Sheet filled the Strait of Georgia by ca. 34–32 cal. ka BP (Dyck and Fyles, 1963) and expanded into the Puget Lowland, where a detailed radiocarbon chronology documents ice expansion and contraction (Porter and Swanson, 1998). The ice sheet terminus reached the latitude of the international border by ca. 18.9 cal. ka BP, the latitude of Seattle by 17.6 cal. ka BP, and its terminus near the latitude of Olympia by ca. 17.0 cal. ka BP. The advance rate of the ice sheet was very rapid, estimated at ca. 135 m a$^{-1}$ (Porter and Swanson, 1998). The ice sheet terminus had retreated north of the latitude of Seattle by 16.6 cal. ka BP. It subsequently retreated into the northern Puget Lowland by 15.6 cal. ka BP, when glacial-marine deposition commenced (Dethier et al., 1995).

The retreat rate was yet more rapid than the advance rate, and was likely augmented by marine and freshwater calving margins.

Late-glacial advances of the Puget Lobe occurred around the latitude of the international border, but the chronology is controversial and is not represented in Fig. 2. Clague et al. (1997) described evidence of two advances in southern British Columbia prior to the Younger Dryas stadial, with no subsequent readvance, while Kovanen and Easterbrook (1997) described evidence of two advances in southern British Columbia. Thus, these ages from older dates to ca. 95 ka (MIS 5b), while the Mountain advance dates to ca. 10 ka. Recalculation of these ages using revised CRONUS production rates produces ages of ca. 21.5, 16.9 and 11.2 ka (J. Licciardi, pers. comm., 2008).

These results indicate that the maximum or near-maximum Late Pleistocene advances in the Wallowa Mountains occurred during MIS 2. Licciardi et al. (2004) reported no ages representing earlier Late Pleistocene advances, but they did note older, more subdued moraines downvalley of the dated moraines. Those older moraines, originally mapped by Crandell (1967), might represent an advance that occurred earlier in Late Pleistocene time. In general, however, it appears that the pattern of MIS 2 ice advances in the Wallowa Mountains correlated broadly with those of the Canadian ice sheets. Significantly, the ca. 17 ka advance correlates with Heinrich event 1 in the North Atlantic region and with prominent ice advances in several locations in this region.

Sawtooth Mountains

The Sawtooth Mountains are a NW-SE trending, linear range in central Idaho. This range rises to 3300 m a.s.l. and has an extensive moraine belt in the adjoining basin. That moraine belt is mantled locally with numerous boulders amenable to...
CRN exposure dating, and hosts numerous lakes, ponds and marshes amenable to sediment coring and $^{14}$C dating.

While the Sawtooth Mountain glaciation has been the subject of repeated study over the last several decades, the only chronological data have been obtained in the past decade. Thackray et al. (2004) delineated two moraine groups on the basis of moraine morphometric data and produced a partial moraine chronology based on $^{14}$C dating of sediment cores from moraine-dammed lakes and marshes in four southern valleys of the Sawtooth Mountains. They determined that two moraines in the younger moraine group date to ca. 14 and 17 cal. ka BP. Significantly, they determined that extensive ice volume was sustained until ca. 14 cal. ka BP. From the moraine morphometric data, they further suggested that the older moraine group dates to MIS 3 or earlier.

Sherard (2006) produced a provisional $^{10}$Be CRN chronology from moraine boulders in the Redfish Lake area, in the northeastern portion of the range. Her dates on a set of intermediate moraines range from ca. 14 to 16 ka, correlating closely with the $^{14}$C ages of Thackray et al. (2004) on similar moraines to the south. However, sparse dates on geomorphically older moraines correlate broadly with the global LGM, contradicting the suggestion of Thackray et al. (2004) that the older moraine group pre-dates the global LGM. Ongoing research utilising CRN dating of the Pleistocene moraine sequence and on low- and high-elevation lacustrine sediment cores should help clarify these relationships.

A minor Younger Dryas-synchronous readvance also appears to have occurred in the Sawtooth Mountains. Sherard (2006) obtained four ages ranging from 11.4 to 11.7 ka on moraines at Bench Lakes and inferred that those ages represent a Younger Dryas-synchronous advance. However, production rate and analytical uncertainties limit that inference. Mijal (2008) used a radiocarbon chronology on proglacial lacustrine sediments to infer a minor cirque glacier ice advance ca. 12.9–12.2 cal. ka BP, correlating with the early portion of the Younger Dryas stadial. The equilibrium line altitude depression implied by the two studies is quite different, however, and further work will be required to resolve those differences.

Several studies have placed age constraints on Yellowstone Plateau glaciation. Notably, Sturchio et al. (1994) used U-series dating of travertine deposits to constrain the chronology of inferred ice thickness for the northern Yellowstone outlet glacier. They concluded that the outlet glacier had achieved maximum thickness twice – ca. 40 and 26 ka – with a readvance ca. 18 ka. The link between travertine deposition and ice thicknesses is based upon inferences of hydraulic head fluctuations due to ice thickness, so these results do not necessarily reflect the maximum position of the ice front.

Licciardi et al. (2001) and Licciardi and Pierce (2008) have constructed extensive CRN chronologies from moraine boulders and other glacial deposits, documenting much more recent maximum ice extent than did Sturchio et al. (1994). Those chronologies also reveal a complex spatial and temporal pattern of ice build-up. The eastern outlet glacier reached its maximum position ca. 19 ka, the northern outlet glacier ca. 16.5 ka and the southern outlet glacier ca. 15 ka (the latter inferred from stratigraphic relationships with dated moraines from adjacent Teton Range mountain glaciers). Thus, the maximum extent of the greater Yellowstone glacial system post-dated the global LGM by 2–6 ka and the maximum extent of nearby Wind River Range mountain glaciers (see below) by 4–6 ka, and was not uniform across the plateau.

Licciardi and Pierce (2008) attribute these variable maximum ice timings to two possible causes: true regional climatic contrasts and ice dynamic influences. They suggest that the ice sheet anticyclone inferred from palaeoclimate modelling may have limited Yellowstone ice build-up at the global LGM by creating dry conditions at Yellowstone, but that the ice sheet anticyclone may not have influenced the Wind River Range glaciers in the same fashion. With respect to the variability in the timing of ice build-up across the Yellowstone Plateau they favour an interpretation invoking ice dynamics, suggesting that ice build-up proceeded from north-east to south-west across the Beartooth Uplift and Yellowstone Plateau. That ice build-up caused maximum ice positions to be reached later in the western portion of the plateau, where ice exited the plateau via outlet glaciers.

Yellowstone Plateau

The Yellowstone Plateau lies at the northeastern end of the Snake River Plain, rising to an average elevation of 2800 m a.s.l., with surrounding ranges as high as 3400 m a.s.l. As the plateau is underlain by the Yellowstone Hotspot, the rocks are dominantly rhyolite with lesser basalt. The bounding mountain ranges and the northern portion of the plateau include Eocene volcanic rocks, Archaean granitic rocks, Proterozoic–Palaeozoic sedimentary rocks and other bedrock. This plateau has hosted extensive ice caps during at least the last two glaciations (Richmond, 1964; Pierce, 1979), but its earlier history of glaciation and, in fact, of uplift, remain unknown. During the last glaciation, the ice cap covered ~30 000 km$^2$ of the plateau and surrounding ranges, with lobes flowing off the plateau and into surrounding lowlands to the north, west, south and east. Despite its continental setting, the western and southern portions of the plateau receive abundant winter moisture funnelled into the region along the Snake River Plain, a subsided volcanic province stretching several hundred kilometres to the south-west. Orographically focused snowfall is especially prevalent on the western edge of the Plateau, and was likely enhanced by growth of the extensive and thick plateau ice cap (Pierce et al., 2007).

Wind River Range

The Wind River Range lies in west-central Wyoming. These mountains lie squarely in a region dominated by continental climatic conditions, largely isolated from moist Pacific air masses. Thus, snowline elevations are high, but with peak elevations as high as 4200 m a.s.l., the Wind River Range has hosted extensive Pleistocene glaciers and maintains substantial modern ice cover. Wind River Range glaciation has a long history of study, and contains the type localities for the Rocky Mountain glacial units Pinedale (MIS 2) and Bull Lake (MIS 6) (Blackwelder, 1915).

Gosse et al. (1995) produced a detailed $^{10}$Be CRN chronology for the moraine sequence at Fremont Lake. Lying near the town of Pinedale, this is the type moraine sequence for the Pinedale glaciation. On the basis of 44 boulder exposure ages, Gosse et al. (1995) determined that the glaciers reached their maximum extent ca. 21.5 ka and remained at or near that extent for ca. 6 ka (recalculated to ca. 24–17 ka by Licciardi and Pierce, 2008, using new production rates and several scaling schemes). Thus, the Fremont Lake chronology shows close correlation with the Laurentide ice sheet maximum and the global LGM. However, as the ice terminal appears to have been maintained at near-maximum position for several
thousand years, a small increase in ice extent around 17 ka would have rendered the Wind River chronology very similar to chronologies elsewhere, such as that of the northern Yellowstone outlet glacier. The climatic factors affecting ice extent at Yellowstone and elsewhere immediately following the global LGM may well have affected the Wind River Range as well.

Uinta Mountains

The Uinta Mountains are a 300 km long, east–west trending range straddling the Utah–Wyoming border, ~250 km south of the Wind River Range. These mountains rise to 4100 m a.s.l. and hosted an extensive system of Late Pleistocene mountain glaciers. Significantly, the Uintas lie 100 km downwind of the palaeoshorelines of Pleistocene Lake Bonneville, a ~51 000 km² pluvial lake.

Uinta Mountain glaciers reached maximum extent after the global LGM. Munroe et al. (2006) constructed a $^{10}$Be CRN chronology from 21 moraine boulders in two valleys cutting the southern flank of the Uinta Mountains. They determined weighted mean moraine ages of ca. 18, 17 and 15 ka. The 18 and 17 ka ages represent two nested moraines in the Lake Fork drainage. CRN dates from moraine boulders in the Provo River headwaters, western Uinta Mountains, indicate maximum ice extent ca. 17 ka (Refsnider et al., 2008). Similarly, glaciers in the upper Bear River drainage reached maximum extent ca. 18 ka, although glacial flour in Bear Lake indicates ice expansion 32–24 cal. ka BP (Laabs et al., 2007). Together, the moraine ages indicate that glaciers in the Uinta Mountains reached their maximum extent after the LGM, remaining near maximum position until ca. 17–18 ka. These ages correlate with the dated highstand of Lake Bonneville, and Munroe et al. (2006) infer that Lake Bonneville acted as a local moisture source for the glaciers, helping to maintain ice volume even as regional temperatures ameliorated. Ongoing dating efforts suggest that the chronology of glaciation varied across the Uinta Mountains, with the pattern of chronologies suggesting that southwesterly flow across Lake Bonneville influenced southern-slope Uinta glaciers more so than northern-slope Uinta glaciers (Laabs et al., 2007).

Munroe et al. (2006) also found that reconstructed equilibrium line altitudes (ELAs) reflect a proximal western moisture source. ELAs decline smoothly from 3100 m a.s.l. in the eastern portion of the range to 2500 m a.s.l. in the western portion of the range, over a distance of ~120 km. Laabs et al. (2006) conducted coupled two-dimensional mass balance and ice flow modelling to constrain climatic conditions for Uinta Mountain glaciers. They determined that temperatures were 6–7°C lower than modern temperatures, and that precipitation in the southern Uinta Mountains was one to two times modern precipitation. They found that glaciers in the Wasatch Mountains, which lay immediately adjacent to Lake Bonneville, required two to three times modern precipitation.

Thus, while the Uinta Mountains lie squarely in a continental climatic setting, similar to that of the Wind River Range, the presence of Lake Bonneville influenced regional climate strongly enough to drive mountain glaciers to post-global LGM maximum positions. The post-global LGM timing of maximum ice extent in the Uinta Mountains ultimately appears to be a result of the reorganisation of continental climatic systems, in particular the southerly diversion of the winter jet stream due to the presence of continental ice sheets to the north (e.g. Kutzbach et al., 1993; Thompson et al., 1993).

Summary of glacial chronologies

Clearly, the chronology of Late Pleistocene ice build-up and shrinkage across the western United States was inconsistent. Glaciers in the more maritime Olympic and Cascade Mountains reached their maximum extent during MIS 5 or 4, and had global LGM advances of lesser extent. Their detailed fluctuations during MIS 4–2 are quite dissimilar, however. Glaciers in interior ranges and plateaus also varied, but there appear to be two primary timings of glacial maxima. In the Wallowa Mountains and Wind River Range (and possibly the Sawtooth Mountains), maximum ice extent correlated with the global LGM. In contrast, maximum ice extent in the Yellowstone Plateau region and in the Uinta Mountains occurred after the global LGM, around 19–17 ka, and significant readvances of glaciers in the Wallowa and Sawtooth Mountains also occurred at that time.

The term ‘global Last Glacial Maximum’ has little meaning for these mountain glacier systems when viewed at a regional scale. Thus, as has become apparent from many regional studies, the term should only be applied to the peak of $\delta^{18}O$ in marine sediment core records (21 ± 2 ka; Mix et al., 2001), and, by correlation, to maximum global ice volume and, less directly, the maximum ice extent of the Laurentide ice sheet. It is far more useful to consider the ‘local ice maximum’ or ‘local last glacier maximum’ for mountain glacier systems, and to infer their palaeoclimatic relevance independent of that of the ice sheet maximum. However, it is clear that mountain glacier and ice sheet records are linked in multiple ways. Clearly, growth of both types of ice systems was influenced by the same general climatic cooling. Furthermore, through ice sheet effects on atmospheric circulation, the growth of the ice sheets influenced mountain glacier extent to a substantial degree.

Discussion

Glaciers across the northwestern United States varied widely in their response to climatic fluctuations. These variations must reflect a combination of two factors: (1) spatial variability of palaeoclimatic fluctuations; and (2) contrasting glacier mass balance characteristics between regions and ranges. In the following paragraphs, interpretations of the broad chronological pattern of glaciation are proposed and hypotheses are presented for further testing.

Milankovitch orbital forcing has long been considered a dominant influence on climate, and Milankovitch insolation variations have served as guideposts for explaining diverse climate proxy records. In particular, Northern Hemisphere ice sheet variations and, thus, global ice volume have clear linkages with Northern Hemisphere insolation minima. However, as mountain glacier fluctuations reflect regional rather than global or hemispheric climatic fluctuations, mountain glacier records are likely to be influenced by Milankovitch variations in a complex fashion. Glaciers that respond more readily to temperature reduction (‘temperature-dependent glaciers’) should exhibit maximum ice extent around strong insolation minima (e.g., ca. 116, 95, 72 and 23 ka; Berger and Loutre, 1991), due to reduced summer ablation. These glacier systems tend to be in more continental climatic regimes.

There also exists a mechanism by which mountain glaciers could reach maximum extent during hemispheric insolation maxima. Mid-latitude westerly wind strength is influenced

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by the seasonal pole-equator temperature gradient (e.g. Dodson, 1998; Shulmeister et al., 2004, for the Southern Hemisphere). That gradient is influenced strongly by obliquity and precession cycles. As westerly flow is the main force delivering moisture to western North America, glaciers strongly influence local and with mass balance characteristics dependent upon maritime moisture flux ('moisture-dependent glaciers') will be very sensitive to westerly flow variations – more so than to dramatic temperature reduction. Even modest temperature decline in concert with strong moisture delivery can drive rapid ice expansion (e.g. Rother and Shulmeister, 2006, for New Zealand; Thackray, 2001, for the Olympic Mountains), with bigger glaciers during wetter, less cold Late Pleistocene stades. Thus, moisture-dependent glaciers in the northwestern United States might be expected to expand at times of hemispheric insolation maxima (ca. 105, 84, 59, 34 and 11 ka; Berger and Loutre, 1991) when westerly flow is strong, rather than at times of insolation minima.

Naturally, the foregoing describes end-member glacier mass balance styles, whereas glaciers in reality relate to climate in complex fashion. Nonetheless, this framework bears some relevance to the mountain glacier chronologies described herein. The Olympic Mountains chronology, in particular, appears to correlate strongly with insolation maxima, and is clearly influenced by moisture variability (Thackray, 2001). The Hoh Oxbow 2 advance, the most extensive since 50 ka, occurred 34–31 cal. ka BP, in close correspondence with an insolation maximum, and exhibited the rapid build-up and retreat characteristic of moisture influenced glaciers. The Lyman Rapids advance may also correlate with an insolation maximum (59 ka), but dating of that advance is tenuous and needs to be strengthened through other dating methods. Conversely the global LGM-equivalent Twin Creeks 1 advance (23–22 cal. ka BP) occurred during the strongest Late Pleistocene insolation minimum and was far less extensive. Westerly flow at that time was likely reduced by two factors – minimum insolation and the development of an ice-sheet-driven anticyclone – that should have brought dry easterly winds across the region (e.g. Kutzbach et al., 1993; Thompson et al., 1993).

Just 200 km to the east, the eastern Washington Cascades glaciers behaved quite differently. The Icicle Creek glaciers constructed moraines during insolation minima, suggesting a stronger link to summer temperature depression (Kaufman et al., 2004; Porter et al., 2008). However, the global LGM-correlative advances – occurring during the strongest insolation minimum – were less extensive than were earlier advances, so the relationship to insolation minima was complex. Lying east of the Cascade range crest, the Icicle Creek glaciers receive abundant Pacific moisture in their source areas but extend into an area of drier, colder, more continental climate. The continental climate setting suggests that reduced ablation, and thus insolation minima, may play a stronger role in glacier mass balance there than in the fully maritime Olympic Mountains. One distinct similarity between the Olympic and Cascade glacial systems is the limited extent of their global LGM-correlative advances. It is likely that dry, cold conditions at the global LGM affected glacier mass balance similarly in the two areas. On the whole, the contrast of the Olympic and Cascade chronologies reflects the complexity of climate–glacier relationships, even on sub-regional scales.

Chronologies of ice systems further inland also appear to reflect their continental climatic setting, but locally bear the signature of Pacific moisture influences. For example, in the Wind River Range (Gosse et al., 1995) and Wallowa Mountains (Licciardi et al., 2004), glaciers reached their maximum extent at the time of the Northern Hemisphere insolation minimum around the global LGM, ca. 21 ka. The Sawtooth Mountain glaciers may also have reached their maximum extent at that time (Sherard, 2006), but dating remains equivocal. On the whole, this pattern is not surprising. Global LGM cooling was likely far more severe in the continental interior, and ice chronologies at these temperature-dependent glacier systems reflect that cooling. In maritime-influenced areas, global LGM cooling was more limited, and moisture influences could be more strongly expressed.

Several ice systems reached maximum or near-maximum Late Pleistocene extent after the global LGM, broadly correlative with Heinrich event 1 (e.g. Bond et al., 1999). The northern Yellowstone ice cap outlet glacier reached its maximum extent ca. 17 10Be ka (Licciardi et al., 2004; Licciardi and Pierce, 2008). Sawtooth Mountain glaciers also reached near-maximum extent ca. 17 cal. ka BP to 16 ka (Thackray et al., 2004; Sherard, 2006). Glaciers in the western portion of the Uinta Mountains reached maximum extent ca. 18–17 ka (Munroe et al., 2006). Eastern Cascade Mountain glaciers exhibited a major advance 18–15 ka (Porter et al., 2008), and the Puget Lobe of the Cordilleran Ice Sheet reached its maximum Late Pleistocene extent ca. 17 cal. ka BP (Porter and Swanson, 1998).

There are several possible reasons why several glacial systems reached their maximum or near maximum extent at ca. 17 ka. First, a widespread cooling event affected many areas of the globe around 17 ka, in concert with Heinrich event 1. The exact mechanism of cooling transfer to the northwestern United States is not clear, but it is apparent that a prominent cooling event affected the region (see Licciardi et al., 2004). Secondly, moisture delivery into the region was apparently limited around the global LGM by an ice sheet anticyclone, suppressing glacier growth (see above; Thackray, 2001; Licciardi et al., 2004; Licciardi and Pierce, 2008), but may have increased shortly thereafter. Licciardi et al. (2004) discerned a latitudinal pattern in responses of glacier systems to the ice sheet effects. Glacier systems closer to the ice sheet were more strongly affected by the ice sheet anticyclone, while those further removed from the ice sheet were less strongly affected, and reached maximum extent around the global LGM. As the Canadian ice sheets began to shrink following the global LGM, the glacial anticyclone likely began to weaken, and the Northern Hemisphere insolation gradient became more favourable to vigorous westerly atmospheric flow. Thus the advance of several ice systems around 17 ka may have been enhanced by reorganized atmospheric flow and reinvigorated moisture transport into the region. The effects of enhanced moisture transport around 17 ka would have been limited to ice systems that were more strongly influenced by moisture variability in the first place ('moisture-dependent glaciers').

In various ways, each of these ice source areas is strongly dependent upon Pacific storm systems. For the Olympic Mountains, Cordilleran Ice Sheet and Cascade Mountains, the effects are direct, as the glaciers originated in areas of maritime climate. For the other systems, those influences are strong despite locations in the continental interior: the Wallowa Mountains receive abundant moisture funnelled through the Columbia River Gorge, and the Sawtooth Mountains also lie at the distal end of that moisture influence (e.g. Meyer et al., 2004). The Yellowstone Plateau lies at the distal end of the Snake River Plain, another prominent pathway for moist weather systems (Pierce et al., 2007). Glacier growth in the Uinta Mountains, conversely, appears more strongly influenced by a local moisture source – Lake Bonneville – and glaciers reached their maximum extent around the time of the Bonneville highstand (Munroe et al., 2006).

An unresolved paradox remains in these relationships. The highstand of Lake Bonneville, ca. 17 ka, has long been


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interpreted as a response to diversion of the winter jet stream into the Great Basin region (e.g. Thompson et al., 1993). However, extensive mountain glacier growth far to the north at around the same time (and inferred here to have resulted from increased moisture delivery) suggests that the average jet stream position had retreated to the north by that time. Clearly, improved glacial chronologies, additional palaeoclimatic proxy data and an improved understanding of Late Pleistocene atmospheric dynamics are necessary to resolve this problem.

Conclusions

Glaciers across the northwestern United States varied widely in the timing of their major advances. That variability appears to be related to two factors: (1) the spatial heterogeneity of precipitation and temperature fluctuations; and (2) the mass balance characteristics of individual glacial systems. Glacial systems in more continental climatic regimes (Wind River Range, Wallowa Mountains and possibly the Sawtooth Mountains) reached their maximum extent at the time of maximum temperature decline around the global LGM. Glaciers in more moisture-influenced climatic settings (Olympic Mountains, Cordilleran Ice Sheet, Cascade Mountains, northern Yellowstone ice cap, possibly Sawtooth Mountains) reached their maximum extent at times other than the global LGM, but the timing of their advances varied widely. However, chronologies of those glacial systems reveal numerous climatic events, favourable for ice growth, that may have important palaeoclimatic implications, and that may be masked elsewhere by strong global LGM-correlative advances. The global LGM was clearly a major cold event in the region, but its influence was determined by climatic setting and, ultimately, by mass balance characteristics of local ice systems.

Thus it is clearly important for the sake of regional (and in fact global) palaeoclimate models that researchers examine carefully the details of complex glacial chronologies spanning Late Pleistocene time. Those long-term glacial chronologies have the greatest potential to reveal both the spatial and temporal heterogeneity of climate change and the variable response of ice systems. Colman and Pierce (1992) identified several regional glacial sequences with pre-global LGM advances, and those areas may be ripe for renewed dating efforts.

The chronologies reviewed in this article, and our understanding of their implications, are far from perfect. Dating uncertainties, rooted in the dating methods themselves and in the geological complexity of the sampling sites, limit the precision and accuracy of the chronologies. The precipitation and temperature influences on glacier growth and retreat are not thoroughly discernible in the glacial chronologies and remain speculative for many areas. Estimates of temperature or precipitation from independent proxies such as pollen, lacustrine studies and amino acid palaeothermometry (e.g. Kaufman, 2003) are necessary to make palaeoclimatic inferences more robust. Additionally, glacier mass balance modelling (e.g. Plummer and Phillips, 2003; Laabs et al., 2006) could help to elucidate possible combinations of temperature and precipitation responsible for glacier growth.

To improve understanding of regional glaciation and palaeoclimates, a number of steps must be taken. First, ice fluctuation chronologies need to be improved in some areas where the timing of maximum Late Pleistocene glaciation remains undefined. Examples include the Olympic and Sawtooth Mountains, and areas in which calibrated relative-age studies (e.g. Colman and Pierce, 1992) indicate pre-global LGM advances. Second, the density of high-quality glacial chronologies needs to be increased. New Late Pleistocene chronologies in the Oregon Cascade Range, in the desert mountains of eastern Oregon, in western Idaho (currently undertaken by the Idaho Geological Survey), in eastern Idaho, in western Montana and in Wyoming will greatly improve understanding of regional climatic fluctuations. Third, continuous proxy records need to be developed for glaciated areas, particularly through studies of proglacial, volcanic and structurally controlled lakes. While such studies are relatively common for Lateglacial and Holocene time, they rarely extend to the global LGM or earlier. Fourth, in a broader, more disciplinary sense, the linkages between insolation, fluctuations in temperature and precipitation, glacier mass balance and glacial sedimentation need to be more thoroughly understood (e.g. Owen et al., 2008). Improved knowledge of those relationships will encourage hypothesis-driven study of glacial chronologies, not only in this region, but around the globe.

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