

Would North American Paleoindians have Noticed Younger Dryas Age Climate Changes?

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Abstract Paleoindian groups occupied North America throughout the Younger Dryas Chronozone. It is often assumed that cooling temperatures during this interval, and the impact these would have had on biotic communities, posed significant adaptive challenges to those groups. That assessment of the nature, severity and abruptness of Younger Dryas changes is largely based on ice core records from the Greenland ice sheet where changes were indeed dramatic. This paper reviews climatic and environmental records from this time period in continental North America. We conclude that, on the Great Plains and in the Rocky Mountains, conditions were in reality less extreme. It therefore follows that conditions during the Younger Dryas interval may not have measurably added to the challenge routinely faced by Paleoindian groups who, during this interval, successfully (and perhaps rapidly) dispersed across the diverse habitats of Late Glacial North America.

Keywords Younger Dryas · Paleoindian · North America · Hunter-gatherer adaptations

The Younger Dryas interval was first recognized in European macrofossil and pollen records by the (re)appearance of a cold-tolerant Arctic flowering shrub, *Dryas octopetala* (mountain avens), following a long period of warming (the Bølling-Allerød) after the Last Glacial Maximum (LGM). (There was an ‘Older Dryas,’ which originally separated the Bølling and Allerød, but it is now known to have only been a rather minor cool phase, hence the conjunction ‘Bølling-Allerød’).

It soon became apparent that the cooling of the Younger Dryas was not restricted to northwestern Europe where it was first defined, but was more widespread, and in places involved a return to near-glacial climatic conditions. Not only did changing climatic conditions during the Younger Dryas Chronozone (YDC)—originally dated to 11,000–

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10,000 ^{14}C years BP (Mangerud et al. 1974)—mark a significant detour from what was thought to have been uninterrupted warming after the LGM, it appeared that those changes occurred rapidly, making this episode a prime illustration of abrupt climate change (Alley 2000b; Peteet 2000).

In the last two decades, there has been considerable research across disciplines into the cause(s) and especially the consequences and extent of YDC climate change. This includes inquiry into its geographic extent, timing and rapidity, severity and seasonality, impact on atmospheric, oceanic and terrestrial environments, and effects on reservoirs of greenhouse gases (including carbon dioxide and methane) and, in turn, on radiocarbon dating during this period. The literature is substantial (e.g. Alley 2000a, b, 2007; Alley and Clark 1999; Bondevik et al. 2006; Broecker 1995, 2003, 2006a, b; Broecker and Denton 1990; Broecker et al. 1989; Clark et al. 2002; Denton et al. 2005; Firestone et al. 2007; Goslar et al. 1995, 2000; Hughen et al. 2000; Lie and Paasche 2006; Lowell et al. 2005; Mayewski et al. 1993; Muscheler et al. 2008; Peteet 1995, 2000; Rind et al. 1986; Robinson et al. 2005; Severinghaus and Brook 1999; Shuman et al. 2002, 2005; Williams et al. 2001, 2002; Wright 1989).

More recently, attention has also focused on the possible adaptive responses of humans and their animal prey to Younger Dryas conditions. In North America, for instance, Younger Dryas age cooling is said to have played a significant role in Paleoindian migration patterns (Faught 2008, p. 689); led to regional abandonment, ‘the demise of the Clovis way of life, the emergence of subregional cultural traditions, [and] the extinction of the megafauna’ (Anderson and Faught 2000, p. 512); and even obliged Florida Paleoindians to start wearing furs (Dunbar and Vojnovski 2007, p. 179) (see also D.G. Anderson et al. 2002, p. 16; Dansie and Jerrems 2005; Fiedel 1999; Hall et al. 2004; Haynes 2005, 2008; Haynes 2002; Lovvorn et al. 2001; Newby et al. 2005). Geneticists seeking to explain reduced mitochondrial DNA diversity in Native American populations have likewise characterized Younger Dryas age climatic deterioration as ‘drastic enough to extinguish [ancestral] northern populations,’ creating a population bottleneck in the peopling of the Americas (Forster et al. 1996, p. 941; Malhi et al. 2007; Smith et al. 2005; but see Kemp et al. 2007).

But was the impact of cooling on Pleistocene North Americans that pronounced and/or widespread? Such claims are usually rooted in the assumption that climate changes during the YDC came on abruptly, involved a severe plunge in temperature, and triggered equally rapid changes in the plant and animal communities on which people depended, an assumption that is based largely on climate changes evident in Greenland ice core data. Was cooling during the Younger Dryas similarly abrupt and severe, and in the same direction across North America? There are reasons to be skeptical. Even if it was, to claim that changes impacted Paleoindians would require the demonstration that changes in the archaeological record of a region *directly* resulted from changes in climate or environment documented at that archaeological moment and location (Meltzer 1991, p. 237). After all, not all climate changes had cultural consequences, and not all cultural changes had a climatic trigger.

Still, it is important to understand the climatic and ecological conditions during the YDC, for this was the stage for Paleoindian dispersal and adaptations across the continent, and there were significant changes over its course (D.G. Anderson et al. 2002; Meltzer 2009). And so we explore here what we do and do not know about the climate and environment of North America during that span, and its possible impact on Paleoindians.

We begin by summarizing the causes and consequences of YDC climate change, to assess the degree to which those changes, prominently expressed in Greenland ice core

data, can be generalized to other parts of the northern hemisphere. We then turn to patterns and variation in YDC climate and environments across North America, followed by a more detailed examination of how this episode played out on the Great Plains and the Rocky Mountains (based on pollen, isotope, and stratigraphic records), regions where a Paleoindian presence is well-dated and well-documented.

A Younger Dryas Overview: Cause(s), Consequences, Chronology

The Late Glacial was generally a time of increasing insolation (incoming solar radiation), warming climates, retreating ice sheets, rising sea levels, and a host of related and local environmental responses. These changes were mostly unidirectional, but were punctuated by massive meltwater releases from under/behind the ice sheets, and occasional reversals of the overall warming trend. Younger Dryas-age cooling represents the last and most pronounced of these reversals. Although the YDC was first recognized as ‘a distinct vegetational setback in large parts of Europe’ (Björck 2007, p. 1986; Watts 1980), it took on more global significance in the late 1980s and early 1990s, when a comparable signal was recognized in chronomid (midges), pollen, and sediment records from high-latitude northeastern North American lake cores (e.g. Engstrom et al. 1990; Mathewes 1993; Mayle et al. 1993; Mayle and Cwynar 1995a, b; Mott et al. 1986; Peteet 1987; Walker et al. 1990; Wright 1989).

A distinct ‘cold spike’ was likewise observed in Greenland ice cores, and in foraminifera and isotope sequences in North Atlantic deep sea sediments (e.g. Alley et al. 1993; Broecker et al. 1989; Mayewski et al. 1993; Ruddiman and McIntyre 1981; Taylor et al. 1997; Yu and Wright 2001; for summaries, including more recent evidence, see Alley 2007; Hald et al. 2007). The Greenland ice core record gained particular attention because it provided the first temperature estimates for Younger Dryas cooling and allowed very precise dating of the interval, which suggested its onset and (especially) its demise were rapid, perhaps occurring over decades (Steffensen et al. 2008; see also Alley 2000a; Dansgaard et al. 1989; Taylor et al. 1997). (It has even been suggested that the term ‘Younger Dryas’ ought to be replaced by the acronym GS-1, which stands for ‘Greenland [Isotope] Stadial 1’ (Björck et al. 1998). Although that term may be more accurate or at least better defined, for now it is less confusing to stay with Younger Dryas). The emerging data on the YDC initially suggested cooling might have even taken place on a hemispheric, if not a global, scale (Peteet 1995, 2000; Steig et al. 1998). Given its implications for potential future changes in climate, the rapid cooling and subsequent warming, and its possible causes, became a hot topic.

A longstanding hypothesis—first proposed by Broecker et al. (1985, 1989)—was that it was triggered by disruption of the Atlantic ocean’s meridional overturning circulation (MOC) or, more colloquially, thermohaline circulation (THC; but see terminological objections in Wunsch 2002). Today, ocean water warmed in the tropics (the Gulf Stream) travels northward along the North American east coast at relatively shallow depths (<1 km below the surface). It rises to the surface in the North Atlantic and cools, releasing the water vapor and heat that today brings wet, mild winters to northwestern Europe. The cooler, saltier, and hence denser water then slowly sinks back into the abyss (the so-called North Atlantic Deep Water), and flows south back toward the Equator at depths of 2–5 km.

But if the surface water in the North Atlantic is freshened by heavy runoff from rain or a significant pulse of glacial meltwater, it can freeze in winter (saltwater has a lower freezing point than freshwater). If substantially cooled or even frozen, the surface water will fail to

sink, and in so doing effectively slow or even stop MOC (Alley 2000c; Church 2007). Shutting down that circulation, Broecker et al. (1989) proposed, would cool the North Atlantic, which in turn would ripple through the earth's climate system. Alley (2000c, p. 142) uses the analogy of winters to the east of the Great Lakes: when the lakes are open, cold Arctic winds blow across and become warmer and wetter, then dump snow on the downwind region; when the lakes freeze in especially harsh winters, the downwind region becomes much colder and drier.

Evidence for a disruption of MOC is now available from multiple physical and geochemical proxies (Alley 2007, pp. 256–259). These include comparison of ^{14}C in benthic (bottom-dwelling) and planktonic (water-column) foraminifera and in deep-sea corals (this on the reasoning that waters will be enriched or depleted in ^{14}C depending on surface temperatures, source, and degree of mixing). Additional evidence comes from $^{231}\text{Pa}/^{230}\text{Th}$ ratios in foraminifera, which have different residence times in the ocean, and are accordingly more/less likely to be removed by deep water, and hence during times when MOC is shut down, the ratio of $^{231}\text{Pa}/^{230}\text{Th}$ will rise. The documented instances of MOC disruption in the terminal Pleistocene are strongly correlated with episodes of 'punctuated' climate change, including during the YDC (Lynch-Steiglitz et al. 2007), and provide compelling evidence that changes in ocean circulation are an important driver for abrupt climate change (Clark et al. 2002; Gherardi et al. 2005; Keigwin et al. 1991; McManus et al. 2004; Muscheler et al. 2004; Robinson et al. 2005; see summaries in Alley 2007; Broecker 2003).

But what would cause a shutdown of MOC? Broecker et al. (1989) initially pointed to glacial Lake Agassiz, the vast freshwater lake that developed on the southwestern margin of the melting and retreating Laurentide Ice Sheet, beginning c. 13,500 cal BP. At its maximum, Lake Agassiz covered 840,000 km² and contained ~163,000 km³ of meltwater that in places was ~770 m deep (Teller 2004). At different stages in its ~4,000 year history, the lake drained episodically via multiple outlets in different directions, depending on the water level, topography, and the relative position of Laurentide ice. Arranged in what appears to be chronological order, the meltwater drainage outlets were as follows: south down the Mississippi River into the Gulf of Mexico; east into the Lake Superior basin and then out through the St. Lawrence Valley into the North Atlantic; northwest into the Arctic Ocean via the Athabasca and Mackenzie valleys; east again down the St. Lawrence; and, finally, a pulse north into Hudson's Bay and then into the North Atlantic (Leverington and Teller 2003; Teller 2004).

The prime suspect in triggering the Younger Dryas changes was the meltwater pulse that sent Lake Agassiz's waters flooding eastward, presumably when retreat of the Superior Lobe of the Laurentide ice sheet uncovered low-lying regions around Lake Superior (Dyke 2004, p. 395), opening a drainage outlet that would have suddenly diverted Agassiz meltwater eastwards into the Great Lakes, and then out into the North Atlantic via the St. Lawrence seaway (Leverington and Teller 2003; Leverington et al. 2000; Teller 2004). It was argued that a sudden injection of an estimated ~9,500 km³ of cold fresh water from Lake Agassiz into the northeastern North Atlantic could have shut down the MOC, with all the attendant climatic consequences (Broecker 2003; see also Alley 2000a; Alley and Clark 1999; Broecker et al. 1989; Clark et al. 2001, 2002; Dyke 2004; Teller et al. 2002).

In this scenario, the Younger Dryas climate changes were the instrument of their own demise: the colder temperatures that came to the upper Midwest, coupled with 'lake-effect' snows from nearby Lake Agassiz, triggered the Marquette ice re-advance that pushed far enough south that Lake Agassiz's eastern outlet was once again blocked (Clark et al. 2001, p. 284; Dyke 2004). With that, Lake Agassiz's meltwater ceased to flow into the North

Atlantic, the MOC was re-established, and air temperatures on the Greenland ice sheet rose rapidly (Alley 2000a). Or so it appeared.

Although there remains considerable confidence in the hypothesis that an MOC shut-down was the *proximate* cause of the Younger Dryas climate changes (Alley 2007; Broecker and Kunzig 2008, p. 129; McManus et al. 2004), there is now a great deal of controversy over its *ultimate* cause. The chronology of an eastward drainage of Lake Agassiz's water is proving to be complicated and uncertain: the low-stand thought to mark the level of the lake after it flushed eastward into the North Atlantic (the Moorhead Phase) is younger than it should be to support this hypothesis, and no outlet channel or debris from what must have been an outflow of considerable magnitude has been found in the Lake Superior Basin (Broecker 2006a, b; Fisher et al. 2006, 2008; Lowell et al. 2005; Teller et al. 2005). One of the long-suspected outlet regions—Thunder Bay—was still glaciated at the time of the Younger Dryas, and the outlet drainages and flood debris that have been found there are too young to be related to the Younger Dryas interval (Fisher et al. 2006, 2008). In addition to the Lake Superior Basin outflow, meltwater drainage down the Mississippi—thought to have ceased during the Younger Dryas when Agassiz's water was diverted to the east—apparently continued throughout (e.g. Aharon 2003).

In the face of these discrepancies, alternative hypotheses have been advanced to account for an MOC-stopping freshwater flux, including a large meltwater discharge from the Keewatin ice dome into the Arctic Ocean via the McKenzie River (Tarasov and Peltier 2005, 2006); thick Arctic sea-ice flushed out into the North Atlantic as global sea levels rose (Bradley and England 2008); and, far more controversially, an extra-terrestrial impact (Firestone et al. 2007; see Kerr 2008; Pinter and Ishman 2008a, b; cf. Bunch et al. 2008; Firestone and West 2008). For the moment, Lake Agassiz remains in the mix (Carlson 2008; Bradley and England 2008), because the chronology of deglaciation and drainage is still too poorly known to preclude a meltwater pulse down the St. Lawrence, and significantly, there is compelling geochemical evidence—a Younger Dryas-age spike in tracers such as Mg/Ca, U/Ca, and $^{87}\text{Sr}/^{86}\text{Sr}$, elements that could only have had a western Canadian bedrock source—in sediment cores from the St. Lawrence estuary (Carlson et al. 2007; also Brand and McCarthy 2005).

Whatever the cause, the climatic consequences varied. The clearest and arguably the most pronounced signal of climate change comes from the Greenland ice cores. Recorded there in multiple proxies is a mean annual Younger Dryas air temperature about 15–16°C colder than present (Denton et al. 2005, pp. 1161, 1168; Severinghaus et al. 1998). That's cold, though not quite as cold as the LGM, when Greenland temperatures were as much as ~23°C colder than present (Denton et al. 2005, p. 1168).

However, the Younger Dryas may not have been cold year-round. Denton et al. (2005) observed that Younger Dryas-age glacial moraines in Greenland extended only about 12 km beyond moraines formed in the Little Ice Age (nineteenth century). That was well short of the LGM ice-marginal position, which reached more than 120 km beyond the Little Ice Age moraine. And yet the average annual temperature during the YDC (15°C colder than present, as just noted) was much colder than it was during the Little Ice Age—the latter is estimated at just ~1°C colder than present. Denton et al. (2005, p. 1168) reasoned that if YDC temperatures were that cold year round, glacial ice should have extended a much greater distance than it did, far beyond the Little Ice Age moraines, and perhaps even approaching the LGM extent. That it did not, and given that glacial expansion is limited by *summer* temperatures, Denton et al. (2005, p. 1169) concluded that 'a large fraction ... of temperature lowering ... during the Younger Dryas occurred in the winter season'. Accordingly, in order to account for its average *annual* air temperature, Younger

Dryas winters must have been a great deal colder—of the order of 24°C cooler than present—with Younger Dryas summer temperatures not more than 5–6°C colder (although Lie and Paasche [2006, p. 406], in a re-consideration of the same evidence, put Younger Dryas summer temperatures at 7.5–10°C colder). Younger Dryas climate, they infer, was strongly seasonal, with an ~18–20°C swing between summer and winter. Their inference is in keeping with calculated summer and winter insolation values that likewise indicate sharp seasonal swings (Berger and Loutre 1991). Those very cold winters, linked to the influx of freshwater and the shutdown of Younger Dryas MOC, would have combined to cause the formation of extensive sea-ice over the North Atlantic (Denton et al. 2005). By Broecker's estimate, ice must have covered 'the entire Norwegian Sea, extending perhaps to the southern end of the British Isles' (Broecker 2006b, p. 213).

In portions of the northern hemisphere, glaciers stopped retreating and in places even began to re-advance (as discussed below; see also Benson et al. 2007; Dyke 2004; Licciardi et al. 2004). As a result, postglacial rising seas stalled (Edwards et al. 1993; Fairbanks 1989; Lambeck et al. 2002), and atmospheric methane sharply declined from 680 to 460 parts per billion (MacDonald et al. 2006). Methane rose during the earlier, warmer Bølling-Allerød, probably as a result of several processes, including peatlands that developed on deglaciated terrain, the warming of soils south of the ice margin, and the flooding of organic soils as sea level rose. The subsequent methane decrease, in turn, may have come as tropical regions cooled, and northern peatlands froze or were covered over by re-advancing ice sheets. There are other explanations—though no consensus—for the rise and fall of methane through the Late Glacial period.

There is also evidence in the Greenland ice cores of higher winds and greater storminess over northern hemisphere land and sea, with a YDC atmosphere that carried ~5–7 times more dust and ~3 times more salt than in the millennia that preceded or followed (the atmosphere during the latter period was itself twice as dusty and salty as the atmosphere in the pre-Industrial era [Alley 2000c, p. 112–115]). That dust has been geochemically sourced to the deserts of western China, suggesting that the effects of continental Younger Dryas winters were propagated some distance into the temperate zone, perhaps shifting the Atlantic intertropical convergence zone southward (Lea et al. 2003), and weakening the Asian summer monsoon system, for which there is independent isotopic evidence (Wang et al. 2001; also Denton et al. 2005, p. 1178).

The Younger Dryas has been described as a return to full glacial climatic conditions, but even in the core areas of YDC cooling (Greenland and the North Atlantic), it was not a LGM rerun. Moreover, as Alley and Clark observed, the changes accompanying the Younger Dryas might have been 'large and rapid around the North Atlantic, [but were] probably smaller and slower elsewhere' (Alley and Clark 1999, p. 174). Indeed, some regions show no significant Younger Dryas-age climate change, while others showed warming. Nor were global changes synchronous. In Antarctic ice cores, for example, cooling precedes the YDC (the Antarctic Cold Reversal, as its known), suggesting there was a 'bipolar seesaw' in global climates, at least on a millennial time scale (e.g. Alley and Clark 1999; Blunier and Brook 2001; Broecker 1998; Clark et al. 2002; Lea et al. 2003; Morgan et al. 2002; Steig 2001; Stocker 1998, 2002; Turney et al. 2003; Wang et al. 2001; cf. Severinghaus et al. 2004 and Wunsch 2003, 2004). As recently summarized by Alley:

The YD is seen to have been a cold event across much of the Northern Hemisphere, but especially centered on the northeastern Atlantic, with anomalously dry conditions in broad monsoonal regions of Africa and Asia, tropical atmospheric circulation

shifted southward especially in the Americas, and southern [hemisphere] warmth (Alley 2007, p. 247).

We explore in greater detail below Younger Dryas-age changes in North America.

The Younger Dryas, as earlier noted, was originally identified as a pollen zone marked by the re-appearance of *Dryas* pollen. It was subsequently codified as a ‘chronozone,’ a stratigraphic interval with time-synchronous boundaries which, in pre-calibration days, were set at 11,000 and 10,000 ^{14}C years BP (Mangerud et al. 1974; Anderson 1997). It was initially assumed that the Younger Dryas Chronozone and ‘biozone’ were synchronous. However, chronostratigraphic entities by definition have fixed temporal boundaries, while biostratigraphic units are inherently time-transgressive (or at least temporally slippery) over large areas. Not surprisingly, at some sites the appearance of vegetation indicative of cooler climates did not always coincide with the timing of the Younger Dryas Chronozone (Björck et al. 1998, p. 286).

Obscuring matters further, the chronological details of the YDC are difficult to calibrate, owing to distortions in the radiocarbon time scale during this interval caused by variation in atmospheric ^{14}C (Björck et al. 1998, p. 286–287; Muscheler et al. 2008). The amount of ^{14}C in the atmosphere at a moment in time is a function of (1) solar activity and geomagnetic intensity driving ^{14}C production, and (2) the carbon cycle, particularly through the oceans, which harbor ~ 60 times as much carbon as the atmosphere. Even a small perturbation in the uptake or release of oceanic ^{14}C will cause large changes in atmospheric ^{14}C (Robinson et al. 2005, p. 1469; also Hua et al. 2009, p. 2988–2989). The shutdown of MOC was much more than a small perturbation.

As a result, atmospheric ^{14}C —measured as change in $^{14}\text{C}/^{12}\text{C}$ ratio ($\Delta^{14}\text{C}$) relative to a standard—varied during the YDC. Precisely how much it varied is unknown, since tree-ring calibration falls short of spanning the entire period, and the calibration currently used—the marine ^{14}C record from the Cariaco basin (Hughen et al. 2004, p. 1062)—assumes a reservoir age that is constant with time, which may not be the case (Hua et al. 2009, p. 2983; Kromer et al. 2004, p. 1205). Comparison of ^{14}C with ^{10}Be —the cosmogenic isotope of beryllium (see below)—indicates that there was sufficient variance in marine reservoir ages around the end of the Allerød and beginning of the Younger Dryas to call into question the accuracy of that calibration curve over this interval (Muscheler et al. 2008, p. 265–266; see also Blackwell et al. 2006, p. 411; Hua et al. 2009, p. 2983; Kromer et al. 2004, p. 1205). (That variance can be detected because although ^{14}C and ^{10}Be are both produced by cosmic rays, ^{10}Be is unaffected by the carbon cycle and so it can be used to control for the relative contribution of solar activity to ^{14}C production: Muscheler et al. 2000, 2004).

Moreover, the Atlantic Ocean MOC shutdown in the first few centuries of the Younger Dryas, and consequent decreased uptake of ^{14}C in the deep ocean, led to an increase in atmospheric $\Delta^{14}\text{C}$ (Hua et al. 2009; see also Hughen et al. 2000; Alley 2007; Bondevik et al. 2006; Marchitto et al. 2007; Marotzke 2000). Estimates of the precise timing and amount of that enrichment vary, but recent work based on an extended atmospheric $\Delta^{14}\text{C}$ record identifies an abrupt increase of $\sim 40\%$ starting at c. 12,760 cal BP (Hua et al. 2009, p. 2986). This is later (by ~ 240 years) and smaller (cf. 65%) than the rise in $\Delta^{14}\text{C}$ previously reported from the Cariaco basin (Hughen et al. 2000). It also comes later than the MOC shutdown, for reasons suggested by Hua et al. (2009, pp. 2987–2988). Rising atmospheric $\Delta^{14}\text{C}$ during this time results in radiocarbon dates that appear too young (the radiocarbon clock ‘speeds up,’ in Broecker’s [2003, p. 159] terms), with the result that the calibrated age of a sample spans a comparatively wider ^{14}C age range. This is why that

portion of the calibration curve (e.g. Reimer et al. 2004, p. 1052) tends toward the vertical (Hajdas et al. 2006, p. 17; Muscheler et al. 2008, p. 266).

Atmospheric $\Delta^{14}\text{C}$ peaked c. 12,630 cal BP (at $\sim 235\text{‰}$), diminished slightly and then fluctuated (around 220‰) for the next ~ 300 years. Atmospheric $\Delta^{14}\text{C}$ then began to decline c. 12,300 cal BP, presumably associated with the resumption of MOC (Hua et al. 2009, p. 2988; Hughen et al. 2000). Falling $\Delta^{14}\text{C}$ ‘slowed down’ the ^{14}C clock, since similar ^{14}C ages would apply across a range of calendar years (Hajdas et al. 2006, p. 17), forming a radiocarbon age ‘plateau’. Here, the calibration curve tends more toward the horizontal (Reimer et al. 2004, p. 1052). This was followed at the end of the YDC by a more abrupt drop in $\Delta^{14}\text{C}$, which brought atmospheric $\Delta^{14}\text{C}$ to near its modern levels, and allows more secure calibration of ^{14}C years to calendar years.

Accordingly, bracketing ages for the Younger Dryas Chronozone are based not on radiocarbon ages but rather on counting annual layers of ice in Greenland ice cores (e.g. the new GICC05 time scale), with the onset of Younger Dryas cooling and subsequent warming identified in the ice core proxies (Steffensen et al. 2008, p. 680–681; also Alley 2007; Blunier and Brook 2001; Stocker 2002). Although there remains discussion as to which proxies and which ice core chronology should be used to bracket this episode (Muscheler et al. 2008; Southon 2004; Steffensen et al. 2008), the most recent, high-resolution analysis—which uses isotopic analyses of deuterium excess and $\delta^{18}\text{O}$ as indicators of past ocean surface and air temperatures, respectively—indicates that Younger Dryas cooling began 12,900 calendar years before present, with the warming starting 11,700 calendar years before present (Steffensen et al. 2008, Table 1; in that study, the year 2000 was designated as the ‘present’).

These data further reveal something of the relative speed of those climatic transitions: the onset of Younger Dryas cooling spanned two centuries, reflecting ‘collapsing ecosystems as the extreme YD winter conditions ... set in’ (Muscheler et al. 2008, p. 263; also Steffensen et al. 2008), while post-Younger Dryas warming was faster, taking place in a matter of decades (Steffensen et al. 2008, p. 681; see also Blackwell et al. 2006; Muscheler et al. 2008, p. 263; cf. Haynes 2008, p. 6520).

Importantly, not all climate proxies changed simultaneously: atmospheric and ocean temperature changes largely outpaced changes in other indicators, such as dust and salt (Steffensen et al. 2008; see also Björck et al. 1998; Muscheler et al. 2008). That lagged response is attributable to ‘the inertia of land surfaces drying out, and vegetation dying off in the dust source regions before large fluxes of dust can be re-established’ (Steffensen et al. 2008, p. 683). Such differences in timing are not unexpected, and are relevant to the discussion (below) of the varying response times of plants, animals and landscapes to Younger Dryas climate changes (Benson 2004; Grimm and Jacobson 2004; Shuman et al. 2005; Webb et al. 2004).

Although the 1,200 year span of the YDC was marked by abrupt, severe, sharply seasonal cooling over Greenland, this is ‘one point, and, although some of the variables measured in the ice cores may be representative of large spatial scales, even global, it is nevertheless recording effects on top of one point on the Greenland ice cap. Based on its location, there are simply limits within which it can vary’ (Viau and Gajewski 2007, p. 264). We cannot assume Younger Dryas age climate and climate changes in Greenland characterize regions elsewhere, especially not in the unglaciated, temperate mid-latitudes of North America which provided the adaptive stage for Paleoindians.

The Younger Dryas in North America

Several baseline parameters are critical to understanding climatic and environmental conditions in North America during Younger Dryas time: for one, the orographic and thermal effects of the North American ice sheets were much diminished compared to LGM times, but nonetheless had an influence on climate (Dyke 2004; Yu and Wright 2001). Second, and to a degree counter-acting that influence, northern hemisphere insolation and atmospheric CO₂ had by Younger Dryas times both climbed to near interglacial levels (Berger and Loutre 1991; Carlson 2008; Licciardi et al. 2004, p. 83). In regard to the former, although summer insolation would not peak until the Early Holocene, YDC summer insolation was still higher than at any time in mid-latitude North America (30°–50°N) over the preceding 70,000 years (data from Berger and Loutre 1991, ncdc.noaa.gov/pub/data/paleo/insolation). Younger Dryas-age CO₂ values were rising as well, and approaching 265 ppm, compared to ~190 ppm during the LGM (Monnin et al. 2001). Finally, seasonality was amplified, compared to LGM or even later Holocene times (Denton et al. 2005; Shuman et al. 2002), as YDC winter insolation lows, coupled with summer highs, created a highly seasonal insolation signal (20% farther apart than at present), and conditions that were essentially without modern equivalent (Cole and Arundel 2005, p. 715; Grigg and Whitlock 1998, p. 296; Shuman et al. 2002, p. 1786; Webb et al. 2004, pp. 473–474).

Physical conditions during Younger Dryas times were thus, arguably, unique (Shuman et al. 2002, p. 1786), unlike what occurred even in previous deglaciations (Broecker 2006a; Carlson 2008). Although beyond the scope of this paper, we would note that this putative uniqueness could be relevant to the question of why a suite of mammals that had previously survived multiple glacial-interglacial cycles failed to survive this one.

There is now an extensive literature on Younger Dryas age conditions in unglaciated North America south of the (still lingering) continental ice sheets (see Fig. 1). From it we can draw a number of generalizations but, as always, the devil is in the detail, and for that reason we highlight patterns as well as variation.

First, there was cooling across northeastern North America during this period, although far less than in Greenland. Estimates of Younger Dryas mean annual temperature based on data from a variety of proxies (e.g., chironomids, pollen, oxygen isotopes) indicate that mean annual temperatures were no more than ~5°C cooler than at present, and often of the order of just 3–4°C cooler (Denniston et al. 2001; Peteet 2000; Yu 2000, 2007; Yu and Eicher 1998). In a few areas, notably Atlantic Canada and northern New England, conditions were colder—one estimate puts the summer temperature decline at ~5°C (Peteet 2000); more cooling in this region is not unexpected, given its proximity to the cold North Atlantic (Yu and Wright 2001, p. 354). Cooling attenuated westward into the continental interior (Cwynar and Spear 2001; Shuman et al. 2002; Walker and Cwynar 2006; Yu and Wright 2001). To describe the Younger Dryas in North America as ‘a long-lasting deep freeze’ (Haynes 2008, p. 6524) is an overstatement.

Indeed, not all of North America experienced cooling: some regions saw relative *warming* or at least temperate climates during the YDC (Grimm et al. 2006; Shuman et al. 2002; Yu and Wright 2001). This was particularly so across the southeastern United States and portions of the mid-continent east of the Rocky Mountains. Warmer conditions in these areas were probably due to a convergence of factors, among them: distance from the North Atlantic; that cold Arctic air continued to be trapped north of the Laurentide ice sheet, which allowed an uninterrupted northward penetration of warm Caribbean air into the continental interior (Yu and Wright 2001, p. 361; also Shuman et al. 2002, p. 1786; cf.



Fig. 1 Base map of North America, showing localities for which there is evidence of climate and climate change during the Younger Dryas Chronozone, as well as the location of pollen sites listed in Table 1. See the legend for symbol information. References for information on the localities shown on the map: Alta Babciora (Metcalfe et al. 2000); Barrett Lake (MacDonald et al. 2008); Berry East, Black Gum, Blood Pond (Lindbladh et al. 2007); Bolan Lake (Briles et al. 2005); Boriack Bog (Bryant 1977); Brown's Pond (Kneller and Peteet 1999); Chatsworth Bog, Crystal Lake, and Nelson Lake (Grimm 2008); Chicago Lakes (Benson et al. 2007); Como Lake (Shafer 1989); Copley Lake (Fall 1997); Cottonwood Lake (Barnosky et al. 1987); Crawford Lake (Yu and Eicher 1998); Devil's Lake (Grimm and Jacobson 2004); Folsom (Balakrishnan et al. 2005); Gordon Lake (Grigg and Whitlock 1998); Grand Canyon (Cole and Arundel 2005); Great Salt Lake (Oviatt et al. 2005); Guadalupe Mountains (Polyak et al. 2004); Guardipee Lake (Barnosky 1989); Head Lake (Shafer 1989); Hershkop Bog (Larson et al. 1972); Klamath Mountains (Vacco et al. 2005); Lake Estancia (R.S. Anderson et al. 2002); Lake Tulane (Grimm et al. 2006); Little Lake (Grigg and Whitlock 1998); Makepeace Cedar Swamp and Pequot Cedar Swamp (Newby et al. 2000); Marmes (Huckleberry and Fadum 2007); McCall Fen (Doerner and Carrara 2001); Medicine Lake (Allen 1992; Radle 1981); Misty Lake (Lacourse 2005); Moon Lake (Laird et al. 1996); Mt. Baker (Kovnen and Slaymaker 2005); Mt. Rainier (Heine 1998); Northern Rocky Mountains (Brunelle et al. 2005); Onondaga Cave (Denniston et al. 2001); Owens Lake (Mensing 2001); Patschke Bog (Camper 1991); Pickerel Lake (Watts and Bright 1968); Pond Safety, Lost Pond, Lake of Clouds (Cwynar and Spear 2001); Pyramid Lake (Briggs et al. 2005); Rattle Lake (Björck 1985; Grimm and Jacobson 2004); San Gorgonio Mountains (Owen et al. 2003); San Pedro Valley (Haynes and Huckell 2007); Sherd Lake (Burkart 1976); Sierra Nevada Mountains (James et al. 2002); Sky Pond (Menounos and Reasoner 1997); Southern Great Basin (Quade et al. 1998); Southern High Plains (Holliday 2000a); Spiritwood Lake (McAndrews, n.d.); Splains Lake (Fall 1997); Squamish River (Friele and Clague 2002); Starkweather Lake (MacDonald et al. 2008); Stewart Bog (Jiménez-Moreno et al. 2008); Sunshine (Huckleberry et al. 2001); Twiss Marl Pond (Yu 2000); White Lake (Yu 2007); Woodworth Lake (McAndrews et al. 1967); Yellowstone (Licciardi and Pierce 2008)

Nordt et al. 2008, pp. 1607–1608); and, specifically in regard to the southeast, that the shutdown of the MOC reduced the export of heat to the north, warming the southeastern Atlantic coast (Grimm et al. 2006, p. 2206; Grimm and Jacobson 2004). It seems doubtful that cultural changes in southeastern North America can be attributed to the 'severe cold' of the YDC (e.g. Dunbar and Vojnovski 2007, p. 179).

Precipitation and run-off patterns likewise varied through the YDC, both on a continental and sub-continental scale (e.g. R.S. Anderson et al. 2002; Holliday 2000a; Huckleberry et al. 2001; Mann and Meltzer 2007; Polyak et al. 2004; Quade et al. 1998; Yu and Wright 2001; cf. Haynes 2008, p. 6521). Wetter conditions prevailed across a wide area of the southeast, perhaps as a response to the steep temperature gradient then existing between the cold subpolar waters and warm subtropical waters, which increased the frequency of storms entering the region (Peteet 2000, pp. 1359–1360; also Kneller and Peteet 1999; Grimm and Jacobson 2004, p. 391; Grimm et al. 2006, p. 2208). Elsewhere, precipitation was more variable. Pollen and other proxies from the Pacific Northwest show dry conditions in some areas, but wet conditions in areas not too distant (Briles et al. 2005; Doerner and Carrara 2001; Grigg and Whitlock 1998; Kovanen and Slaymaker 2005; Lacourse 2005). A spatially complex precipitation pattern appears to hold true of the Midwest and New England as well (Lindbladh et al. 2007; Newby et al. 2000), in part depending on proximity to the regional climatic effect of Lake Agassiz (Shuman et al. 2002, p. 1787).

Variation in temperature and precipitation partially explains variability in the responses of glaciers to climate changes during this interval. In Younger Dryas times, ice re-advanced along portions of the Laurentide and Cordilleran ice sheet margins (Dyke 2004; Friele and Clague 2002; Yu and Wright 2001), as well as in the Rocky Mountains, the Coast Mountains of British Columbia, the North Cascade Range in Washington, and even as far south as the San Bernadino Mountains of southern California (e.g. Benson et al. 2007; Kovanen and Easterbrook 2001; Licciardi et al. 2004; Menounos and Reasoner 1997; Owen et al. 2003; Reasoner and Jodry 2000). But this was not a universal pattern. During the YDC, glaciers retreated on Mt. Rainier, or were otherwise—as in Yellowstone and the northwest Sierra Nevada mountains of northern California—unresponsive (Clark and Gillespie 1997; Heine 1998; James et al. 2002; Licciardi and Pierce 2008). Nevertheless, some caution is appropriate here since, as Daniel Mann observes, the absence of a Younger Dryas age moraine does not prove there was no Younger Dryas age glacial response, owing to the inherent and sometimes complicated taphonomy of glacial deposits (D. Mann, personal communication, 2008).

The varied nature of alpine glacier responses probably reflects a complex of causes related to regional variation in precipitation, temperature, seasonality, and snow cover, and the influence of remnant ice sheets on atmospheric circulation, among other factors (Lakeman et al. 2008, pp. 197–198; Licciardi et al. 2004; Oviatt et al. 2005, pp. 280–281; Vacco et al. 2005, p. 254).

Just as climate varied over space, it also changed over time. Great Basin lake records, for example, show a spike in wetter conditions early in the Younger Dryas interval, which within a few centuries gave way to drier conditions (R.S. Anderson et al. 2002; Briggs et al. 2005; MacDonald et al. 2008; Mensing 2001; Oviatt et al. 2005). Changes from wet to dry over the course of the YDC are also apparent elsewhere (e.g. Doerner and Carrara 2001), as are changes in temperature (e.g. Yu and Eicher 1998, p. 2236; Yu 2000, p. 1734). Such, of course, is to be expected given that throughout this period summer insolation and atmospheric CO₂ were increasing, while ice sheet influence on atmospheric circulation was waning (Carlson et al. 2007; Grigg and Whitlock 1998; Kneller and Peteet 1999; MacDonald et al. 2008; Mensing 2001; Oviatt et al. 2005).

Accordingly, we cannot generalize that the Younger Dryas interval was, for instance, strictly cool/dry, or warm/wet. Younger Dryas climates varied spatially, temporally, and seasonally (Yu and Wright 2001), which is to be expected whenever a multifaceted climate change is superimposed over an existing geography that possesses variable topography, complex climatic gradients and physical parameters.

How and to what degree vegetation changed in response to climate change depends on a host of factors, including generation time, location, local and regional climate expression (and severity), dispersal mechanisms and patterns, and response times and thresholds (Grimm and Jacobson 2004, p. 387). Vegetation can track climate change relatively rapidly, ‘consistently <200 year and often <100 year’ by one estimate (Williams et al. 2002). However, this is not necessarily the case for all plant taxa, some of which ‘may significantly lag century-scale and even some millennial-scale climate oscillations’ (Webb et al. 2004, p. 473; Yu 2007; cf. Grimm and Jacobson 2004, p. 394).

Plant dispersal rate and ability—whether into newly available areas or in response to climate change—likewise varies individualistically, as different taxa ‘expand in different directions, at different rates, and at different times’ in response to changes in climate, and depending in part on mechanisms and constraints on seed dispersal (e.g., whether by wind, birds, mammals, or insects) and seed establishment (Grimm and Jacobson 2004, p. 385). For most major forest trees, dispersal rates are typically in the range of 100–500 m/year (Grimm and Jacobson 2004, p. 385), though a few species can move with greater speed. In the northern reaches of eastern North America, for example, pine appears to have shifted its range during this span at rates >300 km/century (3,000 m/year), indicating that at least in this case ‘seed dispersal did not significantly delay the vegetation responses’ (Shuman et al. 2002, p. 1788). In any case, biota in existence coming in to the YDC, already in flux as a consequence of Late Glacial warming, were varied and unique, lacking modern equivalents (Briles et al. 2005; Griggs and Whitlock 1998; Grimm and Jacobson 2004; Kokorowski et al. 2008; Shuman et al. 2002, 2005; Yu and Wright 2001).

In some areas of the Pacific Northwest, Younger Dryas time was marked by dry montane forests, seen in the pollen record as an increase in *Pinus monticola* and a decrease in *Pseudotsuga*, at least at higher elevations. This vegetation type suggests winters were generally cooler and accompanied by abundant precipitation, while summers were warmer and drier, with presumably greater seasonality. In other areas YDC vegetation was marked by increases in spruce and fir, and a decrease in pine (Briles et al. 2005). In general, vegetation response varied across the region (depending, for example, on a site’s proximity to the Pacific Ocean, and its elevation), but in most places climate change ‘was not great enough to result in a large shift in vegetation’ (Grigg and Whitlock 1998, p. 295). Such was also the case further to the north and west in Alaska and Siberia (ancient Beringia), where late glacial vegetation was highly varied, and in some areas showed no apparent changes as a result of climate change during Younger Dryas time (Bigelow and Powers 2001; Hofecker and Elias 2007; Kokorowski et al. 2008).

In eastern North America, biotic communities were likewise seasonally and spatially varied on a ‘subcontinental’ scale (Cannon 2004; Grimm and Jacobson 2004; Shuman et al. 2002). The most significant vegetation response in this region occurred where climate changes were most pronounced: in the more northern latitudes (generally above 40°North) in a swathe from Maritime Canada and New England west into the Great Lakes, and particularly the area immediately adjacent to the North Atlantic (Shuman et al. 2005, p. 2199). In these areas, there were major changes in the distribution and abundance of taxa that led to new, unique and sometimes spatially complex vegetation communities (Grimm and Jacobson 2004, p. 391; Shuman et al. 2002, p. 1784; Yu and Wright 2001). In Illinois, warm-tolerant elm and oak populations expanded during Younger Dryas times. Forests from the Midwest and eastward into New England—which had been ash-dominated during the preceding Bølling-Allerød warm period—were replaced by a westward expanding pine-dominated woodland. Cold- and dry-adapted sedges rapidly returned to prominence in northern areas, while spruce tree range retreated southward (Shuman et al. 2002, pp. 1782,

1785; see also Grimm and Jacobson 2004, p. 391; Shuman et al. 2005; Webb et al. 2004, p. 470). Vegetation changes south and west of those areas were far less significant (Peteet 2000; Shuman et al. 2005, Figure 4; Yu and Wright 2001, pp. 359–360), highlighting the fact that one cannot generalize from one part of the continent to another, even within eastern North America.

Furthermore, although in the northern portion of eastern North America the changes in vegetation are generally attributed to YDC climate changes, that may not be universally true. For example, Lindbladh et al. (2007, p. 506), report that a transition in dominant spruce species from *Picea glauca* to *P. mariana* occurred before the onset of the YDC, driven not by cooling but by an atmospheric CO₂ increase that favored one species over the other.

Vegetation changes in the Pacific Northwest generally started and ended earlier, while overlapping the Younger Dryas Chronozone, but were not synchronous across the region. In some instances, biotic changes seem to have lagged the onset of the YDC by several centuries and/or outlasted it by ~600 years (Grigg and Whitlock 1998, p. 296; also Briles et al. 2005, p. 53; Lacourse 2005). Assuming that the age control is sufficiently precise to constrain these events, this chronological mismatch (which is not unique to this region) could reflect several factors, including differences in when climate changes played out locally or regionally (but see Vacco et al. 2005), the lagged responses of different proxies (such as vegetation) to climate changes, and/or age control problems owing to vagaries in radiocarbon dating during this interval (Briles et al. 2005, p. 53; see also Denniston et al. 2001; Grigg and Whitlock 1998; Grimm and Jacobson 2004, p. 391; Lindbladh et al. 2007; Oviatt et al. 2005; Yu 2007).

In the northern reaches of eastern North America (where, as noted, there was significant vegetation change during the YDC), pollen records provide evidence that vegetation changes were relatively rapid and generally synchronous, at least within the margin of error of radiocarbon dating and the temporal resolution of pollen records (Shuman et al. 2002, p. 1784; Shuman et al. 2005; Yu and Wright 2001). Rapidity in this instance is measured in hundreds of years: ‘much of the [vegetation] change happened during the centuries that mark the beginning and end of the YDC’ (Shuman et al. 2002, p. 1785). Overall, in fact, Shuman and colleagues show that the largest *millennial scale* vegetation shifts in the last 21,000 years occurred at the onset and the termination of the Younger Dryas, although these changes were far more pronounced in eastern North America above 40°N.

Of course, vegetation patterns had been changing since the onset of post-LGM warming, which raises two especially relevant questions: in so far as vegetation changed in response to Younger Dryas climate changes, how rapid was the change(s), and was it rapid enough to have had real-time consequences for Paleoindians or their animal prey? We may assume that animal communities changed in the Younger Dryas as well, as a second-order response to changes in the vegetation and to the niches opened up by mammalian extinctions. However, and unfortunately, the fossil faunal record is not as extensive or detailed as the pollen record (Cannon 2004; FAUNMAP 1996), so less is known of the rate(s) of change.

Rapidity on a multi-century or millennial time scale is assuredly too long to be relevant to real-time human forager adaptations. But were changes in biota occurring more rapidly than that? The ‘effects of rapid changes,’ as Shuman et al. (2005, p. 2195) rightly note, are ‘best isolated’ at shorter time scales (correspondingly, slow changes are only detectable when viewed across long spans of time: Shuman et al. 2005, p. 2195). To address that issue, we turn to the paleoecological record of the Great Plains and adjacent Rocky Mountains, sub-continental areas with a reasonably well-documented and dated Clovis and Folsom Paleoindian presence.

Younger Dryas Vegetation Changes on the Great Plains and Rocky Mountains

As earlier noted, much of the interior of mid-latitude North America remained relatively warm and temperate during the Younger Dryas, since cold Arctic air was still mostly trapped north of the remaining Laurentide ice sheet, and warm Caribbean air was able to penetrate northward and amplify the insolation-driven highs of summer (Yu and Wright 2001, p. 356), leading to greater seasonal extremes than before (or later). Still, there was variation in this region too. In some high elevation areas of the Rockies [e.g. >3,000 m above sea level (masl)], montane glaciers were re-activated, and tundra and timberlines shifted down slope—perhaps 60–120 m (Reasoner and Jodry 2000, p. 52)—and cold-loving trees such as Engelmann spruce and bristlecone pine became more abundant (Fall 1997; Reasoner and Jodry 2000).

At the lower elevation (2,109 masl) of the Folsom type site in New Mexico, the Paleoindian bison bone bed yielded nearly a dozen species of snails that today only live at or well above the elevation of the site, and whose oxygen isotopes hint at summer temperatures lower by several degrees (Balakrishnan et al. 2005; Meltzer 2006). Although temperatures were cooler, it was not cold enough long enough for alpine or subalpine plants to move into the area. Younger Dryas-age pollen and charcoal recovered from the bone bed and from nearby Bellisle Lake on Johnson Mesa came from trees that grow in the region today (Meltzer 2006). Likewise, stable isotopic evidence from bison and snail remains point to drier conditions, as well as to a strong C₄ signature in regional grasslands (Meltzer 2006, pp. 200–203; see also Connin et al. 1998; Koch et al. 2004; Nordt et al. 2007, 2008). These data support the inference of cooler and generally drier conditions on the Great Plains (Mann and Meltzer 2007), and likely also reflect higher summer insolation, among other factors. Also relevant, but not considered in any detail here, are Younger Dryas atmospheric CO₂ values, which can differentially favor C₄ over C₃ plants, even at temperatures cooler than present (Ehleringer et al. 1997; Koch et al. 2004), as well as changes in the seasonality of precipitation (Paruelo and Lauenroth 1996).

Of course, that is cooler and drier relative to the present. In an effort to better understand what pre-Younger Dryas conditions were like, and the rapidity and degree of vegetation change into (and out of) the YDC, we turned to the North American Pollen Database (NAPD) available from the National Climate Data Center (www.ncdc.noaa.gov/paleo/pollen.html).

The Sample and Analytical Procedure

The NAPD provides longitudinal data on pollen and vegetation change. This database was searched for localities with pollen records older than 10,000 cal BP, and ones which occur within a polygon bounded between 29° and 50° north, and 97°–113° west, a search area that broadly overlaps the Plains and adjacent Rocky Mountains. The 63 pollen records found in that search area were from a slightly smaller number of actual localities (some localities were cored on multiple occasions). However, not all of these 63 localities are on the Great Plains or adjacent Rocky Mountains, even by the most generous definition of that region. This is so because the Great Plains extends northwestward from Texas to Montana at an angle to our latitude/longitude search polygon. Sites in Arizona, for example, fell within the polygon but are not on the Great Plains. Excluding such sites, or ones that did not have pollen and sediments that were reliably Younger Dryas in age, quickly reduced the list of 63 localities to just 16 sites (Table 1; Fig. 1). To put that number in perspective, a comparable search of the Great Lakes region that covers just half the area of latitude and

Table 1 Pollen sites from the Great Plains and adjacent areas from the North American Pollen Database

Site (citation)	Latitude/longitude (elevation in meters)	Total number of ¹⁴ C ages	¹⁴ C ages surrounding YDC	Comments
1. Bortiack Bog, TX (Bryant 1977)	30°21'N 97°07'W 143 masl	n = 4 (all peat)	9,580 ± 160 13,810 ± 210 15,460 ± 250	SCD values >0.15 at: 12,700–12,200, 12,200–11,700, 11,700–11,200 cal BP (500 year scale); 12,400–12,200 and 11,600–11,400 cal BP (200 year scale); 11,600–11,500 cal BP (100 year scale). Significant SCD values during and after YDC
2. Como Lake, CO (Shafer 1989)	37°33'N 105°30'W 3,523 masl	n = 5 (1 AMS, 2 wood and 2 gyttja)	8,715 ± 65 10,260 ± 230 11,730 ± 290	No SCD values >0.15
3. Copley Lake, CO (Fall 1997)	38°52'N 107°04'W 3,250 masl	n = 7 (all bulk sediment)	8,210 ± 355 10,325 ± 465 12,810 ± 750	No SCD values >0.15
4. Cottonwood Lake, SD (Barnosky et al. 1987)	44°50'N 99° 54'W 549 masl	n = 11 (AMS only)	9,770 ± 90 10,080 ± 80 10,300 ± 80 11,690 ± 180	SCD values >0.15 at 13,400–13,200 cal BP (200 year scale) and 11,400–11,300 cal BP (100 year scale), prior to and long after YDC
5. Guardipee Lake, MT (Barnosky 1989)	48°33'N 112°43'W 1,233 masl	n = 3 (2 rejected; tephra ages used)	9,220 ± 290	Cannot use—insufficient chronological control
6. Head Lake, CO (Shafer 1989)	37°42'N 105°30'W 2,300 masl	n = 6 (all bulk sediment; 2 basal ages rejected, other 4 corrected)	8,990 ± 290 9,560 ± 160	Cannot use—insufficient chronological control
7. Hershkop Bog, TX (Larson et al. 1972)	29°35'N 97°36'W 124 masl	n = 15 (all peat, many averaged)	6,004 (aver. 5) 10,573 (aver. 5)	SCD value >0.15 at 12,700–12,200 (500 year scale), during YDC
8. Medicine Lake, SD (Allen 1992)	44°49'N 97°21'W 519 masl	n = 5 (only 1 wood date used; others interpolated from Radle 1981)	10,690 ± 90	Cannot use—insufficient chronological control
9. Medicine Lake, SD (Radle 1981)	44°49'N 97°21'W 519 masl	n = 6 (1 wood and 5 gyttja; 1 rejected, others corrected and used)	7,400 ± 100 10,940 ± 135	Cannot use—insufficient chronological control
10. Moon Lake, ND (Laird et al. 1996)	46°51'N 98°09'W 444 masl	n = 14 (all AMS, 2 averaged)	9,530 ± 130 10,780 ± 100 11,770 ± 80	SCD values >0.15 at 11,600–11,400 cal BP (200 year scale) and 12,500–12,400 cal BP (100 year scale), during and long after YDC

Table 1 continued

Site (citation)	Latitude/longitude (elevation in meters)	Total number of ¹⁴ C ages	¹⁴ C ages surrounding YDC	Comments
11. Patschke Bog, TX (Camper 1991)	30°22'N 97°07'W 142 masl	<i>n</i> = 4 (all peat or sediment)	1,590 ± 60 10,090 ± 100 13,470 ± 170	SCD values >0.15 at 11,700–11,200 cal BP (500 year scale), after YDC
12. Pickerel Lake, SD (Watts and Bright 1968)	45°30'N 97°19'W 395 masl	<i>n</i> = 1 (wood; others interpolated)	10,670 ± 140	Cannot use—insufficient chronological control
13. Sherd Lake, WY (Burkart 1976)	44°16'N 107°00'W 2,665 masl	<i>n</i> = 2 (gyttja)	9,060 ± 265 12,910 ± 400	No SCD values >0.15
14. Spiritwood Lake, ND (McAndrews n.d.)	47°04'N 98°35'W 437 masl	<i>n</i> = 3 (gyttja)	8,295 ± 135 10,970 ± 160	SCD values >0.15 at: 13,700–13,200 and 13,200–12,700 cal BP (500 year scale). Significant SCD values prior to and at onset of YDC
15. Splains Lake, CO (Fall 1997)	38°49'N 107°04'W 3,165 masl	<i>n</i> = 8 (all gyttja or clay)	8,750 ± 220 11,040 ± 240 12,020 ± 530	No SCD values >0.15
16. Woodworth Pond, ND (McAndrews et al. 1967)	47°09'N 99°14'W 580 masl	<i>n</i> = 0 (chronology based on pollen correlation with Moon Lake)		Cannot use—insufficient chronological control

Table 1 pollen data from <http://www.ncdc.noaa.gov/paleo/paleo.html>; citation in column 1 refers to original publication

two-thirds of longitude (i.e. 40°–50° north, 80°–90° west)—a large portion of which is water—yielded 76 localities with pollen records predating 9,000 BP.

That list of 16 sites in our sample had to be further reduced in order to eliminate localities that were poorly constrained chronologically. Some of these localities were cross-dated to other pollen cores based on like taxa, or had arbitrarily ‘corrected’ radiocarbon ages, or had dates that were otherwise imprecise. We sought only those sites that had a sufficient number of *reliable* radiocarbon ages from the cores themselves to allow the construction of the calibrated age–depth curve necessary to correctly estimate the ages at particular depths in the core and the error estimates attached to them (Telford et al. 2004, p. 1), as well as to adjust for the effects of changes in sedimentation rate through the core.

To determine how many ages were necessary to have that degree of chronological control, we experimented with several cores in which more than a dozen radiocarbon dates were available, constructing for each an age–depth model with *MCAgeDepth* (Higuera 2008), a program that uses a cubic spline technique and Monte Carlo simulation to generate an age–depth model and associated confidence intervals based on the relative weight (standard deviation) and probability distribution of each date. In selectively removing dates from the data on which the model was based, it quickly became apparent that the age–depth curves became unreliable when only half a dozen dates were available, and became utterly unrealistic when fewer than three radiocarbon dates were available. As Telford et al. (2004) observe, in order ‘to achieve accurate, high precision, radiocarbon-based chronologies, researchers will have to use many more dates than is currently the norm, and uncertainties may always be high during radiocarbon plateaux’—which, of course, is precisely the case for the period of interest here. To this, it is important to add that because deposition rates can vary, particularly during periods when climate was potentially different, such as during the YDC, it is important to ensure that core segments that fell within the YDC are radiocarbon dated, and that their ages have not been determined by extrapolation from ages well above or below that section of the core (D. Mann, personal communication, 2008).

As a result of these data-cleaning steps, our sample includes only nine sites which are within the area of interest and which provide reliable chronologies. Admittedly, this is not an ideal data set: the number of sites is relatively small, and most come from the southern or eastern margins of the Plains, or from higher elevations on the western edge. One of the few sites located squarely on the central Plains, Cheyenne Bottoms (KS), unfortunately had to be rejected because of chronology problems. And, finally, the pollen data are in percentages, and have all the liabilities of closed-arrays (although the latter are routinely used in such analyses). Nonetheless, such are the data that are available.

For each of these localities, percentages of the fifteen most abundant pollen types were tallied for levels dated between ~13,400 and ~11,200 cal BP, to track changes before, during, and after the YDC. Naturally, given the size of the region and differences in climate, elevation, and vegetation, the 15 top pollen taxa were not always the same. (Other approaches have compared the same suite of taxa across a broad area (e.g. Overpeck et al. 1991; Shuman et al. 2005). Some taxa (e.g. *Artemisia*, *Chenopodiaceae*, *Pinus*, and *Poaceae*) were present in many of the cores, while others (e.g. *Picea*, *Quercus*) occurred in fewer, but are useful indicator taxa of Younger Dryas-age changes, if primarily on a regional or local scale.

To assess the degree of vegetation change over time, we determined squared chord distance values (SCD), which are commonly used for assessing differences in pollen percentage data (Overpeck et al. 1985; see also Jacobson et al. 1987; Overpeck et al. 1991; Shuman et al. 2005; Webb et al. 2004; Williams et al. 2001; but see comments by Viau and Gajewski 2007, and reply by Shuman et al. 2007) That statistic is calculated using the following formula from Overpeck et al. (1985): $d_{ij} = \sum_k (\sqrt{p_{ik}} - \sqrt{p_{jk}})^2$ where d_{ij} is the

distance between samples i and j ; p_{ik} is the proportion of pollen type k in sample i ; p_{jk} is the proportion of pollen type k in sample j . SCD values were calculated for intervals of 100, 200, and 500 years (Fig. 2), using a linear model to interpolate pollen percentages between radiocarbon dated points to derive values for those intervals, following the analytical lead of Shuman et al. (2002, 2005), among others (the model was necessary in order to compare cores on the same temporal scale). We used these several intervals because, as noted, slow changes in vegetation will not be apparent across shorter time periods, while rapid changes in vegetation will be difficult to detect in (if not masked by) longer time intervals (Shuman et al. 2005, p. 2195; Viau and Gajewski 2007, p. 266). Ideally, we would examine vegetation changes across even shorter time periods, but recognizing the radiocarbon plateau during the YDC and that standard deviations accompanying radiocarbon ages are often greater than, say, 50 years, any ‘precision’ in doing so would be more apparent than real.

As has been shown by comparing modern pollen samples across different vegetation types, SCD values >0.15 indicate significantly different biomes and, thus, presumably, different climate conditions (Overpeck et al. 1985; Shuman et al. 2005, p. 2195). The >0.15 cutoff is routinely used as a benchmark to identify significant changes in fossil pollen spectra (e.g. Shuman et al. 2002, 2005), and is so used here. It is also important to

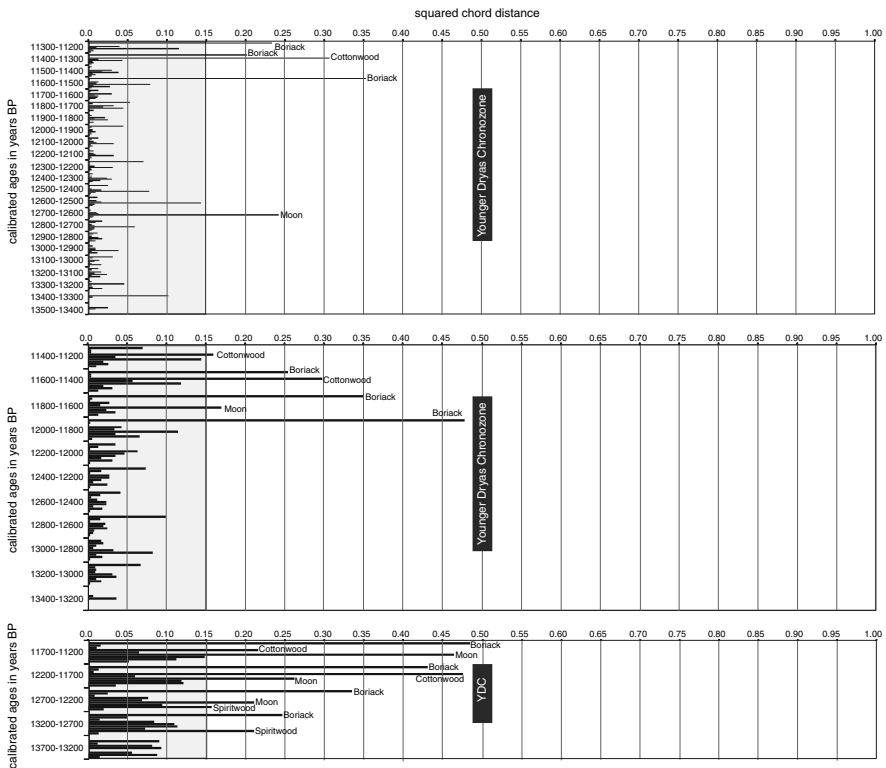


Fig. 2 Squared chord distance (SCD) values for pollen localities (Table 1) on three time scales: 500 years (bottom), 200 years (middle), and 100 years (top) before, during, and after the Younger Dryas Chronozone (in calibrated years). The horizontal bars indicate SCD values for each of the sites within that temporal interval. Bars that fail to extend outside the shaded area (SCD values <0.15) are not significant; only those sites for which the SCD value is >0.15 are significant, and labeled by name. Vertical bars indicate the span of the YDC. See the text for a fuller explanation

explore which taxa appear to be driving any significant SCD values (Shuman et al. 2007; Viau and Gajewski 2007).

Results

Our results are shown in Fig. 2, which displays SCD values occurring at the three different time scales. Localities in which SCD values >0.15 occur are labeled. The greatest number of localities and time periods displaying significant SCD values, not surprisingly, occur on the coarsest temporal scale—500 year intervals (Fig. 2a), but in only four localities: Boriack Bog (on the southeastern Plains), and Cottonwood, Moon and Spiritwood lakes (all on the northeastern Plains). The changes in these two regions are very different in timing and taxa.

At Spiritwood Lake, the 500-year scale changes are driven by a decline in spruce pollen (*Picea*) that begins long before the onset of the YDC: spruce dominates (at 62.6%) the pollen spectrum at 13,400 cal BP, but by the onset of the YDC 500 years later had dropped to just 14.8%; it continues in a slower but steady decline over the next 1,000 years of the YDC. At nearby Moon Lake declining spruce percentages likewise drive the significant SCD values, though the most significant changes appear toward the end of the span (especially post 11,600 cal BP) when spruce declines relatively abruptly. In both sites, the spruce decline corresponds with an increase in pine. At Cottonwood Lake, some 240 km to the south of Moon and Spiritwood, spruce and pine were less consequential elements of the vegetation, declined earlier, and the significant SCD values appear to be driven by the appearance of vegetation that signifies an increasingly open landscape (*Artemisia*, *Ambrosia*).

On the southern Plains, the significant changes that took place at Boriack Bog are quite different from those on the northern Plains. They were driven by fluctuations in alder (*Alnus*), which is generally on the decline after the first few centuries of the YDC, then rises toward the latter half of the YDC, only to decline sharply after the YDC. The more open post Younger Dryas age landscape sees an increase in weedy plants (especially *Asteraceae* and *Ambrosia*).

These changes are significant, but across a broad temporal span (500 years). In fact, significant changes at one locality on one temporal scale may not carry over at that same locality on a finer temporal scale, for that depends on the abruptness and rate of change. Some significant SCD values, in fact, disappear altogether when viewed at a finer temporal scale, though in other cases and at other localities significant vegetation changes come into better temporal focus.

Thus, vegetation changes at Boriack Bog that were significant on a 500 year scale throughout the sequence are only significant on a 200 year time scale after 12,000 cal BP (Fig. 2b). During those centuries alder has a steady rise (from 38.8% to 60.8% at 11,700 cal BP), followed by a sharp decline (to 6.6% at 11,500 cal BP).

On the northern Plains, significant changes are visible on the 200 year time scale at two localities—Cottonwood Lake and Moon Lake. At Cottonwood Lake the significant changes in SCD values only occur after the YDC, and are driven by a sharp rise in *Ambrosia*, especially after 11,400 cal BP. To the north, and near the close of the Younger Dryas Chronozone, Moon Lake saw significant changes in surrounding vegetation, driven by a relatively brief spike in spruce abundance. This taxon, as noted, had been declining over time but rose between 11,800 and 11,700 cal BP, only to diminish sharply thereafter (from 45% to 12.3%). The spruce decline corresponded with the rise of an increasingly open, grass dominated landscape, and with increased abundance of birch and willow trees.

Only three localities underwent significant changes on a 100 year scale (Fig. 2a), and at two of those (Boriack Bog and Cottonwood Lake) these changes took place centuries after the YDC (and these Holocene-age changes involved the increasing opening of the landscape). As was also evident at longer time scales, the changes at Boriack were driven by changes in alder, Asteraceae and Ambrosia; and at Cottonwood, the late spike in Ambrosia. At Moon Lake, the most significant change occurs in the century between 12,700 and 12,600 cal BP which appears to be driven by a very sharp, steep early-YDC decline in spruce (from 59.1% to 21.1%), which then rises briefly—although not significantly—at 12,400 cal BP (to 62.4%) and then again steadily declines until—as noted above—after 11,600 cal BP, when it plummets and then virtually disappears.

Granting the obvious limitations of this data set, here are some general observations on the results obtained from it. First, although vegetation changed throughout the YDC, these changes were more subtle than significant (SCD values <0.15), or at least less abrupt. The greatest number of significant changes occurred on a 500-year temporal scale (at four localities), and significant instances of vegetation change were less frequent on the 200 and 100 year time scales.

Most of the significant changes, however, did not coincide with the onset or end of the Younger Dryas Chronozone, save when viewed across a longer (e.g. 500 year) span of time [which could be a function of our ability to resolve short-term changes in the paleoecological record (C. Whitlock, personal communication, 2008)]. Instead, the majority of the significant changes in vegetation occur in the *middle* centuries of the Younger Dryas Chronozone or the centuries *after* the YDC—an apparent temporal lag in vegetation response (suggesting that critical thresholds for change took longer to emerge), or a response to general insolation-driven growing-season warming [comparable to that seen in the Pacific Northwest (Briles et al. 2005)].

It is telling that there were no significant changes in vegetation biomes in the Rocky Mountains where, by virtue of the sensitivity of alpine vegetation, we might have expected to see more rapid responses to climate change, had YDC cooling, for example, been significant. This is not to say, of course, there was neither cooling nor vegetation change, or that more subtle trends in vegetation change cannot be discerned. But these are more apparent on an individual species level, since different taxa, by virtue of their autoecology, will differ in the time and magnitude of their responses.

For example, at the several high elevation ($>3,000$ masl) pollen localities in the Rocky Mountains (Como, Copley, and Splains lakes) there are visible, albeit statistically insignificant, trends. Most noticeably, spruce levels are relatively constant through the YDC at the lowest of those lakes (Splains Lake, at 3,165 masl), but during that same time period are on the increase at Copley Lake (3,250 masl) while decreasing at Como Lake (3,523 masl). That pattern suggests a down-slope movement of spruce in response to cooler alpine temperatures (see also Fall 1997), which would reverse itself in the centuries after the YDC, as spruce (along with pine) regained dominance at higher elevations.

Likewise, subtle responses to cooler temperatures are visible at Cottonwood Lake and Moon Lake on the northeastern Plains: spruce trees, which were at their maximum in the millennia prior to the YDC, but had begun to decline in the centuries preceding it, increased somewhat over the course of the YDC, peaking in the centuries just before 12,000 cal BP, before again starting to decline. However, at nearby Spiritwood Lake the decline in spruce was significant and monotonic. A mid-YDC rise is not evident at this locality. Ultimately, of course, spruce became an inconsequential (and relict) element of Great Plains vegetation by Early Holocene times, with the long term trend toward fewer trees and more open, grassy environments. Within that span, there was an increase in pine,

which reached a maximum earlier in more southerly localities, later in more northerly ones. But pine was always a relatively minor component of the vegetation (never approaching the amounts seen in the mountains). Given that pine is an over-producer of pollen, and that these trees were then expanding westward into the southern Great Lakes (Shuman et al. 2002), its low level presence in the northeastern Plains may record only long-distance pollen transport. Finally, it is perhaps most important to observe that the changes that were occurring were not necessarily detectable on the timescale of a human lifetime, and would not necessarily have required a rapid or radical adaptive response or change in lifeways.

Overall, the long term *trends* of vegetation change were more significant than the short term ones. Depending on the area, these trends began before the onset of the Younger Dryas, or became evident only well into or after this period. Such general conclusions, however, are based on the limited pollen records available in this region. To supplement these data, we follow Yu and Wright's (2001, p. 351) suggestion of also using 'paleoclimatic information from other deposits, such as loess, sand dunes, and soils' (see also Denniston et al. 2001, p. 148).

Younger Dryas-Age Sediments and Stratigraphy, and Possible Climatic Implications

Of particular interest here is whether the Younger Dryas Chronozone can be recognized as a discrete stratigraphic entity, especially in archaeological contexts. Certainly, strata dating between 12,900 and 11,700 cal BP are widely recognized, but are there deposits or soils that can be directly linked to local climatic conditions during this time period? Some suggest that cooling during Younger Dryas time is apparent stratigraphically (e.g., Fiedel 1999; Huckell and Haynes 2003; Wagner and McAvoy 2004), especially on the Great Plains (Haynes 2008). Yet, given the considerable variation in Younger Dryas age temperature and precipitation, and the highly variable biotic responses, an abrupt, continent-wide, lockstep stratigraphic response seems unlikely. The examples presented below indicate the apparent absence of such a response, although Haynes (2008) presents an alternative view.

Younger Dryas-age stratigraphic records on the Great Plains and in neighboring regions are preserved in valleys, closed basins, and on uplands: settings where most Paleoindian sites are reported (Holliday and Mandel 2006). The valleys well illustrate the variable character of depositional environments on the Great Plains during the YDC. The South Platte River in Colorado northeast of Denver yielded some of the earliest-discovered Paleoindian sites, including the Dent Site, which produced Clovis fluted points associated with mammoth bones (Figgins 1933). The Dent mammoth bones, dated to ~12,900 cal BP (Waters and Stafford 2007), are in the upper alluvium of the Kersey-Broadway terrace (Haynes et al. 1998). That date provides an approximation for the end of Kersey alluviation. If the bones were redeposited (Brunswig 2007), the alluviation must have continued somewhat later. A date of ~11,650 cal BP from fill in the next lower terrace (Haynes et al. 1998) shows that abandonment of the Kersey terrace, incision, and the start of the next alluviation cycle, was between ~12,900 and ~11,650 cal BP. This suggests the YDC was marked by geomorphic instability. The headwaters of the South Platte and its tributaries in the Front Range may have been subjected to Younger Dryas glaciation (Menounos and Reasoner 1997; Benson et al. 2007), but the South Platte out on the Plains was apparently undergoing geomorphic adjustment to tremendous quantities of meltwater and sediment more typical of deglaciation, perhaps due to *pre*-Younger Dryas conditions in the mountains.

A very different geomorphic setting just before and during the YDC is in evidence at the Aubrey Clovis site along the upper Trinity River on the Rolling Plains of north Texas (Ferring 2001a). There, deep valley incision (~ 20 m) began some time after $\sim 26,000$ cal BP (Ferring 2001a, pp. 30, 37), followed by aggradation of coarse-grained mainstream alluvium. Alluviation ceased $>18,200$ cal BP. Landscape stability with localized ponding and deflation occurred during the Clovis occupation. Fine-grained overbank alluviation and burial of the occupation surface began some time between $\sim 12,900$ and $\sim 12,500$ cal BP and continued through the early Holocene (Ferring 2001a, pp. 37–54). The shift in style of alluvial processes from the late Pleistocene into the early Holocene is clearly indicative of decreasing discharge through the late Pleistocene (i.e., high energy to low energy deposits) and is also probably related to increased sediment yield from the drainage basin following occupation and on into the Holocene (i.e., the long term aggradation of fine sediments). These sedimentological and hydrological characteristics are typical of many large streams (Brown 1997, pp. 17–44; Ferring 2001b; Gladfelter 2001; Huckleberry 2001), but there is no obvious chronostratigraphic or climate-stratigraphic link to climates of the YDC.

To the north, the terminal Pleistocene has yet another stratigraphic manifestation. Mandel (2008) reports 49 dated localities from 37 stream valleys, draws, and fans in the Kansas and Arkansas drainage systems, which head on and are incised into the High Plains. Similar to the Trinity River example, these drainages were deeply incised in the late Pleistocene, then began aggrading. Streams were initially quasi-stable in incised positions in the late Pleistocene, meandering across the floors of their respective valleys. Active meandering eventually waned and the floodplains became more stable except for incremental additions of flood deposits. The result was development of an over-thickened (up to 2 m) A horizon that forms a distinct stratigraphic marker. Stabilization and soil cumulization began as early as $\sim 15,600$ cal BP but was underway in most sections between 13,300 and 12,900 cal BP; hence the onset of this process was time-transgressive. The cumelic soils are buried by flood deposits, interpreted as representing the onset of high-magnitude flooding. This shift in depositional environment was likewise time-transgressive, varying from $\sim 11,400$ to $\sim 10,200$ cal BP. Stable carbon isotopes indicate that a warming trend started before 12,900 cal BP and continued into the Holocene. As on the Trinity River, changes in alluvial styles along the Kansas and Arkansas drainages in the late Pleistocene were almost certainly related to declining discharge or at least declines in stream energy. The period of alluvial stability, concomitant soil cumulization, and warming includes the YDC, but is not exclusive to it. Shifting geomorphic responses along these valleys, therefore, cannot be linked to the onset or end of Younger Dryas age cooling.

Lithostratigraphic manifestations of the terminal Pleistocene also vary in lower order drainages with no perennial flow. This is illustrated in two areas with well-known stratified Paleoindian sites: the Southern High Plains of west/northwest Texas and eastern New Mexico, and the neighboring Clayton-Raton volcanic field of northeast New Mexico. The dry valleys or 'draws' of the Southern High Plains are tributaries of the Brazos and Colorado rivers. Holliday (1995) gathered stratigraphic data from 110 localities along 10 draws. These localities included the Clovis type site along Blackwater Draw and the Lubbock Lake site along Yellowhouse Draw. As on the Central Plains, the valleys of the Southern High Plains contained streams in equilibrium (neither incising nor aggrading) in the terminal Pleistocene ($\sim 21,300$ to $\sim 12,900$ cal BP), meandering across the valley floors. Locally, this process ended abruptly at $\sim 12,900$ cal BP when water ceased to flow and valley floors switched to palustrine and lacustrine settings. These lake and marsh conditions locally persisted until as late as $\sim 9,500$ cal BP.

In other reaches of some draws, water continued to flow for centuries or even millennia after 12,900 cal BP [e.g. the base of the overlying lacustrine sediments at the Mustang Springs site is dated to $\sim 11,800$ cal BP (Meltzer 1991)]. Where adequate radiocarbon control is available, the data suggest that the contact between the alluvium and the standing water deposits is time-transgressive downstream, perhaps following a water table that was dropping due to the onset of aridity (Holliday 1995). Stable carbon isotopes and microvertebrates (Johnson 1986, 1987a, b; Holliday 1995, 2000) suggest that a warming and drying trend along the draws occurred through *and after* Younger Dryas time.

The Clayton-Raton volcanic field is on the Great Plains and forms the headwaters of the Dry Cimarron River. Stratigraphic studies include geoarchaeological investigations at the Folsom type site along Wild Horse Arroyo (Meltzer 2006), and in the far upstream headwaters as well as along other tributary reaches of the Cimarron (Mann and Meltzer 2007). The drainages were dominated by aggradation from $\sim 12,900$ to $\sim 11,400$ cal BP, but the style and chronology of filling varied dramatically from site to site. Along the mainstream, alluviation occurred before 11,900 cal BP (Mann and Meltzer 2007). However, at the Folsom site, alluviation or colluviation is dated before 13,200 cal BP followed by eolian sedimentation until $\sim 11,400$ cal BP (Meltzer 2006, pp. 112–153). Along channel floors, incision ensued after $\sim 11,400$ cal BP, while uplands were stable (Mann and Meltzer 2007). Late Pleistocene (and Younger Dryas-age) geomorphic processes in the upper Cimarron varied through time and space. Thus, the stratigraphic manifestation of the Younger Dryas in the upper Cimarron varies significantly depending on location even within a drainage system. As noted above, analyses of carbon, oxygen, and nitrogen isotopes from snail shell and bison bone suggest that the region was dominated by cool, dry summers during at least the middle portion of Younger Dryas time (Balakrishnan et al. 2005; Meltzer 2006). Wetter conditions developed toward the end of the Younger Dryas.

Upland settings on the Great Plains also illustrate an array of geomorphic responses to late-Pleistocene environmental changes and are marked by varied stratigraphic records. LGM silt referred to as Peoria Loess forms the most extensive loess deposit in North America and the thickest LGM loess in the world (Bettis et al. 2003; Roberts et al. 2003). The Peoria Loess is locally buried by Holocene silt referred to as the Bignell Loess (Bettis et al. 2003). At the top of the Peoria Loess where it is buried is the distinctive Brady Soil (Bettis et al. 2003). Where the Peoria Loess is not buried this soil is the regional Mollisol of the Great Plains (Jacobs and Mason 2005, 2007), and represents the Paleoindian landscape of the central Great Plains uplands.

Formation of the Brady Soil has been linked to Younger Dryas age climate (Haynes 2008), but dating clearly shows that pedogenesis began before the YDC and continued afterwards (Mason et al. 2008). Soil formation was initiated by landscape stability beginning $\sim 14,600$ to $\sim 13,300$ cal BP as loess deposition waned. Stability may also have been linked to higher effective precipitation, possibly from a somewhat cooler climate. Dating suggests that stability and the beginning of Brady Soil formation were time-transgressive across the region, 'ranging from just before the B-A [Bølling-Allerød] into the early YD' (Mason et al. 2008, p. 1776). The soil was buried by silt remobilized from nearby dune fields during episodes of early Holocene aridity (generally $\sim 10,700$ to $\sim 10,200$ cal BP) (Mason et al. 2003).

Stable carbon isotopes show that a warming trend began as Peoria Loess deposition ended and the warming continued through pedogenesis (Johnson and Willey 2000; Feggestad et al. 2004; Miao et al. 2007). Some time between 12,900 and 11,400 cal BP effective precipitation declined and drying began, probably due to overall warming. The aridity of the early Holocene represented a culmination of the drying trend (recall this was

the summer insolation maximum) and crossing of a geomorphic threshold (Mason et al. 2008).

This geomorphic and paleoenvironmental scenario for development of the Brady Soil has been applied to the Leonard Paleosol buried within the Oahe Formation on the Northern Great Plains (Mason et al. 2008). The Brady and Leonard soils are distinct stratigraphic entities due to eolian processes, driven by abundant sediment supply in the late Pleistocene, aridity in the early Holocene, and ‘a broad peak of high effective moisture across the late Pleistocene to Holocene boundary, rather than well-defined climatic episodes corresponding to the Bolling–Allerod and Younger Dryas in the North Atlantic region’ (Mason et al. 2008, p. 1772). Thus, the Peoria–Brady–Bignell stratigraphic sequence is a result of a series of factors having to do with regional sedimentological and geomorphic processes, with essentially *no* direct relationship to the climate of the YDC.

The uplands of the Southern High Plains contain two different stratigraphic records that encompass the Younger Dryas Chronozone: dune fields and playa basins. Several dune fields cross the region: the oldest layers of sand are Folsom (i.e. Younger Dryas age) or late-Paleoindian age (Holliday 2000a; 2001). Evidence for Clovis-age sedimentation is rare. The implied record of wind erosion beginning ~12,900 cal BP is believed to be an upland geomorphic manifestation of the drying trend recorded in the draws, described above (Holliday 2000a).

Playa basins are relatively small (most <1.5 km²), roughly circular depressions on the High Plains surface (Holliday et al. 1996, 2008). All radiometric dating shows that the basins predate the YDC by at least a few millennia, and were filling before Clovis time (Holliday et al. 1994, 1996, 2008). Most formed by wind deflation and contain several meters of dark gray mud that accumulated continuously after the basins formed. The fills encompass the YDC. Sedimentology and stable carbon isotopes are indicative of a warming and perhaps drying trend through the YDC (Holliday et al. 2008), but exhibit no obvious Younger Dryas age marker bed.

The data demonstrate that a host of geomorphic processes produced the terminal Pleistocene and early Holocene stratigraphic records of the Great Plains and adjacent areas. Moreover, the Younger Dryas interval is not necessarily manifest as a distinct lithostratigraphic or biostratigraphic entity in all of these different types of deposits and soils. The various geomorphic systems of the Great Plains did not behave synchronously in response to any common climate driver. These stratigraphic records reflect local environmental conditions and probably a complex response to the reorganization of mid-latitude climates in the terminal Pleistocene and early Holocene. For example, once the playa basins formed, the depositional environments within them favored continuous accumulation of one dominant lithology (carbonaceous mud) more or less independent of regional climate changes. The resulting basin fill is essentially homogeneous with no lithostratigraphic indicator of the YDC. In contrast, a wide array of geomorphic processes and depositional environments characterized stream valleys at this time, varying from floodplain stability and terrace formation along the upper South Platte to rapid filling of a deeply incised channel along the Trinity River, and floodplain soil cumulation in central and eastern Kansas and Nebraska.

On the Great Plains uplands, the vagaries of wind erosion and eolian deposition likewise created a varied stratigraphic record. On the Central and Northern Great Plains, an abundant supply of glacial outwash in nearby river valleys resulted in a thick pile of Peoria Loess, while the Southern Great Plains, far removed from outwash streams, were stable. A warming and probably drying trend ensued throughout the Plains at the close of the Pleistocene. The surface of the Southern Great Plains was destabilized ~12,900 cal BP,

resulting in Folsom-age (and Younger Dryas-age) wind erosion and eolian deposition, while to the north, as sediment supply in the valleys waned, the loess stabilized and became modified by pedogenesis through the warming/drying trend. Stability persisted throughout the YDC, perhaps due to a north–south temperature gradient that kept the cooler Central Great Plains relatively more moist. This trend continued until the early Holocene, when even that region was subjected to sufficient aridity and local wind erosion that the Bignell loess was deposited, burying the terminal Pleistocene soil at the top of the Peoria Loess (locally creating the Brady soil-stratigraphic unit).

If a generalization can be made, it is that local stratigraphic records depend on a variety of factors, most importantly local depositional environments, but are also linked to regional depositional settings and local and regional geomorphic processes. These factors, of course, can be linked to climate, but not always, and even the linkages can be indirect (e.g. Schumm 1965; Bull 1991, pp. 104–121, 282–285; Brown 1997, pp. 252–253; Ferring 2001b, pp. 88–89; Gladfelter 2001, pp. 100–109; Huckleberry 2001, p. 71). Attributing these depositional environments (and the resulting stratigraphic sequences) to more distant high-latitude climatic oscillations is even more problematic.

Several researchers have noted the formation of organic-rich ‘black mats’ during Younger Dryas time. These ‘mats’ include black to dark gray to light gray and even white deposits and soil A horizons (Haynes 2008, p. 6520). The Murray Springs Clovis site in southeastern Arizona has perhaps the classic example of this deposit (the ‘Clanton clay’). It is an algal mat dated from ~12,800 to ~11,200 cal BP (Haynes and Huckell 2007, p. 237). Based on data from this site and others across North America, Haynes (2008) argues black mats are a by-product of Younger Dryas climate, a consequence of ‘increased precipitation and/or colder climates when rainfall is more effective in recharging water tables because of reduced evapotranspiration’ (Haynes 2008, p. 6520).

Quade et al. (1998), however, show that black mats are common in the terminal Pleistocene, but also in early and late Holocene sections in the Southwest. They represent recharge of aquifers, and not any particular climate ‘event’. Moreover, they represent relatively slow response to recharge and, in the case of those dating to the terminal Pleistocene, the oldest represent recharge that began by 13,700 cal BP, well before the onset of the YDC. Across the continent various ‘black mats’, such as the lake and marsh deposits at the Clovis type site and Lubbock Lake or the Brady Soil, appear to result from several different processes (e.g. shifts from alluvial to lacustrine or palustrine conditions at Clovis and Lubbock Lake, or upland landscape stability and formation of the Brady Soil, beginning prior to the YDC).

Rather than an abrupt, continent-wide, lockstep stratigraphic response to YDC conditions, there was geomorphic equifinality. That is, a variety of processes were underway on and around Great Plains drainages and uplands during Bølling-Allerød and Younger Dryas times, some of which resulted in carbonaceous sediments or soils. There is little evidence for a causal link between Younger Dryas climate in its ‘type’ area and these geomorphic processes.

Sampling is also a complicating factor in using stratigraphic sections to reconstruct the climate of the YDC, or any specific climate perturbation. Many of the localities described above [and those reported by Haynes (2008)], are archaeological sites, and hence not necessarily representative of regional stratigraphic records. They may reflect local and unique environmental or hydrological conditions, such as spring activity (e.g. Clovis, Lubbock Lake, Mustang Springs), that naturally attracted humans including, eventually, archaeologists. Regional stratigraphic studies both in and away from archaeological sites (e.g. Holliday 1997; Mandel 2008) provide a more complete picture of environmental

changes and also a perspective on the subtleties of such changes (e.g. the time-transgressive nature of changes in valley filling reported by Holliday [1997] and Mandel [2008]).

Would Paleoindians have Noticed Younger Dryas Climate Changes?

We close by returning to the question we asked at the outset. In seeking to answer it, it is perhaps useful to exclude such stereotypic claims as that the North America megafauna were denied access to water and became extinct because the climate ‘suddenly reversed and became glacially cold during the Younger Dryas’ (Haynes 2005, p. 128), causing the freezing of ‘surface water over vast areas’ (Dansie and Jerrems 2005, pp. 56, 72). Neither scenario is supported. Likewise, there was no ‘unexpected onset of ... harsh conditions’ that were part ‘of intense cold worldwide’ (D.G. Anderson et al. 2002, p. 16), let alone characteristic of North America. The notion that YDC cooling was sudden, extensive, severe, and somehow took Paleoindians by surprise seems overly dramatic, if not simplistic.

As discussed above, North American climates cooled in many places (except in southeastern North America) during Younger Dryas times, but there was *not* a return to LGM temperatures, and cooling was most apparent in landmass regions alongside the North Atlantic, and at higher elevations. Yet, even at 2,630 masl on the western slope of the Rocky Mountains, a Folsom group established a winter camp at the Mountaineer site that dates squarely to the middle of the YDC (Stiger 2006). Moreover, Younger Dryas *winters* may have been very cold, but they were followed by Younger Dryas *summers*. And those summers were likely not deep freezes, given insolation and seasonality highs, increasing CO₂, and the relatively spotty record of glacial re-advance during this period (again with the caveat, expressed earlier, regarding the use of the absence of glacial moraines as a climatic indicator). Would climatic and environmental changes of the YDC been noticed or had an impact on Paleoindians? The answer is likely a qualified ‘it depends’—it depends on *where* they were, and *when* they were there.

The answer to the *when* question is complicated, not least by the general dearth of radiocarbon ages on Paleoindian sites across North America. Table 2 lists available radiocarbon dates from Paleoindian sites that fall before, during, and slightly after the Younger Dryas Chronozone (data from Buchanan et al. 2008; Holliday 2000b; Waters and Stafford 2007; we included pre- and post-YDC age Paleoindian sites—the latter were sites postdating 11,700 but predating 11,500 cal BP [10,000 ¹⁴C years BP] in order to see any larger chronological trends). These are radiocarbon ages for which the association between the radiocarbon date and the archaeological occupation is reasonably secure. This group of 45 sites includes only those for which the age uncertainty on the original radiocarbon determination was equal to or less than 100 radiocarbon years; 17 sites with standard deviations >100 radiocarbon years were excluded, so as not to confound sample or laboratory dating uncertainties with potential YDC effects. Calibration was done using CALIB 5.0.1.

A plot of the calibrated ages and standard deviations (Fig. 3) shows that virtually all of the sites fall within the YDC, at least at 1 standard deviation. Since a number of Clovis and Clovis-like fluted point sites postdate the onset of the YDC by centuries, there is little reason to pursue the suggestion that the onset of Younger Dryas climates caused ‘the fragmentation of Clovis culture’ (D.G. Anderson et al. 2002, p. 16). And given the large number of sites that date to the YDC, it is apparent that conditions did not preclude human habitation of North America at this time.

Table 2 Radiocarbon ages on North American Paleoindian sites, in chronological order by decreasing calibrated age

Site name	Component	Radiocarbon years		Calibrated years			
		Age	SD	–1 SD	Age	+1 SD	SD span
Aubrey	Clovis	11,570	70	13,307	13,409	13,473	166
Lubbock	No diagnostic	11,100	60	12,949	13,012	13,071	122
Lange-Ferguson	Clovis	11,080	40	12,941	12,997	13,045	104
Sloth Hole	No diagnostic	11,050	50	12,912	12,973	13,026	114
Anzick	Clovis	11,040	35	12,901	12,956	12,994	93
Bonneville Estates	No diagnostic	11,010	40	12,886	12,931	12,959	73
Paleo Crossing	EFP	10,980	75	12,857	12,924	12,968	111
Dent	Clovis	10,990	25	12,888	12,913	12,932	44
Arlington Springs	No diagnostic	10,960	80	12,848	12,912	12,955	107
Kanorado	No diagnostic	10,980	40	12,887	12,910	12,934	47
Domebo	Clovis	10,960	30	12,837	12,898	12,917	80
Colby	Clovis	10,950	30	12,869	12,893	12,911	42
Shawnee-Minisink	EFP	10,935	15	12,867	12,885	12,899	32
Sheridan Cave	Clovis?	10,915	30	12,855	12,877	12,893	38
Murray Springs	Clovis	10,885	50	12,838	12,865	12,887	49
Lehner	Clovis	10,875	40	12,838	12,860	12,878	40
Mill Iron	Goshen	10,840	60	12,818	12,847	12,873	55
Jake Bluff	Clovis/Folsom	10,760	25	12,800	12,816	12,831	31
Agate Basin	Folsom	10,700	70	12,704	12,756	12,822	118
Black Mountain	Folsom	10,631	85	12,414	12,679	12,806	392
Debert	EFP	10,590	50	12,412	12,656	12,763	351
Blackwater Loc 1	Folsom	10,590	80	12,403	12,627	12,782	379
Lindenmeier	Folsom	10,570	60	12,402	12,613	12,746	344
Cooper	Folsom	10,530	45	12,397	12,542	12,668	271
Sunshine	No diagnostic	10,500	27	12,395	12,518	12,619	224
Folsom	Folsom	10,490	20	12,393	12,518	12,598	205
Charlie Lake	No diagnostic	10,500	42	12,393	12,513	12,620	227
Vail	EFP	10,530	100	12,255	12,509	12,729	474
12 Mile Creek	Folsom	10,500	66	12,386	12,495	12,651	265
Upper Twin Mtn	Goshen	10,470	50	12,244	12,473	12,628	384
Barger Gulch	Folsom	10,470	70	12,240	12,449	12,637	397
Wilson Butte	No diagnostic	10,440	67	12,169	12,372	12,589	420
Lookingbill	Late Paleo	10,405	95	12,098	12,308	12,572	474
Mountaineer	Folsom	10,420	20	12,171	12,306	12,392	221
Waugh	Folsom	10,390	60	12,126	12,275	12,386	260
Wasden	Folsom	10,330	50	12,049	12,167	12,347	298
Sheaman	Goshen	10,290	20	12,038	12,069	12,112	74
Hanson	Folsom	10,255	95	11,767	12,010	12,227	460
Kincaid	Folsom	10,245	80	11,776	11,989	12,133	357
Lubbock	Folsom	10,205	65	11,772	11,907	12,046	274
Dilts	Goshen	10,170	50	11,763	11,866	11,975	212

Table 2 continued

Site name	Component	Radiocarbon years		Calibrated years			
		Age	SD	-1 SD	Age	+1 SD	SD span
Jim Pitts	Goshen	10,160	50	11,755	11,846	11,975	220
Marmes	No diagnostic	10,120	82	11,416	11,734	11,974	558
Bonfire Shelter	Plainview	10,080	100	11,397	11,650	11,948	551
Two Moon	Folsom	10,060	60	11,404	11,598	11,751	347

Table 2 data from Buchanan et al. (2008), Holliday (2000b), Waters and Stafford (2007) and Waters et al. (2009)

The chronological record of Paleoindians is complicated, however, especially for sites that fall in the YDC. Sites that date before or within the first century of the YDC—that is, all sites older than 12,800 cal BP (Aubrey to Jake Bluff in Fig. 3)—have calibrated standard deviations that span just 75 years on average. Likewise, all of these sites have radiocarbon ages that are in a portion of the calibration curve for which there is only a single intercept. In contrast, all sites that postdate 12,760 cal BP (starting with Agate Basin at $12,756 \pm 118$ cal BP), have radiocarbon ages that—because of the ‘wiggles’ in this portion of the calibration curve (Reimer et al. 2004)—have multiple intercepts (average 2.21). As a result, one standard deviation for these calibrated ages spans an average of 324 years. The difference in the size of the standard deviation between the two groups of sites (pre- and post-12,760 cal BP) is statistically significant (as measured by G scores and Freeman-Tukey deviates).

In effect, the chronological resolution of the age of Paleoindian sites degrades considerably during the YDC. This ‘degradation’ may be attributable to at least two factors: variation in YDC atmospheric ^{14}C (as discussed above), an obvious possibility given that the degradation neatly coincides with the abrupt $\sim 40\%$ rise in atmospheric $\Delta^{14}\text{C}$ that began at 12,760 cal BP (Hua et al. 2009, p. 2985–2986), and/or shortfalls in the precision of the radiocarbon calibration curve, which at the moment only reaches back with secure dendrochronological control to 12,410 cal BP (Reimer et al. 2004). The former factor seems the more likely explanation, given the impact of variation in atmospheric $\Delta^{14}\text{C}$ on radiocarbon ages (Muscheler et al. 2008). It seems less likely that inadequate calibration is to blame, since the YDC centuries within reach of dendrochronological calibration (e.g. 11,700–12,410 cal BP) still have broad standard deviations. When the dendrochronologically based calibration is extended, it will be of interest to see whether standard deviations for the period from 12,410 to 12,760 cal BP become narrower than they are now, and those earlier than 12,760 cal BP (and 12,900 cal BP) become broader than they are now.

This pattern in the radiocarbon dates has several implications, and not just for pinning down the age of sites occupied during the YDC. For one, Waters and Stafford (2007, p. 1124) interpret the tight cluster of Clovis radiocarbon ages as indicating that Clovis technology ‘appeared synchronously across the United States.’ Such a pattern could result from either the rapid spread of Clovis people across an empty continent, or the fast diffusion of Clovis technology ‘through a preexisting but culturally and genetically undefined human population in North America’ (Waters and Stafford 2007, p. 1124; for a fuller discussion, see Storck 1991). They concluded in favor of the latter. But to select one over the other at this moment seems premature, since the synchronicity of Clovis radiocarbon ages might be more apparent than real, reflecting a lack of high precision calibration of radiocarbon ages during this period of time.

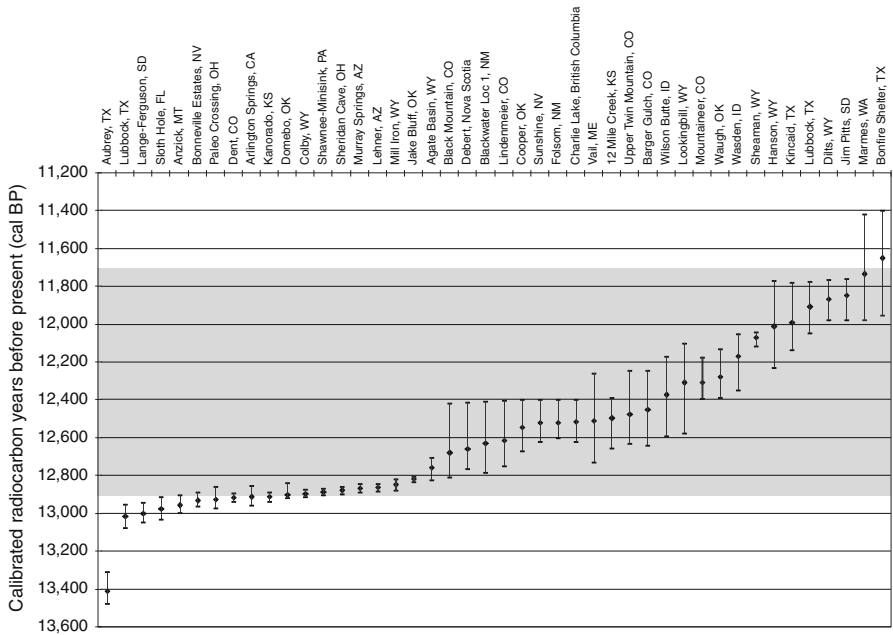


Fig. 3 Boxplot showing calibrated ages (*center diamond*) and 1 standard deviation (*vertical bars*) for Paleoindian sites, ordered by decreasing calendar age (data from Table 2). The *shaded area* represents the Younger Dryas Chronozone; at 1 standard deviation virtually all sites intersect the YDC. Note the increasing span of years represented by 1 standard deviation for sites younger than ~12,700 cal BP (i.e. Black Mountain and sites to the right of it). As described in the text, although the span of calibrated years increases significantly for those sites, the standard deviation of their original radiocarbon dates is not significantly different from those sites older than ~12,700 cal BP

Likewise, there has been considerable recent debate regarding a hypothesized post-Clovis population collapse owing to a supposed comet impact (the very same comet that allegedly triggered the YDC; Firestone et al. 2007). In the absence of direct empirical evidence of such a population collapse (e.g. catastrophic site abandonment or skeletal evidence of widespread mortality), this supposition appears to be based primarily on a change in projectile point styles, and a chronological gap between Clovis and post-Clovis sites. In regard to the latter, Fig. 3 indicates there is no significant gap or discontinuity in the radiocarbon record of Paleoindian sites around the onset of the Younger Dryas Chronozone (the supposed moment of impact). There are, however, temporal discontinuities at several later times. Unless one is prepared to argue these also represent episodes of demographic collapse (we are not), Occam’s Razor (the principle of parsimony) points to such gaps being a function of either ambiguity in the calibration curve, or spotty sampling, and *not* Paleoindian population changes. On a related note, using the frequency of radiocarbon ages through time as a record of Paleoindian ‘demography’ (Buchanan et al. 2008), as opposed to looking at continuity and discontinuity, seems even more fraught with problems. ‘Peaks’ or ‘valleys’ of radiocarbon dates over time can be caused by a host of extraneous factors beyond those already noted, such as whether material from a site was available for dating, whether the samples were run (which is often largely an issue of an investigator’s budget), whether the results were reliable, and so on. For a fuller discussion and critique of the alleged comet-caused Clovis termination, see Holliday and Meltzer (2010).

For Paleoindian sites that fall within the YDC, it will be difficult to demonstrate that a particular change in the archaeological record was a direct response to climatic or environmental change at a specific archaeological moment and location (Meltzer 1991, p. 237), since such changes cannot be dated more precisely than a several-century span. Linking climate and culture change requires both tight chronological correlation and (more important) strong evidence of causation. It will be a challenge, with sites dating to the YDC, to even achieve correlation.

As to *where* to look for such changes, Paleoindian groups occupied much of North America during the YDC, inhabiting and adapting to a wide variety of landscapes. Over that space and span, an increasingly diverse range of artifact styles, tool kits, subsistence and settlement strategies developed (Meltzer 2009). But one cannot a priori assume that such changes were caused by YDC conditions, rather than merely coincident with them. To tie such archaeological developments to the consequences of Younger Dryas age climate or environmental changes will require—given how subtle some of the latter are—a nuanced demonstration of how the climate and environment of the YDC (1) triggered fluctuations in prey populations or plant foods on which people depended, or (2) reduced surface water, and/or (3) caused other changes on the landscape that would have been *detectable* to and directly impacted foragers. It also requires specific archaeological signatures of that response (again, no easy task, given the mismatch between the temporal and spatial scales of the archaeological and paleoecological records on the one hand, and the on-the-ground adaptive activities of hunter-gatherers on the other [Meltzer 2004]).

The most promising places to look for a YDC ‘impact’ are the northern areas of North America, especially northern New England and Maritime Canada, where Younger Dryas climate changes were relatively severe (Peteet 2000; Shuman et al. 2005) and occurred on a faster time scale. Newby et al. (2005) observe that during the Younger Dryas the vegetation of New England and Maritime Canada was largely open tundra, as spruce parkland (and the northern tree line) had shifted south in response to colder and drier climates. As the Younger Dryas interval came to an end, tundra and spruce parkland tracked back to the north, and were replaced by closed, mixed pine forests. They believe this vegetation change had an impact on Paleoindians, for ‘local mammal resources were bound to have been different’ across that vegetation divide (Newby et al. 2005, p. 152). If northeastern Paleoindian groups were caribou hunters (which seems a reasonable assumption, even if direct evidence is meager), the availability of their prey would have changed dramatically as open landscapes gave way to closed mixed pine forest, and herds of caribou were replaced by non-gregarious large mammals such as moose and deer.

Although a reasonable inference, there is no direct evidence of a local post-Younger Dryas age shift in faunal communities (this may reflect on the paucity of evidence, for such changes are occurring on a larger geographic scale, e.g. Cannon 2004). Nor is it evident that the change in the style of projectile point hafting—from fluted to non-fluted points—reflects (or need reflect) that shift in prey suggested by Newby et al. (2005, pp. 142, 150). In fact, there is little reason why fluting should disappear if the targets were moose or deer, as opposed to caribou, and the chronological correlation between the disappearance of fluting and a shift to hunting moose/deer is not secure (point styles were changing throughout eastern North America at this time; it may just be coincidental that one landed on this moment of ecological change, especially when that ‘moment’ extends over several centuries due to the lagged responses of plant and animal communities). More needs to be known about other changes in the tool kit and adaptive strategies of these groups, in order to link presumed changes in prey type to ‘altered hunting techniques or butchering practices’ (Newby et al. 2005, p. 152; see also Ellis and Deller 1997; Eren 2008).

The main lesson to be drawn from the Newby et al. (2005) example is that it is not about a Paleoindian response to the *onset* of Younger Dryas age climate and environmental changes, or even conditions during this period, but instead to changes that *followed* the YDC (see also Lovvorn et al. 2001). As Newby et al. (2005) rightly note, Paleoindian groups were present in this area during the YDC (at sites like Debort and Vail, for example), despite the fact that this was a time (and place) of relatively colder climates and longer term significant changes in vegetation (Shuman et al. 2005).

Even though it is not the focus of our discussion here, Alaska and the high Arctic may be another region to test for a human response to YDC changes in climate and environment. There is some evidence that conditions during this period were harsh. Hoffecker and Elias (2007, p. 228), report that the YDC climate and biota of Beringia were undergoing ‘rapid and dramatic change, and the archaeology reflects this dynamism.’ That said, there is considerable disagreement as to just how severe conditions were, and how or whether changes in the archaeological record—a presumed hiatus of sites and radiocarbon ages—reflect those conditions (cf. Bever 2006; Bigelow and Powers 2001; Goebel et al. 2003; Kokorowski et al. 2008; Kunz and Reanier 1994; Mann et al. 2001; Potter 2008).

All things considered, it is likely that across most of North America south of the retreating ice sheets Paleoindians were not constantly scrambling to keep up with Younger Dryas age climate *changes*. Even so, longer-term *cumulative* trends over the millennium of the YDC unquestionably (if invisibly, from the perspective of a human lifetime) changed the adaptive stage for Paleoindian groups.

Consider, for example, the observation that bison populations on the Great Plains were apparently larger in post-YDC times than earlier, and this despite the fact that the continent-wide bison distribution was considerably reduced (Hill 2007; McDonald 1981). It is difficult to establish whether that increase was a consequence of vegetation changes driven by shifts in climate during the YDC (such as an increase in open, C4-dominated grassland, as discussed above; see also Meltzer 2006), or of the competitive release that followed the extinction of mammoth, horse, camel and other Pleistocene grazers, or perhaps a combination of the two (which seems the most likely explanation). Regardless, this trend, which began in the YDC, created increasingly favorable conditions for bison and their human predators.

Bison are represented in increasing frequency in Clovis, Folsom, and later Paleoindian sites: there are approximately one dozen known to have been killed by Clovis hunters (virtually all of those met their demise at the Murray Springs site: Haynes and Huckell 2007), with Folsom-age bison kills having typically netted an average of 15 animals, while later Paleoindian bison kills averaged nearly 60 animals, a fourfold increase (Meltzer 2009). Whether that increase was the result of hunters becoming more adept, bison becoming more abundant or, once again, both, the net effect was that bison became a significant resource for Paleoindians (Cannon and Meltzer 2004; Grayson and Meltzer 2002; Hill 2007).

In sum, there is reason to expect that the ecological challenges posed to ancestral Paleoindians at the outset of the YDC, were quite different than those facing their descendants many centuries and a millennium later. Of course, we must also be mindful that Paleoindians were highly mobile, still relatively new to North America when the YDC began, had yet to experience relatively *stable* climates or environments, and still were not adaptively tied to a specific region or dependent on particular resources in specific environments. That would come later. Even were they in regions where YDC climatic and ecological changes were occurring on a scale of multiple decades or centuries, they still might not have noticed, since people respond more directly to daily, weekly, and seasonal

conditions. Besides, adapting to changing climatic and environmental conditions was nothing new to them. It was what they did.

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