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Holocene climate variability

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Abstract

Although the dramatic climate disruptions of the last glacial period have received considerable attention, relatively little has been directed toward climate variability in the Holocene (11,500 cal yr B.P. to the present). Examination of ~50 globally distributed paleoclimate records reveals as many as six periods of significant rapid climate change during the time periods 9000–8000, 6000–5000, 4200–3800, 3500–2500, 1200–1000, and 600–150 cal yr B.P. Most of the climate change events in these globally distributed records are characterized by polar cooling, tropical aridity, and major atmospheric circulation changes, although in the most recent interval (600–150 cal yr B.P.), polar cooling was accompanied by increased moisture in some parts of the tropics. Several intervals coincide with major disruptions of civilization, illustrating the human significance of Holocene climate variability.

Keywords: Climate; Rapid climate change; Holocene; Solar variability

Introduction

Although the climate of the Holocene (11,500 cal yr B.P. to the present) has sustained the growth and development of modern society, there is surprisingly little systematic knowledge about climate variability during this period. Many paleoclimate studies over the last decade have highlighted the extreme climate fluctuations of the last glacial interval. If we are to understand the background of natural variability underlying anthropogenic climate change, however, it is important to concentrate on climate of the more recent past. To seek a more comprehensive view of natural climate variability during the present Holocene interglacial. We present in this paper a selection of globally distributed high-resolution climate proxy records. Examination of these records demonstrates that, although generally weaker in amplitude than the dramatic shifts of the last glacial cycle, Holocene climate variations have been larger and more

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frequent than is commonly recognized. Comparison of paleoclimate records with climate forcing time series suggests that changes in insolation related both to Earth's orbital variations and to solar variability played a central role in the global scale changes in climate of the last 11,500 cal yr.

The timing of Holocene climate change events at intervals of approximately 2800–2000 and 1500 yr is well established in the literature (Allen and Anderson, 1993; Bond et al., 1997, 1999, 2001; Bray, 1971, 1972; Dansgaard et al., 1971; Denton and Karlén, 1973; Johnsen et al., 1972; Mayewski et al., 1997; Naidu and Malmgren, 1996; Noren, 2002; O'Brien et al., 1995; Pisias et al., 1973; Sonett and Finney, 1990; Stager et al., 1997; Stuiver and Braziunas, 1989,1993). At least in the northern North Atlantic region, Holocene climate change events recorded in different paleoclimate archives have been demonstrated to be correlated in time, based on the comparison of glacier fluctuation records (Denton and Karlén, 1973), ice core records (O'Brien et al., 1995), and marine sediments (Bond et al., 1997, 1999). This correlation has also been extended to marine sediment records off West Africa (deMenocal et al., 2000a,b).

As a framework for the examination of Holocene climate variability, we utilize the results of Denton and Karlén (1973) that show globally distributed changes in glacier extent. We choose this record as our basis for investigation because it contains geographically broad evidence for change in Holocene climate. Glacier extent is directly related to changes in climate, as indicated by the modern example of widespread glacier retreat coincident with warming over the last century. We recognize that much work has been done since the Denton and Karlén paper, but we are unaware of any work that substantially challenges it. Indeed, a great number of researchers presenting Holocene climate records since 1973 have placed their new data in the context of this pioneering work, as we do here. Validation that Holocene climate variability reflected in the Denton and Karlén (1973) study is of significant enough magnitude and frequency to be identified in a globally distributed array of paleoclimate proxies (e.g., temperature, atmospheric circulation, and moisture balance), however, remains to be demonstrated. It is this demonstration that forms the focus for our paper. Through the multiparameter paleoclimate proxy records assembled for this study, we make the case that Holocene climate has not been stable, but rather that it has been dynamic at scales significant to humans and ecosystems.

Methods

Paleoclimate records used in our study were selected on the basis of length (preference given to full Holocene coverage), sampling resolution (highly resolved), dating quality (uncertainty <500 yr), published interpretation (records that specify a climate variable assigned to proxy data), and geographic distribution (diverse regions). The records are grouped into three regions: Northern Hemisphere (mid- to high-latitudes), low latitudes, and Southern Hemisphere (mid- to high-latitudes) in Figures 1, 2, and 3, respectively. Figure 4 shows globally distributed glacier fluctuation records and climate forcing time series (cosmogenic isotopes reflecting solar variability, orbital insolation changes, volcanic aerosols, and greenhouse gases). Not every record that suits the foregoing requirements is included, but this selection represents a substantial first approximation that can serve as a framework for the inclusion of additional records. Records with annual- to decadal-scale resolution were smoothed with a 200-yr Gaussian filter to facilitate comparison with more coarsely sampled records.

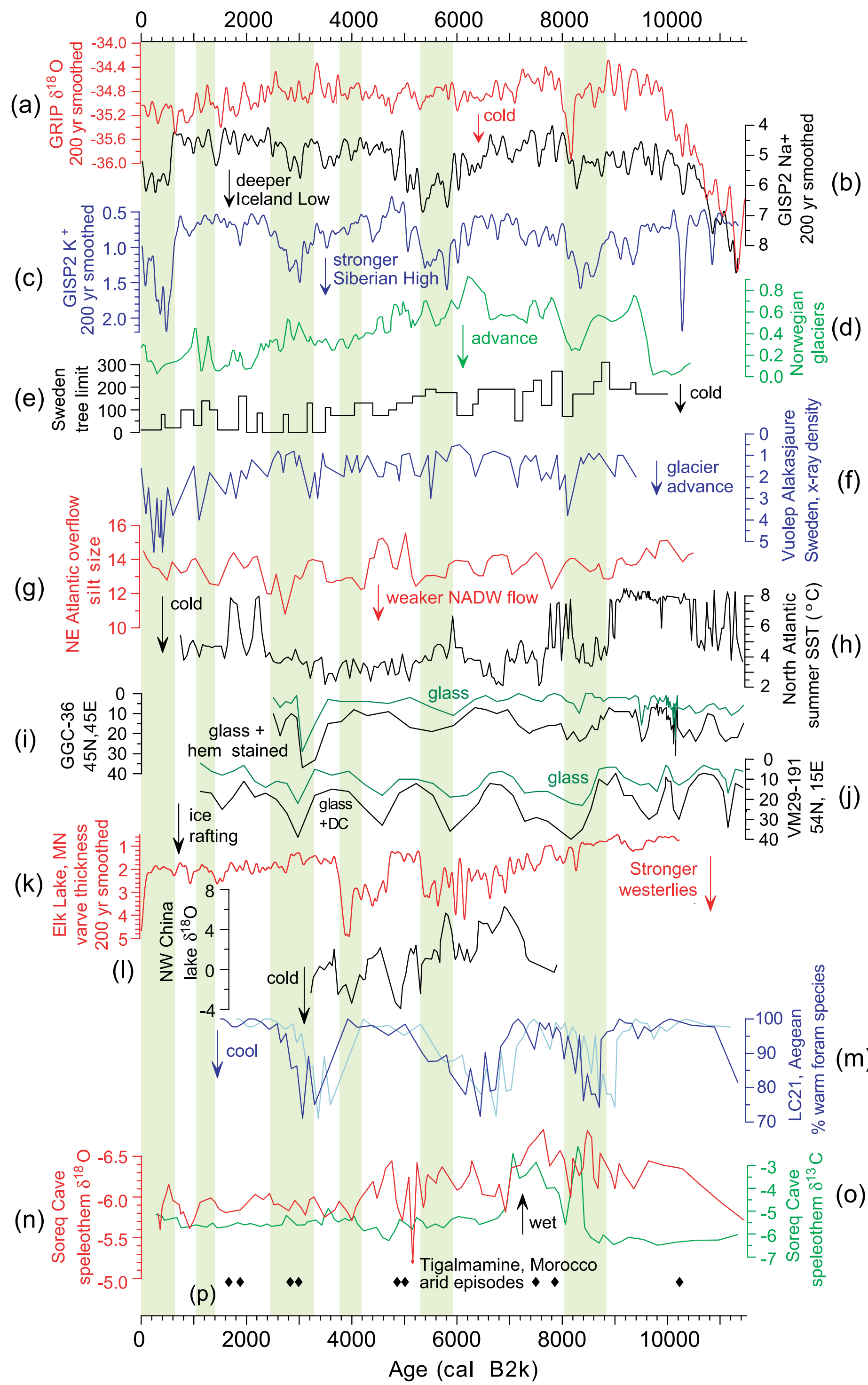
Results

Major periods of Holocene rapid climate change (RCC)

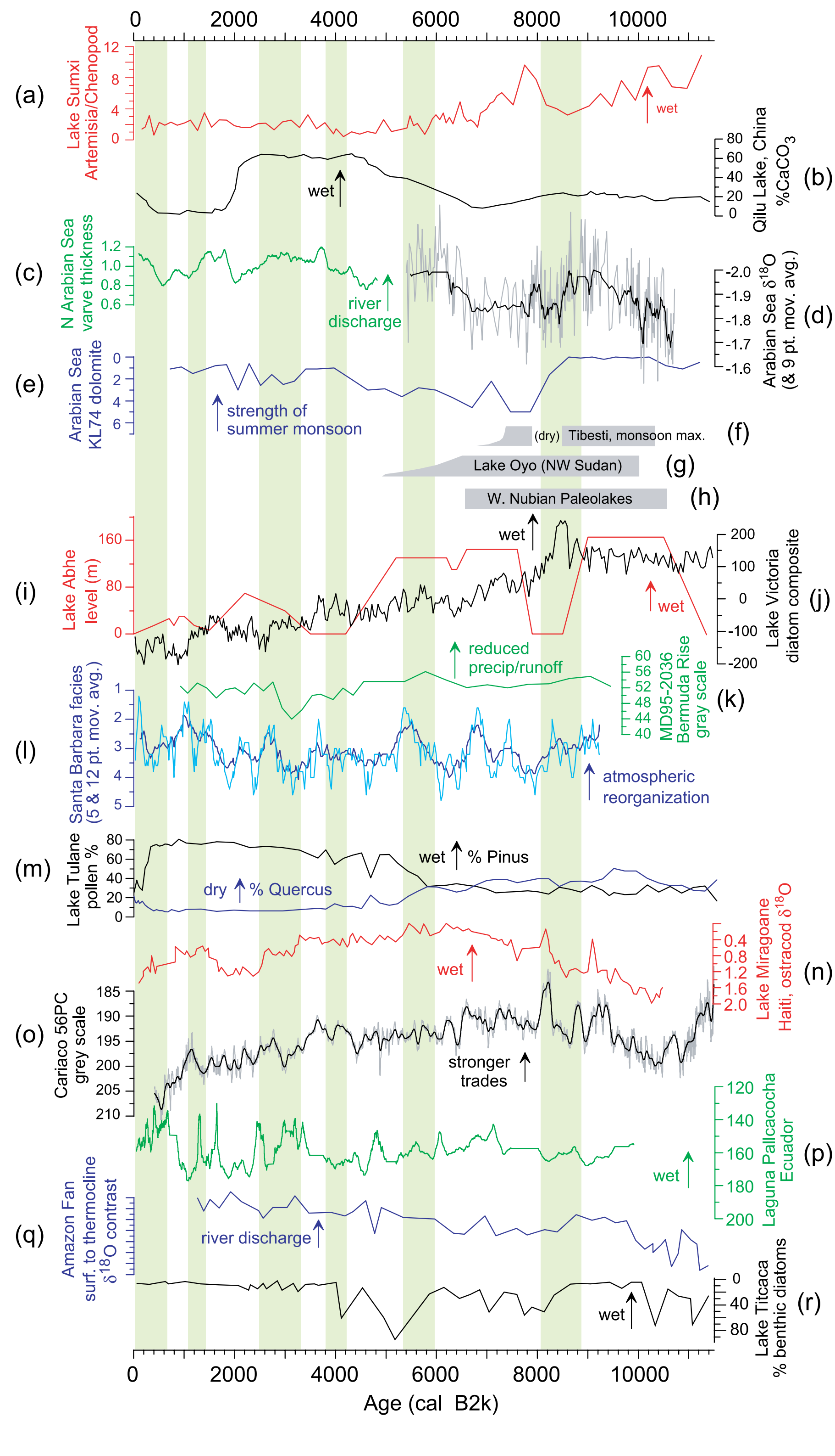
We use the term rapid climate change (RCC) for the intervals of climate change observed in the Denton and Karlén (1973) record, rather than more geographically or temporally restrictive terminology such as "Little Ice Age" and "Medieval Warm Period." We do not mean to imply with this terminology that these changes are comparable in

Figure 1. Northern Hemisphere paleoclimate series, arranged generally by latitude (north, top), with state of climate proxy noted. Green bands represent timing of RCC, tuned to high-resolution GISP2 record. (a) Gaussian smoothed (200 yr) GRIP $\delta^{18}O$ (‰) proxy for temperature (Johnsen et al., 1992). (b) Gaussian smoothed (200 yr) GISP2 sodium (Na^+; parts per billion, ppb) ion proxy for the Icelandic Low (Mayewski et al., 1997; Meeker and Mayewski, 2002). (c) Gaussian smoothed (200 yr) GISP2 potassium (K^+; ppb) ion proxy for the Siberian High (Mayewski et al., 1997; Meeker and Mayewski, 2002). (d) Norwegian glacier advance record (units) (Nesje et al., 2001). (e) Treeline limit shifts in Sweden (units relative to the present) (Karlén and Kuylenstierna, 1996). (f) X-ray density measurements [relative scale of increasing density (i.e., increased silt influx, downward)] for sediments in Lake Vuolep Alakasjaure, northern Sweden (Karlén and Larsson, in review). (g) Northeast Atlantic overflow recorded in silt-sized particles (10–64 m) for NEAP-15K with Gaussian interpolation using a 300-yr window (Bianchi and McCave, 1999). (h) Summer sea surface temperatures (°C) for the North Atlantic (Irminger Sea) from a planktonic foraminiferal modern analogue function (this study). (i) Abundances of volcanic glass particles and hematite-stained grains in sediment core GGC-36 from 45°N, 45°W (Bond et al., 1997, 1999). (j) Abundances of volcanic glass particles and hematite stained grains in sediment core VM29-191 from 54°N, 15°W (Bond et al., 1997). (k) Gaussian smoothed (200 yr) varve thickness (mm) record from Elk Lake (Minnesota, USA) (Bradbury et al., 1993). (l) Isotopic temperature (°C) reconstruction based on $\delta^{18}O$ (‰) of lacustrine carbonates, lake section from Hongshui River, northwest China (Zang et al., 2000). (m) Relative abundance (%) of Aegean core LC21 planktonic foraminiferal species with warm-water affinities (Rohling et al., 2002). Light line represents original calibrated AMS ^{14}C chronology, and heavy line indicates maximum (three to four centuries) correction required to match the Minoan eruption of Santorini to its actual age. (n) $\delta^{18}O$ (‰) for speleothem in Soreq Cave, Israel (Bar-Matthews et al., 1999). (o) $\delta^{13}C$ record (‰) for speleothem in Soreq Cave, Israel (Bar-Matthews et al., 1999). (p) Arid episodes identified in Moroccan Lake Tigalmammine (van Campo and Gasse, 1993).

magnitude or rapidity to the abrupt climate changes of the last glacial period. Nevertheless, as we will show, many of these changes are sufficiently fast from the point of view of human civilization (i.e., a few hundred years and shorter) that they may be considered "rapid." To verify the age brackets for these RCCs, we utilize the well-dated, high-resolution Greenland Ice Sheet Project Two (GISP2) chemistry series (Mayewski et al., 1997) previously correlated to the globally distributed glacier fluctuation record by O'Brien et al. (1995). We do not assume that the glacier fluctuation record or the GISP2 chemistry series capture every possible RCC in the Holocene. We do suggest, however, that our approach provides a useful framework in which the character of Holocene climate variability can be



assessed. Utilizing the annual layer dating of the GISP2 record, RCCs in the Denton and Karlén (1973) glacier fluctuation record can be identified at 9000–8000, 6000–5000, 4200–3800, 3500–2500, 1200–1000, and since 600 cal yr B.P. (green shading in Figures 1–4). The global distribution and proxy climate interpretations for these anomalies appear in Figure 5. Differences in climate from region to region and differences in the sensitivity of the climate proxies from record to record preclude the likelihood that every RCC event would be captured or necessarily should be present in every record. We assert that a globally distributed signature for these RCCs is sufficient to



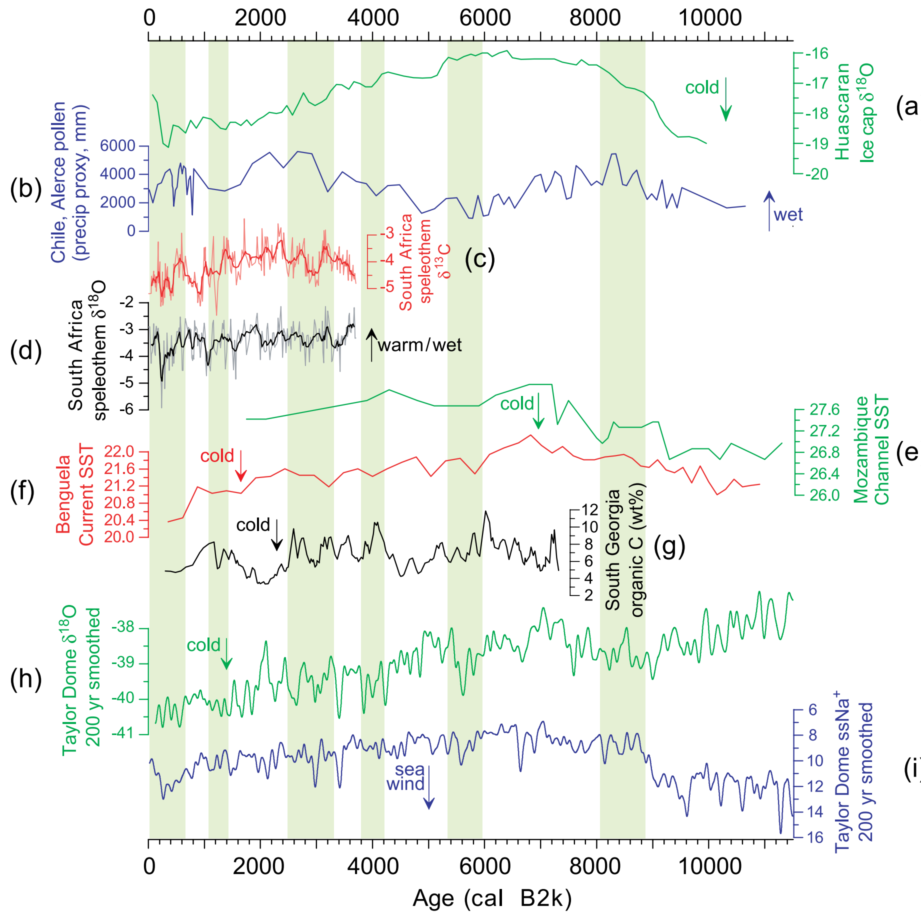


Figure 3. Southern Hemisphere paleoclimate series, arranged generally by latitude (north, top), with state of climate proxy noted. Green bands represent timing of RCC, tuned to high-resolution GISP2 record. (a) δ^{18}O record (‰) for Huascaran ice-cap, Peru (Thompson et al., 1995). (b) Pollen-ratio based reconstruction of precipitation (mm) for Lake Alerce, Chile (Heusser and Streeter, 1980). (c) δ^{13}C record (‰) for speleothem in Cold Air Cave, South Africa (Lee-Thorp et al., 2001). (d) δ^{18}O record (‰) for speleothem in Cold Air Cave, S Africa (Lee-Thorp et al., 2001). (e) Alkenone-based SST record (°C) for core from the Mozambique Channel (MD79257) (Bard et al., 1997). (f) Alkenone-base sea surface temperature record (°C) for core from the Benguela Current (Kim et al., 2002). (g) Organic carbon (%) in a core from Block Lake South Georgia (Rosqvist and Schuber, in press). (h) Gaussian smoothed (200 yr) δ^{18}O record (‰) for Taylor Dome, Antarctica (Steig et al., 2000). Taylor Dome Holocene time scale (Monnin et al., in press). (i) Gaussian smoothed (200 yr) sea-salt Na^+ (ppb) record for Taylor Dome, Antarctica (Mayewski et al., 1996). Taylor Dome Holocene time scale (Monnin et al., in press).

demonstrate that they are of worldwide significance. In the following, we present descriptions of climate change during each of the six RCCs directly developed from information available in Figures 1–5. References for these descriptions appear in the figure captions. Information not apparent from Figures 1–5 is separately referenced in the text of the paper.

Figure 2. Low-latitude paleoclimate series with state of climate proxy noted. Green bands represent timing of RCC, tuned to high-resolution GISP2 record. (a) Artemisia/Chenopodiacea pollen abundance ratio for a core from Lake Sumxi, Tibet (Hodell et al., 1999; van Campo and Gasse, 1993). (b) $CaCO_3$ percentages in a core from Qilu Lake, southeast China (Hodell et al., 1999). (c) Average (200 yr) of varve thickness record (mm) in a core from the Makran margin, north Arabian Sea recording discharge from western Pakistan (Lückge et al., 1999; von Rad et al., 1999). (d) Planktonic foraminiferal δ^{18}O record (‰) for Arabian Sea core 63KA (light line) with 9-pt moving average (heavy line) indicative of Indus River discharge (this study). (e) Dolomite abundance (%) in core KL74 from the Arabian Sea (Sirocko et al., 1993). (f) Humid phases (African monsoon maximum) in the central Saharan Tibesti Mountains, with dry interruption (Maley, 1982). (g) Presence of a lake in the presently hyper-arid Oyo depression, northwest Sudan. Tapered end indicates desiccation phase (Ritchie et al., 1985). (h) Presence of West Nubian paleolakes, indicating humid conditions in an area that today is extremely arid (Hoelzmann et al., 2000). Gray shade indicates uncertainty in date of final desiccation. (i) Lake levels (m, relative to the present) in Lake Abhe, Ethiopia (Gasse, 1977). (j) P:E or lake level proxy based on calibration of a diatom ratio. The time period from 11,500 to 1000 cal yr B.P. is based on diatom series composite series and 1000 cal yr B.P. to the present is based on new littoral diatom series from Pilkington Bay core P2K-1, Lake Victoria (this paper). (k) Gray-scale record for core MD95-2036 from Bermuda Rise (Keigwin and Boyle, 1999). (l) Descriptive facies classification (integrated magnetic susceptibility, physical properties, sediment color, and other data for Santa Barbara Basin ODP Site 893 (Behl and Kennet, 1996; Kennett and Ingram, 1995) suggestive of disruptions in North Pacific atmospheric circulation that affect ocean circulation. (m) Relative pollen abundances (%) for pinus (pine) and quercus (oak) from Lake Tulane, Florida (Grimm et al., 1993). (n) Ostracod-based δ^{18}O record (‰) from Lake Miragoane, Haiti (Hodell et al., 1991). (o) Gray-scale record (light line) for core 56PC from Cariaco Basin (Hughen et al., 1996) along with 200-yr Gaussian smoothing (heavy line). (p) A lake sediment record from Laguna Pallcacocha, Ecuador (gray scale) (Rodbell et al., 1999). (q) Contrast between δ^{18}O values (‰) for surface and thermocline dwelling planktonic foraminifera in core from the Amazon Fan (Maslin and Burns, 2000). (r) Sediment core from Lake Titicaca, Bolivia, and Peru (%benthics) (Baker et al., 2001).

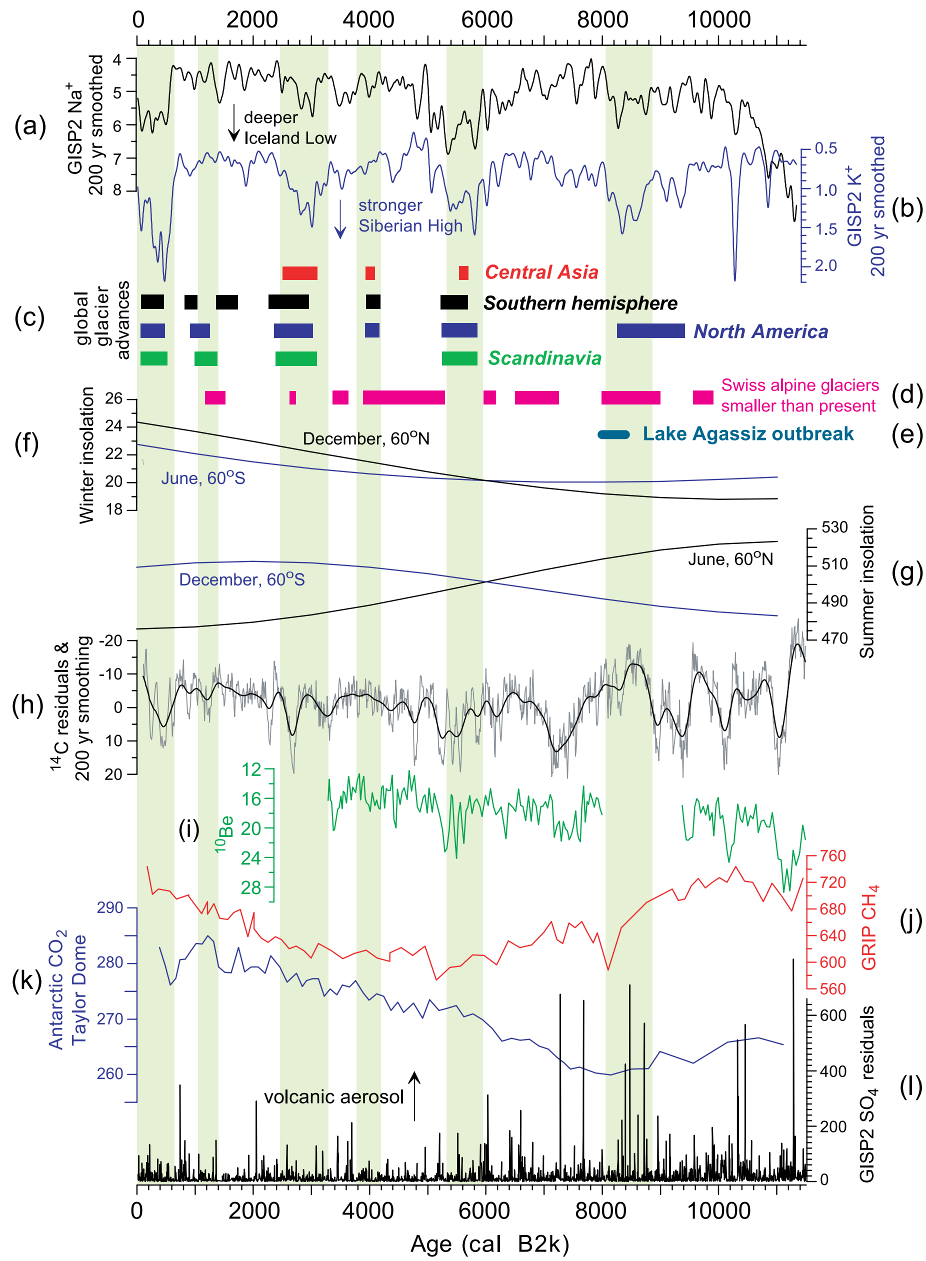


Figure 4. Climate forcing series and globally distributed discontinuous glacier advance records plus GISP2 proxy for atmospheric circulation, included as a continuous record example. Green bands represent timing of RCC tuned to high-resolution GISP2 record. (a) Gaussian smoothed (200 yr) GISP2 Na^+ (ppb) ion proxy for the Icelandic Low (Mayewski et al., 1997; Meeker and Mayewski, 2002). (b) Gaussian smoothed (200 yr) GISP2 K^+ (ppb) ion proxy for the Siberian High (Mayewski et al., 1997; Meeker and Mayewski, 2002). (c) Episodes of distinct glacier advances: European, North American, and Southern Hemisphere (Denton and Karlén, 1973), and central Asia (Haug et al., 2001). (d) Episodes during which Swiss alpine glaciers were smaller than today, derived from dating of emerging tree-stumps (Hormes et al., 2001). (e) Timing of the Holocene outburst of the North American meltwater from Lake Agassiz (Barber et al., 1999). (f) Winter insolation values (W m^{-2}) at 60°N (black curve) and 60°S latitude (blue curve) (Berger and Loutre, 1991). (g) Summer insolation values (W m^{-2}) at 60°N (black curve) and 60°S latitude (blue curve) (Berger and Loutre, 1991). (h) $\Delta^{14}C$ residuals (Stuiver et al., 1998): raw data (light line) and with 200-yr Gaussian smoothing (bold line). (i) ^{10}Be concentrations in the GISP2 ice core (10^3 atoms g^{-1}) (Finkel and Nishizumi, 1997). (j) Atmospheric CH_4 (ppbv) concentrations in the GRIP ice core, Greenland (Chappellaz et al., 1993). (k) Atmospheric CO_2 (ppmv) concentrations in the Taylor Dome, Antarctica, ice core (Indermühle et al., 1999). (l) SO_4^{2-} residuals (ppb) from the GISP2 ice core, Greenland (Zielinski et al., 1996).

"*Glacial Aftermath*" RCC (9000–8000 cal yr B.P.)

The widespread, severe climatic disruption from 9000 to 8000 cal yr B.P. is unique among the Holocene RCC intervals because it occurs at a time when large Northern Hemisphere ice sheets were still present. In the North Atlantic (Fig. 1), there is a significant short-lived cooling called the "8200 yr" event (Alley et al., 1997). It also appears to have been generally cool over much of the Northern Hemisphere throughout this interval, as evidenced by major ice rafting, strengthened atmospheric circulation over the North Atlantic and Siberia, and more frequent polar north-

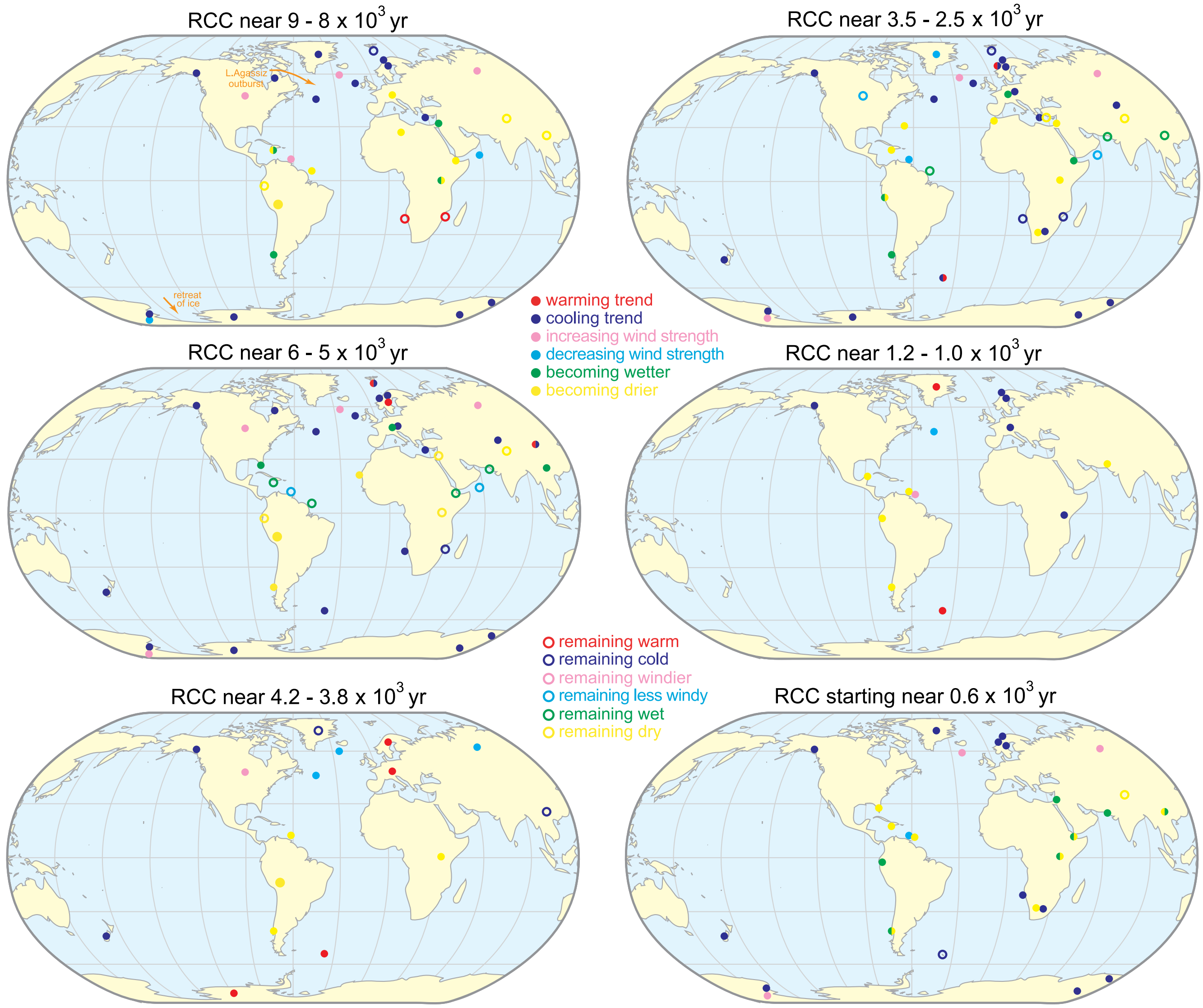


Figure 5. Map displaying state of climate proxies during RCCs near 9000–8000, 6000–5000, 4200–3800, 3500–2500, 1200–1000, and since 600 cal yr B.P.

westerly (winter) outbreaks over the Aegean Sea. Mountain glacier advances occur in northwestern North America and Scandinavia, and treeline limit is lower in Sweden. Glacier retreat occurs in the European Alps, perhaps reflecting the influence of dry northerly winds.

At low latitudes (Fig. 2), this is a period of widespread aridity that occurs midway through a prolonged humid period that began in the early Holocene (deMenocal et al., 2000a). Additionally, this time period is followed by a change to more seasonal and torrential rainfall regimes throughout tropical Africa (Gasse, 2000; Kendall, 1969; Maley, 1982; Nicholson and Flohn, 1980). Summer monsoons over the Arabian Sea and tropical Africa weaken dramatically during this RCC, and trade wind strength and/or rainfall fluctuates dramatically over the Caribbean. Widespread, persistent drought occurs in Haiti, the Amazon basin, Pakistan, and Africa. Lake Titicaca levels decline through this period. Precipitation increases in the Middle East (Fig. 1).

In the Southern Hemisphere (Fig. 3), polar atmospheric circulation over East Antarctica is weak, snow accumulation rates in this region decrease (Steig et al., 2000), and the direction of temperature change is different in different areas of East and West Antarctica are variable (Ciais et al., 1994; Masson et al., 2000). Grounded ice in the Ross Sea retreats (Conway et al., 1999), continuing a trend that began earlier in the Holocene. This is paralleled by sea surface temperature (SST) warming on both the eastern and western flanks of southern Africa. Precipitation generally increases in Chile, most likely due to the intensification of southern mid-latitude westerlies.

Classic "cool poles, dry tropics" RCCs

The RCCs following the 9000–8000 cal yr B.P. interval varied in their intensity and geographic extent, but most generally involved the co-occurrence of high-latitude cooling and low-latitude aridity. This cool poles, dry-tropics pattern is typical of long-term climate trends during the Pleistocene (deMenocal et al., 2000a; Gasse, 2000; Kendall, 1969; Maley, 1982; Nicholson and Flohn, 1980). The most extensive of these RCCs occurred from 6000 to 5000 and from 3500 to 2500 cal yr B.P., and a less widespread RCC also occurred from 4200 to 3800 cal yr B.P. and from 1200 to 1000 cal yr B.P.

In the Northern Hemisphere, the 6000–5000 and 3500–2500 cal yr B.P. RCC intervals feature North Atlantic ice-rafting events (Bond et al., 1997), alpine glacier advances (Denton and Karlén, 1973), and strengthened westerlies over the North Atlantic and Siberia (Meeker and Mayewski, 2002). In Scandinavia, the treeline limit rises in elevation and mountain glaciers advance in the first interval (6000–5000 cal yr B.P.), but the situation reverses in the second interval (3500–2500 cal yr B.P.). Cooling over the northeast Mediterranean is related to winter-time continental/polar air outbreaks. Westerly winds over central North America strengthen from 6000 to 5000 cal yr B.P., but are weak from 3500 to 2500 cal yr B.P.

At lower latitudes (Fig. 2), the RCC interval from 6000 to 5000 cal yr B.P. marks the end of the early to mid Holocene humid period in tropical Africa, beginning a long-term trend of increasing rainfall variability and aridification (Gasse, 2000; 2001), although some areas (e.g., Pakistan, Florida, and the Caribbean) become wetter. Rainfall decreases in northwest India (Enzel et al., 1999) and southern Tibet, and Lake Titicaca levels drop during the period 6000–5000 cal yr B.P. Rainfall in Ecuador and trade wind strength over the Cariaco Basin are relatively stable from 6000–5000 cal yr B.P. but highly erratic from 3500 to 2500 cal yr B.P. The interval 3500–2500 cal yr B.P. also includes pronounced aridity in East Africa, the Amazon Basin, Ecuador, and the Caribbean/Bermuda region (Haug et al., 2001), but Southeast Asia is wet despite a dramatic weakening of winds associated with the summer East Asian Monsoon (Zang et al., 2000).

In the Southern Hemisphere, glaciers advance in New Zealand, and polar ice core records reveal intensified atmospheric circulation and generally lowered temperatures that are superimposed on a long-term trend of increasing summer insolation. Cooling also affects South Georgia Island and SSTs off southern Africa, and eastern South Africa is generally cool. Mid-latitude Chile is drier during 6000–5000 cal yr B.P., but wetter during 3500–2500 cal yr B.P. (van Geel et al., 2000) when discontinuous lake sediment records from Antarctica suggest conditions warmer than today due to increased southern summer insolation (Ingolfsson et al., 1998).

Evidence for the RCC events at 4200–3800 and 1200–1000 cal yr B.P. appear in fewer of the records, but the apparent synchrony and wide spatial distribution of those records that do contain such evidence still suggest global-scale teleconnections as for the earlier intervals. In the Northern Hemisphere, winds over the North Atlantic and Siberia are generally weak during the 4200–3800 and 1200–1000 cal yr B.P. intervals, and temperatures fall in western North America (Scuderi, 1993) and Eurasia (Briffa et al., 1992). Other climatic disruptions, however, while generally synchronous, are highly variable in their distributions, signs, and intensities. For example, glaciers advance in western North America, but retreat in Europe from 4200–3800 cal yr B.P., and Scandinavian ice seems largely unaffected. North Atlantic Deep Water (NADW) production is weak from 4200 to 3800 cal yr B.P., but it increases over the period 1200–1000 cal yr B.P., while westerlies over North America are exceptionally strong from 4200 to 3800 cal yr B.P., but are nearly unchanged during 1200–1000 cal yr B.P.

At low latitudes, these two RCC include variable but generally dry conditions in much of tropical Africa (Gasse, 2000, 2001) and monsoonal Pakistan. Lake Titicaca levels drop, but Haiti is generally wet. In the Cariaco Basin (Haug et al., 2001), trade winds intensify. During the RCC interval 1200–1000 cal yr B.P., aridity extends to Ecuador and glaciers advance on Mt. Kenya (Karlén et al., 1999).

In the Southern Hemisphere, little change occurs in polar wind strength. Temperatures fluctuate over Taylor Dome and mid-latitude Chile is dry during both of these RCCs. Warming from 4200 to 3800 cal yr B.P. occurs at South Georgia Island and is also indicated in lake sediment records from the Antarctic Peninsula and Victoria Land (Hjort et al., 1998; Ingolfsson et al., 1998). New Zealand glaciers advance and eastern South Africa is cool and dry from 1200 to 1000 cal yr B.P.

"*Cool poles, wet tropics*" *RCC starting at ~600 cal yr B.P.*

Both polar regions are cold and windy in this RCC interval, but the low latitude aridity that prevailed during earlier intervals does not generally characterize the tropics during this most recent interval. Unfortunately, determining the nature and duration of later stages of this interval is difficult because high-resolution records for this time are relatively scarce and because several records are missing recent sections as an artifact of sampling. Moreover, interpretation is complicated by potential anthropogenic influences. As a consequence, we investigate the characteristics of this event only from 600 to 150 cal yr B.P.

In the Northern Hemisphere (Fig. 1), glaciers advance and proxy evidence for strengthened westerlies over the North Atlantic and Siberia suggest that climate changes in this interval have the fastest and strongest onset of any in the Holocene (O'Brien et al., 1995), with the possible exception of the short-lived 8200 yr B.P. event. At low latitudes, the Cariaco Basin becomes more arid (Haug et al., 2001), as do Haiti and Florida. Conversely, equatorial East Africa experiences variable but generally humid conditions (Verschuren et al., 2000) in a negative association between tropical African humidity and northern temperatures that is unusual for the late Quaternary. Increasing river discharge in Pakistan and Ecuador suggests that both Indian monsoon and El Niño-Southern Oscillation (ENSO) systems are affected.

In the Southern Hemisphere, portions of the Antarctic Peninsula are warm (Mosley-Thompson, 1996), but East Antarctica is cold (Jouzel et al., 1983; Morgan et al., 1997) in a situation similar to recent bimodal conditions in temperature on the continent (Comiso, 2000; Schneider and Steig, 2002). Winds strengthen over East Antarctica and the Amundsen Sea (Kreutz et al., 1997). South Georgia is generally cool, New Zealand glaciers advance, and precipitation in Chile is highly variable but generally high. Benguela SSTs are cool, and southern Africa has a prominent cool, dry episode.

Discussion

Possible causes of Holocene RCCs

There are numerous potential controls on climate change and varying local- to global-scale boundary conditions (e.g., changes in the hydrologic cycle, sea level, sea ice extent, forest cover) that may account for the observed climate variability in the Holocene. In the following, very basic associations are explored between the paleoclimate response records presented in this paper and several climate forcing time series (Fig. 4): volcanic aerosols, greenhouse gases CO_2 and CH_4, ^{10}Be and Δ^{14}C residual proxies for solar variability (Beer, 2000; Stuiver and Braziunas, 1989, 1993), and examples of winter and summer insolation (we use 60°N and 60°S for illustration). Through this comparison, we attempt to focus on the most likely climate controls for the RCCs.

"*Glacial Aftermath*" *RCC (9000–8000 cal yr B.P.)*

This RCC interval occurs when the Northern Hemisphere was still significantly more glaciated than today, and during the decline in summer insolation since its early Holocene maximum. The 9000–8000 cal yr B.P. interval may thus be interpreted as a partial return toward glacial conditions following an orbitally driven delay in Northern Hemisphere deglaciation. At this time, changes in ice sheet extent and mass balance would still have played a major role in climate change. At least one large pulse of glacier meltwater into the North Atlantic (Barber et al., 1999) probably enhanced production of sea ice, providing an additional positive feedback on climate cooling. This RCC interval represents the last major stage of deglacial climate affecting the Northern Hemisphere. Continued deglaciation in Antarctica during this period was a consequence of the lagged response of the ice sheet to orbitally driven changes in insolation before the Holocene (Conway et al., 1999). Because there is no clear evidence for any ^{10}Be change at this time, the pronounced depression in Δ^{14}C recorded during the first half of this RCC interval more likely reflects reduced oceanic ventilation because enhanced meltwater production may have changed thermohaline circulation in the North Atlantic (Barber et al., 1999; Clark et al., 2001).

This RCC also coincides with a period of unusually high volcanic SO_4 production in the Northern Hemisphere. Volcanic CO_2 devoid of Δ^{14}C may have contributed to the Δ^{14}C minimum noted above, but it is unlikely to have been its primary cause. Volcanic aerosols associated with eruptions during this RCC could have significantly cooled the Northern Hemisphere, perhaps also weakening Afro-Asian monsoon circulation, thus contributing to tropical aridity. Atmospheric CH_4 concentrations dropped sharply during this RCC (Blunier et al., 1995) as the extent of biogenic methane sources declined, probably in response to aridity in the low- to mid-latitudes.

"*Cool poles, dry tropics*" *RCCs (6000–5000, 4200–3800, 3500–2500, 1200–1000 cal yr B.P.)*

There is no evidence for massive freshwater releases into the North Atlantic or for significant Northern Hemisphere ice growth or decay to explain the post 9000–8000 cal yr B.P. RCCs. There are also no systematic changes in the concentrations of volcanic aerosols or atmospheric CO_2. Atmospheric methane concentrations decline after the 9000–8000 cal yr B.P. RCC, and steadily rise after ~5000 cal yr B.P., but this is probably the result rather than the cause of roughly synchronous changes in the global hydrological cycle. Solar variability is a more plausible forcing. In particular, the major RCC events at 6000–5000 and 3500–2500 cal yr B.P. that coincide with maxima in the Δ^{14}C and ^{10}Be records suggest a decline in solar output at these times. It is more difficult to attribute the less widely distributed RCCs at 4200–3800 and 1200–1000 cal yr B.P. to specific forcing mechanisms. The former coincides with a maximum in ^{10}Be, but there is little change in Δ^{14}C at this time to suggest a solar association. Southward migration of the Inter-Tropical Convergence Zone (ITCZ) may explain the low latitude aridity associated with this RCC (Hodell et al., 2001) and would be consistent with the increase in strength of the westerlies over the North Atlantic and consequent glacier advance in northwestern North America. Intensified westerly flow may have resulted in more intense upwelling, hence relatively low Δ^{14}C values. There is a slight increase in atmospheric CO_2 from 1200 to 1000 cal yr B.P., and changes in solar output are linked to drought in the Yucatan at this time (Hodell et al., 1991, 2001).

"*Cool poles, wet tropics*" *RCC starting at ~600 cal yr B.P.*

This most recent RCC interval has a drop in CO_2 and a rise in CH_4, suggestive of wet conditions in the tropics. High levels of volcanic aerosols occur at early stages in the event, perhaps contributing to its onset. A distinct peak in both Δ^{14}C, ^{10}Be, and sunspot records (Beer, 2000; Stuiver and Braziunas, 1989, 1993) strongly suggests that solar variability had a major influence on climate during this interval (Bond et al., 2001; Denton and Karlén, 1973; Mayewski et al., 1997; O'Brien et al., 1995). There is no evidence of NADW production changes, and trade wind intensity is low, suggesting these contributed negligibly to atmospheric $\Delta^{14}C$ changes.

Summary and conclusions

The most important conclusions to be drawn from our compilation of proxy records are that Holocene climate has been highly variable, and that there are multiple controls that must have been responsible for this variability. Furthermore, the RCCs described occur in fairly regular quasi-periodic patterns, and the frequency of these RCC events appears to have increased since the middle Holocene. Finally, not all sites respond synchronously or equally during the RCC events despite their global extent. This latter point emphasizes the complexity of Holocene climate, further highlighting the importance of having widely distributed site-specific paleoclimatic data, to avoid the risk of using data series from one area to extrapolate to another.

As revealed by our synthesis, Holocene climate change can be quite abrupt, even in the absence of the large, unstable ice sheets that so dramatically disrupted Pleistocene climate. Further, Holocene RCCs have been large enough to have significant effects on ecosystems and humans. The short-lived 1200–1000 cal yr B.P. RCC event coincided with the drought-related collapse of Maya civilization and was accompanied by a loss of several million lives (Hodell et al., 2001; Gill, 2000), while the collapse of Greenland's Norse colonies at ~600 cal yr B.P. (Buckland et al., 1995) coincides with a period of polar cooling that is minor by glacial standards. Even the less extensive event from 4200 to 3800 cal yr B.P. coincided with major low-latitude drought and the collapse of the Akkadian Empire (deMenocal et al., 2000a).

The RCC interval 9000–8000 cal yr B.P. is the only event that coincides with a significant increase in volcanic aerosol production and it occurred when bipolar ice sheet dynamics still had the potential for substantial effects on global climate. Therefore, the early Holocene climate probably has more in common with the glacial world than with more recent historical times. This is an important point in the light of recent suggestions that the 8200 yr BP event may be thought of as an analog for future climate change (e.g., National Academy of Sciences, 2002).

All but the 9000–8000 cal yr B.P. RCC and the most recent RCC are characterized in general by bipolar cooling and an intensification of atmospheric circulation in the high latitudes and drying aridity at low latitudes. When the poles cool and polar atmospheric circulation intensifies, the low-latitude band of atmospheric circulation may well be compressed. This could dramatically alter the distribution of moisture bearing winds in the monsoon regions of the world and the carrying capacity for moisture in the atmosphere. Bipolar expansion of high latitude atmospheric circulation systems and subsequent redistribution of low-latitude atmospheric circulation begs a symmetrical, global forcing such as solar variability. Under cooler conditions, tropical aridity may result from a variety of factors, including the weakening of monsoon systems, reduced evaporation from cooler oceans, and weakened thermal convection over tropical landmasses. The most recent RCC (<600 cal yr B.P.) features bipolar cooling but a more variable response in humidity at low latitudes. This interval appears to be more complex than the classic "cool poles, dry tropics" pattern that typified the Pleistocene and most of earlier Holocene RCCs.

Of all the potential climate forcing mechanisms, solar variability superimposed on long-term changes in insolation (Bond et al., 2001; Denton and Karlén, 1973; Mayewski et al., 1997; O'Brien et al., 1995) seems to be the most likely important forcing mechanism for the RCCs except perhaps those at 9000–8000 and 4200–3800 cal yr B.P. We therefore suggest that significantly more research into the potential role of solar variability is warranted, involving new assessments of potential transmission mechanisms to induce climate change (e.g., Bard et al., 2000; Beer, 2000; Bray, 1971) and potential enhancement of natural feedbacks that may amplify the relatively weak forcing related to fluctuations in solar output (Saltzman and Moritz, 1980).

The hydrological cycle that governs the latent heat distribution in the atmosphere through water vapor transport clearly plays a major role in the distribution of Holocene climate variability, as indicated by the large fluctuations in lake levels, monsoon activity, and regional humidity registered in these paleoclimate records. Ocean–atmosphere numerical modeling experiments reveal long-term changes in moisture balance and ENSO strength. During the mid-Holocene, for example, these changes appear to be related to orbitally driven changes in the seasonal cycle of solar radiation (Clement et al., 2000). Short-term, RCC-style moisture balance events are superimposed on this orbital-driven behavior.

Negligible forcing roles are played by CH_4 and CO_2 during most of the Holocene, although it should be noted that the changes in concentration of these trace gases are minor compared to those experienced during the glacial–interglacial transition and over the last century. Few large shifts in greenhouse gases occur during the pre-anthropogenic Holocene apart from a few notable exceptions such as the CH_4 depression at 8200 cal yr B.P. and the CO_2 decline at 1200 cal yr B.P. Thus, changes in the concentrations of CO_2 and CH_4 appear to have been more the result than the cause of the RCCs.

The global distribution of changes in moisture balance, temperature, and atmospheric circulation during the RCCs seen in Figure 5 is suggestive of a global-scale climate phenomenon on the order of ENSO in magnitude. During ENSO events, the Earth is subjected to massive redistributions of moisture and heat. Although this is speculative, persistent shifts in ENSO frequency may provide a modern, shorter-term analogue for Holocene RCC events.

We emphasize that the present effort is only a first cut at investigating global climate variability within the Holocene. Ultimately, it would be ideal to further quantify our qualitative interpretations with statistical analysis of the time series, using spatiotemporal empirical orthogonal function (EOF) analysis, for example. However, it is premature to do so because the current dating controls for most of the records are sufficiently low to cause the most interesting parts of the records (i.e., the RCCs) to average out in such an analysis. Furthermore, the comparison of multiple kinds of variables (i.e., temperature, precipitation, atmospheric circulation, etc.) requires assumptions about relative weighting of such variables that need to be further investigated before such an analysis would be useful. Determining the appropriate way to objectively blend these records is a desirable research goal. We also fully anticipate that future research will identify additional aspects of the Holocene climate record that are of equal or greater interest to the development of a comprehensive view of climate variability within the current interglacial period. Future advances will require more paleoclimate records, notably in the Southern Hemisphere, and more precise examination of the timing of RCC intervals and teleconnections within those intervals. A sound understanding of the nature and causes of Holocene RCCs, particularly those post-dating the Northern Hemisphere deglaciation, will be of considerable relevance to the modeling and prediction of present and future climate. These events offer our only glimpses of real-world climate responses to natural forcing mechanisms in the absence of significant human influences.

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