Paleoclimate and Amerindians: Evidence from stable isotopes and atmospheric circulation

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Two Amerindian demographic shifts are attributed to climate change in the northwest plains of North America: at ≈11,000 calendar years before present (yr BP), Amerindian culture apparently split into foothills–mountains vs. plains biomes; and from 8,000–5,000 yr BP, scarce archaeological sites on the open plains suggest emigration during xeric "Altithermal" conditions. We reconstructed paleoclimates from stable isotopes in prehistoric bison bone and relations between weather and fractions of C_4 plants in forage. Further, we developed a climate-change model that synthesized stable isotope, existing qualitative evidence (e.g., palynological, erosional), and global climate mechanisms affecting this midlatitude region. Our isotope data indicate significant warming from ≈12,400 to 11,900 yr BP, supporting climate-driven cultural separation. However, isotope evidence of apparently wet, warm conditions at 7,300 yr BP refutes emigration to avoid xeric conditions. Scarcity of archaeological sites is best explained by rapid climate fluctuations after catastrophic draining of the Laurentide Lakes, which disrupted North Atlantic Deep Water production and subsequently altered monsoonal inputs to the open plains.

The influence of Holocene warming on cultural change and landscape use among Amerindians on the northwest plains of North America is strongly debated. Numerous archaeological sites on the northwest Great Plains (hereafter High Plains; Fig. 1) and adjacent Rocky Mountains indicate that Paleoindians ranged continuously over both plains and mountain biomes before the end of the cool Younger Dryas period at ≈11,600 years before present (yr BP). However, after ≈11,000 yr BP (all dates are in calendar years), shifts in biota and the final extinction of megafauna (1) apparently prompted Paleoindians to separate into two concurrent but distinct cultural groups: one favoring the foothills–mountains region and the other depending heavily on bison across the open plains (2). Archaeological evidence indicates a decline in bison kills on the plains later, from 9,000 to 8,000 yr BP. After that, although habitation continued in the mountains, from ≈8,000 to 5,000 yr BP archaeological sites on the High Plains became quite rare (3). Initially, caliche soils (sometimes xeric indicators) and lack of archaeological finds were interpreted to mean complete cultural abandonment because of prolonged "Altithermal" desertification (4). However, calcareous soils can also form from water-borne solutes deposited during periods of increased precipitation. Thus, mid-Holocene summers might have been wet as well as hot and, indeed, limited seasonal hunting and gathering apparently occurred on the open plains during the mid-Holocene (3, 5). Nevertheless, the relative and absolute number of archaeological sites on the entire Great Plains decreased significantly between 8,000 and 5,000 yr BP (6).

Evidence from the High Plains for the timing and intensity of Holocene warming is fragmentary and based mainly on pollen and erosional data that provide only qualitative estimates of climate. Consequently, we took two new approaches to investigate the chronology and intensity of Holocene warming. First, we measured stable carbon isotopic values (expressed as $\delta^{13}C$) in bison bone from archaeological sites and related the relative dominance of C_3 vs. C_4 plants in bison forage to weather variables in Wyoming. Second, to provide alternative evidence for climatic chronology, we developed a climate model that synthesized our stable isotope evidence with existing qualitative data (e.g., palynological, erosional) and current information on the oceanic and atmospheric mechanisms of global climate change as applied to the High Plains.

Weather on the High Plains varied with relative atmospheric inputs from the Pacific Ocean, from the Gulf of Mexico [Fig. 1; black arrows (*Inset*)], and from the northeast before the draining of Lakes Agassiz and Ojibway (the Laurentide Lakes) (Fig. 1 *Inset*, white arrow). Because of orographic precipitation, Pacific westerly winds conveyed little moisture to the Plains. Thus, with loss of moisture from the northeast (7) after the Laurentide Lakes drained at ≈8,200 yr BP (8) and effects of the Laurentide Ice Sheet declined greatly (9), monsoonal rains from the Gulf of Mexico (7, 10) determined the severity of summer warming on the High Plains.

Methods

Bison and Archaeological Sites. Bison bones were collected from archaeological sites that encompassed the two periods of interest (11). These sites spanned 5,000 years within a 90-km radius (Fig. 1). The relative amounts of ^{13}C and ^{12}C ($\delta^{13}C$ values) in bone collagen reflect the relative dominance of C_3 and C_4 plants in the diet in adult bison (12–14). From mean $\delta^{13}C$ values in bone collagen of bison excavated at each archaeological site and/or cultural stratum, we estimated the fraction of C_4 plants in concurrent vegetation [Table 1; see also Table 3 for individual $\delta^{13}C$ values (published as supplemental data on the PNAS web site, www.pnas.org)] (11). Individual $\delta^{13}C$ values are listed in Supplementary Material. Because bison are unselective grazers (17), $\delta^{13}C$ values in their bone collagen directly reflect the composition of vegetation communities and in adult bison do not detectably deviate from dietary input with age (14). Before invasion by humans, herds apparently grazed within a radius of ≈160 km determined by the distribution and abundance of food and water (17, 18). Conservative estimates of collagen turnover are 13–20 years (16). Communal bison kills seldom contain more than one or two individuals from older age classes (>10 years old) (4). Except for one subadult from the Carter/Kerr–McGee (CKM) site, we sampled adult bison.

From the CKM site (48CA12), we report $\delta^{13}C$ values from the Folsom cultural level (19). From the Agate Basin site (48NO201), three distinct strata represent hunts from three cultural periods: the Folsom (ABF), the Agate Basin (ABAB),

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Abbreviations: yr BP, calendar years before present; $\delta^{13}C$, stable carbon isotope values; CKM, Carter/Kerr–McGee archaeological site; ABF, Folsom level at the Agate Basin archaeological site; ABAB, Agate Basin level at the Agate Basin archaeological site; ABHG, Hell Gap level at the Agate Basin archaeological site; C_3 and C_4, plants that fix three or four carbon molecules, respectively, in their first stable photosynthetic compound; NWS, National Weather Service; NADW, North Atlantic Deep Water; HK, Hawken site.

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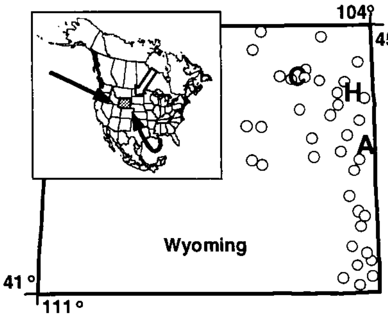

Fig. 1. Map of archaeological sites (letters) and NWS reporting stations (circles) on the High Plains of Wyoming. (*Inset*) Inputs of major air masses that dominated climatic circulation on the High Plains at various periods during the Holocene. Archaeological sites are Carter/Kerr-McGee (C), Hawken (H), and Agate Basin (A).

and the Hell Gap cultures (ABHG) (20). From the Hawken site (HK) (48CK303), $\delta^{13}C$ values were collected from the Early Plains Archaic Period (21). Both CKM (elevation 1,384 m) and Agate Basin (elevation 1,189 m) lie on low scarps and rolling terraces presently vegetated by short grass prairie dominated by *Bouteloua gracilis* (C_4), *Agropyron smithii* (C_3), and *Stipa comata* (C_3). HK (elevation 1,524 m) lies within the Red Valley at the edge of the Black Hills among short and mixed grasses dominated by *A. smithii* (C_3) and *S. comata* (C_3) with some *Andropogon scoparius* (C_4).

C_3/C_4 Plant Distribution. C_3 or C_4 designates whether plants fix three or four carbon molecules, respectively, in their first stable organic compound of photosynthesis. The distribution of C_4 plants generally varies in relation to altitude (temperature minima) and seasonality (22, 23). In general, cool and moist conditions favor C_3-dominated communities, because some C_3 plants tolerate lower temperatures and have lower water-use efficiencies than C_4 plants. In contrast, C_4 plants are favored in open environments with high irradiance, which are often warm and dry (24). An important phenological distinction strongly influences C_3/C_4 ratios in plant communities. Although mature C_4 plants withstand water stress better than mature C_3 plants, dry soils and low temperatures are more limiting for germination and seedling establishment of C_4 than of C_3 plants (e.g., ref. 25). Thus, increased C_4 production generally indicates a climate shift not simply to xeric conditions, as is generally emphasized, but rather denotes wet warm springs and dry summers. Such climate change is incorporated into the $\delta^{13}C$ values of the bones of grazing bison (14).

C_4 Fractions and Isotope Values. C_3 plants discriminate more strongly against ^{13}C than do C_4 plants, resulting in mean $\delta^{13}C$ values of $-27‰$ in C_3 plants and $-12.5‰$ in C_4 plants (24). Stable carbon isotope values are expressed relative to a fossil belemnite, *Belemnitilla americana*, as $\delta^{13}C$ values in parts per thousand (‰): $(^{13}C/^{12}C\ \text{sample}/^{13}C/^{12}C\ \text{standard} - 1) \times 1{,}000$. Prehistoric atmosphere was enriched in ^{13}C with by $\approx 1‰$ (26). Additionally, $\delta^{13}C$ values are biochemically enriched in bone collagen by 5‰. Consequently, we calculated the fraction of dietary C_4 plants by direct interpolation of $\delta^{13}C$ values from bone collagen (27) between 100% C_4 plants ($-12.5‰ + 6‰ = -6.5‰$) and 100% C_3 plants ($-27‰ + 6‰ = -21‰$).

Collagen Extraction. Nonmineral components of bone (here referred to as "collagen") were isolated and purified by first mechanically removing the outer 3–4 mm of compact bone ($\approx 0.5 \times 1 \times 5$ cm and ≥ 5 g). Bones were treated with 2:1:0.8 methanol/chloroform/deionized water solution to remove lipids, then demineralized in 0.3 M HCl (paired samples from CKM were demineralized in ethylene diamine triacetic acid, with no difference in the quality of collagen matrices or in their $\delta^{13}C$ values) at 4°C on a gyratory shaker table (100 rpm). Humic substances were removed by soaking the collagen in 0.125 M NaOH at 4°C for 24 h on a gyratory shaker table (100 rpm). A final 30 min immersion of collagen in the methanol/chloroform/deionized water solution removed remaining lipids. Collagen was solubilized in 100 ml of deionized water held at pH 3.0 in a 90°C water bath for 12 h, which also removed diagenetic carbonates. Solution pH was maintained by additions of 0.3 M HCl. When the solution was completely solubilized, the pH was adjusted to 5 with 0.0125 M NaOH. After evaporation at 70°C to 20 ml volume, samples were shell-frozen and freeze-dried (11, 28). Samples were combusted in a Carlo Erba CHN (carbon hydrogen nitrogen) analyzer coupled to a SIRA-10 Isotope Ratio Mass Spectrometer fitted with a special triple trap to isolate cyrogenically and purify the CO_2 (analyses performed at Tieszen Laboratory Augustana College, Sioux Falls, SD). Lab standards and random replicates ensured that precision was better than $\pm 0.2‰$. Any sample failing to meet criteria for minimal diagenesis was rejected. Criteria are (*i*) percent nitrogen $>1\%$; (*ii*) carbon/nitrogen = 2.9 to 3.5; (*iii*) percent of initial mass $>4\%$; (*iv*) percent carbon $>1.7\%$; and (*v*) collagen matrix maintained structural cohesion with loss of shape only at edges (refs. 11 and 14; L.L.T., unpublished material).

C_4 Fractions vs. Climate. On the basis of plant samples taken by the Natural Resources Conservation Service near 37 National Weather Service (NWS) reporting stations in Wyoming, we developed linear regressions (29) relating the fraction of above-ground net primary production of C_4 plants in average years to 15 climatic variables for the maximum period of record from the NWS (20–80 years) (30, 31). The NWS stations (Fig. 1) lie between 1,188 and 1,850 m altitude. The Natural Resources Conservation Service quantifies above-ground net primary pro-

Table 1. $\delta^{13}C$ values in bison bone from archaeological sites and resulting estimates of %C_4 plants in past plant communities (see text)

Site	*n*	MNI	Species	^{14}C yr BP	yr BP	$\delta^{13}C$	%C_4
CKM	7	4	*Bison antiquus*	10,612 (74)	12,590 (100)	−18.18 (0.88)	19.72 (6.07)
ABF	11	9	*B. antiquus*	10,612 (74)	12,590 (100)	−17.72 (1.02)	22.62 (7.03)
ABAB	14	14	*B. antiquus*	10,430 (570)	12,390 (960)	−18.53 (0.56)	17.03 (3.86)
ABHG	11	11	*B. antiquus*	10,157 (325)	11,910 (706)	−16.62 (1.66)	30.21 (11.45)
HK	12	12	*Bison occidentalis*	6370 (150)	7280 (147)	−15.49 (2.25)	38.00 (15.52)

We calculated the minimum number of individual bison (MNI) sampled from each cultural component by sampling the right calcanei, non-conjoining astragali, or fused 2/3 tarsals (88% of samples), and by significant size or morphological contrasts in bones. Standard deviations are shown in parentheses. For ABF and ABHG, radiocarbon dates were derived from combined regional radiocarbon dates weighed by their standard deviations (11). The ABAB ^{14}C date (RL-557) was the most reliable of four radiocarbon dates. The HK ^{14}C date is the average of two optimal dates (RL-437, RL-185). Radiocarbon dates are converted to calendar years (15).

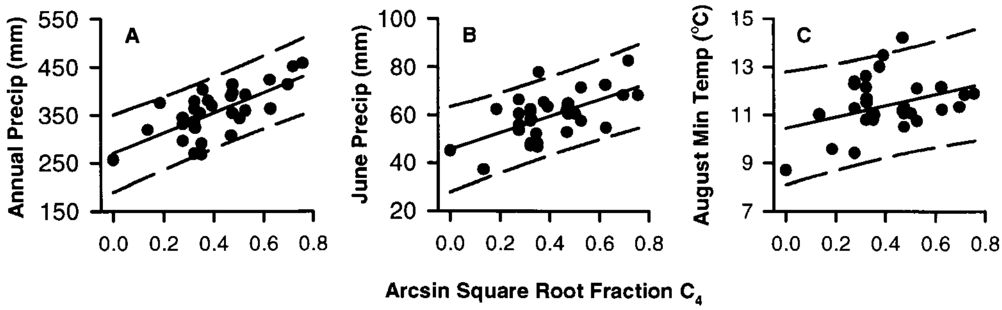

Fig. 2. Relationships between the fraction of above-ground net primary production of C_4 plants in average years (X = arcsine of $\sqrt{}$ fraction C_4) and (*A*) normal annual precipitation (Annual Precip = 271.28 + 217.06 X, $r^2 = 0.47$, $P = 0.0001$), (*B*) normal June average precipitation (June Precip = 45.18 + 36.99 X, $r^2 = 0.35$, $P = 0.0006$), and (*C*) normal August minimum temperature (August Min Temp = 10.30 + 2.39 X, $r^2 = 0.17$, $P = 0.02$). Dashed lines show 95% prediction intervals.

duction of all plant species with >5% contribution either inside exclosures at disturbed sites or at pristine sites. We calculated the fraction of total biomass comprised of C_4 plant species for normal production on the three major soil types (11).

Among several significant correlations with C_4 vegetation, the strongest were normal annual precipitation, normal June average precipitation, and normal August minimum temperature (Fig. 2). Several modern sites have zero C_4 production. Consequently, we performed an arcsine square-root transformation on the fraction of C_4 production, which makes the variances independent of the mean and produces a better fit. We substituted the transformed (arcsine$\sqrt{}$) average fraction of C_4 vegetation (Table 1) from each sampled population into regression equations (Fig. 2) to hindcast paleoclimate conditions (Table 2).

Tests of Linear Regression Models: Predicted vs. Observed. To test goodness of fit for each linear regression model, we compared predicted vs. observed climatic values for NWS reporting stations. Ninety-seven percent (for normal annual precipitation) to 100% (for both normal June average precipitation and normal August minimum temperature) of predictions generated from regression models fell within a 95% prediction interval, which was calculated from observed values reported at modern NWS stations. Prediction interval is the appropriate test, because it includes variability caused by uncertainty about a single (vs. repeated) prediction (32). As a further test of the model, we predicted values for four sites that were located within the geographic region but not used to develop the models. One hundred percent of those new predictions fell within a 95% prediction interval for normal annual precipitation, June average precipitation, and normal August minimum temperature.

Although multilinear regressions (general linear models procedure, step-wise entry) explained as much as 92% of the variation in C_4 production (29), we rejected multilinear regression as a reliable method for estimating paleoenvironmental conditions for several reasons. Multilinear regressions included some independent variables that, despite improving the overall fit of the model, lacked a significant individual relationship with C_4 production. Moreover, multilinear regressions relating several climatic variables to C_4 production have entirely different slopes and errors about the estimation than does a simple regression predicting climate. Consequently, back-calculating climate with multilinear regressions by inserting Y (transformed C_4 production) and solving for X (a climatic variable) propagates large errors. Fewer than 40% of predictions from multilinear regressions fell within a 95% prediction interval for reported conditions from NWS stations.

Temporal Variation in C_4 Fractions. To compare changes in the fraction of C_4 plants over time, we tested for differences between each archaeological site or stratum (Bonferroni procedure, α = 0.05). However, for estimates of paleoclimate, we compared conditions at each archaeological site or stratum to present conditions at the nearest NWS station (Table 2). We did this because we have a time series of bison samples only at the Agate Basin site for the period from 12,600 to 11,900 yr BP and cannot directly measure climate change during the period separating ABHG (11,900 yr BP) and HK (7,200 yr BP) archaeological sites.

Table 2. Past climatic variables for archaeological sites estimated from $\delta^{13}C$ values in bison bone, associated dietary fractions of C_4 plants, and correlations of C_4 fractions with weather variables

*NWS Station/			Annual precipitation		June precipitation		August minimum temperature	
Archaeological Site	yr BP	MNI	mm	*P*	mm	*P*	°C	*P*
*Gillette (2 east)	Present		391.2 (95.5)		71.4 (45.7)		12.1 (1.5)	
CKM	12,590	4	371.2 (12.0)	0.09	62.2 (20.4)	0.07	**11.4 (0.1)**	<0.001
*Morrisey	Present		306.1 (65.5)		52.6 (25.9)		14.2 (1.5)	
ABF	12,590	9	**378.9 (20.3)**	<0.001	**63.5 (3.5)**	0.02	**11.5 (0.2)**	<0.001
ABAB	12,390	14	**363.6 (10.7)**	<0.001	60.9 (1.8)	0.05	**11.3 (0.1)**	<0.001
ABHG	11,910	11	**397.6 (26.0)**	<0.001	**66.7 (4.4)**	<0.01	**11.7 (0.3)**	<0.001
*Upton	Present		367.3 (77.7)		65.3 (37.9)		11.5 (1.6)	
HK	7,280	12	**415.5 (34.0)**	<0.01	69.8 (5.8)	0.26	11.9 (0.4)	0.09

Values for present weather are from NWS reporting stations of closest geographic proximity (*) (numbers of reporting years ranged from 64 to 69 at Gillette, 20 to 33 at Morrisey, and 39 to 42 at Upton). For archaeological sites, we calculated the standard deviation of dietary C_4 from standard deviations in $\delta^{13}C$ values of each population of bison sampled. The standard error about the regression ($s_{y.x}$) is ±37.35 mm for normal annual precipitation, ±8.25 mm for normal June average precipitation, and ±0.86°C for normal August minimum temperature. Past values in bold are significantly different from current weather at the nearest NWS station (two-tailed t test). Standard deviations are in parentheses. MNI, minimum number of individual bison sampled.

Thus, we interpret the relative climatic conditions at different sites in the past by comparing them to the present.

Results

As others have found (7), our isotope data suggest that during the Younger Dryas (≈12,900–11,600 yr BP), the High Plains received more effective moisture than present-day NWS stations (Table 2). All three cultural periods at Agate Basin experienced cooler August minimum temperatures (+3°C) and were wetter than reported currently (+74 mm normal annual precipitation, +11 mm normal June average precipitation). Although the CKM site (12,600 yr BP) was not wetter than at present, it had cooler August minimum temperatures (≈1°C). The similarity between CKM, ABF, and ABAB (Table 2) indicates that similar plant communities surrounded the CKM and Agate Basin sites during these periods (≈12,600–12,400 yr BP).

However, around the close of the Younger Dryas, the proportion of C_4 plants increased significantly, from 20 to 30%, between the ABAB (12,400 ± 960 yr BP) and ABHG cultural periods (11,900 ± 706 yr BP) at the Agate Basin site (Table 1). Compared with ABAB, $\delta^{13}C$ values in ABHG bison bones show increased June (+6 mm) and annual precipitation (+34 mm) as well as slightly increased August minimum temperatures (+0.4°C). Although the ABHG cultural period was warmer than the preceding ABAB, it remained cooler (−2.5°C August minimum temperature) and wetter (+14 mm June and +90 mm annual precipitation) than present-day conditions (Table 2).

Our data show that, several thousand years later, the High Plains experienced conditions that were again wetter but had become as warm as or slightly warmer than the modern climate (Table 2). $\delta^{13}C$ values from HK (7,300 yr BP) indicate slightly warmer August minimum temperature (+0.5°C), similar June precipitation, and 50 mm more annual precipitation than at present (Table 2). Given that Pacific westerly winds carry little moisture to the High Plains, greater annual precipitation suggests that the season of monsoonal inputs from the Gulf of Mexico started earlier in the spring and lasted later in the fall compared with present-day conditions.

Our stable isotope data indicate rapid warming on the High Plains from 12,400 to 11,900 yr BP, which is consistent with the apparent split into foothills–mountains vs. open plains cultures in the late Pleistocene. In contrast, our data from 7,300 yr BP show that conditions were wet as well as warm during the supposed Altithermal period, which conflicts with the hypothesis that dry conditions compelled humans to avoid the open plains. To provide further insight into climatic variations in this region, we synthesized information about the mechanisms and patterns of oceanic and atmospheric circulation that drove global climate during the late Pleistocene and mid-Holocene. On the basis of this information (see below), we propose that indirect forcing of monsoonal inputs by factors affecting deep water formation in the North Atlantic Ocean determined both precipitation regimes and human-occupancy patterns on the High Plains.

Discussion

Monsoonal inputs to midcontinental North America occur when warming over the land draws moist air from the Gulf of Mexico over the continent, displacing dry westerly winds in summer. The magnitude of this advection (and the northern border of summer monsoons) depends on thermal differences between terrestrial regions of hot rising air vs. the dense humid high-pressure air mass over the Gulf (the Bermuda High) (7, 10). The strength of the Bermuda High is determined by warm transequatorial ocean currents flowing northward into the Gulf of Mexico and Atlantic Ocean (33). These northward surface currents run counter to deep southward flows in the Atlantic (the "conveyor belt"). The conveyor belt begins in the North Atlantic with the submergence of cold saline North Atlantic Deep Water (NADW) (34).

At least twice during the Late Pleistocene and Holocene, disruption of NADW production coincided with fresh-water input to the north Atlantic from either separation and melting of ice shelves (ice-rafting events) (35) or from influx of glacial meltwater (8) (Fig. 3; arrows along the *x* axis indicate these events). These episodes occurred at ≈12,900 yr BP (ice-rafting at the onset of the Younger Dryas episode) and at ≈8,200 yr BP (catastrophic draining of the Laurentide Lakes) (8, 35). In turn, these disruptions of NADW production caused loss of warm surface inflows into the Gulf of Mexico (countercurrents to the conveyor belt). This shift decreased evaporation and weakened the Bermuda High, effectively halting monsoonal rains to the midcontinent (35).

These episodes had important impacts on the climate and culture of the High Plains. According to this model (Fig. 3), loss of NADW production at ≈12,900 yr BP initiated the Younger Dryas episode, returning the High Plains to cool wet glacial conditions reminiscent of conditions during the Last Glacial Maximum. After the Younger Dryas ended (≈11,600 yr BP), Earth's orbital eccentricities shifted the timing of the perihelion to summer, increased the axial tilt from 23.5° to 24.5°, and should have caused warmer summers and cooler winters than at present (Fig. 3; maximal solar insolation in summer by ≈9,000 yr BP) (10). However, continued atmospheric circulation (the glacial anticyclone) over the Laurentide Lakes and Laurentide Ice Sheet maintained the southerly position of the polar front and introduced moisture from the Laurentide Lakes southward (7, 9). This situation created cool wet conditions on the northern portions of the Great Plains, the High Plains, and the eastern slopes of the adjacent Rocky Mountains (40, 41, 46–48). In contrast, regions dominated by dry Pacific westerly winds, such as northwest Wyoming and Idaho (40, 42), apparently received little moisture from either the glacial anticyclone or Gulf monsoons. West of the continental divide in southwestern Colorado, lack of glacial influence coincident with increasing insolation prompted higher treelines and monsoonal inputs from the Pacific (7, 49–51). East of the continental divide, the glacial anticyclone maintained cooler conditions. Despite cool temperatures, which decreased the thermal gradient that drove transport of water vapor, monsoonal precipitation from the Gulf of Mexico apparently reached northward to at least 35° N (Fig. 3) (7, 52).

At ≈8,200 yr BP, catastrophic draining of the Laurentide Lakes through the Hudson Strait into the North Atlantic probably caused the next disruption of NADW production and concomitant global cooling (Fig. 3) (8, 35). Initially, the High Plains became cold, dry, and windy (33, 39), first because of loss of northern inputs of moisture from the weakened glacial anticyclone, and second because of northward retreat of the polar front, which allowed northward expansion of westerly winds (10). Sediment cores (43) and oxygen isotope values (^{18}O) from ice cores (36, 37, 39) [Fig. 3: (^{18}O) trace from the Greenland Ice Sheet Project Two GISP2] at 8,200 yr BP suggest that cooling effects lasted <200 yr. We propose that continental warming outpaced the reestablishment of NADW circulation. With declining influence of the last remnants of the Laurentide Ice Sheet, the polar front shifted northward. Because of high insolation (10), terrestrial summer temperatures not only rebounded but increased to an estimated 2.1°C warmer than present (7). That is, midcontinental temperatures rebounded rapidly, but a strong Bermuda High and accompanying summer monsoonal rains, which depend on a functioning conveyor-belt countercurrent, did not. Subsequently, after ≈8,000 yr BP, increased July insolation, strong dry westerly winds from the Pacific (7), and lack of precipitation from either the north or south created intensely dry, hot, and windy conditions across the midcontinent—the "Altithermal" drought (Fig. 3; vertical bar at 8,000 yr BP).

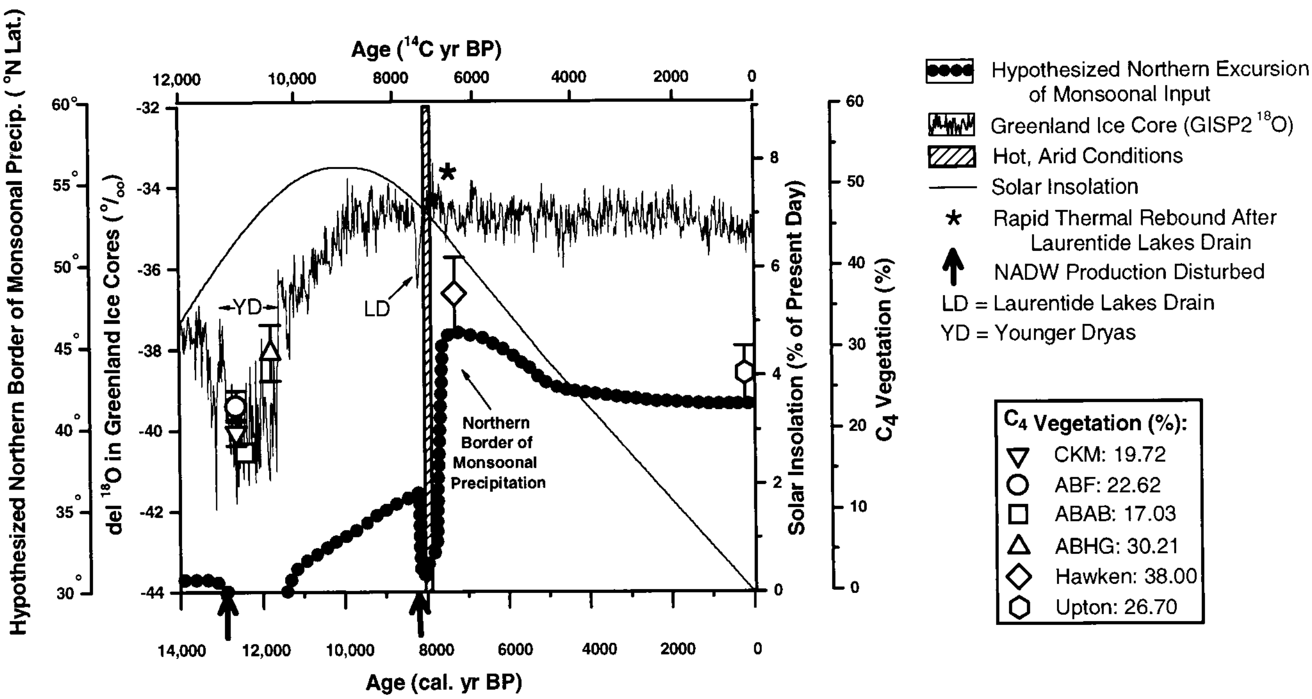

Fig. 3. Interplay of climate change mechanisms and hypothesized northern excursion of monsoonal input from the Gulf of Mexico. Summer insolation is shown as percent difference from present-day radiation for the northern hemisphere because of Earth's axial tilt and summer perihelion (10). The hypothesized northern extension of monsoons is based on reef-drowning events (38), ice cores (36, 37, 39), deep sea cores (8, 35), sediment cores (40–44), climate models (7, 33, 45), and bone $\delta^{13}C$ in this paper.

Abundant evidence supports this interpretation of windy, hot, dry conditions across midcontinent North America around 8,000 yr BP [Fig. 3; extreme GISP2 temperature rebound (*) after the Laurentide Lakes drain]. Large (orders of magnitude) spikes of particulates in Minnesota (43) and Illinois (53) at ≈8,200 yr BP suggest arid conditions and strong westerly winds in the Midwest. Pollen samples from Colorado (49, 50, 52) indicate hot arid conditions at ≈8,000 yr BP. In Wyoming, powerful westerly winds sharply increased eolian deposition on dunes in central Wyoming from ≈8,600 to 7,300 yr BP (54). Dunal accretion does not always suggest arid conditions, but in the eastern mountains of Wyoming, forests of mixed *Pinus* and *Picea* abruptly shifted entirely to more drought-tolerant *Pinus* from ≈8,900 to 8,100 yr BP (46), and in central Wyoming, dry conditions starting at ≈8,200 yr BP no longer supported dense stands of *Artemisia* (47).

Sometime after 7,800 yr BP, we propose that full restoration of NADW circulation (34, 35) prompted northward expansion of the monsoonal boundary and brought wet as well as warm conditions to the High Plains (Fig. 3; latitudinal variation in monsoons). In the summer, higher midcontinent temperatures continued (2.1°C; see above) (7) with estimates of 2.5°C greater than the present at 6,000 yr BP (45). Warm terrestrial temperatures caused strong advection of precipitation from the Bermuda High that propelled monsoonal rains northward beyond their present-day border at ≈42° N to ≈45° N (Fig. 3; northern border of monsoonal precipitation after 7,800 yr BP) (42, 44). These wet warm conditions are indicated by pollen cores in southwestern Colorado and the central Colorado Rocky Mountains (48, 49, 55). Sedimentation rates, pollen, and ostracode distributions in an eastern Wyoming playa also suggest wetter conditions from 10,000 to 5,000 yr BP than after (41). All these findings support our isotope data from HK that indicate wet warm conditions at 7,300 yr BP (Fig. 3; C_4 vegetation for HK at 7,300 yr BP).

Conclusions

Accompanying climate change, Amerindian cultures on the High Plains changed profoundly during the Holocene. Both our isotope data and our integrated model of global circulation show that separation of Amerindian cultures into foothills–mountains vs. open plains regions coincided with the close of the Younger Dryas (≈11,600 yr BP). However, our isotopic data at 7,300 yr BP indicate mild summer conditions, which contrasts with the relative and absolute decline in archaeological sites on the High Plains (2, 6). $\delta^{13}C$ values at HK might reflect mesic conditions specific to the 20,000-km (2) Black Hills region. Nonetheless, given the mild conditions in this region of the High Plains, we would expect to find repeated communal bison kills. However, HK is the only known instance of a bison kill from ≈8,000 to 5,500 yr BP in either Montana or Wyoming (4), suggesting the influence of other factors on human habitation, such as strong climate fluctuations.

Intense desiccation during an abbreviated Altithermal from ≈8,000 to 7,800 yr BP, followed by dry cold winters and warm wet summers, altered plant communities across the region. On the High Plains, C_4 grasses became dominant and/or sagebrush (*Artemisia*) invaded grasslands (40, 42). During the Altithermal, herbivores apparently sought refuge in the western mountains or in the low-altitude grasslands of the eastern Great Plains where bison hunting continued uninterrupted (3). After a short Altithermal, monsoonal influx northward and ensuing warm wet summers should have prompted rapid rebound of grasslands on the High Plains. However, these same conditions caused eastern grasslands to expand into previously forested areas, thus providing good forage (56) and delaying migration of bison onto the High Plains. We propose that because of abundant resources on the eastern Great Plains after the brief Altithermal, herbivore reinvasion lagged behind grassland rebound on the High Plains. Consequently, large game animals and hunting parties pursuing them ventured infrequently onto the High Plains between 8,000 and 5,000 yr BP. After ≈5,000 yr BP, declining insolation, cooler summers, and warmer winters established conditions similar to the present (7), so that year-round use of the High Plains by both large game animals and hunter–gatherers increased (6).

By combining $\delta^{13}C$ values in bison bone with relations between climate and C_4 plant abundance and integrating information about regional qualitative evidence with global mechanisms of past climate, we have clarified climate changes and demographic responses by hunter–gatherers during the Holocene. Our analysis indicates that disruption of NADW production, which interrupted summer monsoonal input, radically influenced plant communities, hunting opportunities, and prehistoric patterns of human habitation on the High Plains of North America.

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