Chapter 3. Hydrological Variability and Change

Lead Author: Edward R. Cook,* Columbia University, Palisades, NY

Contributing Authors: Patrick J. Bartlein,* University of Oregon, Eugene, OR

Noah Diffenbaugh, Purdue University, West Lafayette, IN

Richard Seager,* Columbia University, Palisades, NY

Bryan N. Shuman, University of Wyoming, Laramie, WY

Robert S. Webb,* NOAA Earth System Research Lab, Boulder, CO

John W. Williams, University of Wisconsin, Madison, WI

Connie Woodhouse, University of Arizona, Tucson, AZ

* SAP 3.4 Federal Advisory Committee Member

Key Findings

- Protracted droughts, and their impacts on agricultural production and water supplies, are among the greatest natural hazards facing the United States and the globe today and in the foreseeable future.

- Floods predominantly reflect both antecedent conditions and meteorological events and are often more localized relative to drought in both time and space. On subcontinental-to-continental scales, droughts occur more frequently than floods and can persist for decades and even centuries.

- On interannual to decadal time scales, droughts can develop faster than the time scale needed for human societies to adapt to the change. Thus, a severe drought lasting several years can be regarded as an abrupt change, although it may not reflect a permanent change of state of the climate system.

- Droughts and episodes of regional-scale flooding can both be linked to the large-scale atmospheric circulation patterns over North America, and often occur simultaneously in different parts of the country, compounding their impact on human activities.

- Empirical studies and climate model experiments conclusively show that droughts over North America have been significantly influenced by the state of tropical sea surface temperatures (SSTs). Of particular relevance to North America, cool La

Niña-like SSTs in the eastern equatorial Pacific frequently cause development of droughts over the southwestern United States and northern Mexico. Warm subtropical North Atlantic SSTs play a secondary role in forcing drought in southwestern North America.

- Historic droughts over North America have been severe, the "Dust Bowl" drought of the 1930s being the canonical example, but those droughts were not nearly as prolonged as a series of "megadroughts" reconstructed from tree rings since Medieval times (ca. 1,000 years ago) up to about A.D. 1600. Modeling experiments indicate that these megadroughts were likely partly forced by cool SSTs in the eastern equatorial Pacific as well. However, their exceptional duration has not been adequately explained nor has any involvement in forcing from SST changes in other oceans.
- These megadroughts are significant because they occurred in a climate system that was not being perturbed in a major way by human activity (i.e., the ongoing anthropogenic changes in greenhouse gas concentrations, atmospheric dust loadings, and land-cover changes).
- Even larger and more persistent changes in hydroclimatic variability worldwide are indicated throughout the Holocene (the past 11,500 years) by a diverse set of paleoclimatic indicators including some with annual-to-decadal resolution (e.g., speleothems, varved-lake records, high-resolution lake-sediment records). The global-scale controls associated with those changes were quite different from those of the past millennium and today, but they show the additional range of natural variability and abrupt hydroclimatic change that can be expressed by the climate system, including widespread and protracted (multi-century) droughts.
- There is no clear evidence to date of human-induced global climate change on North American precipitation amounts. However, since the IPCC AR4 report, further analysis of climate model scenarios of future hydroclimatic change over North America and the global subtropics indicate that subtropical aridity is likely to intensify and persist due to future greenhouse warming. This projected drying extends poleward into the United States Southwest, potentially increasing the

likelihood of severe and persistent drought there in the future. If the model results are correct then this drying may have already begun, but currently cannot be definitively identified amidst the considerable natural variability of hydroclimate in Southwestern North America.

Recommendations

- Research is needed to improve existing capabilities to forecast short- and long-term drought conditions and to make this information more useful and timely for decision making. In the future, drought forecasts should be based on an objective multimodel ensemble prediction system to enhance their reliability and the types of information should be expanded to include soil moisture, runoff, and hydrological variables (See also the *Western Governors' Association (2004)* National Integrated Drought Information System Report).
- The trend toward increasing subtropical aridity indicated by climate model projections needs to be investigated further to determine the degree to which it is likely to happen. If the model projections are correct, strategies for response to this pending aridity, on both regional and global scales, are urgently needed.
- Improved understanding of the dynamical causes of long-term changes in oceanic conditions, the atmospheric responses to these ocean conditions, and the role of soil moisture feedbacks are needed to advance drought prediction capabilities. Ensemble drought prediction is needed to maximize forecast skill and downscaling is needed to bring coarse-resolution drought forecasts from General Circulation Models down to the resolution of a watershed. (See also the *National Integrated Drought Information System Implementation Team, 2007.)*
- High-resolution paleoclimatic reconstructions of past drought have been fundamental to the evaluation of causes over North America in historic times and over the past millennium. This research should be expanded geographically to encompass as much of the global land masses as possible for the development and testing of predictive models.

- The record of past drought from tree rings and other proxies has revealed a succession of megadroughts prior to A.D. 1600 that easily eclipsed the duration of any droughts known to have occurred over North America since that time. Understanding the causes of these extraordinary megadroughts is vitally important.
- An understanding of the seasonality of drought and the relationships between winter and growing season droughts during periods of megadroughts documented in paleoclimatic records is needed. In particular, knowledge about the North American monsoon and how its variability is linked to SSTs and winter precipitation variability over decadal and longer time scales in the Southwestern United States and northern Mexico is critical.
- On longer time scales, significant land-cover changes have occurred in response to persistent droughts, and the role of land-cover changes in amplifying or damping drought conditions should be evaluated.
- Improved understanding of the links among gradual changes in climate (e.g., Meridional Overturning Circulation, or MOC), the role of critical environmental thresholds, and abrupt hydrologic changes is needed to enhance society's ability to plan and manage risks.
- The relationship between climate changes and abrupt changes in water quality and biogeochemical responses is not well understood and needs to be a priority area of study for modern process and paleoclimate research.
- The integration of high-resolution paleoclimate records with climate model experiments requires active collaboration between paleoclimatologists and modelers. This collaboration should be encouraged in future research on drought and climatic change in general.
- In order to reduce uncertainties in the response of floods to abrupt climate change, improvements in large-scale hydrological modeling, enhanced data sets for documenting past hydrological changes, and better understanding of the physical processes that generate flooding are all required.

1. Introduction—Statement of the Problem

A reliable and adequate supply of clean fresh water is essential to the survival of each human being on Earth and the maintenance of terrestrial biotic systems worldwide. However, rapidly growing human populations worldwide are increasing the stresses on currently available water supplies even before we factor in anticipated effects of a changing climate on the availiability of a clean and reliable fresh water supply. Changes in the frequency, intensity, and duration of droughts would have a significant impact on water supplies both for human societies and for terrestrial and inshore marine or estruarine ecosystems. Droughts are defined by the international meteorological community: the "prolonged absence or marked deficiency of precipitation", a "deficiency of precipitation that results in water shortage for some activity or for some group", or a "period of abnormally dry weather sufficiently prolonged for the lack of precipitation to cause a serious hydrological imbalance" (*Heim, 2002*; see also *Peterson et al., 2008* (CCSP SAP 3.3, Box 1.3)). Flooding is another important class of hydrologic variability that tends to affect smaller geographic regions and to last for shorter periods of time compared to drought. Consequently, floods generally have smaller impacts on human activities compared to droughts in North America. See the section on floods in the latter part of this chapter for more details.

Much of the research on climatic change, and most of the public's understanding of that work, has concerned temperature and the term "global warming." Global warming describes ongoing warming in this century by a few degrees Celsius, in some areas a bit more and in some a bit less. In contrast, changes in water flux between the surface of the Earth and the atmosphere are not expected to be spatially uniform but to vary much like the current daily mean values of precipitation and evaporation (*IPCC, 2007*). Although projected spatial patterns of hydroclimatic change are complex, many already wet areas are likely to get wetter and already dry areas are likely to get drier, while some intermediate regions on the poleward flanks of the current subtropical dry zones are likely to become increasingly arid. These anticipated changes will increase problems at both extremes of the water cycle, stressing water supplies in many arid and semi-arid regions while worsening flood hazards and erosion in many wet areas. Changes in precipitation intensity – the proportion of the total precipitation falling in events of different magnitude – have the potential to further challenge the management of water in the future. Moreover, the instrumental, historical, and prehistorical record of hydrological variations indicates that transitions between extremes can occur rapidly relative to the time span under consideration. Within time spans of decades, for example, transitions between wet conditions and dry conditions may occur within a year and can persist for several years.

Hydroclimatic changes are likely to affect all regions in the United States. Semi-arid regions of the Southwest are projected to dry further, and model results suggest that the transition may already be underway (*Hoerling and Kumar, 2003; Seager et al., 2007d*). Intensity of precipitation is also expected to increase across most of the country, continuing its recent trend (*Kunkel et al., 2008,* CCSP SAP 3.3, Sec. 2.2.2.2). The drying in the Southwest is a matter of great concern because water resources in this region are already stretched, new development of resources will be extremely difficult, and the population (and thus demand for water) continues to grow rapidly (see Fig. 3.1). This situation raises the politically charged issue of whether the allocation of around 90% of the region's water to agriculture is sustainable and consistent with the course of regional development. Mexico is also expected to dry in the near future, turning this feature of hydroclimatic change into an international and cross-border issue with potential impacts on migration and social stability. The U.S. Great Plains, where deep aquifers are being rapidly depleted, could also experience changes in water supply that affect agricultural practices, grain exports, and biofuel production. Other normally well-watered regions of the United States may also face water shortages caused by short-term droughts when demand outstrips supply and access to new water supplies is severely limited (e.g., Atlanta, GA). Other regions of the United States, while perhaps not having to face a climatic change-induced water shortage, may also have to make changes to infrastructure to deal with the erosion and flooding implications of increases in precipitation intensity.

Increases in the frequency of droughts in response to climate change can in turn produce further climate changes. For example, increased drought frequency may reduce forest growth, decreasing the sequestration of carbon in standing biomass, and increasing its release from the soil (*King et al., 2007* (CCSP SAP 2.2)). Similarly, increasing temperatures and drought will likely promote increased disturbance by fire and insect pathogens, with a consequent impact on ecosystems and their carbon balances (*Backlund et al., 2008* (CCSP SAP 4.3))

In addition, the United States could be affected by hydroclimatic changes in other regions of the world if global climate change becomes a global security issue. Security, conflict, and migration are most directly related to economic, political, social, and demographic factors. However environmental factors, including climate variability and climate change, can also play a role, even if secondary (*Lobell et al., 2008; Nordas and Gleditsch, 2007*). Two recent examples of a quantitative approach to determine the links between conflict and climate are *Raleigh and Urdal (2007)* and *Hendrix and and Glaser (2007)*. Raleigh and Urdal, basing their arguments on statistical relations between late 20th century conflict data and environmental data, find that the influence of water scarcity is at best weak. Hendrix and Glaser focused on sub-Saharan Africa and found that climate variability (e.g. a transition into a dry period) could foster conflict when other conditions (political, economic, demographic, etc.) favored conflict anyway. Hendrix and Glaser also examined a climate projection for sub-Saharan Africa from a single model and found that this led to no significant increase in conflict risk because the year-to-year climate variability did not change. Such quantitative methods need to be applied to other regions where changes in the mean state and variability of climate are occurring now and also to regions where climate change is robustly projected by models. Across different regions of the world, projected increases in flooding risk, potential crop damage and declines in water quality, combined with rising sea level, have the potential to force migration and cause social, economic, and political instability. However, currently there are no comprehensive assessments of the security risk posed by climate change that take account of all the available climate-change projection information and also take into account the multiple causes of conflict and migration. Consequently, no conclusions can yet be drawn on the climate-change impact on global or national security.

The paleoclimatic record reveals dramatic changes in North American hydroclimate over the last millennium that had nothing to do with human-induced changes in greenhouse gases and global warming. In particular, tree ring reconstructions of the Palmer Drought

Severity Index (PDSI, see *Kunkel et al., 2008,* CCSP SAP 3.3, Box 2.1) show vast areas of the Southwest and the Great Plains were severely affected by a succession of megadroughts between about A.D. 800 and 1600 that lasted decades at a time and contributed to the development of a more arid climate during the Medieval Period (A.D. 800 to 1300) than in the last century. These megadroughts have been linked to La Niña-like changes in tropical Pacific SSTs, changes in solar irradiance, and explosive volcanic activity. They are dynamically distinct from projected future drying, which is associated with a quite spatially uniform surface warming, based on model projections. However, the paleoclimatic records differ enough from climate model results to suggest that the models may not respond correctly to radiative forcing. The climate system dynamics associated with these prehistoric megadroughts need to be better understood, modeled, and related to the processes involved in future climate change.

Over longer time spans, the paleoclimatic record indicates that even larger hydrological changes have taken place, in response to past changes in the controls of climate, that rival in magnitude those expected during the next several decades and centuries. For example, the mid-continent of North America experienced conditions that were widespread and persistently dry enough to activate sand dunes, lower lake levels, and change the vegetation from forest to grassland for several millennia during the mid-Holocene (roughly 8,000 to 4,000 years ago). These changes were driven primarily by variations in the Earth's orbit that altered the seasonal and latitudinal distribution of incoming solar radiation. Superimposed on these Holocene variations were variations on centennial and shorter time scales that also were recorded by aeolian activity, and by geochemical and paleolimnological indicators.

The serious hydrological changes and impacts known to have occurred in both historic and prehistoric times over North America reflect large-scale changes in the climate system that can develop in a matter of years and, in the case of the more severe past megadroughts, persist for decades. Such hydrological changes fit the definition of abrupt change because they occur faster than the time scales needed for human and natural systems to adapt, leading to substantial disruptions in those systems. In the Southwest, for example, the models project a permanent drying by the mid-21st century that reaches the level of aridity seen in historical droughts, and a quarter of the projections may reach this level of aridity much earlier. It is not unreasonable to think that, given the complexities involved, the strategies to deal with declining water resources in the region will take many years to develop and implement. If hardships are to be minimized, it is time to begin planning to deal with the potential hydroclimatic changes described here.

2. Causes and Impacts of Hydrological Variability Over North America in the Historical Record

After the 1997-98 El Niño, the Western United States entered a drought that has persisted until the time of writing (July 2007). The driest years occurred during the extended La Niña of 1998-2002. Although winter 2004-05 was wet, dry conditions returned afterwards and even continued through the modest 2006-07 El Niño. In spring 2007 the two massive reservoirs on the Colorado River, Lakes Powell and Mead, were only half full. Droughts of this severity and longevity have occurred in the West before and Lake Mead (held back by Hoover Dam which was completed in 1935) was just as low for a few years during the severe 1950s drought in the Southwest. Studies of the instrumental record make clear that western North America is a region of strong meteorological and hydrological variability in which, amidst dramatic year-to-year variability, there are extended droughts and pluvials (wet periods) running from a few years to a decade. These dramatic swings of hydroclimatic variability have tremendous impacts on water resources, agriculture, urban water supply, and terrestrial and aquatic ecosystems. Drought and its severity can be numerically defined using indices that integrate temperature, precipitation, and other variables that affect evapotranspiration and soil moisture. See *Heim (2002)* for details.

Box 3.1—Impacts of Hydrologic Change: An Example From the Colorado River

An example of the potential impacts of a rapid change to more drought-prone conditions can be illustrated by the recent drought and its effect on the Colorado River system. The Colorado River basin, as well as much of the WesternUnited States, experienced extreme drought conditions from 1999 to 2004, with inflows into Lake Powell between 25% and 62% of average. In spring 2005, the basin area average reservoir storage was at about 50%, down from over 90% in 1999 (*Fulp,* *2005*). Although this most recent drought has caused serious water resource problems, paleoclimatic records indicate droughts as, or more, severe occurred as recently as the mid-19th century (*Woodhouse et al., 2005*). Impacts of the most recent drought were exacerbated by greater demand due to a rapid increase in the populations of the seven Colorado River basin States of 25% over the past decade (*Griles, 2004*). Underlying drought and increases in demand is the fact that the Colorado River resources have been over-allocated since the 1922 Colorado River Compact, which divided water supplies between upper and lower basin States based on a period of flow that has not been matched or exceeded in at least 400 years (*Stockton and Jacoby, 1976; Woodhouse et al., 2006*).

During the relatively short (in a paleoclimatic context) but severe 1999-2004 drought, vulnerabilities of the Colorado River system to drought became evident. Direct impacts included a reduction in hydropower and losses in recreation opportunities and revenues. At Hoover Dam, hydroelectric generation was reduced by 20%, while reservoir levels were at just 71 feet above the minimum power pool at Glen Canyon Dam in 2005 (*Fulp, 2005*). Hydroelectric power generated from Glen Canyon Dam is the source of power for about 200 municipalities (*Ostler, 2005*). Low reservoir levels at Lakes Powell and Mead resulted in the closing of three boat ramps and $10 million in costs to keep others in operation, as well as an additional $5 million for relocation of ferry services (*Fulp, 2005*). Blue ribbon trout fishing and whitewater rafting industries in the upper Colorado River basin (Upper Basin) also suffered due to this drought. In the agricultural sector, depletion of storage in reservoirs designed to buffer impacts of short-term drought in the Upper Basin resulted in total curtailment of 600,000 to 900,000 acre feet a year during the drought (*Ostler, 2005*). As a result of this drought, in combination with current demand, reservoir levels in Lake Mead, under average runoff and normal reservoir operations, are modeled to rise to only 1,120 feet over the next two decades (*Maguire, 2005*). Since the reservoir spills at 1221.4 feet (*Fulp, 2005*), this means the reservoir will not completely fill during this time period.

The Colorado River water system was impacted by the 5-year drought, but water supplies were adequate to meet most needs, with some conservation measures enacted (*Fulp, 2005*). How much longer the system could have handled drought conditions is uncertain, and at some point, a longer drought is certain to have much greater impacts. Under the Colorado River Compact and subsequent legal agreements, the Upper Basin provides 8.25 million acre feet to the Lower Basin each year (although there are some unresolved issues concerning the exact amount). If that amount is not available in storage, a call is placed on the river, and Upper Basin junior water rights holders must forgo their water to fulfill downstream and senior water rights. In the Upper Basin, the junior water rights are held by major water providers and municipalities in the Front Range, including Denver Water, the largest urban water provider in Colorado. Currently, guidelines that deal with the management of the Colorado River system under drought condition are being developed, because supplies are no longer ample to meet all demands during multi-year droughts (*USBR, 2007*). However, uncertainties related to future climate projections make planning difficult.

2.1 What Is Our Current Understanding of the Historical Record?

Instrumental precipitation and temperature data over North America only become extensive toward the end of the 19th century. Records of sea surface temperatures (SSTs) are sufficient to reconstruct tropical and subtropical ocean conditions starting around A.D. 1856. The large spatial scales of SST variations (in contrast to those of precipitation) allow statistical methods to be used to “fill in” spatial and temporal gaps and provide near-global coverage from this time on (*Kaplan et al., 1998; Rayner et al., 2003*). A mix of station data and tree ring analyses has been used to identify six serious multiyear droughts in western North America during this historical period (*Fye et al., 2003; Herweijer et al., 2006*). Of these, the most famous is the “Dust Bowl” drought that included most of the 1930s decade. The other two in the 20th century are the severe drought in the Southwest from that late 1940s to the late 1950s and the drought that began in 1998 and is ongoing. Three droughts in the mid to late 19th century occurred (with approximate dates) from 1856 to 1865, from 1870 to 1876, and from 1890 to 1896.

In all of these droughts, dry conditions prevailed over most of western North America from northern Mexico to southern Canada and from the Pacific Coast to the Mississippi River and sometimes farther east, with wet conditions farther north and farther south. The pattern of the Dust Bowl drought seemed unique in that the driest conditions were in the central and northern Great Plains and that dry conditions extended into the Pacific Northwest, while anomalies in the Southwest were modest.

Early efforts used observations to link these droughts to mid-latitude ocean variability. Since the realization of the powerful impacts of El Niño on global climate, studies have increasingly linked persistent, multiyear North American droughts with tropical Pacific SSTs and persistent La Niña events (*Cole and Cook, 1998; Cole et al., 2002; Fye et al., 2004*). This can be appreciated through the schematic maps shown in Figure 3.2 that show the teleconnection patterns of temperature and precipitation over North America commonly associated with the warm and cold phases of the ENSO cycle over the tropical Pacific. Warm ENSO episodes (El Niños) result in cool-wet conditions from the Southwestern over to the Southeastern United States during the winter season. In contrast, cold ENSO episodes (La Niñas) result in the development of warm-dry (i.e., drought) conditions over the same U.S. region, again primarily for the winter season. In contrast, the importance of ENSO on summer season climate is much stronger elsewhere in the world, like over Southeast Asia and Australasia. However, new research suggests a teleconnection between Pacific SSTs and the North American monsoon as well (e.*g., Castro et al., 2007b*). The North American monsoon (June through September) is a critical source of precipitation for much of Mexico (up to 70% of the annual total) and the Southwestern United States (30%-50%). The means whereby tropical SST anomalies impact climate worldwide are reasonably well understood. The SST anomalies lead to anomalies in the patterns and magnitude of convective heating over the tropical oceans which drive atmospheric circulation anomalies that are transmitted around the world via stationary Rossby waves. The stationary waves then also subsequently impact the propagation of transient eddies thereby altering the patterns of storm tracks, which feeds back onto the mean flow. For a review see Trenberth et al. (1998).

Box 3.2—Waves in the Westerlies, Weather, and Climate Anomalies

Maps of winds in the upper atmosphere (e.g., at 30,000 feet), shown daily as the jet-stream in newspaper, television, and web-based accounts of current weather, typically show a meandering pattern of air flow with three to five "waves" in the westerly winds that circle the globe in the mid-latitudes. These "Rossby waves" in the westerlies consist of sets of *ridges*, where (in the Northern Hemisphere) the flow pattern in the upper atmosphere broadly curves in the clockwise direction, and *troughs*, where the curvature is in a counter-clockwise direction. Rossby waves are ultimately generated by the temperature and pressure gradients that develop between the tropics and high-latitude regions, and in turn help to redistribute the energy surplus of the tropics through the movement of heat and moisture from the tropics toward the middle and high latitudes. Over North America, an upper-level ridge is typically found over the western third of the continent, with a trough located over the region east of the RockyMountains.

Distinct surface-weather conditions can be associated with the ridges and troughs. In the vicinity of the ridges, air sinks on a large scale, becoming warmer as it does so, while high pressure and diverging winds develop at the surface, all acting to create fair weather and to suppress precipitation. In the vicinity of troughs, air tends to converge around a surface low-pressure system (often bringing moisture from a source region like the subtropical North Pacific or Atlantic Oceans or the Gulf of Mexico) and to rise over a large area, encouraging precipitation.

From one day to the next, the ridges and troughs in the upper-level circulation may change very little, leading to their description as *stationary waves*. Meanwhile, smaller amplitude, shorter wavelength waves, or *eddies*, move along the larger scale stationary waves, again bringing the typical meteorological conditions associated with clockwise turning (fair weather) or counter-clockwise turning (precipitation) air streams, which may amplify or damp the effects of the larger scale waves on surface weather. Standard weather-map features like cold and warm fronts develop in response to the large-scale horizontal and vertical motions. Although uplift (and hence cooling, condensation, and precipitation) may be enhanced along the frontal boundaries between different airmasses, fronts and surface low- and high-pressure centers should be thought of as the symptoms of the large-scale circulation as opposed to being the primary generators of weather. The persistence or frequent recurrence of a particular wave pattern over weeks, months, or seasons then imparts the typical weather associated with the ridges and troughs, creating monthly and seasonal climate anomalies.

The particular upper-level wave pattern reflects the influence of fixed features in the climate system, like the configuration of continents and oceans and the location of major mountain belts like the Cordillera of western North America and the Tibetan Plateau (which tend to anchor ridges in those locations), and variable features like sea-surface temperature patterns, and snow-cover and soil-moisture anomalies over the continents.

On longer time scales during the Holocene (roughly the past 11,000 years), climatic variations in general, and hydrologic changes in particular, exceeded in both magnitude and duration those of the instrumental period or of the last millennium. In the mid-continent of North America, for example, between about 8,000 and 4,000 years ago, forests were replaced by steppe as the prairie expanded eastward, and sand dunes became activated across the Great Plains. These Holocene paleoclimatic variations occurred in response to the large changes in the controls of global and regional climates that accompanied deglaciation, including changes in ice-sheet size (area and elevation), the latitudinal and seasonal distribution of insolation, and atmospheric composition, including greenhouse gases and dust and mineral aerosols (*Wright et al., 1993*). Superimposed on these orbital-time-scale variations were interannual to millennial time-scale variations, many abrupt in nature (*Mayewski et al., 2004; Viau et al., 2006*), arising from variations in solar output, volcanic aerosols, and internally generated covariations among the different components of the climate system. On longer, or "orbital" time scales, the ice sheets, biogeochemically determined greenhouse gas concentrations, and dust and aerosol loading should be regarded as internal components of the climate system, but over the past 11,000 years, they changed slowly enough relative to other components of the climate system, such as the atmosphere and surface ocean, that they are most appropriately considered as external controls of regional-scale climate variations (*Saltzman, 2002*).

2.1.1 Coupled Ocean-Atmosphere Forcing of North American Hydrological Variability

The standard approach that uses models to demonstrate a link between SSTs and observed climate variability involves forcing an Atmospheric General Circulation Model (AGCM) with observed SSTs as a lower boundary condition (see *Hoerling et al., in prep* (CCSP SAP 1.3, Box 3.2) for further discussion of this approach). Ensembles of simulations are used with different initial conditions such that the internally generated atmospheric weather in the ensemble members is uncorrelated from one member to the next and, after averaging over the ensemble, the part of the model simulation common to all - the part that is SST forced - is isolated. The relative importance of SST anomalies in different ocean basins can be assessed by specifying observed SSTs only in some areas and using climatological SSTs (or SSTs computed with a mixed layer (ML) ocean) elsewhere.

Schubert et al. (2004a,b) performed a climate model simulation from 1930 to 2004, which suggested that both a cold eastern equatorial Pacific and a warm subtropical Atlantic were the underlying forcing for drought over North America in the 1930s. *Seager et al. (2005b)* and *Herweijer et al. (2006)* performed ensembles that covered the entire period of SST observations since 1856. These studies conclude that cold eastern equatorial Pacific SST anomalies in each of the three 19th century droughts, the Dust Bowl, and the 1950s drought were the prime forcing factors. *Seager (2007)* has made the same case for the 1998-2002 period of the current drought, suggesting a supporting role for warm subtropical Atlantic in forcing drought in the West. During the 1930s and 1950s droughts, the Atlantic was warm, whereas, the 19th century droughts seem to be more solely Pacific driven. Results for the Dust Bowl drought are shown in Figure 3.3, and time series of modeled and observed precipitation over the Great Plains are shown in Figure 3.4. *Hoerling and Kumar (2003)* instead emphasize the combination of a La Niña-like state and a warm Indo-west Pacific Ocean in forcing the 1998-2002 period of the most recent drought. On longer time scales, *Huang et al. (2005)* have shown that models forced by tropical Pacific SSTs alone can reproduce the North American wet spell between the 1976-77 and 1997-98 El Niños. The Dust Bowl drought was unusual in that it did not impact the Southwest. Rather, it caused reduced precipitation and high temperatures in the northern Rocky Mountain States and the western Canadian prairies, a spatial pattern that models generally fail to simulate (*Seager et al., 2007c*).

The SST anomalies prescribed in the climate models that result in reductions in precipitation are small, no more than a fraction of a degree Celsius. These changes are an order of magnitude smaller than the SST anomalies associated with interannual El Niño/Southern Oscillation (ENSO) events or Holocene SST variations related to insolation (incoming solar radiation) variations (~0.50°C; *Liu et al., 2003, 2004*). It is the persistence of the SST anomalies and associated moisture deficits that create serious drought conditions. In the Pacific, the SST anomalies presumably arise naturally from ENSO-like dynamics on time scales of a year to a decade (*Newman et al., 2003*). The warm SST anomalies in the Atlantic that occurred in the 1930s and 1950s (and in between), and usually referred to as part of an Atlantic Multidecadal Oscillation (AMO; Kushnir, 1994; Enfield et al., 2001), are of unknown origin. *Kushnir (1994), Sutton and Hodson (2005),* and *Knight et al. (2005)* have linked them to changes in the Meridional Overturning Circulation (see Chapter 4), which implies that a stronger overturning and a warmer North Atlantic Ocean would induce a drying in southwestern North America. However, others have argued that the AMO-related changes in tropical Atlantic SSTs are actually locally forced by changes in radiation associated with aerosols, rising greenhouse gases, and solar irradiance (*Mann and Emanuel, 2005*).

The dynamics that link tropical Pacific SST anomalies to North American hydroclimate are better understood and, on long time scales, appear as analogs of higher frequency phenomena associated with ENSO. The influence is exerted in two ways: first, through propagation of Rossby waves from the tropical Pacific polewards and eastwards to the Americas (*Trenberth et al., 1998*) and, second, through the impact that SST anomalies have on tropospheric temperatures, the subtropical jets, and the eddy-driven mean meridional circulation (*Seager et al., 2003, 2005a,b; Lau et al., 2006*). During La Niñas both mechanisms force air to descend over western North America, which suppresses precipitation. Although models, and analysis of observations (*Enfield et al., 2001; McCabe et al., 2004; Wang et al., 2006*), support the idea that warm subtropical North Atlantic SSTs can cause drying over western North America, the dynamics that underlay this have not been so clearly diagnosed and explained within model experiments.

The influence of Pacific SSTs on the North American monsoon has been documented at interannual and decadal time scales. In particular, a time-evolving teleconnection response in the early part of the summer appears to influence the strength and position of the monsoon ridge. In contrast to winter precipitation/ENSO relationships, La Niña-like conditions in the eastern and central tropical Pacific favor a wet and early monsoon and corresponding dry and hot conditions in the Central United States (Castro et al., 2001; Schubert et al., 2004a; Castro et al., 2007b). In contrast, El Niño-like conditions favor a dry and delayed monsoon and corresponding wet and cool conditions in the Central United States (op. cit.).

2.1.2 Land Surface Feedbacks on Hydroclimate Variability

The evidence that multiyear North American droughts appear systematically together with tropical SST anomalies and that atmospheric models forced by these anomalies can reproduce some aspects of these droughts indicates that the ocean is an important driver. In addition to the ocean influence, some modeling and observational studies estimate that soil moisture feedbacks also influence precipitation variability (*Oglesby and Erickson, 1989; Namias, 1991; Oglesby, 1991*). *Koster et al. (2004)* used observations to show that on the time scale of weeks, precipitation in the Great Plains is significantly correlated with antecedent precipitation. *Schubert et al. (2004b)* compared models run with average SSTs, with and without variations in evaporation efficiency, and showed that multiyear North American hydroclimate variability was significantly reduced if evaporation efficiency was not taken into account. Indeed, their model without SST variability was capable of producing multiyear droughts from the interaction of the atmosphere and deep soil moisture. This result needs to be interpreted with caution since *Koster et al. (2004)* also show that the soil moisture feedback in models seems to exceed that deduced from observations. In a detailed analysis of models, observations and reanalyses, *Ruiz-*

Barradas and Nigam (2005) and *Nigam and Ruiz-Barradas (2006)* conclude that interannual variability of Great Plains hydroclimate is dominated by transport variability of atmospheric moisture and that the local precipitation recycling, which depends on soil moisture, is overestimated in models and provides a spuriously strong coupling between soil moisture and precipitation.

Past droughts have also caused changes in vegetation. For example, during the Dust Bowl drought there was widespread failure of non-drought-resistant crops that led to exposure of bare soil. Also, during the Medieval megadroughts there is evidence of dune activity in the Great Plains (*Forman et al., 2001*), which implies a reduction in vegetation cover. Conversions of croplands and natural grasses to bare soil could also impact the local hydroclimate through changes in surface energy balance and hydrology. Further, it has been argued on the basis of experiments with an atmosphere model with interactive dust aerosols that the dust storms of the 1930s worsened the drought, and moved it northward, by altering the radiation balance over the affected area (*Cook et al., 2008*). The widespread devegetation caused by crop failure in the 1930s could also have impacted the local climate. These aspects of land-surface feedbacks on drought over North America need to be examined further with other models, and efforts need to be made to better quatify the land-surface changes and dust emissions during the Dust Bowl.

2.1.3 Historical Droughts Over North America and Their Impacts

According to the National Oceanic and Atmospheric Administration (NOAA; see http://www.ncdc.noaa.gov/oa/reports/billionz.html for periodically updated economic information regarding U.S. weather diasters), over the period from 1980 to 2006, droughts and heat waves are among the most expensive natural disaster in the United States along with tropical storms (including the devastating 2005 hurricane season) and wide-spread or regional flooding episodes. The annual cost of drought to the United States is estimated to be in the tens of billions of dollars.

The above describes the regular year-in, year-out costs of drought. In addition, persistent multiyear droughts have had important consequences in national affairs. The icon of drought impacts in North America is the Dust Bowl of the 1930s. In the early 20th

century, settlers transferred large areas of the Great Plains from natural prairie grasses, used to some extent for ranching, to wheat farms. After World War I, food demand in Europe encouraged increased conversion of prairie to crops. This was all possible because these decades were unusually wet in the Great Plains. When drought struck in the early 1930s, the non-drought-resistant wheat died, thus exposing bare soil. Faced with a loss of income, farmers responded by planting even more, leaving little land fallow. When crops died again there was little in the way of "shelter belts" or fallow fields to lessen wind erosion. This led to monstrous dust storms that removed vast amounts of top soil and caused hundreds of deaths from dust inhalation (*Worster, 1979; Hansen and Libecap, 2004; Egan, 2006*). As the drought persisted year after year and conditions in farming communities deteriorated, about a third of the Great Plains residents abandoned the land and moved out, most as migrant workers to the Southwest and California, which had not been severely hit by the drought.

The Dust Bowl disaster is a classic case of how a combination of economic and political circumstances interacted with a natural event to create a change of course in national and regional history. It was in the 1930s that the Federal Government first stepped in to provide substantial relief to struggling farm communities heralding policies that remain to this day. The Dust Bowl drought also saw an end to the settlement of the semi-arid lands of the United States based on individual farming families acting independently. In addition, wind erosion was brought under control via collective action, organized within Soil Conservation Districts, while farm abandonment led to buyouts and a large consolidation of land ownership (*Hansen and Libecap, 2004*). Ironically, the population migration to the West likewise provided the manpower needed in the armaments industry after 1941 to support the U.S. World War II effort.

Earlier droughts in the late 19th century have also tested the feasibility of settlement of the West based on provisions within the Homestead Act of 1862. This act provided farmers with plots of land that may have been large enough to support a family in the East but not enough in the arid West, and it also expected them to develop their own water resources. The drought of the early to middle 1890s led to widespread abandonment in the Great Plains and acceptance, contrary to frontier mythology of "rain follows the plow"

(*Libecap and Hansen, 2002*), that if the arid lands were to be successfully settled and developed, the Federal Government was going to have to play an active role. The result was the Reclamation Act of 1902 and the creation of the U.S. Bureau of Reclamation, which in the following decades developed the mammoth water engineering works that sustain agriculture and cities across the West from the Great Plains to the Pacific Coast (*Worster, 1985*).

On a different level, the Great Plains droughts of the 1850s and early 1860s played a role in the combination of factors that led to the near extinction of the American bison (*West, 1995*). Traditionally, bison tried to cope with drought by moving into the better watered valleys and riparian zones along the great rivers that flowed eastward from the Rocky Mountains. However, by the mid-19th century, these areas had become increasingly populated by Native Americans who had recently moved to the Great Plains after being evicted from their villages in more eastern regions by settlers and the U.S. Army, thereby putting increased hunting pressure on the bison herds for food and commercial sale of hides. In addition, the migration of the settlers to California after the discovery of gold there in 1849 led to the virtual destruction of the riparian zones used by the bison for over-wintering and refuge during droughts. The 1850s and early 1860s droughts also concentrated the bison and their human predators into more restricted areas of the Great Plains still suitable for survival. Drought did not destroy the bison, but it did establish conditions that almost lead to the extinction of one of America's few remaining species of megafauna (*West, 1995; Isenberg, 2000*).

The most recent of the historical droughts, which began in 1998 and persists at the time of writing, has yet to etch itself into the pages of American history, but it has already created a tense situation in the West as to what it portends. Is it like the 1930s and 1950s droughts and, therefore, is likely to end relatively soon? Or is it the emergence of the anthropogenic drying that climate models project will impact this region - and the subtropics in general -within the current century and, quite possibly, within the next few years to decades? *Breshears et al. (2005)* noted that the recent Southwest drought was warmer than 1950s drought and the higher temperatures exacerbated drought impacts in ways that are consistent with expectations for the amplification of drought severity in response to greenhouse forcing. If this drying comes to pass it will impact the future economic, political, and social development of the West as it struggles to deal with declining water resources.

2.1.4 Impacts of Change in the Atmospheric Branch of the Hydrological Cycle for Ground Water and River Flow

The nature of these impacts ranges from reductions in surface-water supplies affecting reservoir storage and operations, and delivery and treatment of water, to drawdown of aquifers, increased pumping costs, subsidence, and reductions of adjacent or connected surface-water flows. A multitude of water uses, including irrigated and unirrigated agriculture, hydroelectric and thermoelectric power (cooling), municipal and industrial water uses, transportation, and recreation (National Assessment, 2000), can be severely impacted by rapid hydroclimatic changes that promote drought. Reductions in water supplies that affect these uses can have profound impacts on regional economies. For example, drought in the late 1980s and early 1990s in California resulted in a reduction in hydropower and increased reliance on fossil fuels, and an additional $3 billion in energy costs (*Gleick and Nash, 1991*).

Rapid changes in climate that influence the atmospheric part of the hydrological cycle can affect the amount, form, and delivery of precipitation, which in turn influence soil moisture, runoff, ground water, surface flows, and lake levels, as well as atmospheric features such as clouds. Changes can take the form of shifts in state to overall wetter or drier conditions, more persistent drought or flood-causing events, and/or a greater frequency of extreme events. All of these types of rapid changes can have serious societal impacts with far-reaching effects on water availability, quality, and distribution (National Assessment, 2000).

Shifts in the climate background state may modulate, and either constructively or destructively influence, the “typical” hydrologic impacts of seasonal to interannual climate variability. For example, the Southwestern United States, which tends to receive higher than average winter-time precipitation during an El Niño event, and relies on these events to refill water supply reservoirs, could benefit from changes that increase or enhance El Niño events, but suffer from increased droughts if La Niña events, which tend to result in dry winters here, become more frequent (Fig. 3.2).

The impacts of these changes can exacerbate scarce water supplies in regions that are already stressed by drought, greater demand, and changing uses. The Department of Interior analysis of Western U.S. water supply issues (USBR, 2005) identifies a number potential water supply crises and conflicts by the year 2025 based on a combination of technical and other factors, including population trends and potential endangered species' needs for water, but under an assumption of a statistically stationary climate (Fig. 3.1). Any transient change in climate conditions that leads to an abrupt regime shift to more persistent or more severe drought will only compound these water supply conflicts and impact society.

Abrupt changes in hydroclimate that lead to sustained drought can have enormous impacts on the management of water systems, in particular, the large managed river systems in western areas of the Western United States Many of these managed systems are facing enormous challenges today, even without abrupt changes, due to increased demands, new uses, endangered species requirements, and tribal water-right claims. In addition, many of these systems have been found to be extremely vulnerable to relatively small changes in runoff (e.g., Nemec and Schaake, 1982; Christensen and Lettenmaier, 2006).

2.2 Global Context of North American Drought

When drought strikes North America it is not an isolated event. In "The Perfect Ocean for Drought," *Hoerling and Kumar (2003)* noted that the post-1998 drought that was then impacting North America extended from the western subtropical Pacific across North America and into the Mediterranean region, the Middle East, and central Asia. There was also a band of subtropical drying in the Southern Hemisphere during the same period. It has long been known that tropical SST anomalies give rise to global precipitation anomalies, but the zonal and hemispheric symmetry of ENSO impacts has only recently been emphasized (*Seager et al., 2005a*).

Hemispheric symmetry is expected if the forcing for droughts comes from the tropics. Rossby waves forced by atmospheric heating anomalies in the tropics propagate eastward and poleward from the source region into the middle and high latitudes of both hemispheres (*Trenberth et al., 1998*). The forced wave train will, however, be stronger in the winter hemisphere than the summer hemisphere because the mid-latitude westerlies are both stronger and penetrate farther equatorward, increasing the efficiency of wave propagation from the tropics into higher latitudes. The forcing of tropical tropospheric temperature change by the tropical SST and air-sea heat flux anomalies will also tend to create globally coherent hydroclimate patterns because (1) the temperature change will be zonally uniform and extend into the subtropics (*Schneider, 1977*) and (2) the result will require a balancing change in zonal winds that will potentially interact with transient eddies to create hemispherically and zonally symmetric circulation and hydroclimate changes.

In the tropics the precipitation anomaly pattern associated with North American droughts is very zonally asymmetric with reduced precipitation over the cold waters of the eastern and central equatorial Pacific and increased precipitation over the Indonesian region. The cooler troposphere tends to increase convective instability (*Chiang and Sobel, 2002*), and precipitation increases in most tropical locations outside the Pacific with the exception of coastal East Africa, which dries, possibly as a consequence of cooling of the Indian Ocean (*Goddard and Graham, 1999*).

North American droughts are therefore a regional realization of persistent near-global atmospheric circulation and hydroclimatic anomalies orchestrated by tropical atmosphere-ocean interactions. During North American droughts, dry conditions are also expected in mid-latitude South America, wet conditions in the tropical Americas and over most tropical regions, and dry conditions again over East Africa. Subtropical to mid-latitude drying should extend across most longitudes and potentially impact the Mediterranean region. However, the signal away from the tropics and the Americas is often obscured by the impact of other climate phenomena such as the North Atlantic Oscillation (NAO) impact on precipitation in the Mediterranean region (*Hurrell, 1995; Fye et al., 2006*). In a similar fashion, the Holocene drought in the mid-continent of

North America (Sec. 4) can be shown to be embedded in global-scale energy balance and atmospheric circulation changes.

2.2.1 The Perfect Ocean for Drought: Gradual Climate Change Resulting in Abrupt Impacts

The study of the 1998-2002 droughts that spread across the United States, Southern Europe, and Southwest Asia provides an example of a potential abrupt regime shift to one with more persistent and/or more severe drought in response to gradual changes in global or regional climate conditions. Research by *Hoerling and Kumar (2003)* provides compelling evidence that these severe drought conditions were part of a persistent climate state that was strongly influenced by the tropical oceans.

From 1998 through2002, prolonged below-normal precipitation and above- normal temperatures caused the United States to experience drought in both the Southwest and Western States and along the Eastern Seaboard. These droughts extended across southern Europe and Southwest Asia, with as little as 50% of the average rainfall in some regions (Fig. 3.5). The *Hoerling and Kumar (2003)* study used climate model simulations to assess climate response to altered oceanic conditions during the 4-year interval. Three different climate models were run a total of 51 times, and the responses averaged to identify the common, reproducible element of the atmosphere's sensitivity to the ocean. Results showed that the tropical oceans had a substantial effect on the atmosphere (Fig. 3.6). The combination of unprecedented warm sea-surface conditions in the western tropical Pacific and 3-plus consecutive years of cold La Niña conditions in the eastern tropical Pacific shifted the tropical rainfall patterns into the far western equatorial Pacific.

Over the 1998 through 2002 period, the cold eastern Pacific tropical sea surface temperatures, though unusual, were not unprecedented. However, the warmth in the tropical Indian Ocean and the west Pacific Ocean was unprecedented during the 20th century, and attribution studies indicate this warming (roughly 1°C since 1950) is beyond that expected of natural variability. The atmospheric modeling results suggest an important role for tropical Indian Ocean and the west Pacific Ocean sea surface conditions in the shifting of westerly jets and storm tracks to higher latitudes with a nearly continuous belt of high pressure and associated drying in the lower mid-latitudes. The tropical ocean forcing of multiyear persistence of atmospheric circulation not only increased the risk for severe and synchronized drying of the mid-latitudes between 1998 and 2002 but may potentially do so in the future, if such ocean conditions occur more frequently.

The *Hoerling and Kumar (2003)* analysis illustrates how changes in regional climate conditions such as slow increases in Indo-Pacific “Warm Pool” SSTs, when exceeding critical environmental thresholds, can lead to abrupt shifts in climate regimes (e.g., the anomalous atmospheric circulation patterns), which in turn alter the hydrologic response to natural variability. The study points out that the overall pattern of warmth in the Indian and west Pacific Oceans was both unprecedented and consistent with greenhouse gas forcing of climate change. Could similar abrupt shifts in climate regimes explain the persistence of droughts in the past? From a paleoclimatic perspective, simulations by *Shin et al. (2006)* using an Atmospheric General Circulation Model (AGCM) with a “slab” ocean, and by *Liu et al. (2003)* and *Harrison et al. (2003)* with a fully coupled Atmosphere-Ocean General Circulation Model (AOGCM) indicate that a change in the mean state of tropical Pacific SSTs to more La Niña-like conditions can explain North American drought conditions during the mid-Holocene. An analysis of Medieval hydrology by *Seagar et al. (2007b)* suggests the widespread drought in North America occurred in response to cold tropical Pacific SSTs and warm subtropical North Atlantic SSTs externally forced by high irradiance and weak volcanic activity (see *Mann et al., 2005; Emile-Geay et al., 2007*).

2.3 Is There Evidence Yet for Anthropogenic Forcing of Drought?

Analyses by *Karoly et al. (2003)* and *Nicholls (2004)* suggest that 2002 drought and associated heat waves in Australia were more extreme than the earlier droughts because the impact of the low rainfall was exacerbated by high potential evaporation. *Zhang et al. (2007)* have suggested that large-scale precipitation trends can be attributed to anthropogenic influences. However there is no clear evidence to date of human-induced global climate change on North American precipitation amounts. The Fourth Assessment Report (AR4) of the IPCC (*IPCC, 2007*) presents maps of the trend in precipitation over the period 1901 to 2005 that shows mostly weak moistening over most of North America and a weak drying in the Southwest. This is not very surprising in that both the first two decades and the last two decades of the 20th entury were anomalously wet over much of North America (*Swetnam and Betancourt, 1998; Fye et al., 2003; Seager et al., 2005b; Woodhouse et al., 2005*). The wettest decades between the 1976/77 and 1997/98 El Niños may have been caused by natural Pacific decadal variability (*Huang et al., 2005*). In contrast to the 20th- century record, the southern parts of North America are projected to dry as a consequence of anthropogenic climate change. After the 1997/98 El Niño, drought has indeed settled into the West, but since it has gone along with a more La Niña-like Pacific Ocean this makes it difficult to determine if some part of the drying is anthropogenic.

Trends based on the shorter period of the post-1950 period show a clear moistening of North America, but this period extends from the 1950s drought to the end of the late-20th century wet period (or pluvial). The 1950s drought has been linked to tropical Pacific and Atlantic SSTs and is presumed to have been a naturally occurring event. Further, the trend from 1950 to the end of the last century is likely to have been caused by the multidecadal change from a more La Niña-like tropical Pacific before 1976 to a more El Niño-like Pacific from 1976 to 1998 (*Zhang et al., 1997*), a transition usually known as the 1976-77 climate or regime shift, which caused wet conditions in the mid-latitude Americas (*Huang et al., 2005*). Again, this change in Pacific SSTs is generally assumed to have been a result of natural Pacific variability, and it has been shown that simple models of the tropical Pacific alone can create multidecadal variations that have this character (*Karspeck et al., 2004*). The warm phase of tropical Pacific decadal variability may have ended with the 1997/98 El Niño, after which La Niña-like conditions prevailed until 2002 followed by weak El Ninos and a return to La Nina in 2007. In these post-1998 years, drought conditions have also prevailed across the West as in previous periods of persistent La Niñas. Consequently, it would be very premature to state that the recent drought heralds a period of anthropogenic drying as opposed to the continuation of natural decadal and multidecadal variations. Detailed analysis of not only precipitation patterns but also patterns of stationary and transient atmospheric circulation, water vapor transports, and SSTs may be able to draw a distinction, but this has not yet been done.

A different view is offered by *Vecchi et al. (2006)*, who used sea level pressure (SLP) data to show a weakening of the along-Equator east-to-west SLP gradient from the late-19th century to the current one. The rapid weakening of this gradient during the 1976-77 climate shift contributes to this trend. *Vecchi et al. (2006)* showed that coupled climate model simulations of the 20th century forced by changes in CO_2, solar irradiance, and other factors also exhibit a weakening of the SLP gradient - a weaker Walker Circulation - which could be taken to mean that the 1976-77 shift, and associated wetting of North America, contained an anthropogenic component. However, as noted in the previous paragraph, it would be very premature to state that the post-2002 period heralds a period of anthropogenic drying as opposed to the continuation of natural decadal and multidecadal variations.

3. North American Drought Over the Past Millennia

Historical climate records provide considerable evidence for the past occurrence of exceptional multi-year droughts on the North American continent and their impacts on American history. In addition, modeling experiments have conclusively demonstrated the importance of large-scale tropical SSTs on forcing much of the observed hydroclimatic variability over North America and other global land areas. What is still missing from this narrative is a better understanding of just how bad droughts can become over North America. Is the 1930s Dust Bowl drought the worst that can conceivably occur over North America? Or, is there the potential for far more severe droughts to develop in the future? Determining the potential for future droughts of unprecedented severity can be investigated with climate models (*Seager et al., 2007d*), but the models still contain too much uncertainty in them to serve as a definitive guide. Rather, what we need is an improved understanding of the past occurrence of drought and its natural range of variability. The instrumental and historical data only go back about 130 years with an acceptable degree of spatial completeness over the United States (see the 19th century instrumental data maps in *Herweijer et al., 2006*), which does not provide us with enough time to characterize the full range of hydroclimatic variability that has occurred in the past and could conceivably occur in the future independent of any added effects due to greenhouse warming. To do so, we must look beyond the historical data to longer natural archives of past climate information.

3.1 Tree Ring Reconstructions of Past Drought Over North America

In the context of how North American drought has varied over the past 2,000 years, an especially useful source of "proxy" climate information is contained in the annual ring-width patterns of long-lived trees (*Fritts, 1976*). A tree can provide information about past climate in its annual ring widths because its growth rate is almost always climate-dependent to some degree. Consequently, the annual ring-width patterns of trees provide proxy expressions of the actual climate affecting tree growth in the past and these expressions can therefore be used to reconstruct past climate. The past 2,000 years is also particularly relevant here because the Earth's climate boundary conditions are not markedly different from those of today, save for the 20th century changes in atmospheric trace gas composition and aerosols that are thought to be responsible for recent observed warming. Consequently, a record of drought variability from tree rings in North America over the past two millennia would provide a far more complete record of extremes for determining how bad conditions could become in the future. Again, this assessment would be independent of any added effects due to greenhouse warming.

An excellent review of drought in the Central and Western United States, based on tree rings and other paleoproxy sources of hydroclimatic variability, can be found in *Woodhouse and Overpeck (1998)*. In that paper, the authors introduced the concept of the "megadrought," a drought that has exceeded the intensity and duration of any droughts observed in the more recent historical records. They noted that there was evidence in the paleoclimate records for several multidecadal megadroughts prior to 1600 that "eclipsed" the worst of the 20th century droughts including the Dust Bowl. The review by *Woodhouse and Overpeck (1998)* was limited geographically and also restricted by the lengths of tree-ring records of past drought available for study. At that time, a gridded set of summer drought reconstructions, based on the Palmer Drought Severity Index (PDSI; *Palmer, 1965*), was available for the conterminous United States, but only back to 1700 (*Cook et al., 1999*). Those data indicated that the Dust Bowl was the worst drought to have hit the U.S. over the past three centuries. However, a subset of the PDSI reconstructions in the western, southeastern, and Great Lakes portions of the United States also extended back to 1500 or earlier. This enabled *Stahle et al. (2000)* to describe in more detail the temporal and spatial properties of the late 16th century megadrought noted earlier by *Woodhouse and Overpeck (1998)* and compare it to droughts in the 20th century. In concurrence with those earlier findings, *Stahle et al. (2000)* showed that even the past 400 years were insufficient to capture the frequency and occurrence of megadroughts that clearly exceeded anything in the historical records in many regions.

3.2 The North American Drought Atlas

Since that time, great progress has been made in expanding the spatial coverage of tree-ring PDSI reconstructions to cover most of North America (*Cook and Krusic, 2004a,b; Cook et al., 2004*). The grid used for that purpose is shown in Figure 3.7. It is a 286-point 2.5° by 2.5° regular grid that includes all of the regions described in *Woodhouse and Overpeck (1998), Cook et al. (1999),* and *Stahle et al. (2000)*. In addition, the reconstructions were extended back 1,000 or more years at many locations. This was accomplished by expanding the tree-ring network from the 425 tree-ring chronologies used by *Cook et al. (1999)* to 835 series used by *Cook et al. (2004)*. Several of the new series also exceeded 1,000 years in length, which facilitated the creation of new PDSI reconstructions extending back into the megadrought period in the Western United States prior to 1600. Extending the reconstructions back at least 1,000 years was an especially important goal. *Woodhouse and Overpeck (1998)* summarized evidence for at least four widespread multi-decadal megadroughts in the Great Plains and the Western United States during the A.D. 750-1300 interval. These included two megadroughts lasting more than a century each during “Medieval” times in California’s Sierra Nevada (*Stine, 1994*). Therefore, being able to characterize the spatial and temporal properties of these megadroughts in the Western United States was extremely important.

Using the same basic methods as those in *Cook et al. (1999)* to reconstruct drought over the conterminous United States, new PDSI reconstructions were developed on the 286-point North American grid (Fig. 3.7) and incorporated into a North American Drought Atlas (NADA; *Cook and Krusic, 2004a,b; Cook et al., 2007*). The complete contents of NADA can be accessed and downloaded at http://iridl.ldeo.columbia.edu/SOURCES/.LDEO/.TRL/.NADA2004/.pdsi-atlas.html. In Figure 3.7, the irregular polygon delineates the boundaries of the area we refer to as southwestern North America and includes the southwestern United States and northern

Mexico. It encompasses all grid points on and within 27.5°-50°N. latitude and 97.5°-125°W. longitude and was the area used by *Cook et al. (2004).* The dashed line along the 40th parallel separates the West into northwest and southwest sectors, which will be compared later.

3.3 Medieval Megadroughts in the Western United States

Cook et al. (2004) examined the NADA contents back to A.D. 800 for the West to place the current turn-of-the-century drought there *(Seager, 2007)* in a long-term context. In so doing, a period of elevated aridity was found in the A.D. 900-1300 period that included four particularly widespread and prolonged multi-decadal megadroughts (Fig. 3.8). This epoch of large-scale elevated aridity was corroborated by a number of independent, widely scattered, proxy records of past drought in the West (*Cook et al., 2004*). In addition, the four identified megadroughts agreed almost perfectly in timing with those identified by *Woodhouse and Overpeck (1998)*, which were based on far fewer data. These findings were rather sobering for the West because they (1) verified the occurrence of several past multidecadal megadroughts prior to 1600, (2) revealed an elevated background state of aridity that lasted approximately four centuries, and (3) demonstrated that there are no modern analogs to the A.D. 900-1300 period of elevated aridity and its accompanying megadroughts. This is clearly a cause for concern because the data demonstrate that the West has the capacity to enter into a prolonged state of dryness without the need for greenhouse gas forcing.

The timing of the A.D. 900-1300 period of elevated aridity is especially worrisome because it occurred during what has historically been referred to as the 'Medieval Warm Period' (MWP; *Lamb, 1965*), a time of persistently above-average warmth over large parts of the Northern Hemisphere (*Esper et al., 2002*), including the Western United States (*LaMarche, 1974*). *Stine (1994)* also noted the association of his prolonged Sierra Nevada droughts with the MWP. Given that his particular climate expression was more related to hydroclimatic variability than to pure temperature change, *Stine (1994)* argued that a more appropriate name for this unusual climate period should be the 'Medieval Climate Anomaly' (MCA) period. We will use MCA from here on out when referring to drought during the Medieval period.

Herweijer et al. (2007) made some detailed examinations of the NADA in order to determine how the megadroughts during the MCA differed from droughts of more modern times. That analysis was restricted to effectively the same spatial domain as that used by *Cook et al. (2004)* for the West, in this case the grid points in the 25°-50°N. latitude, 95°-125°W. longitude box (cf. Fig. 3.7). *Herweijer et al. (2007)* also restricted their analyses to a subset of 106 grid points within this domain with reconstructions available since A.D. 1000. This restriction had no appreciable effect on their results (see also *Cook et al., 2004*). *Herweijer et al. (2007)* compared the average PDSI over the 106 grid points for two distinct periods: A.D. 1000-1470 and 1470-2003. Even without any further analyses, it was clear that the earlier period, especially before 1300, was distinctly more drought-prone than the later period. Of particular interest was the fact that the range of annual drought variability during the MCA was not any larger than that seen after 1470. So, the climate conditions responsible for droughts each year during the MCA were apparently no more extreme than those conditions responsible for droughts during more recent times. This can be appreciated by noting that only 1 year of drought during the MCA was marginally more severe than the 1934 Dust Bowl year. This suggests that the 1934 event may be used as a worst-case scenario for how bad a given year of drought can get over the West.

So what differentiates MCA droughts from modern droughts? As shown by *Herweijer et al. (2007)*, the answer is **duration**. Droughts during the MCA lasted much longer, and it is this characteristic that most clearly differentiates megadroughts from ordinary droughts in the Western United States *Herweijer et al. (2007)* identified four megadroughts during the MCA — A.D. 1021-1051, 1130-1170, 1240-1265, and 1360-1382 — that lasted 31, 41, 26, and 23 years, respectively. In contrast, the four worst droughts in the historic period — A.D. 1855-1865, 1889-1896, 1931-1940, and 1950-1957 — lasted only 11, 8, 9, and 8 years, respectively. The difference in duration is striking.

The research conducted by *Cook et al. (2004), Herweijer et al. (2006, 2007),* and *Stahle et al. (2007)* was based on the first version of NADA. Since its creation in 2004, great improvements have been made in the tree-ring network used for drought reconstruction with respect to the total number of chronologies available for use in the original NADA (up from 835 to 1825) and especially the number extending back into the MCA (from 89 to 195 beginning before A.D. 1300). In addition, better geographic coverage during the MCA was also achieved, especially in the Northwest and the Rocky Mountain States of Colorado and New Mexico. Consequently, it is worth revisiting the results of *Cook et al. (2004)* and *Herweijer et al. (2007)*.

Figure 3.9 shows the updated NADA results now divided geographically into Northwest (Fig. 3.9A), Southwest (Fig. 3.9B), and the Great Plains (Fig. 3.9C). See Figure 3.7 for the sub-areas of the overall drought grid that define these three regions. Unlike the drought area index series shown in Figure 3.8, where more positive values indicate large areas affected by drought, the series shown in Figure 3.9 are simple regional averages of reconstructed PDSI. Thus, greater drought is indicated by more negative values in accordance with the original PDSI scale of *Palmer (1965)*. When viewed now in greater geographic detail, the intensity of drought during the MCA is focused more clearly toward the Southwest, with the Northwest much less affected. This geographic shift in emphasis toward the Southwest during the MCA aridity period is into the region where drought is more directly associated with forcing from the tropical oceans (*Cole et al., 2002; Seager et al., 2005b; Herweijer et al., 2006, 2007*).

Aside from the shift of geographic emphasis in the West during the MCA, the updated version of NADA still indicates the occurrence of multidecadal megadroughts that mostly agree with those of *Herweijer et al. (2007)* and the overall period of elevated aridity as described by *Cook et al. (2004)*. From Figure 3.9B, two of those megadroughts stand out especially strong in the Southwest: A.D. 1130-1158 (~29 years) and 1270-1297 (~28 years). The latter is the "Great Drouth" documented by *A.E. Douglass (1929, 1935)* for its association with the abandonment of Anasazi dwellings in the Southwest. Another prolonged drought in A.D. 1434-1481 (~48 years) is also noteworthy. *Herweijer et al. (2007)* did not mention it because it falls after the generally accepted end of the MCA. This megadrought is the same as the "15th century megadrought" described by *Stahle et al. (2007)*.

3.4 Possible Causes of the Medieval Megadroughts

The causes of the Medieval megadroughts are now becoming unraveled and appear to have similar origin to the causes of modern droughts, which is consistent with the similar spatial patterns of Medieval and modern droughts (*Herweijer et al., 2007*). *Cobb et al. (2003)* have used modern and fossil coral records from Palmyra, a small island in the tropical Pacific Ocean, to reconstruct eastern and central equatorial Pacific SSTs for three time segments within the Medieval period. These results indicate that colder—La Niña-like—conditions prevailed, which would be expected to induce drought over western North America. *Graham et al. (2007)* used these records, and additional sediment records in the west Pacific, to create an idealized pattern of Medieval tropical Pacific SST which, when it was used to force an AGCM, did create a drought over the Southwest. Adopting a different approach, *Seager et al. (2007a)* used the Palmyra modern and fossil coral records to reconstruct annual tropical Pacific SSTs for the entire period of 1320 to 1462 A.D. and forced an AGCM with this record. They found that the overall colder tropical Pacific implied by the coral records forced drying over North America with a pattern and amplitude comparable to that inferred from tree ring records, including for two megadroughts (1360-1400 A.D. and 1430-1460 A.D.). Discrepancies between model and observations can be explained through the combined effect of potential errors in the tropical Pacific SST reconstruction role for SST anomalies from other oceans, other unaccounted external forcings, and climate model deficiencies.

The modeling work suggests that the Medieval megadroughts were driven, at least in part, by tropical Pacific SST patterns in a way that is familiar from studies of the modern droughts. Analyses of the global pattern of Medieval hydroclimate also suggest that it was associated with a La Niña-like state in combination with a warm subtropical North Atlantic and a positive North Atlantic Oscillation (*Seager et al., 2007b; Herweijer et al., 2007*). For example, *Haug et al. (2001)* used the sedimentary record from the Cariaco basin in the Caribbean Sea to argue that northern South America experienced several wet centuries during the Medieval period, which is consistent with a La Nina-like Pacific Ocean. As another example, *Sinha et al. (2007)* used a speleothem (a secondary mineral deposit formed in a cave) record from India to show that at the same time the Indian monsoon was generally strong, especially compared to the subsequent Little Ice Age.

It has been suggested that the tropical Pacific adopted a more La Niña-like mean state during the Medieval period, relative to subsequent centuries, as a response to a relatively strong Sun and weaker volcanic activity (*Mann et al., 2005; Emile-Geay et al., 2007*; see also *Adams et al., 2003*). This follows because a positive radiative forcing warms the western equatorial Pacific by more than the east because in the latter region strong upwelling and ocean heat divergence transports a portion of the absorbed heat toward the subtropics. The stronger east-west gradient then strengthens the Walker Circulation, increasing the thermocline tilt and upwelling in the east such that actual cooling can be induced.

Further support for positive radiative forcing over the tropical Pacific Ocean inducing La Niña-like SSTs and drought over the Southwest comes from analyses of the entire Holocene recorded in a New Mexico speleothem, which shows a clear association between increased solar irradiance (as deduced from the atmospheric ^{14}C content recorded in ice cores) and dry conditions (*Asmerom et al., 2007*). However, the theory for the positive radiative forcing-La Niña link rests on experiments with intermediate complexity models (*Clement et al., 1996, 2000; Cane et al., 1997*). In contrast, the coupled GCMs used in the IPCC process do not, however, respond in this way to rising greenhouse gases and may actually slow the Walker Circulation (*Vecchi et al., 2006*). This apparent discrepancy could arise because the tropical response to changes in solar irradiance is different from the response to rising greenhouse gases or it could be that the coupled GCMs respond incorrectly due to the many errors in simulations of the tropical Pacific mean climate, not the least of which is the notorious double-intertropical convergence zone (ITCZ) problem.

3.5 Megadroughts in the Great Plains and U.S. “Breadbasket”

The emphasis up to now has been on the semi-arid to arid Western United States because that is where the late-20th century drought began and has largely persisted up to the present time. The present drought has therefore largely missed the important crop producing States in the Midwest and Great Plains. Yet, previous studies (*Laird et al., 1996; Woodhouse and Overpeck, 1998; Stahle et al., 2000, 2007*) indicate that megadroughts have also occurred in those regions as well. To illustrate this, we have used the updated NADA to produce an average PDSI series for the Great Plains rectangle indicated in Figure 3.7. That series is shown in Figure 3.9C and it is far more provocative than even the Southwest series. The MCA period shows even more persistent drought, now on the centennial time scale, and the 15th century megadrought stands out more strongly as well. The duration of the MCA megadrought in our record is highly consistent with the salinity record from Moon Lake in North Dakota that likewise shows centennial time scale drought around that time. More ominously, in comparison, the 20th century has been a period of relatively low hydroclimatic variability, with the 1930s Dust Bowl and 1950s southern Great Plains droughts being rather unexceptional when viewed from a paleoclimate perspective. The closest historical analog to the extreme past megadroughts is the Civil War drought (*Herweijer et al., 2006*) from 1855 to 1865 (11 years), followed closely by a multiyear drought in the 1870s. Clearly, there is a great need to understand the causes of long-term drought variability in the Great Plains and the U.S. "Breadbasket" to see how the remarkable past megadroughts indicated in Figure 3.9C developed and persisted. That these causes may be more complicated than those identified with the tropical oceans is suggested by the work of *Fye et al. (2006)*, who found that drought variability in the Mississippi River valley is significantly coupled to variations in the NAO (see also Sec. 2.2).

3.6 Drought in the Eastern United States

Up to this point, the emphasis on drought over the past 2,000 years has been restricted to the Western United States and Great Plains. This choice was intentional because of the current multiyear drought affecting the West, historic droughts of remarkable severity that have struck there (e.g., the 1930s "Dust Bowl" drought), and that region's susceptibility to multi-decadal megadroughts based on the tree-ring evidence (*Herweijer et al., 2007*). Even so, the normally well-watered Eastern United States is also vulnerable to severe droughts, both historically (*Hoyt, 1936; Namias, 1966; Karl and Young, 1987; Manuel, 2008*) and in tree-ring records (*Cook and Jacoby, 1977; Cook et al., 1988; Stahle et al., 1998*), but they have tended to be much shorter in duration compared to those in the West. Does this mean that the Eastern United States has not experienced megadroughts of similar duration as those during the MCA in the West? Evidence from high-resolution sediment core samples from the lower Hudson Valley in New York

(*Pederson et al., 2005*) suggest that there was indeed a period of prolonged dryness there centered around the MCA. *Stahle et al. (1988)* also found tree-ring evidence for unusually persistent drought in North Carolina again during the MCA, as did *Seager et al. (2008)* for the greater Southeast based on the updated NADA. So it appears that megadroughts have also occurred in the Eastern United States, especially during the MCA. The cause of these extended-duration droughts in the Eastern United States during the MCA is presently not well understood.

4. Abrupt Hydrologic Changes During the Holocene

Examination of abrupt climate change during the Holocene (i.e., prior to the beginning of the instrumental or dendroclimatological records) can be motivated by the observation that the projected changes in both the radiative forcing and the resulting climate of the 21st century far exceed those registered by the either the instrumental records of the past century or by the proxy records of the past few millennia (*Jansen et al., 2007; Hegerl et al., 2003, 2007; Jones and Mann, 2004*). In other words, all of the variations in climate over the instrumental period and over the past millennium reviewed above have occurred in a climate system whose controls have not differed much from those of the most of the 20th century. In particular, variations in global-averaged radiative forcing as described in the IPCC Fourth Assessment (*IPCC, 2007*) include:

- values of roughly ±0.5 watts per meter squared (W m^{-2}) (relative to a 1500 to 1899 mean) related to variations in volcanic aerosol loadings and inferred changes in solar irradiance, i.e., from natural sources (*Jansen et al., 2007, Fig. 6.13*);
- total anthropogenic radiative forcing of about 1.75 W m^{-2} from 1750 to 2005 from long-lived greenhouse gases, land-cover change, and aerosols (*Forster et al., 2007, Fig. 2.20b*);
- projected increases in anthropogenic radiative forcing from 2000 to 2100 of around 6 W m^{-2} (*Meehl et al., 2007, Fig. 10.2*).

In the early Holocene, annual-average insolation forcing anomalies (at 8 ka relative to present) range from -1.5 W m^{-2} at the equator to over +5 W m^{-2} at high latitudes in both hemispheres, with July insolation anomalies around +20 W m^{-2} in the mid-latitudes of the

Northern Hemisphere (*Berger, 1978; Berger and Loutre, 1991*). Top-of-the-atmosphere insolation is not directly comparable with the concept of radiative forcing as used in the IPCC Fourth Assessment (*Committee on Radiative Forcing Effects on Climate, 2005*), owing to feedback from the land surface and atmosphere, but the relative size of the anomalies supports the idea that potential future changes in the controls of climate exceed those observed over the past millennium (*Joos and Sphani, 2008*). Consequently, a longer term focus is required to describe the behavior of the climate system under controls as different from those at present as those of the 21st century will be, and to assess the potential for abrupt climate changes to occur in response to gradual changes in large-scale forcing.

The controls of climate during the 21st century and during the Holocene differ from one another, and from those of the 20th century, in important ways. The major contrast in controls of climate between the early 20th, late 20th, and 21st century are in atmospheric composition (with an additional component of land-cover change), while the major contrast between the controls in the 20th century and those in the early to middle Holocene were in the latitudinal and seasonal distribution of insolation. In the Northern Hemisphere in the early Holocene, summer insolation was around 8% greater than present, and winter about 8% less than present, related to the amplification of the seasonal cycle of insolation due to the occurrence of perihelion in summer then, while in the Southern Hemisphere the amplitude of the seasonal cycle of insolation was reduced (*Webb et al., 1993b*). In both hemispheres in the early Holocene, annual insolation was greater than present poleward of 45°, and less than present between 45°N. and 45°S., related to the greater tilt of Earth's axis than relative to today. The energy balance of the Northern Hemisphere during the early Holocene thus features a large increase in seasonality relative to that of the 20th century. This contrast between the past and future will increase throughout the 21st century owing to the ongoing and projected further reduction in snow and ice cover in the Northern Hemisphere winter.

Consequently, climatic variations during the Holocene should not be thought of either as analogs for future climates or as examples of what might be observable under present-day climate forcing if records were longer, but instead should be thought of as a "natural experiment" (i.e., an experiment not purposefully performed by humans) with the climate system that features large perturbations of the controls of climate, similar in scope (but not in detail) to those expectable in the future. In particular, the climates of both the Holocene and the 21st century illustrate the response of the climate system to significant perturbations of radiative forcing relative to that of the 20th or 21st century.

4.1 Examples of Large and Rapid Hydrologic Changes During the Holocene

From the perspective of the present and with a focus on the northern mid-latitudes, the striking spatial feature of Holocene climate variations was the wastage and final disappearance of the middle- to high-latitude North American and Eurasian ice sheets. However, over the much larger area of the tropics and adjacent subtropics, there were equally impressive hydrologic changes, ultimately related to insolation-driven variations in the global monsoon (*COHMAP Members, 1988; Liu et al., 2004*). Two continental-scale hydrologic changes that featured abrupt (on a Holocene time scale) transitions between humid and arid conditions were those in northern Africa and in the mid-continent of North America. In northern Africa, the "African humid period" began after 12 ka with an intensification of the African-Asian monsoon, and ended around 5 ka (*deMenocal et al., 2000; Garcin et al., 2007*), with the marked transition from a "green" (vegetated) Sahara, to the current "brown" (or sparsely vegetated) state. This latter transition provides an example of a climate change that would have significant societal impact if it were to occur today in any region, and provides an example of an abrupt transition to drought driven by gradual changes in large-scale external controls.

In North America, drier conditions than those at present commenced in the mid-continent between 10 and 8 ka (*Thompson et al., 1993; Webb et al., 1993a; Forman et al., 2001*), and ended after 4 ka. This "North American mid-continental Holocene drought" was coeval with dry conditions in the Pacific Northwest, and wet conditions in the south and southwest, in manner consistent (in a dynamic atmospheric circulation sense) with the amplification of the monsoon then (*Harrison et al., 2003*). The mid-Holocene drought in mid-continental North America gave way to wetter conditions after 4 ka, and like the African humid period, provides an example of major, and sometimes abrupt hydrological changes that occurred in response to large and gradual changes in the controls of regional climates.

These continental-scale hydrologic changes obviously differ in the sign of the change (wet to dry from the middle Holocene to present in Africa and dry to wet from the middle Holocene to present in North America), and in the specific timing and spatial coherence of the hydrologic changes, but they have several features in common, including:

- the initiation of the African humid period and the North American Holocene drought were both related to regional climate changes that occurred in response to general deglaciation and to variations in insolation;
- the end of the African humid period and the North American Holocene drought were both ultimately related to the gradual decrease in Northern Hemisphere summer insolation during the Holocene, and to the response of the global monsoon;
- paleoclimatic simulations suggest that ocean-atmosphere coupling played a role in determining the moisture status of these regions, as it has during the 20th century and the past millennium;
- feedback from local land-surface (vegetation) responses to remote (sea-surface temperature, ocean-atmosphere interaction) and global (insolation, global ice volume, atmospheric composition) forcing may have played a role in the magnitude and rapidity of the hydrological changes.

Our understanding of the scope of the hydrologic changes and their potential explanations for both of these regions have been informed by interactions between paleoclimatic data syntheses and climate-model simulations (e.g., *Wright et al., 1993; Harrison et al., 2003; Liu et al., 2007,* see box 3.3). In this interaction, the data syntheses have driven the elaboration of both models and experimental designs, which in turn have led to better explanations of the patterns observed in the data (see *Bartlein and Hostetler, 2004*).

Box 3.3—Paleoclimatic Data/Model Comparisons

Two general approaches and information sources for studying past climates have been developed. *Paleoclimatic observations* (also known as proxy data) consist of paleoecological, geological, and geochemical data, that when assigned ages by various means, can be interpreted in climatic terms. Paleoclimatic data provide the basic documentation of what has happened in the past, and can be synthesized to reconstruct the patterns history of paleoclimatic variations. *Paleoclimatic simulations* are created by identifying the configuration of large-scale controls of climate (i.e. solar radiation, and its latitudinal and seasonal distribution, or the concentration of greenhouse gases in the atmosphere) at a particular time in the past, and then supplying these to a global or regional climate model to generate sequences of simulated meteorological data, in a fashion similar to the use of a numerical weather forecasting model today. (See CCSP SAP 3.1 for a discussion on climate models.) Both approaches are necessary for understanding past climatic variations—the paleoclimatic observations document past climatic variations but can't explain them without some kind of model, and the models that could provide such explanations must first be tested and shown to be capable of simulating the patterns in the data.

The two approaches are combined in *paleoclimatic data/model comparison* studies, in which syntheses of paleoclimatic data from different sources and suites of climate-model simulations performed with different models are combined in an attempt to replicate a past "natural experiment" with the real climate system, such as those provided by the regular changes in incoming solar radiation related to Earth's orbital variations. Previous generations of data/model comparison studies have focused on key times in the paleoclimatic record, such as the Last Glacial Maximum (21,000 years ago) or mid-Holocene (6000 years ago), but attention is now turning to the study of paleoclimatic variability as recorded in high-resolution time series of paleoclimatic data and generated by long "transient" simulations with models.

Paleoclimatic data/model comparisons contribute to our overall perspective on climate change, and can provide critically needed information on how realistically climate models can simulate climate variability and change, what the role of feedbacks in the climate system are in amplifying or damping changes in the external controls of climate, and the general causes and mechanisms involved in climate change.

4.2 The African Humid Period

One of the major environmental variations over the past 10,000 years, measured in terms of the area affected, the magnitude of the overall climatic changes and their rapidity, was the reduction in magnitude around 5,000 years ago of the African-Asian monsoon from its early to middle Holocene maximum, and the consequent reduction in vegetation cover and expansion of deserts, particularly in Africa south of the Saraha. The broad regional extent of enhanced early Holocene monsoons is revealed by the status of lake levels across Africa and Asia (Fig. 3.10), and the relative wetness of the interval is further attested to by similarly broad-scale vegetation changes (*Jolly et al., 1998; Kohfeld and Harrison, 2000*). Elsewhere in the region influenced by the African-Asian monsoon, the interval of enhanced monsoonal circulation and precipitation also ended abruptly, in the interval between 5.0 and 4.5 ka across south and east Asia (*Morrill et al., 2003*), demonstrating that the African humid period was embedded in planetary-scale climatic variations during the Holocene.

A general conceptual model has emerged (see *Ruddiman, 2006*) that relates the intensification of the monsoons to the differential heating of the continents and oceans that occurs in response to orbitally induced amplification of the seasonal cycle of insolation (i.e., increased summer and decreased winter insolation in the Northern Hemisphere) (*Kutzbach and Otto-Bliesner, 1982; Kutzbach and Street-Perrott, 1985; Liu et al., 2004*). In addition to the first-order response of the monsoons to insolation forcing, other major controls of regional climates, like the atmospheric circulation variations related to the North American ice sheets, to ocean/atmospheric circulation reorganization over the North Atlantic (*Kutzbach and Ruddiman, 1993; Weldeab et al., 2007*), and to tropical Pacific ocean/atmosphere interactions (*Shin et al., 2006; Zhao et al., 2007*) likely also played a role in determining the timing and details of the response. In many paleoenvironmental records, the African humid period (12 ka to 5 ka) began rather abruptly (relative to the insolation forcing), but with some spatial variability in its expression (*Garcin et al., 2007*), and similarly, it ended abruptly (*deMenocal et al., 2000*; and see the discussion in *Liu et al., 2007*).

The robust expression of the wet conditions (Fig. 3.10) together with the amplitude of the "signal" in the paleoenvironmental data has made the African humid period a prime focus for synthesis of paleoenvironmental data, climate-model simulations, and the systematic comparison of the two (*COHMAP Members, 1988*), in particular as a component of the Palaeoclimatic Modeling Intercomparison Project (PMIP and PMIP 2; *Joussaume et al., 1999; Crucifix et al., 2005; Braconnot et al., 2007a,b*). The aim of these paleoclimatic data-model comparisons is twofold: (1) to "validate" the climate models by examining their ability to correctly reproduce an observed environmental change for which the ultimate controls are known and (2) to use the mechanistic aspects of the models and simulations produced with them to explain the patterns and variations recorded by the data. Mismatches between the simulations and observations can arise from one or more sources, including inadequacies of the climate models, misinterpretation of the paleoenvironmental data, and incompleteness of the experimental design (i.e., failure to include one or more controls or processes that influenced the real climate) (*Peteet, 2001; Bartlein and Hostetler, 2004*).

In general, the simulations done as part of PMIP, as well as others, show a clear amplification of the African-Asian monsoon during the early and middle part of the Holocene, but one that is insufficient to completely explain the magnitude of the changes in lake status, and the extent of the observed northward displacement of the vegetation zones into the region now occupied by desert (*Joussaume et al., 1999; Kohfeld and Harrison, 2000*). The initial PMIP simulations were "snapshot" or "time-slice" simulations of the conditions around 6 ka, and as a consequence are able to only indirectly comment on the mechanisms involved in the abrupt beginning and end of the humid period. In addition, the earlier simulations were performed using AGCMs, with present-day land-surface characteristics, which therefore did not adequately represent the full influence of the ocean or terrestrial vegetation on the simulated climate.

As a consequence, climate-simulation exercises that focus on the African monsoon or the African humid period have evolved over the past decade or so toward models and experimental designs that (1) include interactive coupling among the atmosphere, ocean, and terrestrial biosphere and (2) feature transient, or time-evolving simulations that, for example, allow explicit examination of the timing and rate of the transition from a green to a brown Sahara. Two classes of models have been used, including (1) Atmosphere Ocean General Circulation Models with interactive oceans (AOGCMs), Atmosphere terrestrial Vegetation General Circulation Models (AVGCMs), or both (AOVGCMs) that typically have spatial resolutions of a few degrees of latitude and longitude and (2) coarser resolution EMICs, or Earth-system models of intermediate complexity, that include representation of components of the climate system that are not amenable to simulation with the higher resolution GCMs (See *Claussen, 2001*, and *Bartlein and Hostetler, 2004*, for a discussion of the taxonomy of climate models.)

The coupled AOGCM simulations have illuminated the role that sea surface temperatures likely played in the amplification of the monsoon. Driven by both the insolation forcing and by ocean-atmosphere interactions, the picture emerges of a role for the oceans in modulating the amplified seasonal cycle of insolation during the early and mid-Holocene in such a way as to increase the summertime temperature contrast between continent and ocean that drives the monsoon, thereby strengthening it (*Kutzbach and Liu, 1997; Zhao et al., 2005*). In addition, there is an apparent role for teleconnections from the tropical Pacific in determining the strength of the monsoon, in a manner similar to the "atmospheric bridge" teleconnection between the tropical Pacific ocean and climate elsewhere at present (*Shin et al., 2006; Zhao et al., 2007; Liu and Alexander, 2007*).

The observation of the dramatic vegetation change motivated the development of simulations with coupled vegetation components, first by asynchronously coupling equilibrium global vegetation models (EGVMs, *Texier et al., 1997*), and subsequently by using fully coupled AOVGCMs (e.g., *Levis et al., 2004; Wohlfahrt et al., 2004; Gallimore et al., 2005; Braconnot et al., 2007a,b; Liu et al., 2007*). These simulations, which also included investigation of the synergistic effects of an interactive ocean and vegetation on the simulated climate (*Wohlfahrt et al., 2004*), produced results that still underrepresented the magnitude of monsoon enhancement, but to a lesser extent than the earlier AGCM or AOGCM simulations. These simulations also suggest the specific mechanisms through which the vegetation and the related soil-moisture conditions (*Levis et al., 2004; Liu et al., 2007*) influence the simulated monsoon.

The EMIC simulations, run as transient or continuous (as opposed to time-slice) simulations over the Holocene, are able to explicitly reveal the time history of the monsoon intensification or deintensification, including the regional-scale responses of surface climate and vegetation (*Claussen et al., 1999; Hales et al., 2006; Renssen et al., 2006*). These simulations typically show abrupt decreases in vegetation cover, and usually also in precipitation, around the time of the observed vegetation change (5 ka), when insolation was changing only gradually. The initial success of EMICs in simulating an abrupt climate and land-cover change in response to a gradual change in forcing influenced the development of a conceptual model that proposed that strong nonlinear feedbacks between the land surface and atmosphere were responsible for the abruptness of the climate change, and, moreover, suggested the existence of multiple stable states of the coupled climate-vegetation-soil system that are maintained by positive vegetation feedback (*Claussen et al, 1999; Foley et al., 2003*). In such a system, abrupt transitions from one state to another (e.g., from a green Sahara to a brown one), could occur under relatively modest changes in external forcing, with a green vegetation state and wet conditions reinforcing one another, and likewise a brown state reinforcing dry conditions and vice versa. The positive feedback involved in maintaining the green or brown states would also promote the conversion of large areas from one state to the other at the same time.

A different perspective on the way in which abrupt changes in the land-surface cover of west Africa may occur in response to gradual insolation changes is provided by the simulations by *Liu et al. (2006, 2007)*. They used a coupled AOVGCM (FOAM-LPJ) run in transient mode to produce a continuous simulation from 6.5 ka to present. They combined a statistical analysis of vegetation-climate feedback in the AOVGCM, and an analysis of a simple conceptual model that relates a simple two-state depiction of vegetation to annual precipitation (*Liu et al., 2006*), and argue that the short-term (i.e.

year-to-year) feedback between vegetation and climate is negative (see also *Wang et al., 2007; Notaro et al., 2008*), such that a sparsely or unvegetated state (i.e., a brown Sahara) would tend to favor precipitation through the recycling of moisture from bare-ground evapotranspiration. In this view, the negative vegetation feedback would act to maintain the green Sahara against the general drying trend related to the decrease in the intensity of the monsoon and amount of precipitation, until such time that interannual variability results in the crossing of a moisture threshold beyond which the green state could no longer be maintained (see *Cook et al., 2006*, for further discussion of this kind of behavior in response to interannual climate variability (i.e., ENSO). In this conceptual model, the transition between states, while broadly synchronous (owing to the large-scale forcing), might be expected to show a more time-transgressive or diachronous pattern owing to the influence of landscape (soil and vegetation) heterogeneity.

These two conceptual models of the mechanisms that underlie the abrupt vegetation change—strong feedback and interannual variability/threshold crossing—are not that different in terms of their implications, however. Both conceptual models relate the overall decrease in moisture and consequent vegetation change to the response of the monsoon to the gradually weakening amplification of the seasonal cycle of insolation, and both claim a role for vegetation in contributing to the abruptness of the land-cover change, either explicitly or implicitly invoking the nonlinear relationship between vegetation cover and precipitation (Fig. 3.11 from *Liu et al., 2007*). The conceptual models differ mainly in their depiction of the precipitation change, with the strong-feedback explanation predicting that abrupt changes in precipitation will accompany the abrupt changes in vegetation, while the interannual variability/threshold crossing explanation does not. It is interesting to note that the *Renssen et al. (2006)* EMIC simulation generates precipitation variations for west Africa that show much less of an abrupt change around 5 ka than did earlier EMIC simulations, which suggests that the strong-feedback perspective may be somewhat model dependent. A recent analysis of a paleolimnological record from the eastern Sahara (*Kröpelin et al., 2008*) shows a more gradual transition from the green to brown state than would be inferred from the marine record of dust flux, which also supports the variability/threshold crossing model.

There is thus some uncertainty in the specific mechanisms that link the vegetation response to climate variations on different time scales, and also considerable temporal-spatial variability in the timing of environmental changes. However, the African humid period and its rapid termination illustrates how abrupt, widespread, and significant environmental changes can occur in response to gradual changes in a large-scale or ultimate control—in this case the amplification of the seasonal cycle of insolation in the Northern Hemisphere and its impact on radiative forcing.

4.3 North American Mid-Continental Holocene Drought

At roughly the same time as the African humid period, large parts of North America experienced drier-than-present conditions that were sufficient in magnitude to be registered in a variety of paleoenvironmental data sources. Although opposite in sign from those in Africa, these moisture anomalies were ultimately related to the same large-scale control - greater-than-present summer insolation in the Northern Hemisphere. In North America, however, the climate changes were also strongly influenced by the shrinking (but still important regionally) Laurentide Ice Sheet. In contrast to the situation in Africa, and likely related to the existence of additional large-scale controls (e.g., the remnant ice sheet, and Pacific ocean-atmosphere interactions), the onset and end of the middle Holocene moisture anomaly was more spatially variable in its expression, but like the African humid period, it included large-scale changes in land cover in addition to effective-moisture variations. Also in contrast to the African situation, the vegetation changes featured changes in the type of vegetation or biomes (e.g., shifts between grassland and forest, *Williams et al., 2004*), as opposed to fluctuations between vegetated and nonvegetated or sparsely vegetated states. There are also indications that, as in Africa and Asia, the North American monsoon was amplified in the early and middle Holocene (*Thompson et al., 1993; Mock and Brunelle-Daines, 1999; Poore et al., 2005*), although as in the case of the dry conditions, there probably was significant temporal and spatial variation in the strength of the enhanced monsoon (*Barron et al., 2005*). The modern association of dry conditions across central North America and somewhat wetter conditions in North Africa during a La Niña phase (*Palmer and Brankovic, 1989*), led *Forman et al. (2001)* to hypothesize that changes in tropical sea surface variability, in particular the persistence of La Niña-type conditions (generally colder and warmer than those at present in the eastern and western parts of the basin, respectively), might have played an important role in modulating the regional impacts of mid-Holocene climate.

A variety of paleoenvironmental indicators reflect the spatial extent and timing of these moisture variations (Figs. 3.12 and 3.13), and in general suggest that the dry conditions increased in their intensity during the interval from 11 ka to 8 ka, and then gave way to increased moisture after 4 ka, and during the middle of this interval (around 6 ka) were widespread. Lake-status indicators at 6 ka indicate lower-than-present levels (and hence drier-than-present conditions) across much of the continent (*Shuman et al., in review*), and quantitative interpretation of the pollen data in *Williams et al. (2004)* shows a similar pattern of overall aridity, but again with some regional and local variability, such as moister-than-present conditions in the Southwestern United States (see also *Thompson et al., 1993*). Although the region of drier-than-present conditions extends into the Northeastern United States and eastern Canada, most of the multiproxy evidence for middle Holocene dryness is focused on the midcontinent, in particular the Great Plains and Midwest, where the evidence for aridity is particularly clear. There, the expression of middle Holocene dry conditions in paleoenvironmental records has long been known, as was the case for the "Prairie Period" evident in fossil-pollen data (see *Webb et al., 1983*), and the recognition of significant aeolian activity (dune formation) on the Great Plains (*Forman et al., 2001; Harrison et al., 2003*) that would be favored by a decrease in vegetation cover.

Temporal variations in the large-scale controls of North American regional climates as well as some of the paleoenvironmental indicators of the moisture changes are shown in Figure 3.13. In addition to insolation forcing (Fig. 3.13A,B), the size of the Laurentide Ice Sheet was a major control of regional climates, and while diminished in size from its full extent at the Last Glacial Maximum (21 ka), the residual ice sheets at 11 ka and 9 ka (Fig. 3.13C) still influence atmospheric circulation over eastern and central North America in climate simulations for those times (*Bartlein et al., 1998; Webb et al., 1998*). In addition to depressing temperatures generally around the Northern Hemisphere, the ice sheets also directly influenced adjacent regions. In those simulations, the development of a "glacial anticyclone" over the ice sheet (while not as pronounced as earlier), acted to diminish the flow of moisture from the Gulf of Mexico into the interior, thus keeping the midcontinent cooler and drier than it would have been in the absence of an ice sheet.

Superimposed on these "orbital time scale" variations in controls and regional responses are millennial-scale variations in atmospheric circulation related to changes in the Atlantic Meriodional Overturning Circulation (AMOC) and to other ocean-atmosphere variability (*Shuman et al., 2005, 2007; Viau et al., 2006*). Of these millennial-scale variations, the "8.2 ka event" (Fig. 3.13D) is of interest, inasmuch as the climate changes associated with the "collapse" of the Laurentide Ice Sheet (*Barber et al., 1999*) have the potential to influence the mid-continent region directly, through regional atmospheric circulation changes (*Dean et al., 2002; Shuman et al., 2002*), as well as indirectly, through its influence on AMOC, and related hemispheric atmospheric circulation changes.

The record of aridity indicators for the midcontinent reveals a more complicated history of moisture variations than does the African case, with some locations remaining dry until the late Holocene, and others reaching maximum aridity during the interval between 8 ka and 4 ka, but in general showing relatively dry conditions between 8 ka and 4 ka. Lake-status records (Fig. 3.13E, *Shuman et al., in review*) show the highest frequency of lakes at relatively low levels during the interval between 8 ka and 4 ka, and a higher frequency of lakes at relatively high levels before and after that interval. Records of widespread and persistent aeolian activity and loess deposition (dust transport) increase in frequency from 10 ka to 8 ka, and then gradually fall to lower frequency in the late Holocene, with a noticeable decline between 5 ka and 4 ka. Pollen records of the vegetation changes that reflect dry conditions (Fig. 3.13G; *Williams, 2002; Williams et al., 2004*) show a somewhat earlier onset of dryness than do the aeolian or lake indicators, reaching maximum frequency around 9 ka. Increased aeolian activity can also be noted during the last 2000 years (Fig. 3.13F, *Forman et al., 2001; Miao et al., 2007*), but was less pronounced than during the mid-Holocene.

The pollen record from Steel Lake, MN, expressed in terms of tree-cover percentages (see *Williams, 2002*, for methods) provides an example to illustrate a pattern of moisture-related vegetation change that is typical at many sites in the Midwest, with an abrupt decline in tree cover at this site around 8 ka, and over an interval equal to or less than the sampling resolution of the record (about 200 years, Fig. 3.13H). This decrease in tree cover and inferred moisture levels is followed by relatively low but slightly increasing inferred moisture levels for about 4,000 years, with higher inferred moisture levels in the last 4,000 years. The magnitude of this moisture anomaly can be statistically inferred from the fossil-pollen data using modern relationships between pollen abundance and climate, as was done for the pollen record at Elk Lake, MN, which is near Steel Lake (Fig. 3.13I; *Bartlein and Whitlock, 1993*; see also *Webb et al., 1998*). Expressed in terms of precipitation, the moisture decrease in the midcontinent needed for these vegetation changes is about 350 millimeters per year (mm y^{-1}), or about 1 millimeter per day (mm d^{-1}), or levels between 50 and 80 percent of the present-day values.

As is the case for the African humid period, the effective-moisture variations recorded by paleoenvironmental data from the midcontinent of North America provide a target for simulation by climate models, and also as was the case for Africa, those simulations have evolved over time toward models with increased coupling among systems. The first generation of simulations with AGCMs featured models that were of relatively coarse spatial resolution, had fixed SSTs, and land cover that was specified to match that of the modern day. These simulations, focusing on 6 ka, revealed some likely mechanisms for developing dry conditions in the midcontinent, such as the impact of the insolation forcing on surface energy and water balances and the direct and indirect effects of insolation on atmospheric circulation (*Webb et al., 1993b; Bartlein et al., 1998; Webb et al., 1998*). However, the specific simulations of precipitation or precipitation minus evapotranspiration (P-E) indicated little change in moisture or even increases in some regions. Given the close link between SST variations and drought across North America at present, and the inability of these early simulations to simulate such mechanisms because they had fixed SSTs, this result is not surprising.

What can be regarded as the current-generation simulations for 6 ka include those done with fully coupled AOGCMs (FOAM and CSM 1, *Harrison et al., 2003*; CCSM 3, *Otto-Bliesner et al., 2006*), and an AGCM with a mixed-layer ocean (CCM 3.10, *Shin et al.,* *2006*). These simulations thereby allow the influence of SST variations to be registered in the simulated climate either implicitly, by calculating them in the ocean component of the models (FOAM, CSM 1, CCSM 3), or explicitly, by imposing them either as present-day long-term averages, or as perturbations of those long-term averages intended to represent extreme states of, for example, ENSO (CCM 3.10). The trade-off between these approaches is that the fully coupled, implicit approach will reflect the impact of the large-scale controls of climate (e.g., insolation) on SST variability (if the model simulates the joint response of the atmosphere and ocean correctly), while the explicitly specified AGCM approach allows the response to a hypothetical state of the ocean to be judged.

These simulations produce generally dry conditions in the interior of North America during the growing season (and an enhancement of the North American monsoon), but as was the case for Africa, the magnitude of the moisture changes is not as large as that recorded by the paleoenvironmental data (with maximum precipitation-rate anomalies on the order of 0.5 mm d^{-1}, roughly half as large as it would need to be to match the paleoenvironmental observations). Despite this, the simulations reveal some specific mechanisms for generating the dry conditions; these include (1) atmospheric circulation responses to the insolation and SST forcing/feedback that favor a “package” of circulation anomalies that include expansion of the subtropical high-pressure systems in summer, (2) the development of an upper-level ridge and large-scale subsidence over central North America (a circulation feature that favors drought at the present), and (3) changes in surface energy and water balances that lead to reinforcement of this circulation configuration. Analyses of the 6 ka simulated and present-day “observed” (i.e., reanalysis data) circulation were used by *Harrison et al. (2003)* to describe the linkage that exists in between the uplift that occurs in the Southwestern United States and northern Mexico as part of the North American monsoon system, and subsidence on the Great Plains and Pacific Northwest (*Higgins et al., 1997*; see also *Vera et al., 2006*).

The summertime establishment of the upper-level ridge, the related subsidence over the middle of the North American continent, and the onshore flow and uplift in the Southwestern United States and northern Mexico are influenced to a large extent by the topography of western North America, which is greatly oversimplified in GCMs (see Fig.

4 in *Bartlein and Hostetler, 2004*). This potential "built-in" source of mismatch between the paleoclimatic simulations and observations can be reduced by simulating climate with regional climate models (RCMs). Summer (June, July, and August) precipitation and soil moisture simulated using RegCM3 (*Diffenbaugh et al., 2006*) is shown in Figure 3.14, which illustrates moisture anomalies that are more comparable in magnitude to those recorded by the paleoenvironmental data than are the GCM simulations. RegCM as applied in these simulations has a spatial resolution of 55 km, which resolves climatically important details of the topography of the Western United States. In these simulations, the "lateral boundary conditions" or inputs to the RCM, were supplied by a simulation using an AGCM (CAM 3), that in turn used the SSTs simulated by the fully coupled AOGCM simulation for 6 ka (and present) by *Otto-Bliesner et al. (2006)*. These SSTs were also supplied directly to RegCM3. The simulations thus reveal the impact of the insolation forcing, as well as the influence of the insolation-related changes on interannual variability in SSTs (over the 30 years of each simulation). The results clearly show the suppression of precipitation over the mid-continent and enhancement over the Southwestern U.S. and northern Mexico, and the contribution of the precipitation anomaly to that of soil moisture (Fig. 3.14). In contrast to the GCM simulations, the inclusion of 6 ka SST variability in the RCM simulations reduces slightly the magnitude of the moisture anomalies, but overall these anomalies are close to those inferred from paleoenvironmental observations and reinforce the conceptual model linking the North American mid-continental Holocene drought to increased subsidence (see also *Shinker et al., 2006; Harrison et al., 2003*).

The potential of vegetation feedback to amplify the middle Holocene drought has not been as intensively explored as it has for Africa, but those explorations suggest that it should not be discounted. *Shin et al. (2006)* prescribed some subjectively reconstructed vegetation changes (e.g., *Diffenbaugh and Sloan, 2002*) in their AGCM simulations and noted a reduction in spring and early summer precipitation (that could carry over into reduced soil moisture during the summer), but also noted a variable response in precipitation during the summer to the different vegetation specifications. *Wohlfahrt et al. (2004)* asynchronously coupled an equilibrium global vegetation model, Biome 4 (*Kaplan et al., 2003*), to an AOGCM and observed a larger expansion of grassland in those simulations than in ones without the vegetation change simulated by the EGVM. Finally, *Gallimore et al. (2005)* examined simulations using the fully coupled AOVGCM (FOAM-LPJ), and while the overall precipitation change for summer was weakly negative, the impact of the simulated vegetation change (toward reduced tree cover at 6 ka), produced a small positive precipitation change.

An analysis currently in progress with RegCM3 suggests that the inclusion of the observed middle Holocene vegetation in the boundary conditions for the 6 ka simulation described above (*Diffenbaugh et al., 2006*) further amplifies the negative summer precipitation anomaly in the core region of the Holocene drought, and also alters the nature of the seasonal cycle of the dependence of soil moisture on precipitation. The magnitude of the drought in these simulations is relatively close to that inferred from the paleoenvironmental data.

The North American midcontinental drought during the middle Holocene thus provides an illustration of a significant hydrologic anomaly with relatively abrupt onset and ending that occurred in response to gradual changes in the main driver of Holocene climate change (insolation), reinforced by regional- and continental-scale changes in atmospheric circulation related directly to deglaciation. As was the case for the African humid period, feedback from the vegetation change that accompanied the climate changes could be important in reinforcing or amplifying the climate change, and work is underway to evaluate that hypothesis.

There are other examples of abrupt hydrological responses to gradual or large-scale climatic changes during the Holocene. For example, the development of wetlands in the Northern Hemisphere began relatively early in the course of deglaciation but accelerated during the interval high summer insolation between 12 ka and 8 ka (*Gajewski et al., 2001; MacDonald et al., 2006*). The frequency and magnitude of floods across a range of different watershed sizes also tracks climate variations during the Holocene (Fig. 3.13J; *Knox 1993, 2000; Ely, 1997*), albeit in a complicated fashion, owing to dependence of flooding on long-term climate and land-cover conditions as well as on short-term meteorological events (see Sec. 6).

4.4 Century-Scale Hydrologic Variations

Hydrologic variations, many abrupt, occur on time scales intermediate between the variations over millennia that are ultimately related to orbitally governed insolation variations and the interannual- to decadal-scale variations documented by annual-resolution proxy records. A sample of time series that describe hydrologic variations on decadal-to-centennial scales over the past 2,000 years in North America appears in Figure 3.15 and reveals a range of different kinds of variation, including:

- generalized trends across several centuries (Fig. 3.15C,F,G);
- step-changes in level or variability (independent of sampling resolution) (Fig. 3.15A,B,F);
- distinct peaks in wet (Fig. 3.15A) or dry conditions (Fig. 3.13F, Fig. 3.15B,G);
- a tendency to remain persistently above or below a long term mean (Fig. 3.15C-F), often referred to as "regime changes"; and
- variations in all components of the hydrologic cycle, including precipitation, evaporation, storage, and runoff, and in water quality (e.g., salinity).

Hydrological records that extend over the length of the Holocene, in particular those from hydrologically sensitive speleothems, demonstrate similar patterns of variability throughout (e.g., *Asmerom et al., 2007*), including long-term trends related to the Holocene history of the global monsoon described above (e.g., *Wang et al., 2005*).

The ultimate controls of these variations include (1) the continued influence of the long-term changes in insolation that appear to be ultimately responsible for the mid-Holocene climate anomalies discussed above, (2) the integration of interannual variations in climate that arise from ocean-atmosphere coupling, and (3) the impact of the variations in volcanism, solar irradiance, long-lived greenhouse gases and aerosols, and land-cover responsible for climatic variations over the past two millennia (*Jansen et al., 2007*, IPCC AR4 WG1, Sec. 6.6) or some combination of these three controls. (See also *Climate Research Committee, National Research Council, 1995*).

No one of these potential controls can account for all of the variations observed in hydrological indicators over the past two millennia. By the late Holocene, the amplitude of the insolation anomalies is quite small (Fig. 3.13A-B), and the impact of deglaciation is no longer significant (Fig. 3.13C-D). Variations in indices that describe decadal-time-scale ocean-atmosphere interactions, often known as "teleconnection" or "climate-mode" indices (e.g., the PDO or "Pacific Decadal Oscillation" or the NAM or "Northern Annular Mode;" see *Trenberth et al., 2007*, IPCC AR4 WG1 Sec. 3.6 for review), are sometimes invoked to explain apparent periodicity or "regime changes" in proxy records (e.g., *Stone and Fritz, 2006; Rasmussen et al., 2006*). However, the observational records that are used to define those indices are not long enough to discriminate among true cyclical or oscillatory behavior, recurrent changes in levels (or regime shifts), and simple red-noise or autocorrelated variations in time series (*Rudnick and Davis, 2003; Overland et al., 2006*), so perceived periodicities in paleoenvironmental records could arise from sources other than, for example, solar irradiance cycles inferred from ^{14}C-production records. Moreover, there are no physical mechanisms that might account for decadal-scale variations over long time spans in, for example, the PDO, apart from those that involve the integration of the shorter time-scale variations (i.e., ENSO; *Newman et al., 2003; Schneider and Cornuelle, 2005*). Finally, although the broad trends global or hemispheric-average temperatures over the past millennium seem reasonably well accounted for by the combinations of factors described in (3) above, there is little short-term agreement among different simulations. Consequently, despite their societal importance (e.g., *Climate Reseach Committee, 1995*), the genesis of centennial-scale climatic and hydrologic variations remains essentially unexplained.

5. Future Subtropical Drying: Dynamics, Paleocontext, and Implications

It is a robust result in climate model projections of the climate of the current century that many already wet areas of the planet get wetter – such as in the oceanic Intertropical Convergence Zone (ITCZ), the Asian monsoonal region, and equatorial Africa - and already dry areas get drier – such as the oceanic subtropical high pressure zones, southwestern North America, the Intra-America Seas, the Mediterranean region, and southern Africa (*Held and Soden, 2006*); see also *Hoerling et al. (2006)*. Drying and wetting as used here refer to the precipitation minus the surface evaporation, or P-E. P-E is the quantity that, in the long-term mean over land, balances surface and subsurface runoff and, in the atmosphere, balances the vertically integrated moisture convergence or divergence. The latter contains components due to the convergence or divergence of water vapor by the mean flow convergence or divergence, the advection of humidity by the mean flow, and the convergence or divergence of humidity by the transient flow. A warmer atmosphere can hold more moisture, so the pattern of moisture convergence or divergence by the mean flow convergence or divergence intensifies. This makes the deep tropical regions of the ITCZ wetter and the dry regions of the subtropics, where there is descending air and mean flow divergence, drier (*Held and Soden, 2006*).

While a warming-induced intensification of hydrological gradients is a good first start for describing hydrological change, there are many exceptions to this simple picture. For example the Amazon is a wet region where models do not robustly predict either a drying or a wetting. Here the models create more El Niño-like tropical Pacific SSTs that tend to make the Amazon drier, highlighting the potential importance of tropical circulation changes in climate change (*Li et al., 2006*). The Sahel region of West Africa dried dramatically in the latter half of the last century (*Nicholson et al., 2000*), which has been attributed to changes in SSTs throughout the tropics (*Giannini et al., 2003*). The models within the IPCC AR4 generally reproduce these changes in SST and Sahel drying as a consequence of anthropogenic climate change during the late-20th century (*Biasutti and Giannini, 2006*). However the same models have widely varying projections for how precipitation will change in the Sahel over the current century with some predicting a return to wetter conditions (*Biasutti and Giannini, 2006; Hoerling et al., 2006*). It is unknown why the modeled response in the Sahel to 20th century radiative forcing is different to the response to current-century forcing. However, it is worth noting that the one climate model that best simulates the 20th century drying continues to dry the Sahel in the current century (*Held et al., 2005*). In this tropical region, as in the Amazon, hydrological change appears to potentially involve nonlocal controls on the atmospheric circulation as well as possible complex land-surface feedbacks.

The greater southwestern regions of North America, which include the American Southwest and northern Mexico, are included within this region of subtropical drying.

Seager et al. (2007d) show that there is an impressive agreement amongst the projections with 19 climate models (and 47 individual runs) (Fig. 3.16). These projections collectively indicate that this region progressively dries in the future and that the transition to a more arid climate begins in the late 20th century and early current century (Fig. 3.17). The increased aridity becomes equivalent to the 1950s Southwest drought in the early part of the current century in about a quarter of the models and half of the models by mid-century. *Seager et al. (2007d)* also showed that intensification of the existing pattern of atmospheric water-vapor transport was only responsible for about half the Southwest drying and that half was caused by a change in atmospheric circulation. They linked this fact to a poleward expansion of the Hadley Cell and dry subtropical zones and a poleward shift of the mid-latitude westerlies and storm tracks, both also robust features of a warmer atmosphere (*Yin, 2005; Bengtsson et al., 2006; Lu et al., 2007*). The analysis of satellite data by *Seidel et al. (2008)* suggests such a widening of Earth's tropical belt over the past quarter century as the planet has warmed. This analysis is consistent with climate model simulations that suggest future subtropical drying as the jet streams and the associated wind and precipitation patterns move poleward with global warming. Note, however, that GCMs are unable to capture the mesoscale processes that underlie the North American monsoon (*e.g., Castro et al., 2007a*), so there is uncertainty regarding the impact of these changes on monsoon season precipitation in the American Southwest and northern Mexico.

The area encompassing the Mediterranean regions of southern Europe, North Africa, and the Middle East dries in the model projections even more strongly, with even less disagreement amongst models, and also beginning toward the end of the last century. Both here and in southwestern North America, the drying is not abrupt in that it occurs over the same time scale as the climate forcing strengthens. However, the severity is such that the aridity equivalent to historical droughts — but as a new climate rather than a temporary state — is reached within the coming years to a few decades. Assessed on the time scale of water-resource development, demographic trends, regional development, or even political change, this could be described as a "rapid" if not abrupt climate change and, hence, is a cause for immediate concern.

The future subtropical drying occurs in the models for reasons that are distinct from the causes of historical droughts. The latter are related to particular patterns of tropical SST anomalies, while the former arises as a consequence of overall, near-uniform warming of the surface and atmosphere and how that impacts water-vapor transports and atmospheric circulation. Both mechanisms involve a poleward movement of the mid-latitude westerlies and similar changes to the eddy-driven mean meridional circulation. However, a poleward expansion of the Hadley Cell has not been invoked to explain the natural droughts. Further future drying is expected to be accompanied by a maximum of warming in the tropical upper troposphere (a consequence of moist convection in the deep tropics), whereas natural droughts have gone along with cool temperatures in the tropical troposphere. Hence, past droughts are not analogs of future drying, which should make identification of anthropogenic drying easier when it occurs.

It is unclear how apt the Medieval megadroughts are as analogs of future drying. As mentioned above, it has been suggested that they were caused by tropical Pacific SSTs being La Niña-like for up to decades at a time during the Medieval period, as well as by the subtropical North Atlantic being warm. The tropical Pacific SST change possibly arose as a response to increased surface solar radiation. If this is so, then future subtropical drying will likely have no past analogs. However, it cannot be ruled out that the climate model projections are wrong in not producing a more La Niña-like state in response to increased radiative forcing. For example, the current generation of models has well known and serious biases in their simulations of tropical Pacific climate, and these may compromise the model projections of climate change. If the models are wrong, then it is possible that the future subtropical drying caused by general warming will be augmented by the impacts of an induced more La Niña-like state in the tropical Pacific. However, the association between positive radiative forcing, a more La Niña-like SST state, and dry conditions in southwestern North America that has been argued for using paleoclimate proxy data is for solar forcing, whereas future climate change will be driven by greenhouse forcing. It is not known if the tropical climate system responses to solar and greenhouse gas forcing are different. These remaining problems with our understanding of, and ability to model, the tropical climate system in response to radiative forcing mean that there remains uncertainty in how strong the projected drying in the Southwest will be, an uncertainty that includes the possibility that it will be more intense than in the model projections.

Future drying in southwestern North America will have significant social impacts in both the U.S. and Mexico. To date there are no published estimates of the impact of reduced P-E on the water-resource systems of the region that take full account of the climate projections. To do so would involve downscaling to the river basin scale from the projections with global models using either statistical methods or regional models, a problem of considerable technical difficulty. However both *Hoerling and Eischeid (2007)* and *Christensen and Lettenmaier (2006)* have used simpler methods to suggest that the global model projections imply that Colorado River flow will drop by between several percent and a quarter. While the exact number cannot, at this point, be known with any certainty at all, our current ability to model hydrology in this region unambiguously projects reduced flow.

Reduced flow in the Colorado and the other major rivers of the Southwest will come at a time when the existing flow is already fully allocated and when the population in the region is increasing. Current allocations of the Colorado River are also based on proportions of a fixed flow that was measured early in the last century at a time of unusual high flow (*Woodhouse et al., 2005*). It is highly likely that it will not be possible to meet those allocations in the projected drier climate of the relatively near future. In this context it needs to be remembered that agriculture uses some 90% of Colorado River water and about the same amount of total water use throughout the region, but even in California with its rich, productive, and extensive farmland, agriculture accounted for no more than 2% of the State economy.

6. Floods: Present, Past, and Future

Like droughts, floods, or episodes of much wetter-than-usual conditions, are embedded in large-scale atmospheric circulation anomalies that lead to a set of meteorological and hydrological conditions that support their occurrence. In contrast to droughts, floods are usually more localized in space and time, inasmuch as they are related to a specific combination of prior hydrologic conditions (e.g., the degree of soil saturation prior to the flood) upon which specific short-term meteorological events (intense rainstorms or rapid snowmelt) are superimposed (*Hirschboeck, 1989; Mosley and McKerchar, 1993; Pilgrim and Codery, 1993*) Floods are also geomorphologically constrained by drainage-basin and floodplain characteristics (*Baker et al., 1988; O'Connor and Costa, 2003*). However, when climatic anomalies are large in scope and persistent, such as those that occurred during 1993 in the Upper Mississippi Valley (*Kunkel et al., 1994; Anderson et al., 2003*) (and again in 2008), regionally extensive episodes of flooding can occur. When climate significantly changes, as it has in the past (*Knox, 2000*), and will likely do in the future, changes in the overall flood regime, including the frequency of different size floods and the areas affected, will also occur (*Kundzewicz et al., 2007*).

6.1 The 1993 Mississippi Valley Floods—Large-Scale Controls and Land-Surface Feedback

The flooding that occurred in the Upper Mississippi Valley of central North America in the late spring and summer of 1993 provides a case study of the control of a major flood event by large-scale atmospheric circulation anomalies. Significant feedback from the unusually wet land surface likely reinforced the wet conditions, which contributed to the persistence of the wet conditions. The 1993 flood ranks among the top five weather disasters in the United States,and was generated by the frequent occurrence of large areas of moderate to heavy precipitation, within which extreme daily total rainfall events were embedded. These meteorological events were superimposed on an above-normal soil-moisture anomaly at the beginning June of that year (*Kunkel et al., 1994*). These events were supported by the occurrence of a large-scale atmospheric circulation anomaly that featured the persistent flow of moisture from the Gulf of Mexico into the interior of the continent (*Bell and Janowiak, 1995; Trenberth and Guillemot, 1996*). The frequency of seasonal (90-day long) excessive (i.e. exceeding a 20-year return period) precipitation anomalies has generally been increasing over time in the United States (*Kunkel et al., 2008* (CCSP SAP 3.3, Sec. 2.2.2.3, Fig. 2.9).

The atmospheric circulation features that promoted the 1993 floods in the Mississippi Valley, when contrasted with the widespread dry conditions during the summer of 1988, provide a "natural experiment" that can be used to evaluate the relative importance of remote (e.g., the tropical Pacific) and local (over North America) forcing, and of the importance of feedback from the land surface to reinforce the unusually wet or dry conditions. For example, *Trenberth and Guillemot (1996)* used a combination of observational and "reanalysis" data (*Kalnay et al., 1996*), along with some diagnostic analyses to reveal the role of large-scale moisture transport into the midcontinent, with dryness ocurring in response to less flow and flooding in response to greater-than-normal flow. *Liu et al. (1998)* used a combination of reanalysis data and simple models to examine the interactions among the different controls of the atmospheric circulation anomalies in these 2 years.

Although initial studies using a regional climate model pointed to a small role for feedback from the wet land surface in the summer of 1993 to increase precipitation over the midcontinent (*Giorgi et al., 1996*), subsequent studies exploiting the 1988/1993 natural experiment using both regional climate models and general circulation models point to an important role for the land surface in amplifying the severity and persistence of floods and droughts (*Bonan and Stillwell-Soller, 1998; Bosilovich and Sun, 1999; Hong and Pan, 2000; Pal and Eltahir, 2002*). These analyses add to the general pattern that emerges for large moisture anomalies (both wet and dry) in the midcontinent of North America to have a) local controls (i.e. atmosoheric circulation and moisture flux over North America), b) remote controls (e.g. Pacific SST anomalies) and c) a significant role for feedback that can reinforce the moisture anomalies. The 1993 floods continue to be a focus for climate model intercomparisons (*Anderson et al., 2003*).

6.2 Paleoflood Hydrology

The largest floods observed either in the instrumental or in the paleorecord have a variety of causes (*O'Connor and Costa, 2004*), for the most part related to geological processes. However, some the largest floods are meteorological floods, which are relevant for understanding the nature of abrupt climate changes (*Hirschboeck 1989; House et al. 2002*) and potential changes in the environmental hazards associated with flooding (*Benito et al., 2004; Wohl, 2000*). Although sometimes used in an attempt to extend the instrumental record for operational hydrology purposes (i.e., fitting flood-distribution probability density functions; *Kochel and Baker, 1982; Baker et al., 1988*), paleoflood hydrology also provides information on the response of watersheds to long-term climatic variability or change (*Ely, 1997; Ely et al., 1993; Knox, 2000*), or to joint hydrological-climatological constraints on flood magnitude (*Enzel et al., 1993*).

Knox (2000, see also *Knox, 1985, 1993*) reconstructed the relative (to present) magnitude of small floods (i.e., those with frequent return intervals) in southwestern Wisconsin during the Holocene using radiocarbon-dated evidence of the size of former channels in the floodplains of small watersheds, and the magnitude (depth) of larger overbank floods using sedimentological properties of flood deposits. The variations in flood magnitude can be related to the joint effects of runoff (from precipitation and snowmelt) and vegetation cover (Fig. 3.13). The largest magnitudes of both sizes of flood occurred during the mid-Holocene drought interval, when tree cover was low, permitting more rapid runoff of flood-generating snowmelt and precipitation (see *Knox, 1972*). As tree cover increased with increasing moisture during the interval from 6 ka to 4 ka, flood magnitudes decreased, then increased again after 3.5 ka as effective moisture increased further in the late Holocene.

The paleoflood record in general suggests a close relationship between climatic variations and the flood response. This relationship may be quite complex, however, inasmuch as the hydrologic response to climate changes is mediated by vegetation cover, which itself is dependent on climate. In general, runoff from forested hillslopes is lower for the same input of snowmelt or precipitation than from less well-vegetated hillslopes (*Pilgrim and Cordery, 1993*). Consequently, a shift from dry to wet conditions in a grassland may see a large response (i.e., an increase) in flood magnitude at first (until the vegetation cover increases), while a shift from wet to dry conditions may see an initial decrease in flood magnitude, followed by an increase as vegetation cover is reduced (*Knox, 1972, 1993*). This kind of relationship makes it difficult to determine the specific link between climate variations and potentially abrupt responses in flood regime without the development of appropriate process models. Such models will require testing under conditions different from the present, as is the case for models of other environmental systems. Paleoflood data are relatively limited relative to other paleoenvironmental indicators, but work is underway to assemble a working database (*Hirschboeck, 2003*).

6.3 Floods and Global Climate Change

One of the main features of climate variations in recent decades is the emergence of a package of changes in meteorological and hydrological variables that are consistent with global warming and its impact on hydrological cycle and the frequency of extreme events (*Trenberth et al., 2007*, IPCC AR4, WG4, Ch. 3). The mechanisms underlying these changes include the increase in atmospheric moisture, the intensity of the hydrologic cycle, and the changes in atmospheric circulation as the atmosphere warms (*Knight et al., 2008*). As described in one of the key findings of *Gutowski et al., (2008*; CCSP SAP 3.3, Ch. 3) "Heavy precipitation events averaged over North America have increased over the past 50 years, consistent with the increased water holding capacity of the atmosphere in a warmer climate and observed increases in water vapor over the ocean." (See also *Easterling et al., 2000, Kunkel, 2003; Kunkel et al., 2003*). In addition, the frequency of season-long episodes of greater-than-average precipitation is increasing (*Kunkel et al., (2008*; CCSP SAP 3.3, Sec. 2.2.2.3), and the timing of snowmelt is changing in many parts of the country (see Sec. 7). All of the meterological controls of flooding (short- and long-duration heavy precipitation, snowmelt) are thus undergoing long-term changes. However, there is considerable uncertainty in the specific hydrologic response and its temporal and spatial pattern, owing to the auxiliary role that atmospheric circulation patterns and antecedent conditions play in generating floods, and these factors experience interannual- and decadal-scale variations themselves (*Kunkel, 2003*).

These changes in the state of the atmosphere in turn lead to the somewhat paradoxical conclusion that both extremely wet events (floods) and dry events (droughts) are likely to increase as the warming proceeds (*Kundzewicz et al., 2007*, IPCC AR4 WG2 Ch. 3). The extreme floods in Europe in 2002, followed by the extreme drought and heatwave in 2003, have been used to illustrate this situation (*Pal et al., 2004*). They compared observed 20th-century trends in atmospheric circulation and precipitation with the patterns of these variables (and of extreme-event characteristics: dry-spell length and maximum 5-day precipitation) projected for the 21st century using a regional climate model, and noted their internal consistency and consistency with the general aspects of anthropogenic global climate changes.

Projections of future hydrological trends thus emphasize the likely increase in hydrological variability in the future that includes less frequent precipitation, more intense precipitation, increased frequency of dry days, and also increased frequency of extremely wet days (CCSP SAP 3.3, Sec. 3.6.6, in prep.). Owing to the central role of water in human-environment interactions, it is also likely that these hydrological changes, and increases in flooding in particular, will have synergistic impacts on such factors as water quality and the incidence of water-borne diseases that could amplify the impact of basic hydrologic changes (*Field et al., 2007*, IPCC AR4, WG2, Ch. 14.4.1, 14.4.9). The great modifications by humans that have taken place in watersheds around the world further complicate the problem of projecting the potential for future abrupt changes in flooding.

6.4 Assessment of Abrupt Change in Flood Hydrology

Assessing the likelihood of abrupt changes in flood regime is a difficult proposition that is compounded by the large range in temporal and spatial scales of the controls of floods, and the consequent need to scale down the large-scale atmospheric and water- and energy-balance controls, and scale up the hillslope- and watershed-scale hydrological responses. Nevertheless, there is work underway to combine the appropriate models and approaches toward this end (e.g. *Jones et al., 2006; Fowler and Kilsby, 2007; Maurer, 2007*). This work could be enhanced by several developments, including:

- Enhanced modeling capabilities. The attempts that have been made thus far to project the impact of global climate change on hydrology, including runoff, streamflow, and floods and low-flows, demonstrate that the range of models and the approaches for coupling them are still in an early developmental stage (relative to, for example, coupled Atmosphere-Ocean General Circulation Models). Sufficient computational capability must be provided (or made available) to facilitate development and use of enhanced models.
- Enhanced data sets. Basic data on the flood response to climatic variations, both presentday and prehistoric, are required to understand the nature of that response across a range of conditions different from those of the present. Although human impacts on watersheds and recent climatic variability have provided a number of

natural experiments that illustrate the response of floods to controls, the impact of larger environmental changes than those found in the instrumental record are required to test the models and approaches than could be used.

- Better understanding of physical processes. The complexity of the response of extreme hydrologic events to climatic variations, including as it does the impacts on both the frequency and magnitude of meteorological extremes, and mediation by land cover and watershed characteristics that themselves are changing, suggests that further diagnostic studies of the nature of the response should be encouraged.

7. Other Aspects of Hydroclimate Change

The atmosphere can hold more water vapor as it warms (as described by the Clausius-Clapeyron equation), to the tune of about 7% per degree Celsius of warming. With only small changes projected for relative humidity (*Soden et al., 2002*), the specific humidity content of the atmosphere will also increase with warming at this rate. This is in contrast to the global mean precipitation increase of about 1-2% per degree Celsius of warming. The latter is caused when evaporation increases to balance increased downward longwave radiation associated with the stronger greenhouse trapping. For both of these constraints to be met, more precipitation has to fall in the heaviest of precipitation events as well explained by *Trenberth et al. (2003)*.

The change in precipitation intensity seems to be a hydrological change that is already evident (*Kunkel et al., 2008* (CCSP SAP 3.3, Secs. 2.2.2.2 and 2.2.2.3). *Groisman et al. (2004)* demonstrate that daily precipitation records over the last century in the United States show a striking increase, beginning around 1990, in the proportion of precipitation within very heavy (upper 1% of events) and extreme (upper 0.1%) of events. In the annual mean there is a significant trend to increased intensity in the southern and central plains and in the Midwest, and there is a significant positive trend in the Northeast in winter. In contrast the Rocky Mountain States show an unexplained significant trend to decreasing intensity in winter.

Groisman et al. (2005) show that the observed trend to increasing precipitation intensity is seen across much of the world, and both they and *Wilby and Wigley (2002)* show that climate model projections of the current century show that this trend will continue. *Groisman et al. (2005)* make the point that the trends in intensity are greater than the trends in mean precipitation, that there is good physical reason to believe that they are related to global warming, and that they are likely to be more easily detected than changes in the mean precipitation.

Increases in precipitation intensity can have significant social impacts as they increase the potential for flooding and overloading of sewers and wastewater treatment plants. See *Rosenzweig et al. (2007)* for a case study of New York City's planning efforts to deal with water-related aspects of climate change. Increasing precipitation intensity can also lead to an increase of sediment flux, including potentially harmful pathogens, into water-supply reservoirs, thus necessitating more careful water-quality management, a situation already being faced by New York City. (See http://www.amwa.net/cs/climatechange/newyorkcity for a useful discussion of how a major metropolitan area is already beginning to address this issue.)

Another aspect of hydroclimatic change that can be observed in many regions is the general decrease in snowpack and snow cover (*Mote et al., 2005; Déry and Brown, 2007; Dyer and Mote, 2006*; see also *Lettenmaier et al., 2008*; CCSP SAP 4.3, Sec. 4.2.4). Winter snowfall and the resulting accumulated snowpack depend on temperature in complicated ways. Increasing temperatures favor greater moisture availability and total precipitation (in much the same way that precipitation intensity depends on temperature) and hence greater snow accumulation (if winter temperatures are cold enough), but greater snowmelt and hence a reduced snowpack if temperatures increase enough. Regions with abundant winter precipitation, and winter temperatures close to freezing could therefore experience an overall increase in winter precipitation as temperatures increase but also an overall decrease in snow cover as the balance of precipitation shifts from snow to rain, along with an earlier occurrence of spring snowmelt. Such trends seem to be underway in many regions (*Moore et al., 2007*), but particularly in the Western United States (*Mote et al., 2005, 2008*).

As a consequence of reduced snowpack and earlier spring snowmelt, a range of other hydrologic variables can be affected, including the amount and timing of runoff, evapotranspiration, and soil moisture (*Hamlet et al., 2007; Moore et al., 2007*). Although gradual changes in snowcover and snowmelt timing could be the rule, the transition from general winter-long snowcover, to transient snowcover, to occasional snow cover, could appear to be quite abrupt, from the perspective of the hydrology of individual watersheds.

Most studies of past and modern impacts on water resources focus on abrupt changes in the physical system such as the duration of ice cover and timing of snow melt, lake thermal structure, evaporation, or water level, with considerably less attention on abrupt changes in water quality (e.g. *Lettenmaier et al., 2008*; CCSP SAP 4.3, Sec. 4.2.5). Assessing recent climate impacts on water quality has been complicated by human land use. For example, analysis of contemporary data in the northern Great Plains suggests that climate impacts are small relative to land use (*Hall et al., 1999*). A similar conclusion has been reached in Europe based on the paleoclimate literature, where humans have been impacting the environment for thousands of years (*Hausmann et al., 2002*). Some of the best evidence for climate changes resulting in changes in water quality and on aquatic biological communities comes from work in the Experimental Lakes Area in Canada where land use changes have been more limited (*Schindler, 1996a,b*). This work showed how climate changes affect ion concentration, nutrients, and dissolved organic carbon concentrations, often amplifying acidification and other external perturbations. Other evidence suggests that that climate warming might affect water quality (phytoplankton biomass and nutrient concentrations) indirectly by affecting lake thermal structure (*Lebo et al., 1994; Gerten and Adrian, 2000*). The climate changes may lead to abrupt changes in salinity and water quality for drinking, irrigation, and livestock. The recent paleolimnological records of abrupt changes in salinity have been inferred from changes in diatoms in the sediments of Moon Lake, ND (*Laird et al., 1996*), and the Aral Sea (*Austin et al., 2007*); however, determining if the magnitude of these abrupt changes represents a significant degradation of water quality is difficult to discern.

8. Conclusions

Drought is among the greatest of recurring natural hazards facing both the people of the United States and humanity worldwide today and in the foreseeable future. Its causes are complex and not completely understood, but its impact on agriculture, water supply, natural ecosystems, and other human needs for survival can be severe and long lasting in human terms, making it one of the most pressing scientific problems to study in the field of climatic change. Floods, though generally more localized in time and space than droughts, are also a major natural hazard, and share with droughts many of same large-scale controls and the potential for experiencing major changes in these controls in the future.

Droughts can develop faster than the time scale needed for human societies and natural systems to adapt to the increase in aridity. Thus, a severe drought lasting several years may be experienced as an abrupt change to drier conditions even though wetter conditions will eventually return. The 1930s Dust Bowl drought, which resulted in a mass exodus from the parched Great Plains to more favorable areas in the West, is one such example. The drought eventually ended when the rains returned, but the people did not. For them it was a truly abrupt and permanent change in their lives. Thus, it is a major challenge of climate research to find ways to help reduce the impact of future droughts through improved prediction and the more efficient use of the limited available water resources.

For examples of truly abrupt and long-lasting changes in hydroclimatic variability over midcontinental North America and elsewhere in the world, we must go back in time to the middle Holocene, when much larger changes in the climate system occurred. The climate boundary conditions responsible for those changes were quite different from those today, so the magnitude of change that we might conceivably expect in the future under “natural” forcing of the climate system might not to be as great. However, the rising level of greenhouse gas forcing that is occurring now and in the foreseeable future is truly unprecedented, even over the Holocene. Therefore, the abrupt hydrologic changes in the Holocene ought to be viewed as useful examples of the magnitude of change that could conceivably occur in the future, and the mechansisms through which that change occurs.

The need for improved drought prediction on time scales of years to decades is clear now. To accomplish this will require that we develop a much better understanding of the causes of hydroclimatic variability worldwide. It is likely that extended periods of anomalous tropical ocean SSTs, especially in the eastern equatorial Pacific ENSO region, strongly influence the development and duration of drought over substantial land areas of the globe. As the IPCC AR4 concluded, “the palaeoclimatic record suggests that multi-year, decadal and even centennial-scale drier periods are likely to remain a feature of future North American climate, particularly in the area west of the Mississippi River.” Multiple proxies indicate the past 2,000 years included periods with more frequent, longer and/or geographically more extensive droughts in North America than during the 20th century. However, the record of past drought from tree rings offers a sobering picture of just how severe droughts can be under natural climate conditions. Prior to A.D. 1600, a succession of megadroughts occurred that easily eclipsed the duration of any droughts known to have occurred over North America since that time. Thus, understanding the causes of these extraordinary megadroughts is of paramount importance. Increased solar forcing over the tropical Pacific has been implicated, as has explosive volcanism, but the uncertainties remain large.

However significant enhanced solar forcing has been in producing past megadroughts, the level of current and future radiative forcing due to greenhouse gases is very likely to be of much greater significance. It is thus disquieting to consider the possibility that drought-inducing La Niña-like conditions may become more frequent and persistent in the future as greenhouse warming increases. We have no firm evidence that this is happening now, even with the serious drought that has gripped the West since about 1998. Yet, a large number of climate models suggest that future subtropical drying is a virtual certainty as the world warms and, if they are correct, indicate that it may have already begun. The degree to which this is true is another pressing scientific question that must be answered if we are to know how to respond and adapt to future changes in hydroclimatic variability.

References

Adams, J.B., M.E Mann, and C.M. Ammann, 2003: Proxy evidence for an El Niño-like response to volcanic forcing. *Nature*, **426**, 274-278.

Anderson, C.J., R.W. Arritt, E.S. Takle, Z.T. Pan, W.J. Gutowski, F.O. Otieno, R. da Silva, D. Caya, J.H. Christensen, D. Luthi, M.A. Gaertner, C. Gallardo, F. Giorgi, S.Y. Hong, C. Jones, H.M.H. Juang, J.J. Katzfey, W.M. Lapenta, R. Laprise, J.W. Larson, G.E. Liston, J.L. McGregor, R.A. Pielke, J.O. Roads, and J.A. Taylor, 2003: Hydrological processes in regional climate model simulations of the central United States flood of June-July 1993. *Journal of Hydrometeorology*, **4(3)**, 584-598.

Anderson, L., M.B. Abbott, B.P. Finney, and S.J. Burns, 2005: Regional atmospheric circulation change in the North Pacific during the Holocene inferred from lacustrine carbonate oxygen isotopes, Yukon Territory, Canada. *Quaternary Research*, **64(1)**, 21-35.

Asmerom, Y., V. Polyak, S. Burns, and J. Rassmussen, 2007: Solar forcing of Holocene climate: New insights from a speoleothem record, southwestern United States. *Geology*, **35**, 1-4.

Austin, P., A.W. Mackay, O. Palagushkina, and M.J. Leng, 2007: A high-resolution diatom-inferred palaeoconductivity and lake level record of the Aral Sea for the past 1600 yr. *Quaternary Research*, **67**, 383-393.

Backlund, P., A. Janetos, D.S. Schimel, J. Hatfield, M. Ryan, S. Archer, and D. Lettenmaier, 2008: Executive Summary. In: *The effects of climate change on agriculture, land resources, water resources, and biodiversity.* A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research. Washington, DC, USA, 362 pp.

Baker, V.R., R.C. Kochel, and P.C. Patton, 1988: Flood Geomorphology. New York, Wiley.

Barber, D.C., A. Dyke, C. Hillaire-Marcel, A.E. Jennings, J.T. Andrews, M.W. Kerwin, G. Bilodeau, R. McNeely, J. Southon, M.D. Morehead, and J.M. Gagnon, 1999: Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes. *Nature*, **400(6742)**, 344-348.

Barron, J.A., D. Bukry, and W.E. Dean, 2005: Paleoceanographic history of the Guaymas Basin, Gulf of California, during the past 15,000 years based on diatoms, silicoflagellates, and biogenic sediments. *Marine Micropaleontology*, **56(3-4)**, 81-102.

Bartlein, P.J, and S.W. Hostetler, 2004: Modeling paleoclimates. In: The Quaternary Period in the United States. [Gillespie, A.R., S.C. Porter, and B.F. Atwater (eds.)]. Elsevier, Amsterdam.

Bartlein, P.J., and C. Whitlock, 1993: Paleoclimatic interpretation of the Elk Lake pollen record. In: Elk Lake, Minnesota: Evidence for Rapid Climate Change in the North-Central United States. [Bradbury. J.P., and W.E. Dean (eds.)]. Geological Society of America, Boulder, CO.

Bartlein, P.J., K.H. Anderson, P.M. Anderson, M.E. Edwards, C.J. Mock, R.S. Thompson, R.S. Webb, T. Webb III, and C. Whitlock, 1998: Paleoclimate simulations for North America over the past 21,000 years: Features of the simulated climate and comparisons with paleoenvironmental data. *Quaternary Science Reviews*, **17(6-7)**, 549-585.

Bell, G.D. and J.E. Janowiak, 1995: Atmospheric circulation associated with the Midwest floods of 1993. *Bulletin of the American Meteorological Society*, **76(5)**, 681-695.

Bengtsson, L., K.I. Hodges, and E. Roeckner, 2006: Storm tracks and climate change. *Journal of Climate*, **19**, 3518-3543.

Benito, G., M. Lang, M. Barriendos, M.C. Llasat, F. Frances, T. Ouarda, V.R. Thorndycraft, Y. Enzel, A. Bardossy, D. Coeur, and B. Bobee, 2004: Use of systematic, palaeoflood and historical data for the improvement of flood risk estimation: Review of scientific methods. *Natural Hazards*, **31(3)**, 623-643.

Benson, L., M. Kashgarian, R. Rye, S. Lund, F. Paillet, J. Smoot, C. Kester, S. Mensing, D. Meko, and S. Lindström, 2002: Holocene multidecadal and multicentennial droughts affecting Northern California and Nevada. *Quaternary Science Reviews*, **21**, 659–682.

Berger, A., 1978: Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quaternary Research*, **9**, 139-167.

Berger, A., and M.F. Loutre, 1991: Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews*, **10**, 297-317.

Biasutti, M. and A. Giannini, 2006: Robust Sahel drying in response to late 20th century forcings. *Geophysical Research Letters*, **33**, doi:10.1029/2006GL026067.

Bonan, G.B. and L.M. Stillwell-Soller, 1998: Soil water and the persistence of floods and droughts in the Mississippi River Basin. *Water Resources Research*, **34**, 2693-2701.

Booth, R.K., M. Notaro, S.T. Jackson, and J.E. Kutzbach, 2006: Widespread drought episodes in the western Great Lakes region during the past 2000 years: Geographic extent and potential mechanisms. Earth and Planetary Science Letters, **242**, 415-427.

Bosilovich, M.G. and W.Y. Sun, 1999: Numerical simulation of the 1993 Midwestern flood: Local and remote sources of water. *Journal of Geophysical Research-Atmospheres*, **104(D16)**, 19415-19423.

Braconnot, P., B. Otto-Bliesner, S. Harrison, S. Joussaume, J.Y. Peterchmitt, A. Abe-Ouchi, M. Crucifix, E. Driesschaert, T. Fichefet, C.D. Hewitt, M. Kageyama, A. Kitoh, A. Laine, M.F. Loutre, O. Marti, U. Merkel, G. Ramstein, P. Valdes, S.L. Weber, Y. Yu, and Y. Zhao. 2007a. Results of PMIP2 coupled simulations of the Mid-Holocene and last glacial maximum—Part 1: Experiments and large-scale features. *Climate of the Past*, **3(2)**, 261-277.

Braconnot, P., B. Otto-Bliesner, S. Harrison, S. Joussaume, J.Y. Peterchmitt, A. Abe-Ouchi, M. Crucifix, E. Driesschaert, T. Fichefet, C.D. Hewitt, M. Kageyama, A. Kitoh, M.F. Loutre, O. Marti, U. Merkel, G. Ramstein, P. Valdes, L. Weber, Y. Yu, and Y. Zhao. 2007b. Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum—Part 2: Feedbacks with emphasis on the location of the ITCZ and mid- and high latitudes heat budget. *Climate of the Past*, **3(2)**, 279-296.

Breshears, D.D., N.S. Cobb, P.M. Rich, K.P. Price, C.D. Allen, R.G. Balice, W.H. Romme, J.H. Kastens, M.L. Floyd, J. Belnap, J.J. Anderson, O.B. Myers, and C.W. Meyer, 2005: Regional vegetation die-off in response to global-change type drought. *Proceedings of the National Academy of Sciences*, **102(42)**, 15144-15148.

Cane, M.A., A.C. Clement, A. Kaplan, Y. Kushnir, D. Pozdnyakov, R. Seager, S.E. Zebiak, and R. Murtugudde, 1997: Twentieth-century sea surface temperature trends. *Science*, **275**, 957-960.

Castro, C.L., T.B. McKee, and R.A. Pielke, Sr., 2001: The Relationship of the North American Monsoon to tropical and North Pacific sea surface temperatures as revealed by observational analyses. *Journal of Climate*, **14**, 4449-4473.

Castro, C.L. R.A. Pielke, Sr., and J.O. Adegoke, 2007a: Investigation of the summer climate of the contiguous U.S. and Mexico using the Regional Atmospheric Modeling System (RAMS). Part I: Model climatology (1950-2002). *Journal of Climate*, **20**, 3866-3887.

Castro, C.L., R.A. Pielke, Sr., J.O. Adegoke, S.D Schubert, and P.J. Pegion, 2007b: Investigation of the summer climate of the contiguous U.S. and Mexico using the Regional Atmospheric Modeling System (RAMS). Part II: Model climate variability. *Journal of Climate*, **20**, 3888-3901.

Chiang, J.C.H., and A.S. Sobel, 2002: Tropical tropospheric temperature variations caused by ENSO and their influence on the remote tropical climate. *Journal of Climate*, **15**, 2616-2631.

Christensen, N., and D.P. Lettenmaier, 2006: A multimodel ensemble approach to assessment of climate change impacts on the hydrology and water resources of the Colorado River basin. *Hydrology and Earth System Sciences Discussion*, **3**, 1-44.

Claussen, M, 2001: Earth system models. In: Understanding the Earth System. [Ehlers, E., and T. Krafft (eds.)]. Springer, Berlin, 147-162.

Claussen, M., P. Hoelzmann, H.J. Pachur, C. Kubatzki, V. Brovkin, and A. Ganopolski, 1999: Simulation of an abrupt change in Saharan vegetation in the mid-Holocene. *Geophysical Research Letters*, **26(14)**, 2037-2040.

Clement, A.C., R. Seager, M.A. Cane, and S.E. Zebiak, 1996: An ocean dynamical thermostat. *Journal of Climate*, **9**, 2190-2196.

Clement, R., R.Seager, and M.A. Cane, 2000: Suppression of El Niño during the mid-Holocene due to changes in the Earth’s orbit. *Paleoceanography*, **15**, 731-737.

Climate Research Committtee, National Research Council. 1995. Natural climate variability on decade-to-century time scales. National Academy Press, Washington, DC, 644 pp.

Cobb, K., C.D. Charles, H. Cheng, and R.L. Edwards, 2003: El Niño/southern oscillation and tropical Pacific climate during the last millennium. *Nature*, **424**, 271-276.

COHMAP Members, 1988: Climatic changes of the last 18,000 years: Observations and model simulations. *Science*, **241**, 1043-1052.

Cole, J.E., and E.R. Cook, 1998: The changing relationship between ENSO variability and moisture balance in the continental United States. *Geophysical Research Letters*, **25(24)**, 4529-4532.

Cole, J.E., J.T. Overpeck, and E.R. Cook, 2002: Multiyear La Niña events and persistent drought in the contiguous United States. *Geophysical Research Letters*, **29(13)**, 10.1029/2001GL013561.

Committee on Radiative Forcing Effects on Climate, 2005: Radiative forcing of climate change. National Academy Press.

Cook, E.R., and G.C. Jacoby, Jr., 1977: Tree-ring drought relationships in the Hudson Valley, New York. Science, **198**, 399-401.

Cook, B.I., G.B. Bonan, and S. Levis, 2006: Soil moisture feedbacks to precipitation in southern Africa. *Journal of Climate*, **19**, 4198-4206.

Cook, B.I., R. Miller and R. Seager, 2008: Dust and sea surface temperature forcing of the 1930s Dust Bowl drought. *Geophys. Res. Lett.*, in press.

Cook, E.R., and P.J. Krusic. 2004a. The North American drought atlas. Lamont-Doherty Earth Observatory and the National Science Foundation.

Cook, E.R., and P.J. Krusic, 2004b: North American summer PDSI reconstructions. *IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series No. 2004-045*. NOAA/NGDC Paleoclimatology Program, Boulder, CO, 24 pp.

Cook, E.R., M. Kablack, and G.C. Jacoby, 1988: The 1986 drought in the southeastern United States: how rare an event was it? Journal of Geophysical Research – Atmospheres, **93(D1)**, 14257-14260.

Cook, E.R., D.M. Meko, D.W. Stahle, and M.K. Cleaveland. 1999. Drought reconstructions for the continental United States. *Journal of Climate*, **12**, 1145-1162.

Cook, E.R., R. Seager, M.A. Cane, and D.W. Stahle, 2007: North American drought: reconstructions, causes and consequences. *Earth Science Reviews*, **81**, 93-134.

Cook, E.R., C. Woodhouse, C.M. Eakin, D.M. Meko, and D.W. Stahle, 2004: Long-term aridity changes in the western United States. *Science*, **306**, 1015-1018.

Crucifix, M., P. Braconnot, S. Harrison, and B. Otto-Bliesner, 2005: Second phase of Paleoclimate Modelling Intercomparison Project. *Eos, Transactions of the American Geophysical Union*, **86(28)**, 264.

Dean, W.E., R.M. Forester, and J.P. Bradbury, 2002: Early Holocene change in atmospheric circulation in the Northern Great Plains: An upstream view of the 8.2 ka cold event. *Quaternary Science Reviews*, **21(16-17)**, 1763-1775.

Déry, S. J., and R. D. Brown, 2007: Recent Northern Hemisphere snow cover extent trends and implications for the snow-albedo feedback. *Geophysical Research Letters*, **34(22)**, doi:1029/2007GL031474

Dyer, J.L., and T.L. Mote, 2006: Spatial variability and trends in obseved show depth over North America. *Geophysical Research Letters*, **33**, doi:10.1029/2006GRL027258.

Dyke, A.S, 2004: An outline of North American deglaciation with emphasis on central and northern Canada. In: Quaternary Glaciations—Extent and Chronology, Part II: North America. [Ehlers, J., and P.L. Gibbard (eds.)]. Elsevier, Amsterdam, 373-424.

deMenocal, P., J. Ortiz, T. Guilderson, J. Adkins, M. Sarnthein, L. Baker, and M. Yarusinsky, 2000: Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews*, **19(1-5)**, 347-361.

Diffenbaugh, N.S., and L.C. Sloan, 2002: Global climate sensitivity to land surface change: The Mid Holocene revisited. *Geophysical Research Letters*, **29(10)**, 1476, doi:10.1029/2002GL014880.

Diffenbaugh, N.S., M. Ashfaq, B. Shuman, J.W. Williams, and P.J. Bartlein, 2006: Summer aridity in the United States: Response to mid-Holocene changes in insolation and sea surface temperature. *Geophysical Research Letters*, **33**, L22712, doi:10.1029/2006GL028012.

Douglass, A.E., 1929: The secret of the southwest solved with talkative tree rings. *National Geographic Magazine*, December, 736-770.

Douglass, A.E., 1935: Dating Pueblo Bonito and other ruins of the southwest. National Geographic Society Contributed Technical Papers. Pueblo Bonito Series 1, 1-74.

Dyke, A.S, 2004: An outline of North American deglaciation with emphasis on central and northern Canada. In: *Quaternary Glaciations—Extent and Chronology, Part II: North America.* [Ehlers, J., and P.L. Gibbard (eds.)]. Elsevier, Amsterdam.

Easterling, D.R., J.L. Evans, P.Y. Groisman, T.R. Karl, K.E. Kunkel, and P. Ambenje, 2000: Observed variability and trends in extreme climate events: A brief review. *Bulletin of the American Meteorological Society*, **81(3)**, 417-425.

Egan, T, 2006: The worst hard time. Houghton Mifflin, 340 pp.

Ely, L., 1997: Response of extreme floods in the southwestern United States to climatic variations in the late Holocene. *Geomorphology*, **19**, 175-201.

Ely, L.L., Y. Enzel, V.R. Baker, and D.R. Cayan, 1993: A 5000-year record of extreme floods and climate change in the southwestern United States. *Science*, **262**, 410-412.

Emile-Geay, J., M.A. Cane, R. Seager, A. Kaplan and P. Almasi, 2007: ENSO as a mediator for the solar influence on climate. *Paleoceanography*, **22**, doi:10.1029/2006PA001304.

Enfield, D.B., A.M. Mestas-Nunez, and P.J. Trimble, 2001: The Atlantic multidecadal oscillation and its relation to rainfall and river flows in the continental U.S. *Geophysical Research Letters*, **28(10)**, 2077-2080.

Enzel, Y., L.L. Ely, P.K. House, V.R. Baker, and R.H. Webb, 1993: Paleoflood evidence for a natural upper bound to flood magnitudes in the Colorado River Basin. *Water Resources Research*, **29(7)**, 2287-2297.

Esper, J., E.R. Cook, and F. Schweingruber, 2002: Low-frequency signals in long tree-ring chronologies for reconstructing past temperature variability. *Science*, **295**, 2250-2253.

Field, C.B., L.D. Mortsch,, M. Brklacich, D.L. Forbes, P. Kovacs, J.A. Patz, S.W. Running and M.J. Scott, 2007: North America. In: *Climate Change 2007: Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change.* [Parry, M.L., O.F. Canziani, J.P. Palutikof, P.J. van der Linden, and C.E. Hanson (eds.). Cambridge University Press, Cambridge, United Kingdom, 617-652.

Foley, J.A., M.T. Coe, M. Scheffer, and G.L. Wang, 2003: Regime shifts in the Sahara and Sahel: Interactions between ecological and climatic systems in northern Africa. *Ecosystems*, **6(6)**, 524-539.

Forman, S.L., R. Oglesby, and R.S. Webb, 2001: Temporal and spatial patterns of Holocene dune activity on the Great Plains of North America: Megadroughts and climate links. *Global and Planetary Change*, **29(1-2)**, 1-29.

Forster, P., V. Ramaswamy, P. Artaxo, T. Berntsen, R. Betts, D.W. Fahey, J. Haywood, J. Lean, D.C. Lowe, G. Myhre, J. Nganga, R. Prinn, G. Raga, M. Schulz and R. Van Dorland, 2007: Changes in atmospheric constituents and in radiative forcing. In: *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York.

Fowler, H.J., and C.G. Kilsby, 2007: Using regional climate model data to simulate historical and future river flows in northwest England. *Climatic Change*, *80*, 337-367.

Fritts, H.C., 1976: Tree rings and climate. Academic Press, London, 567 pp.

Fulp, T, 2005: How low can it go? *Southwest Hydrology*, March/April, 16-17, 28.

Fye, F., D.W. Stahle, and E.R. Cook, 2003: Paleoclimatic analogs to 20th century moisture regimes across the USA. *Bulletin of the American Meteorological Society*, **84(7)**, 901-909.

Fye, F.K., D.W. Stahle, and E.R. Cook, 2004: Twentieth century sea surface temperture patterns in the Pacific during decadal moisture regimes over the USA. *Earth Interactions*, **8(22)**, 1-22.

Fye, F.K, D.W. Stahle, E.R. Cook, and M.K. Cleaveland, 2006: NAO influence on sub-decadal moisture variability over central North America. *Geophysical Research Letters*, **33**, L15707, doi:10.1029/2006GL026656.

Gajewski, K., A. Viau, M. Sawada, D. Atkinson, and S. Wilson, 2001: Sphagnum peatland distribution in North America and Eurasia during the past 21,000 years. *Global Biogeochemical Cycles*, **15(2)**, 297-310.

Gallimore, R., R. Jacob, and J. Kutzbach, 2005: Coupled atmosphere-ocean-vegetation simulations for modern and mid-Holocene climates: role of extratropical vegetation cover feedbacks. *Climate Dynamics*, **25(7-8)**, 755-776.

Garcin, Y., A. Vincens, D. Williamson, G. Buchet, and J. Guiot, 2007: Abrupt resumption of the African Monsoon at the Younger Dryas-Holocene climatic transition. *Quaternary Science Reviews*, **26(5-6)**, 690-704.

Gerten, D., and R. Adrian, 2000: Climate-driven changes in spring plankton dynamics and the sensitivity of shallow polymictic lakes to the North Atlantic oscillation. *Limnology and Oceanography*, **45**, 1058-106.

Giannini, A., R. Saravanan, and P. Chang, 2003: Oceanic forcing of Sahel rainfall on internanual to interdecadal timescales. *Science*, **302**, 1027-1030.

Giorgi, F., L.O. Mearns, C. Shields, and L. Mayer. 1996. A regional model study of the importance of local versus remote controls of the 1988 drought and the 1993 flood over the central United States. *Journal of Climate*, **9(5)**, 1150-1162.

Gleick, P.H., and L. Nash, 1991: The societal and environmental costs of the continuing California drought. Pacific Institute for Studies in Development, Environment, and Security, Oakland, CA.

Goddard, L., and N.E. Graham, 1999: Importance of the Indian Ocean for simulating rainfall anomalies over eastern and southern Africa. *Journal of Geophysical Research*, **104**, 19099-19116.

Graham, N.E., M.K. Hughes. C.M. Ammann, K.M. Cobb, M. P. Hoerling, D. J. Kennett , J.P. Kennett, B. Rein. L. Stott, P.E. Wigand, and T. Xu, 2007: Tropical Pacific-mid latitude teleconnections in medieval times. *Climatic Change*, **83**, 241-285.

Griles, J.S, 2004: Building on success, facing the challenges ahead. Paper presented at Colorado River Water Users Association, Las Vegas NV, December 2004. http://www.usbr.gov/.

Groisman, P.Y., R.W. Knight, D.R. Easterling, T.R. Karl, G.C. Hegerl, and V.N. Razuvaev, 2005: Trends in intense precipitation in the climate record. *Journal of Climate*, **18**, 1326-1350

Groisman, P.Y., R.W. Knight, T.R. Karl, D.R. Easterling, B. Sun, and J.H. Lawrimore, 2004: Contemporary changes of the hydrological cycle over the contiguous United States: Trends derived from in situ observations. *Journal of Hydrometeorology*, **5**, 64-85.

Grootes, P.M., M. Stuiver, J.W.C. White, S. Johnsen, and J. Jouzel, 1993: Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature*, **366(9)**, 552-554.

Gutowski, W.J., G.C. Hegerl, G.J. Holland, T.R. Knutson, L.O. Mearns, R.J. Stouffer, P.J. Webster, M.F. Wehner, and F.W. Zwiers, 2008: Causes of Observed Changes in Extremes and Projections of Future Changes in *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands*. T.R. Karl, G.A. Meehl, C.D. Miller, S.J. Hassol, A.M. Waple, and W.L. Murray (eds.). A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC.

Hales, K., J.D. Neelin, and N. Zeng, 2006: Interaction of vegetation and atmospheric dynamical mechanisms in the mid-holocene African monsoon. *Journal of Climate*, **19(16)**, 4105-4120.

Hall, R.I., P.R. Leavitt, R. Quinlan, A.S. Dixit, and J.P. Smol, 1999: Effects of agriculture, urbanization, and climate on water quality in the northern Great Plains. *Limnology and Oceanography*, **44**, 739-756.

Hamlet, A.F., P.W. Mote, M.P. Clark, and D.P. Lettenmaier, 2007: Twentieth-century trends in runoff, evapotranspiration, and soil moisture in the western United States: *Journal of Climate*, **20**, 1468-1486.

Hansen, Z.K., and G.D. Libecap, 2004: Small farms, externalities and the Dust Bowl of the 1930s. *Journal of Political Economy*, **112**, 665-694.

Harrison, S.P., J.E. Kutzbach, Z. Liu, P.J. Bartlein, B. Otto-Bliesner, D. Muhs, I.C. Prentice, and R.S. Thompson, 2003: Mid-Holocene climates of the Americas: A dynamical response to changed seasonality. *Climate Dynamics*, **20(7-8)**, 663-688.

Haug, G.H., K.A. Hughen, D.M. Sigman, L.C. Peterson, and U. Rohl. 2001. Southward migration of the Intertropical Convergence Zone through the Holocene. *Science* 293:1304-1308.

Hausmann, S., et al., 2002: Interactions of climate and land use documented in the varved sediments of Seebergsee in the Swiss Alps. *The Holocene*, **12**, 279-289.

Hegerl, G.C., T.J. Crowley, M. Allen, W.T. Hyde, H.N. Pollack, J. Smerdon, and E. Zorita, 2007: Detection of human influence on a new, validated 1500-year temperature reconstruction. *Journal of Climate*, **20(4)**, 650-666.

Hegerl, G.C., T.J. Crowley, S.K. Baum, K. Kim, and W.T. Hyde, 2003: Detection of volcanic, solar and greenhouse gas signals in paleo-reconstructions of Northern Hemispheric temperature. *Geophysical Research Letters*, **30(5)**, 1242, doi:10.1029/2002GL016635.

Heim, R.R., Jr., 2002: A review of twentieth-century drought indices used in the United States. *Bulletin of the American Meteorological Society*, **83(8)**, 1149-1165.

Held, I.M., and B.J. Soden, 2006: Robust responses of the hydrological cycle to global warming. *Journal of Climate*, **19**, 5686-5699.

Held, I.M., T.L. Delworth, J. Lu, K.L. Findell, and T.R. Knutson, 2005: Simulation of Sahel drought in the 20^{th} and 21^{st} centuries. *Proceedings of the National Academies of Science*, **102**, 17891-17896.

Herweijer, C., R. Seager, and E.R. Cook, 2006: North American droughts of the mid-to-late nineteenth century: A history, simulation and implication for medieval drought. *The Holocene*, **16(2)**, 159-171.

Hendrix, C.S., and S.M. Glaser, 2007: Trends and triggers: Climate, climate change and civil conflict in sub-Saharan Africa. *Political Geography*, **26**, 695-715.

Herweijer, C., R. Seager, E.R. Cook, and J. Emile-Geay, 2007: North American droughts of the last millennium from a gridded network of tree-ring data. *Journal of Climate*, **20**, 1353-1376.

Higgins, R.W., Y. Yao, and X.L. Wang, 1997: Influence of the North American monsoon system on the U.S. summer precipitation regime. *Journal of Climate*, **10**, 2600-2622.

Hirschboeck, K.K., 1989: Climate and floods. In: *Floods and Droughts: Hydrologic Perspectives on Water Issues*. U.S. Geologic Survey.

Hirschboeck, K.K., 2003: Floods, paleofloods, and drought: Insights from the Upper Tails. CLIVAR/PAGES/IPCC Drought Workshop, November 18-21, Tucson, AZ, 6 pp.

Hoerling, M., J. Hurrell, J. Eischeid, and A. Phillips, 2006: Detection and attribution of twentieth century northern and southern African rainfall change. *Journal of Climate*, **19**, 3989-4008.

Hoerling, M., and A. Kumar, 2003: A perfect ocean for drought. *Science*, **299**, 691-694.

Hoerling, M., and J. Eischeid, 2007: Past peak water in the Southwest. *Southwest Hydrology*, **6**, 18.

Hoerling, M., G. Hegerl, D. Karoly, A. Kumar and D. Rind, in prep. Attribution of the Causes of Climate Variations and Trends over North America during the Modern Reanalysis Period in Reanalysis of Historical Climate Data for Key Atmospheric Circulation Features: Implications for Attribution of Causes of Observed Change. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC. [no editors listed in Public Review Draft]

Hong, S.Y., and Z.T. Pan, 2000: Impact of soil moisture anomalies on seasonal, summertime circulation over North America in a regional climate model. *Journal of Geophysical Research*, **105**, 29625–29634.

House, P.K., R.H. Webb, V.R. Baker, and D.R. Levish, 2002: Ancient floods, modern hazards: Principles and applications of paleoflood hydrology. In: *Water Science and Application 5*. American Geophysical Union, Washington, DC.

Hoyt, J.C., 1936: Droughts of 1930-34. U.S. Geological Survey Water-Supply Paper 680, 106 p.

Huang, H.-P., R. Seager, and Y. Kushnir, 2005: The 1976/77 transition in precipitation over the Americas and the influence of tropical sea surface temperatures. *Climate Dynamics*, **24**, 721-740.

Hurrell, J.W., 1995: Decadal trends in the North Atlantic oscillation: Regional temperatures and precipitation. *Science*, **269**, 676-679.

IPCC, 2007: Climate Change 2007. The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, 996 pp.

Isenberg, A.C, 2000: The destruction of the bison: An environmental history 1750-1920. Cambridge University Press, Cambridge, United Kingdom, 206 pp.

Jansen, E., J. Overpeck, K.R. Briffa, J.-C. Duplessy, F. Joos, V. Masson-Delmotte, D. Olago, B. Otto-Bliesner, W.R. Peltier, S. Rahmstorf, R. Ramesh, D. Raynaud, D. Rind, O. Solomina, R. Villalba, and D. Zhang, 2007: Palaeoclimate. *In: Climate Change 2007. The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor, and H.L. Miller (eds.)]. Cambridge University Press. Cambridge, United Kingdom, pp. 433-497.

Jolly, D., S.P. Harrison, B. Damnati, and R. Bonnefille, 1998: Simulated climate and biomes of Africa during the late Quaternary: Comparison with pollen and lake status data. *Quaternary Science Reviews*, **17(6-7)**, 629-657.

Jones, P.D., and M.E. Mann, 2004: Climate over past millennia. *Reviews of Geophysics*, **42**, RG2002, doi:10.1029/2003RG000143.

Jones, R.N., L. Zhang, F.H.S. Chiew, and W.C. Boughton, 2006: Estimating the sensitivity of mean annual runoff to climate change using selected hydrological models. *Advances in Water Resources*, **29(10)**, 1419-1429.

Joos, F., and R. Spahni, 2008: Rates of change in natural and anthropogenic radiative forcing over the past 20,000 years: *Proceedings of the National Academy of Sciences of the United States of America*, **105**, 1425-1430.

Joussaume, S., K.E. Taylor, P. Braconnot, J.F.B. Mitchell, J.E. Kutzbach, S.P. Harrison, I.C. Prentice, A.J. Broccoli, A. Abe-Ouchi, P.J. Bartlein, C. Bonfils, B. Dong, J. Guiot, K. Herterich, C.D. Hewitt, D. Jolly, J.W. Kim, A. Kislov, A. Kitoh, M.F. Loutre, V. Masson, B. McAvaney, N. McFarlane, N. de Noblet, W.R. Peltier, J.Y. Peterschmitt, D. Pollard, D. Rind, J.F. Royer, M.E. Schlesinger, J. Syktus, S. Thompson, P. Valdes, G. Vettoretti, R.S. Webb, and U. Wyputta, 1999: Monsoon changes for 6000 years ago: Results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP). *Geophysical Research Letters*, **26(7)**, 859-862.

Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, A. Leetmaa, B. Reynolds, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K. Mo, C. Ropelewski, J. Wang, R. Jenne, and D. Joseph, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bulletin of the American Meteorological Society*, **77**, 437–471.

Kaplan, A., M.A. Cane, Y. Kushnir, A.C. Clement, M.B. Blumenthal, and B. Rajagopalan, 1998: Analyses of global sea surface temperature 1856-1991. *Journal of Geophysical Research*, **103(C9)**, 18567-18589.

Kaplan, J.O., N.H. Bigelow, I.C. Prentice, S.P. Harrison, P.J. Bartlein, T.R. Christensen, W. Cramer, N.V. Matveyeva, A.D. McGuire, D.F. Murray, V.Y. Razzhivin, B. Smith, D.A. Walker, P.M. Anderson, A.A. Andreev, L.B. Brubaker, M.E. Edwards, and A.V. Lozhkin, 2003: Climate change and Arctic ecosystems: 2. Modeling, paleodata-model comparisons, and future projections. *Journal of Geophysical Research*, **108(D19)**, 8171, doi:10.1029/2002JD002559.

Karl, T.R., and P.J. Young. 1987. The 1986 Southeast drought in historical perspective. Bulletin of the American Meteorological Society, **68(7)**, 773-778.

Karoly, D.J., A. Risbey, A. Reynolds, and K. Braganza, 2003: Global warming contributes to Australia's worst drought. *Australasian Science*, April, 14-17.

Karspeck, A.R., R. Seager, and M.A. Cane, 2004: Predictability of tropical Pacific decadal variability in an intermediate model. *Journal of Climate*, **17**, 1167-1180.

King, A.W., L. Dilling, G.P. Zimmerman, D.M. Fairman, R.A. Houghton, G. Marland, A.Z. Rose, and T.J. Wilbanks, 2007: What Is the Carbon Cycle and Why Care? In: *The First State of the Carbon Cycle Report (SOCCR): The North American Carbon Budget and Implications for the Global Carbon Cycle.*. A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research [King, A.W., L. Dilling, G.P. Zimmerman, D.M. Fairman, R.A. Houghton, G. Marland, A.Z. Rose, and T.J. Wilbanks (eds.)]. National Oceanic and Atmospheric Administration, National Climatic Data Center, Asheville, NC, USA, pp. 15-20.

Knight, J.R., R.J. Allan, C.K. Folland, M. Vellinga, and M.E. Mann, 2005: A signature of persistent natural thermohaline circulation cycles in observed climate. *Geophysical Research Letters*, **32**, doi:10.1029/2005GL024233.

Knight, D.B., R.E. Davis, S.C. Sheridan, D.M. Hondula, L.J. Sitka, M. Deaton, T.R. Lee, S.D. Gawtry, P.J. Stenger, F. Mazzei, and B.P. Kenny, 2008: Increasing frequencies of warm and humid air masses over the conterminous United States from 1948 to 2005. *Geophysical Research Letters*, **35**, doi:10.1029/2008GL033697.

Knox, J.C., 1972: Valley alluviation in southwestern Wisconsin. *Annals of the Association of American Geographers*, **62(3)**, 401-410.

Knox, J.C., 1985: Responses of floods to Holocene climatic-change in the Upper Mississippi Valley. *Quaternary Research*, **23(3)**, 287-300.

Knox, J.C., 1993: Large increases in flood magnitude in response to modest changes in climate. *Nature*, **361**, 430-432.

Knox, J.C., 2000: Sensitivity of modern and Holocene floods to climate change. *Quaternary Science Reviews*, **19**, 439-457.

Kochel, R.C., and V.R. Baker, 1982: Paleoflood. *Hydrology*, **215(4531)**, 353-361.

Kohfeld, K.E., and S.P. Harrison, 2000: How well can we simulate past climates? Evaluating the models using global palaeoenvironmental datasets. *Quaternary Science Reviews*, **19(1-5)**, 321-346.

Koster, R., P.A. Dirmeyer, Z. Guo, G. Bonan, E. Chan, P. Cox, C.T. Gordon, S. Kanae, E. Kowalczyk, E. Lawrence, et al., 2004: Regions of strong coupling between soil moisture and precipitation. *Science*, **305**, 1138-1140.

Kröpelin, S., Verschuren, D., Lezine, A. M., Eggermont, H., Cocquyt, C., Francus, P., Cazet, J. P., Fagot, M., Rumes, B., Russell, J. M., Darius, F., Conley, D. J., Schuster, M., von Suchodoletz, H., and Engstrom, D. R. (2008). Climate-driven ecosystem succession in the Sahara: the past 6000 years. Science, **320,** 765-768.

Kundzewicz, Z.W., L.J. Mata, N.W. Arnell, P. Döll, P. Kabat, B. Jiménez, K.A. Miller, T. Oki, Z. Sen, and I.A. Shiklomanov, 2007: Freshwater resources and their management. In: *Climate Change 2007. Impacts, Adaptation and Vulnerability. Contribution of Working Group II to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. [Parry, M.L., O.F. Canziani, J.P. Palutikof, P.J. van der Linden, and C.E. Hanson (eds.)]. Cambridge University Press, Cambridge, United Kingdom, 173-210.

Kunkel, K.E, 2003: North American trends in extreme precipitation. *Natural Hazards*, **29(2)**, 291-305.

Kunkel, K.E., S.A. Changnon, and J.R. Angel, 1994: Climatic aspects of the 1993 Upper Mississippi River Basin flood. *Bulletin of the American Meteorological Society*, **75(5)**, 811-822.

Kunkel, K.E., D.R. Easterling, K. Redmond, and K. Hubbard, 2003: Temporal variations of extreme precipitation events in the United States: 1895-2000. *Geophysical Research Letters*, **30(17)**.

Kunkel, K.E., P.D. Bromirski, H.E. Brooks, T. Cavazos, A.V. Douglas, D.R. Easterling, K.A. Emanuel, P.Ya. Groisman, G.J. Holland, T.R. Knutson, J.P. Kossin, P.D. Komar, D.H. Levinson, R.L. Smith, 2008: Observed Changes in Weather and Climate Extremes in *Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands*.

Kushnir, Y., 1994: Interdecadal variations in North Atlantic sea surface temperature and associated atmospheric conditions. *Journal of Climate*, **7**, 141-157.

Kutzbach, J.E., and Z. Liu, 1997: Response of the African monsoon to orbital forcing and ocean feedbacks in the middle Holocene. *Science*, **278**, 440-443.

Kutzbach, J.E., and W.F. Ruddiman, 1993: Model description, external forcing, and surface boundary conditions. *In: Global Climates since the Last Glacial Maximum*. [Wright, H.E., Jr., J.E. Kutzbach, T. Webb III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein (eds.)]. University of Minnesota Press, Minneapolis.

Kutzbach, J.E., and F.A. Street-Perrott, 1985: Milankovitch forcing of flucuations in the level of tropical lakes from 18 to 0 kyr BP. *Nature*, **317**, 130-134.

Kutzbach, J.E., and B.L. Otto-Bliesner, 1982: The sensitivity of the African-Asian monsoonal climate to orbital parameter changes for 9000 yr B.P. in a low-resolution general circulation model. *Journal of the Atmospheric Sciences*, **39(6)**, 1177-1188.

Laird, K.R., S.C. Fritz, K.A. Maasch, and B.F. Cumming, 1996: Greater drought intensity and frequency before A.D. 1200 in the northern Great Plains, U.S.A. *Nature*, **384**, 552-554.

LaMarche, V.C., Jr., 1974: Paleoclimatic inferences from long tree-ring records. *Science*, **183**, 1043-1048.

Lamb, H.H., 1965, The early medieval warm epoch and its sequel. *Palaeogeography, Palaeoclimatology, Palaeoecology*, **1**, 13-37.

Lau, N.C., A. Leetmaa, and M.J. Nath, 2006: Attribution of atmospheric variations in the 1997-2003 period to SST anomalies in the Pacific and Indian Ocean basins. *Journal of Climate*, **19**, 3607-3628.

Lebo, M.E., J.H. Reuter, C.R. Goldman, and C.L. Rhodes, 1994: Interannual variability of nitrogen limitation in a desert lake: influence of regional climate. *Canadian Journal of Fisheries and Aquatic Sciences*, **51**, 862-872.

Lettenmaier, D., D. Major, L. Poff, and S. Running, 2008: Water Resources. In: *The effects of climate change on agriculture, land resources, water resources, and* *biodiversity.* A Report by the U.S. Climate Change Science Program and the subcommittee on Global change Research. Washington, DC., USA, 362 pp.

Levis, S., G.B. Bonan, and C. Bonfils, 2004: Soil feedback drives the mid-Holocene North African monsoon northward in fully coupled CCSM2 simulations with a dynamic vegetation model. *Climate Dynamics*, **23(7-8)**, 791-802.

Libecap, G.D., and Z.K. Hansen, 2002: "Rain Folows the Plough" and dryfarming doctrine: The climate information problem and homestead failure in the upper Great Plains 1890-1925. *Journal of Economic History*, **62**, 86-120.

Li, W., R. Fu, and R.E. Dickinson, 2006: Rainfall and its seasonality over the Amazon in the 21st century as assessed by the coupled models for the IPCC AR4. *Journal of Geophysical Research,* **111**, doi:10.1029/2005JD006355.

Liu, A.Z., M. Ting, and H. Wang, 1998: Maintenance of circulation anomalies during the 1988 drought and 1993 floods over the United States. *Journal of the Atmospheric Sciences*, **55(17)**, 2810-2832.

Liu, Z., and M. Alexander, 2007: Atmospheric bridge, oceanic tunnel, and global climatic teleconnections. *Reviews of Geophysics*, **45**, RG2005, doi:10.1029/2005RG000172.

Liu, Z., E. Brady, and J. Lynch-Steiglitz, 2003: Global ocean response to orbital forcing in the Holocene. *Paleoceanography*, **18(2)**, 1054, doi:10.1029/2002PA000826.

Liu, Z., S.P. Harrison, J. Kutzbach, and B. Otto-Bliesner, 2004: Global monsoons in the mid-Holocene and oceanic feedback. *Climate Dynamics*, **22(2-3)**, 157-182.

Liu, Z., Y. Wang, R. Gallimore, F. Gasse, T. Johnson, P. deMenocal, J. Adkins, M. Notaro, I. C. Prentice, J. Kutzbach, R. Jacob, P. Behling, L. Wang, and E. Ong, 2007: Simulating the transient evolution and abrupt change of Northern Africa atmosphere-ocean-terrestrial ecosystem in the Holocene. *Quaternary Science Reviews*, **26(13-14)**, 1818-1837.

Liu, Z.Y., Y. Wang, R. Gallimore, M. Notaro, and I.C. Prentice, 2006: On the cause of abrupt vegetation collapse in North Africa during the Holocene: Climate variability vs. vegetation feedback. *Geophysical Research Letters*, **33**, L22709, doi:10.1029/2006GL028062.

Lobell, D.B., M.B. Burke, C. Tebaldi, M.D. Mastrandrea, W.P. Falcon, and R.L. Naylor, 2008: Prioritizing climate change adaptation needs for food security in 2030. *Science*, **319**, 607-610.

Lu, J., G. Vecchi, and T. Reichler, 2007: Expansion of the Hadley Cell under global warming. *Geophysical Research Letters*, **34**, doi:10.1029/2006GL028443.

MacDonald, G.M., D.W. Beilman, K.V. Kremenetski, Y.W. Sheng, L.C. Smith, and A.A. Velichko, 2006: Rapid early development of circumarctic peatlands and atmospheric CH_4 and CO_2 variations. *Science*, **314(5797)**, 285-288.

Maguire, R., 2005: The effects of drought in lower basin river operations. *Southwest Hydrology*, March/April, 22-23, 38.

Mann, M.E., and K.A. Emanuel, 2005: Hurricane trends linked to climate change. *Eos*, **87**, 233.

Mann, M.E., M.A. Cane, S.E. Zebiak, and A. Clement, 2005: Volcanic and solar forcing of the tropical Pacific over the past 1000 years. *Journal of Climate*, **18**, 447-456.

Manuel, J., 2008: Drought in the Southeast: Lessons for water management. Environmental Health Perspective, **116**, A168-A171.

Mason, J.A., J.B. Swinehart, R.J. Goble, and D.B. Loope, 2004: Late-Holocene dune activity linked to hydrological drought, Nebraska Sand Hills, USA. *Holocene*, **14(2)**, 209-217.

Maruer, E.P, 2007: Uncertainty in hydrologic impacts of climate change in the Sierra Nevada, California, under two emissions scenarios. *Climatic Change*, **82**, 309-325.

Mayewski, P.A., W. Karlén, K.A. Maasch, L.D. Meeker, E.A. Meyerson, F. Gasse, S. van Kreveld, K. Holmgren, J. Lee-Thorp, G. Rosqvist, F. Rack, M. Staubwasser, R.R. Schneider, E.J. Steig, E.E. Rohling, and J.C. Stager, 2004: Holocene climate variability. *Quaternary Research*, **62(3)**, 243-255.

McCabe, G.J., M.A. Palecki, and J.L. Betancourt, 2004: Pacific and Atlantic Ocean influences on multidecadal drought frequency in the United States. *Proceedings of the National Academies of Science*, **101(12)**, 4136-4141.

Meehl, G.A., T.F. Stocker, W.D. Collins, P. Friedlingstein, A.T. Gaye, J.M. Gregory, A. Kitoh, R. Knutti, J.M. Murphy, A. Noda, S.C.B. Raper, I.G. Watterson, A.J.

Weaver and Z.-C. Zhao, 2007: Global climate projections. In: *Climate Change 2007. The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change.* [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York.

Meko, D.M., C.A. Woodhouse, C.A. Baisan, T. Knight, J.J. Lukas, M.K. Hughes, and M.W. Salzer, 2007: Medieval drought in the upper Colorado River Basin. *Geophysical Research Letters*, **34**, L10705.

Miao, X.D., J.A. Mason, J.B. Swinehart, D.B. Loope, P.R. Hanson, R.J. Goble, and X.D. Liu, 2007: A 10,000 year record of dune activity, dust storms, and severe drought in the central Great Plains. *Geology*, **35(2)**, 119-122.

Mock, C.J., and A.R. Brunelle-Daines. 1999. A modern analogue of western United States summer palaeoclimate at 6000 years before present. *The Holocene*, **9(5)**, 541-545.

Mosley, M.P., and A.I. McKerchar, 1993: Streamflow. In: Handbook of Hydrology. [Maidment, D.R. (ed.)]. McGraw-Hill, New York, 8.1-8.39

Morrill, C., J.T. Overpeck, and J.E. Cole, 2003: A synthesis of abrupt changes in the Asian summer monsoon since the last deglaciation. *Holocene*, **13(4)**, 465-476.

Moore, J.N., J.T. Harper, and M.C. Greenwood, 2007: Significance of trends toward earlier snowmelt runoff, Columbia and Missouri Basin headwaters, western United States. *Geophysical Research Letters*, **34(16)**, doi:10.1029/2007GL032022.

Mote, P., A. Hamlet, and E. Salathé, 2008: Has spring snowpack declined in the Washinton Cascades. *Hydrology and Earth System Sciences*, **12**, 193-206.

Mote, P.W., 2006, Climate-driven variability and trends in mountain snowpack in western North America: *Journal of Climate*, **19**, 6209-6220.

Mote, P.W., A.F. Hamlet, M.P. Clark, and D.P. Lettenmaier, 2005: Declining mountain snowpack in western north America. *Bulletin of the American Meteorological Society*, **86**, 39-49.

Namias, J. 1966. Nature and possible causes of the northeastern United States drought during 1962-65. Monthly Weather Review, **94(9)**, 543-554.

Namias, J., 1991: Spring and summer 1988 drought over the Great Plains: Causes and predictions. *Journal of Climate*, **4**, 54-65.

National Assessment, 2000. Water: The potential consequences of climate variability and change for the water resources of the United States. Report of the Water Sector Assessment Team of the National Assessment of the Potential Consequences of Climate Variability and Change for the U.S. Global Change Research Program. [Gleick, P. (ed.)]. U.S. Global Change Research Program, 151 pp.

National Integrated Drought Information System Implementation Team, 2007: A pathway for national resilience. http://www.drought.gov/pdf/NIDIS-IPFinal-June07.pdf.

Nemec, J., and J.C. Schaake, 1982: Sensitivity of water resource systems to climate variation. *Hydrologic Sciences Journal*, **27**, 327-343.

Newman, M., G.P. Compo, and M.A. Alexander, 2003: ENSO-forced variability of the Pacific Decadal Oscillation. *Journal of Climate*, **16**, 3853-3857.

Nicholls, N., 2004: The changing nature of Australian droughts. *Climatic Change*, **63**, 323-336.

Nicholson, S.E., B. Some, and B. Kone, 2000: An analysis of recent rainfall conditions in West Africa, including the rainy seasons of the 1997 and 1998 La Niña years. *Journal of Climate*, **13**, 2628-2640.

Nigam, S., and A. Ruiz-Barradas, 2006: Great Plains hydroclimate variability: The view from North American regional reanalysis. *Journal of Climate*, **19**, 3004-3010.

Nordas, R., and N.P. Gleditsch, 2007: Climate change and conflict. *Political Geography*, **26**, 627-638.

Notaro, M., Y. Wang, Z. Liu, R. Gallimore, and S. Levis, 2008: Combined statistical and dynamical assessment of simulated vegetation-rainfall interactions in North Africa during the mid-Holocene. *Global Change Biology*, **14(2)**, 347–368, doi:10.1111/j.1365-2486.2007.01495.x

O'Connor, J.E., and J.E. Costa, 2004: The world's largest floods, past and present: Their causes and magnitudes. U.S. Geological Survey. http://purl.access.gpo.gov/GPO/LPS56009.

O'Connor, J.E., J.E. Costa, and Geological Survey (U.S.), 2003: Large floods in the United States where they happen and why. U.S. Geological Survey. http://purl.access.gpo.gov/GPO/LPS37149.

Oglesby, R.J., 1991: Springtime soil moisture, natural climatic variability, and North American drought as simulated by the NCAR community climate model 1. *Journal of Climate*, **4(9)**, 890-897.

Oglesby, R.J., and D.J. Erickson III, 1989: Soil moisture and the persistence of North American drought. *Journal of Climate*, **2(11)**, 1362-1380.

Ostler, D.A, 2005: Upper Colorado River basin perspectives on the drought. *Southwest Hydrology*, March/April, 18, 29.

Otto-Bliesner, B.L., E.C. Brady, G. Clauzet, R. Tomas, S. Levis, and Z. Kothavala, 2006: Last Glacial Maximum and Holocene climate in CCSM3. *Journal of Climate*, **19(11)**, 2526-2544.

Overland, J.E., D.B. Percival, and H.O. Mofjeld, 2006: Regime shifts and red noise in the North Pacific. In: Deep-Sea Research Part I. *Oceanographic Research Papers*, **53(4)**, 582-588.

Pal, J.S., F. Giorgi, and X.Q. Bi, 2004: Consistency of recent European summer precipitation trends and extremes with future regional climate projections. *Geophysical Research Letters*, **31(13)**.

Pal, J.S., and E.A.B. Eltahir, 2002: Teleconnections of soil moisture and rainfall during the 1993 midwest summer flood. *Geophysical Research Letters*, **29(18)**.

Palmer, W.C., 1965: Meteorological drought. Weather Bureau Research Paper 45, U.S. Department of Commerce, Washington, DC, 58 pp.

Palmer, T.N., and C. Brankoviç, 1989: The 1988 US drought linked to anomalous sea surface temperature. *Nature*, **338(6210)**, 54-57.

Peteet, D., 2001: Late glacial climate variability and general circulation model (GCM) experiments: An overview. In: Interhemispheric Climate Linkages. [Markgraf, V. (ed.)]. Academic Press, San Diego.

Pederson, D.C., D.M. Peteet, D. Kurdyla, and T. Guilderson, 2005: Medieval Warming, Little Ice Age, and European impact on the environment during the last millennium in the lower Hudson Valley, New York, USA. Quaternary Research, **63**, 238-249.

Peterson, T.C., D.M. Anderson, S.J. Cohen, M. Cortez-Vázquez, R.J. Murnane, C. Parmesan, D. Phillips, R.S. Pulwarty, J.M.R. Stone, 2008: Why Weather and Climate Extremes Matter in Weather and Climate Extremes in a Changing Climate. Regions of Focus: North America, Hawaii, Caribbean, and U.S. Pacific Islands. T.R. Karl, G.A. Meehl, C.D. Miller, S.J. Hassol, A.M. Waple, and W.L. Murray (eds.). A Report by the U.S. Climate Change Science Program and the Subcommittee on Global Change Research, Washington, DC.

Pilgrim, D.H., and I. Cordery, 1993: Flood runoff. In: Handbook of Hydrology. [Maidment, D.R. (ed.)]. McGraw-Hill, New York, 9.1-9.42.

Poore, R.Z., M.J. Pavich, and H.D. Grissino-Mayer, 2005: Record of the North American southwest monsoon from Gulf of Mexico sediment cores. *Geology*, **33(3)**, 209-212.

Raleigh, C., and H. Urdal, 2007: Climate change, environmental degradation and armed conflict. *Political Geography*, **26**, 674-694.

Rasmussen, J.B.T., V.J. Polyak, and Y. Asmerom, 2006: Evidence for Pacific-modulated precipitation variability during the late Holocene from the southwestern USA. *Geophysical Research Letters*, **33(8)**.

Rayner, N.A., D.E. Parker, E.B. Horton, C.K. Folland, L.V. Alexander, D.P. Rowell, E.C. Kent, and A. Kaplan, 2003: Globally complete analyses of sea surface temperature, sea ice and night marine air temperature, 1871-2000. *Journal of Geophysical Research*, **108**, 4407, doi:10.1029/2002JD002670.

Renssen, H., V. Brovkin, T. Fichefet, and H. Goosse, 2006: Simulation of the Holocene climate evolution in Northern Africa: The termination of the African Humid Period. *Quaternary International*, **150**, 95-102.

Rosenzweig, C., D.C. Major, K. Demong, C. Stanton, R. Horton, and M. Stults, 2007: Managing climate change risks in New York City's water systems: Assessment and adaptation planning. *Mitigation and Adaptation Strategies for Global Change*, DOI 10.1007/s11027-006-9070-5.

Ruddiman, W.F, 2006: Orbital changes and climate. *Quaternary Science Reviews*, **25(23-24)**, 3092-3112.

Rudnick, D.L., and R.E. Davis, 2003: Red noise and regime shifts. *Deep_Sea Research I*, **50**, 691-699.

Ruiz-Barradas, A., and S. Nigam, 2005: Warm season rainfall over the U.S. Great Plains in observations, NCEP and ERA-40 Reanalyses, and NCAR and NASA atmospheric model simulations. *Journal of Climate*, **18**, 1808-1830.

Saltzman, B., 2002: Dynamical paleoclimatology: Generalized theory of global climate change. Academic Press, San Diego.

Schindler, D.W., S.E. Bayley, B.R. Parker, K.G. Beaty, and D.R. Cruikshank, 1996a: The effects of climatic warming on the properties of boreal lakes and streams at the Experimental Lakes Area, northwestern Ontario. *Limnology and Oceanography*, **41**, 1004-1017.

Schindler, D.W., P.J. Curtis, B.R. Parker, and M.P. Stainton, 1996b: Consequences of climate warming and lake acidification for UV-B penetration in North American boreal lakes. *Nature*, **379**, 705-708.

Schneider, E.K., 1977: Axially symmetric steady-state models of the basic state for instability and climate studies. Part II: Non-linear calculations. *Journal of the Atmospheric Sciences*, **34**, 280-296.

Schneider, N., and B.D. Cornuelle, 2005: The forcing of the Pacific decadal oscillation. *Journal of Climate*, **18(21)**, 4355-4373.

Schubert, S.D., M.J. Suarez, P.J. Region, R.D. Koster, and J.T. Bacmeister. 2004a. Causes of long-term drought in the United States Great Plains. *Journal of Climate*, **17**, 485-503.

Schubert, S.D., M.J. Suarez, P.J. Region, R.D. Koster, and J.T. Bacmeister. 2004b. On the cause of the 1930s Dust Bowl. *Science*, **303**, 1855-1859.

Seager, R., 2007: The turn-of-the-century drought across North America: Dynamics, global context and past analogues. *Journal of Climate*, **20**, 5527-5552.

Seager, R., A. Tzanova, and J. Nakamura, 2008: Drought in the Southeastern United States: Causes, variability over the last millennium and the potential for future hydroclimatic change. Journal of Climate (in review).

Seager, R., N. Harnik, Y. Kushnir, W.A. Robinson, and J. Miller, 2003: Mechanisms of hemispherically symmetric climate variability. *Journal of Climate*, **16**, 296-2978.

Seager, R., N. Harnik, W.A. Robinson, Y. Kushnir, M. Ting, H.P. Huang, and J. Velez. 2005a, Mechanisms of ENSO-forcing of hemispherically symmetric precipitation variability. *Quarterly Journal of the Royal Meteorological Society*, **131**, 1501-1527.

Seager, R., Y. Kushnir, C. Herweijer, N. Naik and J. Velez. 2005b, Modeling of tropical forcing of persistent droughts and pluvials over western North America: 1856-2000. *Journal of Climate*, **18**, 4068-4091.

Seager, R., R. Burgman, Y. Kushnir, A. Clement, N. Naik, and J. Velez. 2007a. Tropical Pacific forcing of North American medieval megadroughts: Testing the concept with an atmosphere model forced by coral-reconstructed SSTs. *Journal of Climate*, submitted.

Seager, R., N. Graham, C. Herweijer, A.L. Gordon, Y. Kushnir, and E. Cook. 2007b. Blueprints for Medieval hydroclimate. *Quaternary Science Reviews*, **26**, 2322-2336.

Seager, R., Y. Kushnir, M. Ting, M. Cane, N. Naik, and J. Velez, 2007c. Would advance knowledge of 1930s SSTs have allowed prediction of the Dust Bowl drought? *Journal of Climate*, in press.

Seager, R., M. Ting, I. Held, Y. Kushnir, J. Lu, G. Vecchi, H-P. Huang, N. Harnik, A. Leetmaa, N-C. Lau, C. Li, J. Velez, and N. Naik. 2007d. Model projections of an imminent transition to a more arid climate in southwestern North America. *Science*, **316**, 1181-1184.

Seidel, D.J., Q. Fu, W.J. Randel, and T.J. Reichler, 2008: Widening of the tropical belt in a changing climate. *Nature, Geoscience*, **1**, 21-24. Published online: 2 December 2007 doi:10.1038/ngeo.2007.38.

Shin, S-I., P.D. Sardeshmukh, R.S. Webb, R.J. Oglesby, and J.J. Barsugli, 2006: Understanding the mid-Holocene climate. *Journal of Climate*, **19(12)**, 2801-2818, doi:10.1175/JCLI3733.1.

Shinker, J.J., P.J. Bartlein, and B. Shuman, 2006: Synoptic and dynamic climate controls of North American mid-continental aridity. *Quaternary Science Reviews*, **25**, 1401-1417.

Shuman, B., P. Bartlein, N. Logar, P. Newby, and T. Webb III, 2002: Parallel climate and vegetation responses to the early Holocene collapse of the Laurentide Ice Sheet. *Quaternary Science Reviews*, **21(16-17)**, 1793-1805.

Shuman, B., P.J. Bartlein, and T. Webb III, 2005: The magnitudes of millennial- and orbital-scale climatic change in eastern North America during the Late Quaternary. *Quaternary Science Reviews*, **24(20-21)**, 2194-2206.

Shuman, B., and B. Finney, 2006: Late-Quaternary lake-level changes in North America. In: *Encyclopedia of Quaternary Science*. [Elias, S. (ed.)]. Elsevier, Amsterdam.

Shuman, B., P.J. Bartlein, and T. Webb III, 2007: Response to "Comments on: 'The magnitude of millennial- and orbital-scale climatic change in eastern North America during the Late-Quaternary' by Shuman et al." *Quaternary Science Reviews*, **26(1-2)**, 268-273.

Shuman, B., J.W. Williams, N.S. Diffenbaugh, M. Ashfaq, and P.J. Bartlein, in review: The effects of insolation and soil-moisture behavior on moisture patterns and variability in the central United States during the mid-Holocene. *Quaternary Science Reviews*.

Sinha, A., K. Cannariato, L.D Stott, H. Chaing, R.L. Edwards, M.G. Yadava, R. Ramesh, and I.B. Singh. 2007. A 900-year (600 to 1500 A.D.) record of the Indian summer monsoon precipitation from the core monsoon zone of India. *Geosphysical Research Letters* 34, doi:10.1029/2007GL030431.

Soden, B.J., R.T. Wetherald, L. Stenchikov, and A. Robock, 2002: Global cooling after the reuption of Mount Minatubo: A test of climate feedback by water vapor. *Science*, **296**, 727-730.

Stahle, D.W., M.K. Cleaveland, and J.G. Hehr, 1988: North Carolina climate change reconstructed from tree rings: A.D. 372 to 1985. Science, **240**, 1517-1520.

Stahle, D.W., M.K. Cleaveland, D.B. Blanton, M.D. Therrell, and D.A. Gay, 1998: The lost colony and Jamestown droughts. Science, 280(5363), 564-567.

Stahle, D.W., E.R. Cook, M.K. Cleaveland, M.D. Therrell, D.M. Meko, H.D. Grissino-Mayer, E. Watson, and B.H. Luckman, 2000: Tree-ring data document 16th century megadrought over North America. *Eos, Transactions of the American Geophysical Union*, **81(12)**, 121, 125.

Stahle, D.W., F.K. Fye, E.R. Cook, and R.D. Griffin, 2007: Tree-ring reconstructed megadroughts over North America since AD 1300. *Climatic Change*, doi 10.1007/s10584-006-9171-x (in press).

Stine, S., 1994: Extreme and persistent drought in California and Patagonia during mediaeval time. *Nature*, **369**, 546-549.

Stockton C.W., and G.C. Jacoby, 1976: Long-term surface-water supply and streamflow trends in the Upper Colorado River basin. Lake Powell Research Project Bulletin No. 18. National Science Foundation, 70 pp.

Stone, J.R., and S.C. Fritz, 2006: Multidecadal drought and Holocene climate instability in the Rocky Mountains. *Geology*, **34(5)**, 409-412.

Stuiver, M., P.M. Grootes, and T.F. Braziunas. 1995. The GISP2 & δ^{18}O climate record of the past 16 500 years and the role of the sun, ocean, and volcanoes. *Quaternary Research*, **44(3)**, 341-354.

Sutton, R.T., and D.R.L. Hodson, 2005: Atlantic Ocean forcing of North American and European summer climate. *Science*, **309**, 115-118.

Swetnam, T.W., and J.L. Betancourt, 1998: Mesoscale disturbance and ecological response to decadal climate variability in the American southwest. *Journal of Climate*, **11**, 3128-3147.

Texier, D., N. de Noblet, S.P. Harrison, A. Haxeltine, D. Jolly, S. Joussaume, F. Laarif, I.C. Prentice, and P. Tarasov, 1997: Quantifying the role of biosphere-atmosphere feedbacks in climate change: coupled model simulations for 6000 years BP and comparison with palaeodata for northern Eurasia and northern Africa. *Climate Dynamics*, **13**, 865-882.

Thompson, R.S., C. Whitlock, P.J. Bartlein, S.P. Harrison, and W.G. Spaulding, 1993: Climatic changes in western United States since 18,000 yr B.P. In: Global

Climates since the Last Glacial Maximum. [Wright, H.E., Jr., J.E. Kutzbach, T. Webb III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein (eds.)]. University of Minnesota Press, Minneapolis.

Trenberth, K.E., and C.J. Guillemot, 1996: Physical processes involved in the 1988 drought and 1993 floods in North America. *Journal of Climate*, **9(6)**, 1288-1298.

Trenberth, K.E., G.W. Branstator, D. Karoly, A. Kumar, N.-C. Lau, and C. Ropelewski. 1998. Progress during TOGA in understanding and modeling global teleconnections associated with tropical sea surface temperature. *Journal of Geophysical Research*, **103**, 14291-14324.

Trenberth, K.E., A. Dai, R.M. Rasmussen, and D.B. Parsons, 2003: The changing character of precipitation. *Bulletin of the American Meteorological Society*, **84**, 1205-1217.

Trenberth, K.E., P.D. Jones, P. Ambenje, R. Bojariu, D. Easterling, A. Klein Tank, D. Parker, F. Rahimzadeh, J.A. Renwick, M. Rusticucci, B. Soden and P. Zhai, 2007: Observations: Surface and atmospheric climate change. In: *Climate Change 2007. The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. [Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)]. Cambridge University Press, Cambridge, United Kingdom, and New York.

USBR (U.S. Department of the Interior, Bureau of Reclamation), 2005: Water 2025: Preventing crises and conflict in the west. http://www.doi.gov/water2025/.

USBR (U.S. Department of the Interior, Bureau of Reclamation), 2007: Draft environmental impact statement, Colorado River interim guidelines for lower basin shortages and coordinated operations for Lake Powell and Lake Mead.

Vecchi, G.A., B.J. Soden, A.T. Wittenberg, I.M. Held, A. Leetmaa, and M.J. Harrison, 2006: Weakening of tropical Pacific atmospheric circulation due to anthropogenic forcing. *Nature*, **441**, 73-76.

Vera, C., W. Higgins, J. Amador, T. Ambrizzi, R. Garreaud, D. Gochis, D. Gutzler, D. Lettenmaier, J. Marengo, C.R. Mechoso, J. Nogues-Paegle, P.L.S. Dias, and C.

Zhang, 2006: Toward a unified view of the American Monsoon Systems. *Journal of Climate*, **19(20)**, 4977-5000.

Viau, A.E., K. Gajewski, M.C. Sawada, and P. Fines, 2006: Millennial-scale temperature variations in North America during the Holocene. *Journal of Geophysical Research*, **111**, D09102, doi:10.1029/2005JD006031.

Wang, C., D.B. Enfield, S.K. Lee, and C.W. Landsea, 2006: Influences of the Atlantic warm pool on western hemisphere summer rainfall and Atlantic hurricanes. *Journal of Climate*, **19(12)**, 3011-3028.

Wang, Y.J., H. Cheng, R.L. Edwards, Y.Q. He, X.G. Kong, Z.S. An, J.Y. Wu, M.J. Kelly, C.A. Dykoski, and X.D. Li, 2005: The Holocene Asian monsoon: Links to solar changes and North Atlantic climate. *Science*, **308(5723)**, 854-857.

Wang, Y., M. Notaro, Z. Liu, R. Gallimore, S. Levis, and J.E. Kutzbach, 2007: Detecting vegetation-precipitation feedbacks in mid-Holocene North Africa from two climate models. *Climate of the Past Discussions*, **3**, 961-975.

Webb, T., III, K.H. Anderson, P.J. Bartlein, and R.S. Webb, 1998: Late Quaternary climate change in eastern North America: a comparison of pollen-derived estimates with climate model results. *Quaternary Science Reviews*, **17(6-7)**, 587-606.

Webb, T., III, P.J. Bartlein, S.P. Harrison, and K.H. Anderson, 1993a: Vegetation, lake levels, and climate in eastern North America for the past 18,000 years. In: *Global Climates since the Last Glacial Maximum*. [Wright, H.E., Jr., J.E. Kutzbach, T. Webb III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein (eds.)]. University of Minnesota Press, Minneapolis.

Webb, T., III, E.J. Cushing, and H.E. Wright, Jr., 1983: Holocene changes in the vegetation of the Midwest. In: *Late-Quaternary Environments of the United States, vol. 2*. [Wright, H.E., Jr. (ed.)]. University of Minnesota Press, Minneapolis.

Webb, T., III, W.F. Ruddiman, F.A. Street-Perrott, V. Markgraf, J.E. Kutzbach, P.J. Bartlein, H.E. Wright, Jr., and W.L. Prell. 1993b. Climatic changes during the past 18,000 years; Regional syntheses, mechanisms, and causes. In: *Global Climates Since the Last Glacial Maximum*. [Wright, H.E., Jr., J.E. Kutzbach, T.

Webb III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein (eds.)]. University of Minnesota Press, Minneapolis.

Weldeab, S., D.W. Lea, R.R. Schneider, and N. Andersen, 2007: 155,000 years of West African Monsoon and ocean thermal evolution. *Science*, **316**, doi:10.1126/science.1140461.

West, E., 1995: The way to the west: Essays on the Great Plains. New Mexico Press, 244 pp.

Western Governor's Association, 2004, Creating a drought warning system for the 21st century: The National Integrated Drought Information System. http://www.westgov.org/wga/publicat/nidis.pdf

Wilby, R.L., and T.M.L. Wigley, 2002: Future changes in the distribution of daily precipitation totals across North America. *Geophysical Research Letters*, **29**, doi:10.1029/2001GL013048.

Williams, J.W., 2002: Variations in tree cover in North America since the last glacial maximum. *Global and Planetary Change*, **35**, 1-23.

Williams, J.W., B.N. Shuman, T. Webb III, P.J. Bartlein, and P.L. Leduc, 2004: Late-quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs*, **74(2)**, 309-334.

Wohl, E.E., 2000: Inland flood hazards: Human, riparian and aquatic communities. Cambridge University Press, Cambridge, United Kingdom, and New York.

Wohlfahrt, J., S.P. Harrison, and P. Braconnot, 2004: Synergistic feedbacks between ocean and vegetation on mid- and high-latitude climates during the mid-Holocene. *Climate Dynamics*, **22(2-3)**, 223-238.

Woodhouse, C.A., and J.T. Overpeck, 1998: 2000 years of drought variability in the central United States. *Bulletin of the American Meteorological Society*, **79**, 2693-2714.

Woodhouse C.A., S.T. Gray, and D.M. Meko, 2006: Updated streamflow reconstructions for the Upper Colorado River basin. *Water Resources Research*, **42**, W05415, doi:10.1029/2005WR004455.

Woodhouse, C.W., K.E. Kunkel, D.R. Easterling, and E.R. Cook, 2005: The 20th century pluvial in the western United States. *Geophysical Research Letters*, **32**, L07701, doi:10.1029/2005GL022413.

Worster, D., 1979: Dust Bowl: The Southern Plains in the 1930s. Oxford University Press, New York, 277 pp.

Worster, D., 1985: Rivers of empire: Water, aridity and the growth of the American West. Oxford University Press, New York, 416 pp.

Wright, H.E., Jr., J.E. Kutzbach, T. Webb III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein, 1993: Global climates since the last glacial maximum. [Wright, H.E., Jr., J.E. Kutzbach, T. Webb III, W.F. Ruddiman, F.A. Street-Perrott, and P.J. Bartlein (eds.)]. University of Minnesota Press, Minneapolis.

Wright, H.E., Jr., F.S. Hu, I. Stefanova, J. Tian, and T.A. Brown, 2004: A chronological framework for the Holocene vegetational history of central Minnesota: The Steel Lake pollen record. *Quaternary Science Reviews*, **23(5-6)**, 611-626.

Yin, J., 2005: A consistent poleward shift of the storm tracks in simulations of 21st century climate. *Geophysical Research Letters*, **32**, L06105, doi:10.1029/2004GL022058.

Zhang, X., F.W. Zwiers, G.C. Hegerl, F.H. Lambert, N.P. Gillett, and S. Solomon, 2007: Detection of human influence on 20th century precipitation trends. *Nature*, **448**, 461-466, doi:10.1038.

Zhang, Y., J.M. Wallace, and D.S. Battisti, 1997: ENSO-like decade-to-century scale variability: 1900-93. *Journal of Climate*, **10**, 1004-1020.

Zhao, Y., P. Braconnot, O. Marti, S.P. Harrison, C. Hewitt, A. Kitoh, Z. Liu, U. Mikolajewicz, B. Otto-Bliesner, and S.L. Weber, 2005: A multi-model analysis of the role of the ocean on the African and Indian monsoon during the mid-Holocene. *Climate Dynamics*, **25(7-8)**, 777-800.

Zhao, Y., P. Braconnot, S.P. Harrison, P. Yiou, and O. Marti, 2007: Simulated changes in the relationship between tropical ocean temperatures and the western African monsoon during the mid-Holocene. *Climate Dynamics*, **28(5)**, 533-551.

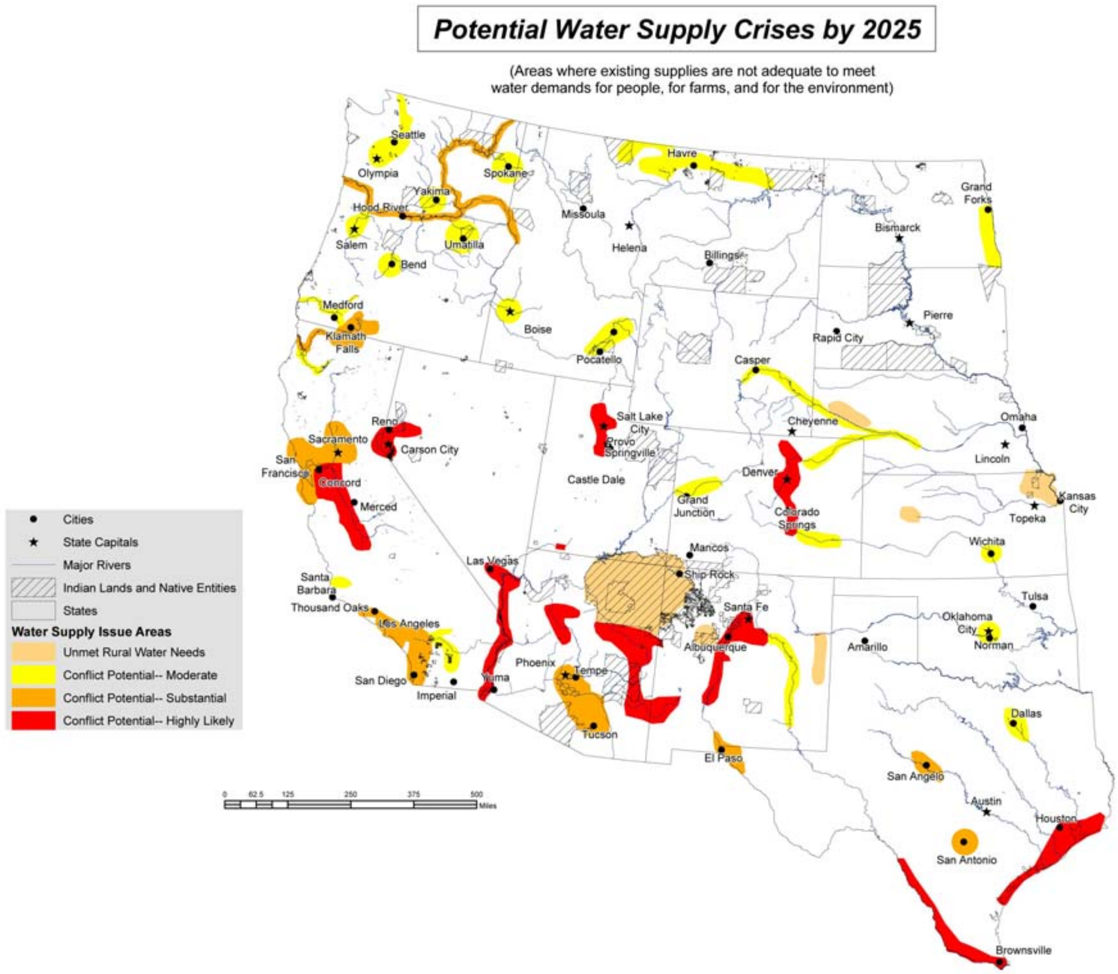

Figure 3.1. Interior Department analysis of regions in the West where water supply conflicts are likely occur by 2025 based on a combination of technical and other factors, including population trends and potential endangered species' needs for water. The red zones indicate areas where the conflicts are most likely to happen. See DOI Water 2025 Status Report (U.S. Department of Interior , *Bureau of Reclamation, 2005*; http://www.usbr.gov/water2025/report.html) for details. Note: There is an underlying assumption of a statistically stationary climate.

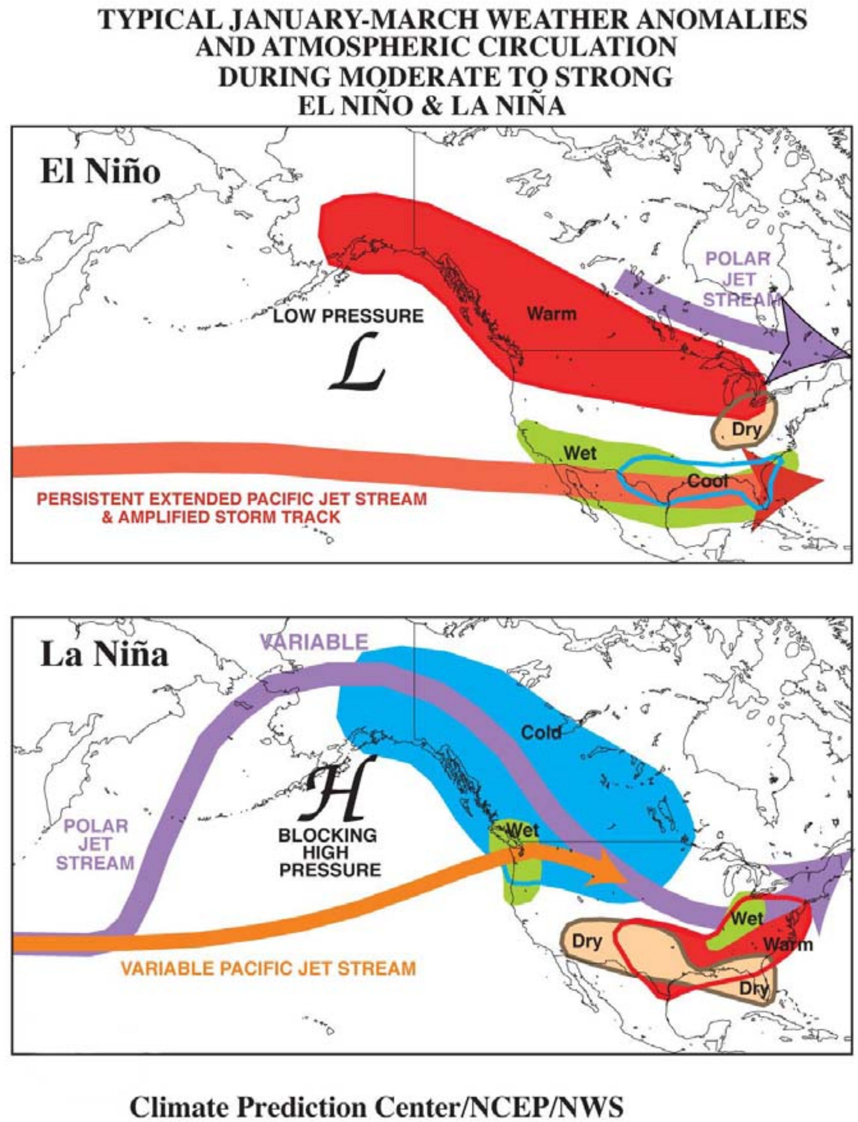

Figure 3.2. Schematic maps showing the influence of El Niño/Southern Oscillation (ENSO) variability on regional climate over North America. Warm episodes (El Niños) result in cooler-wetter conditions from the Southwest to the Southeast during the winter season. Warm-dry conditions prevail over the same region during cold episodes (La Niñas), also during winter. (From http://www.cpc.noaa.gov/products/analysis_monitoring/ensocycle/nawinter.shtml)

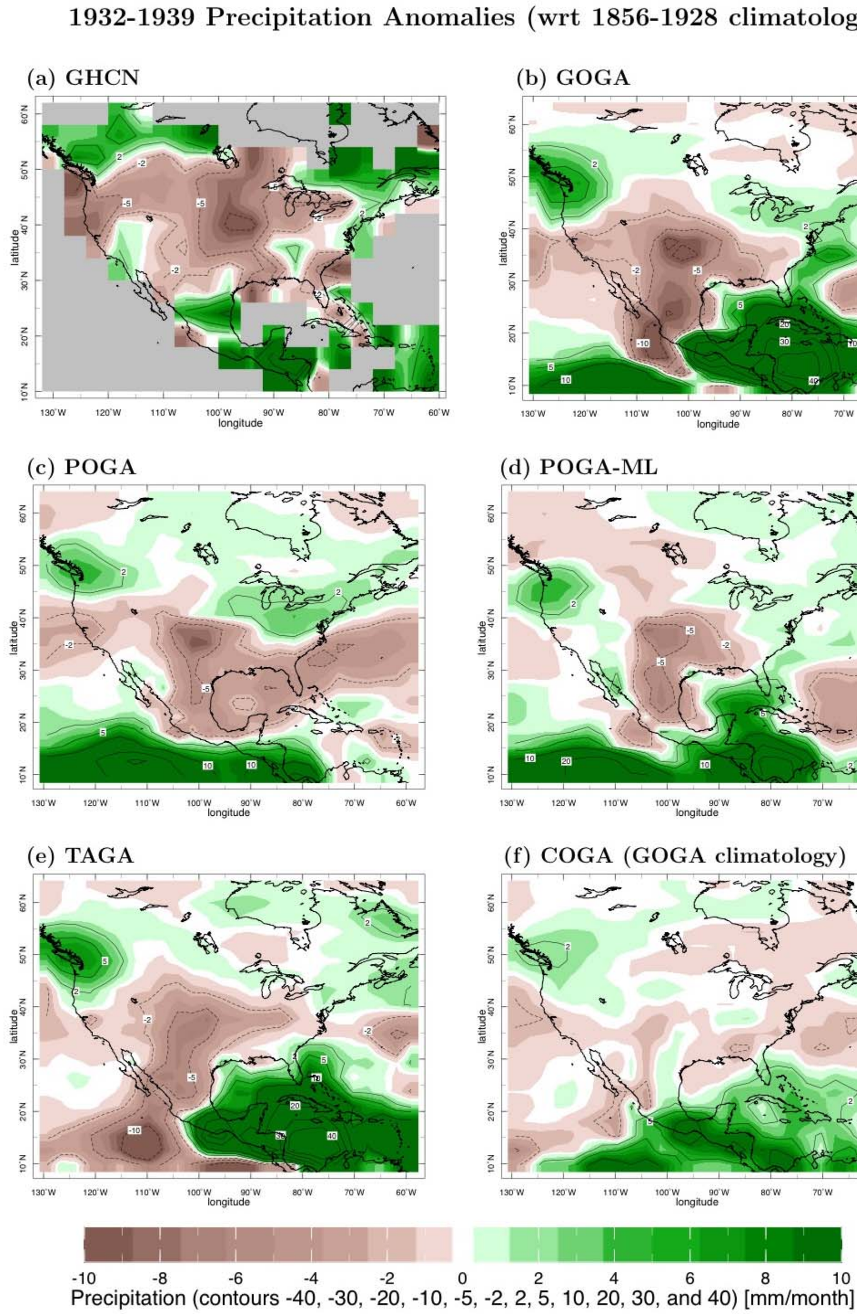

Figure 3.3. The observed (top left) and modeled precipitation anomalies during the Dust Bowl (1932 to 1939) relative to an 1856 to 1928 climatology. Observations are from Global Historical Climatology Network (GHCN). The modeled values are model ensemble means from the ensembles with global sea surface temperature (SST) forcing (GOGA), tropical Pacific forcing (POGA), tropical Pacific forcing and a mixed layer ocean elsewhere (POGA-ML), tropical Atlantic forcing (TAGA), and forcing with land and atmosphere initialized in January 1929 from the GOGA run and integrated forward

with the 1856-1928 climatological SST (COGA). The model is the NCAR CCM3. Units are millimeters (mm) per month. From *Seager et al. (2007c)*.

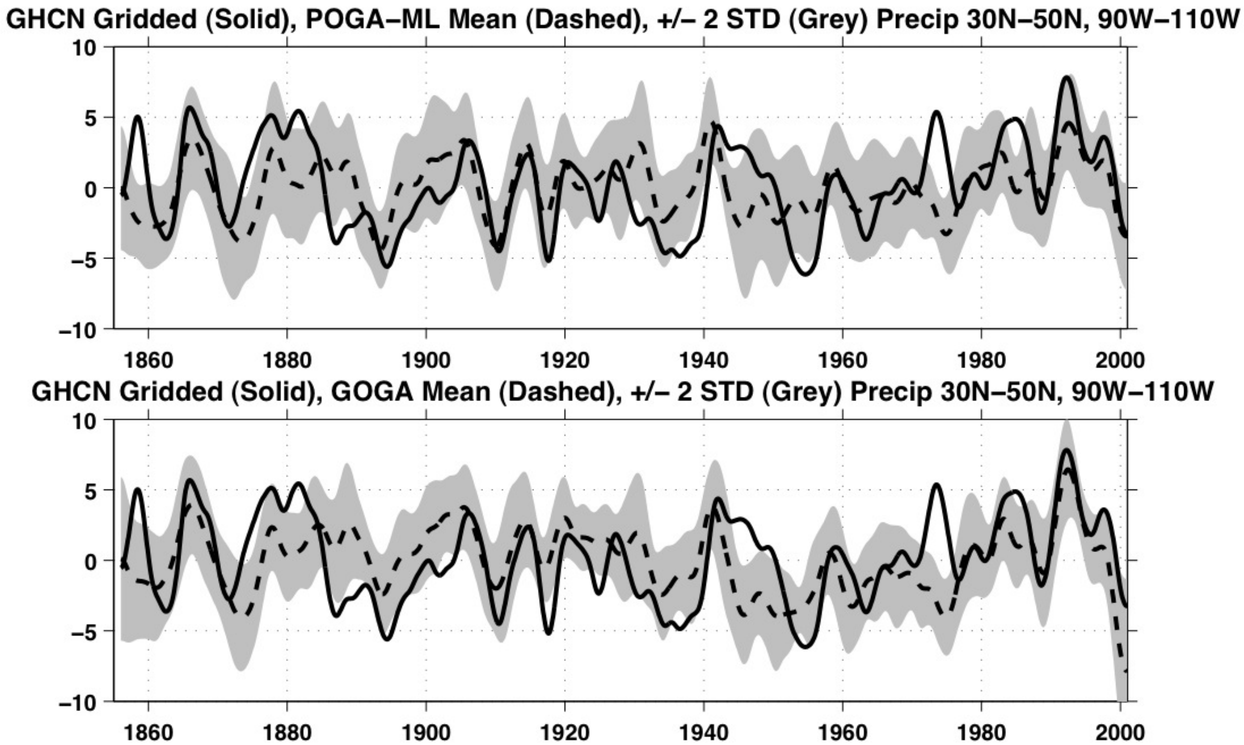

Figure 3.4. (top) The precipitation anomaly (in millimeters per month) over the Great Plains (30°N.-50°N., 90°W.-110°W.) for the period 1856 to 2000 from the POGA-ML ensemble mean with only tropical Pacific sea surface temperature (SST) forcing and from gridded station data. (bottom) Same as above but with GOGA ensemble mean with global SST forcing. All data have been 6-year low-pass filtered. The shading encloses the ensemble members within plus or minus of 2 standard deviations of the ensemble spread at any time. From *Seager et al. (2005b)*. GHCN, Global Historical Climatology Network.

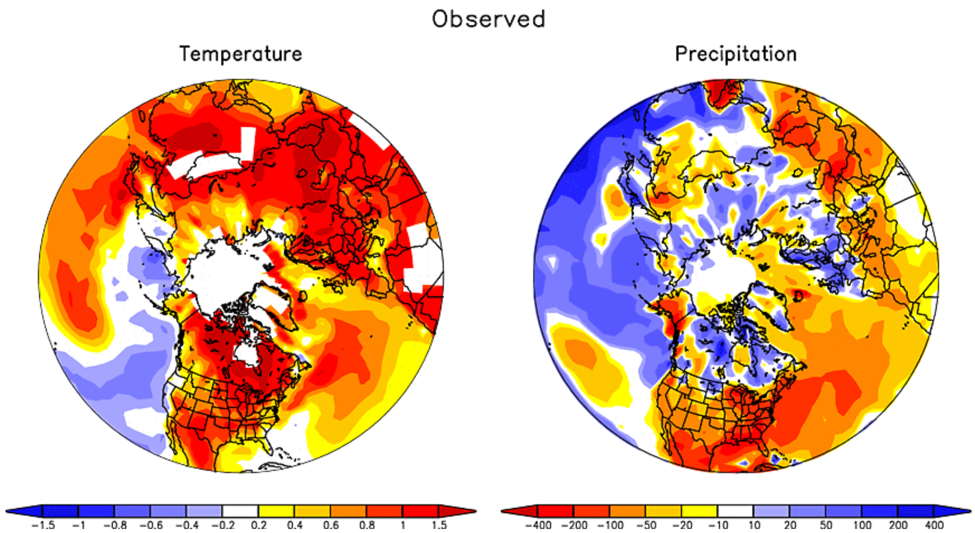

Figure 3.5. Observed temperature (°C) and precipitation (millimeters) anomalies (June 1998-May 2002). Figure from http://www.oar.noaa.gov/spotlite/archive/spot_drought.html.

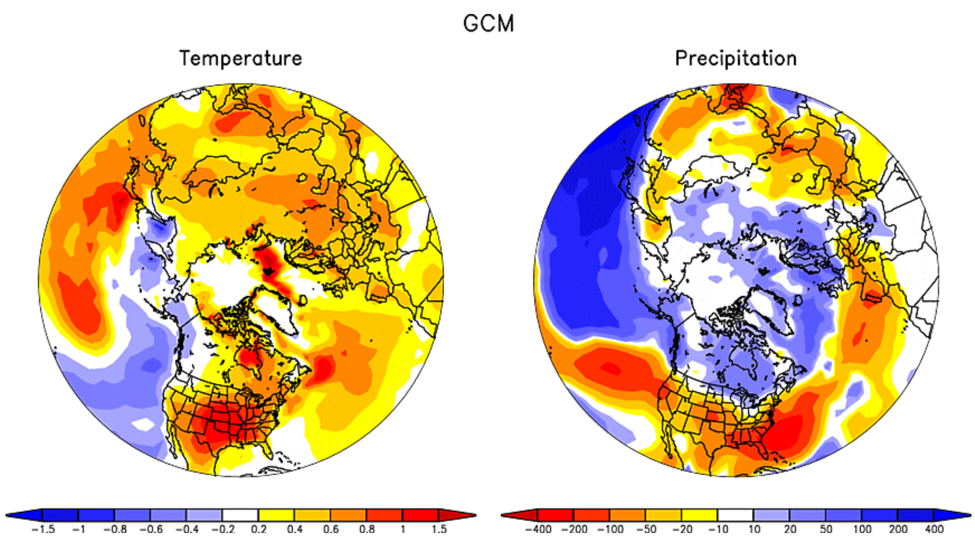

Figure 3.6. Model-simulated temperature (°C) and precipitation (millimeters) anomalies given observed SSTs over the June 1998 – May 2002 period. GCM, General Circulation Model. Figure from http://www.oar.noaa.gov/spotlite/archive/spot_drought.html.

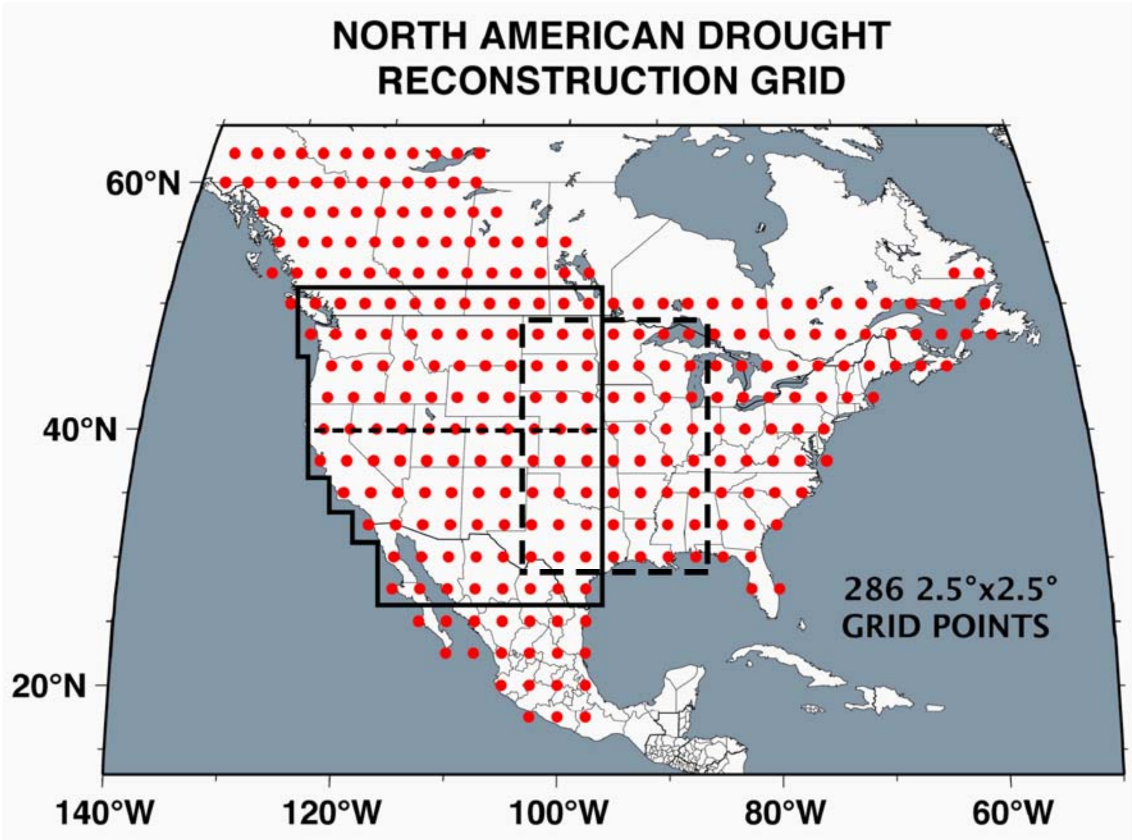

Figure 3.7. Map showing the distribution of 286 grid points of drought reconstructed for much of North America from long-term tree-ring records. The large, irregular polygon over the West is the area analyzed by *Cook et al. (2004)* in their study of long-term aridity changes. The dashed line at 40°N. divides that area into Northwest and Southwest zones. The dashed-line rectangle defines the Great Plains region that is also examined for long-term changes in aridity here.

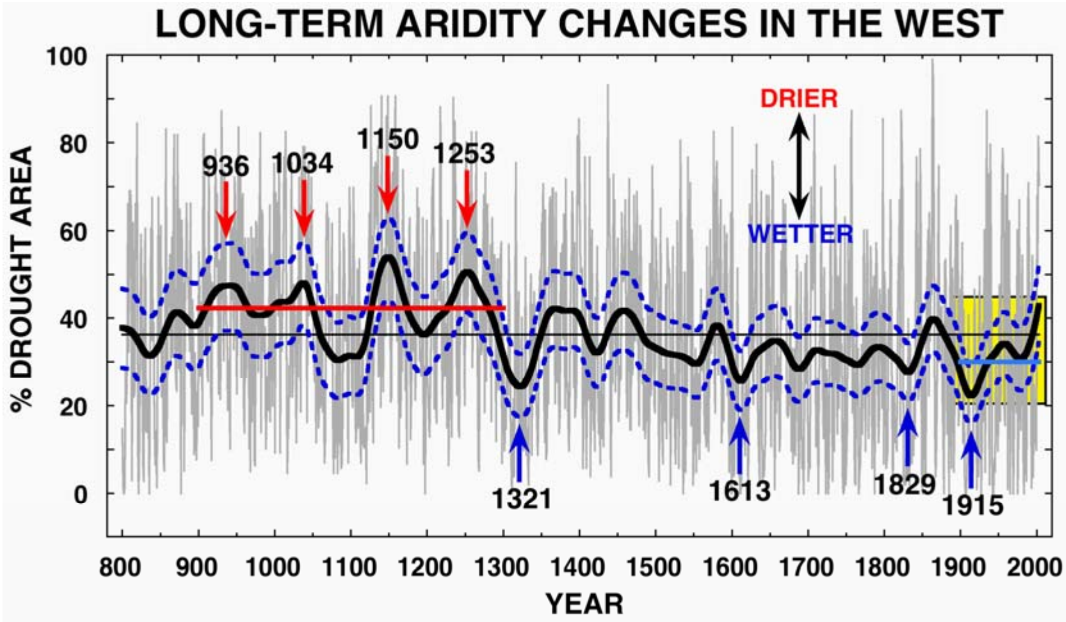

Figure 3.8. Percent area affected by drought (Palmer Drought Severity Index (PDSI) <-1) in the area defined as the West in Figure 3.7 (redrawn from *Cook et al., 2004*). More-positive numbers mean large areas in the West affected by drought. Annual data are in gray and a 60-year low-pass filtered version is indicated by the thick smooth curve. Dashed blue lines are 2-tailed 95% confidence limits based on bootstrap resampling. The modern (mostly 20th century) era is highlighted in yellow for comparison to a remarkable increase in aridity prior to about A.D. 1300.

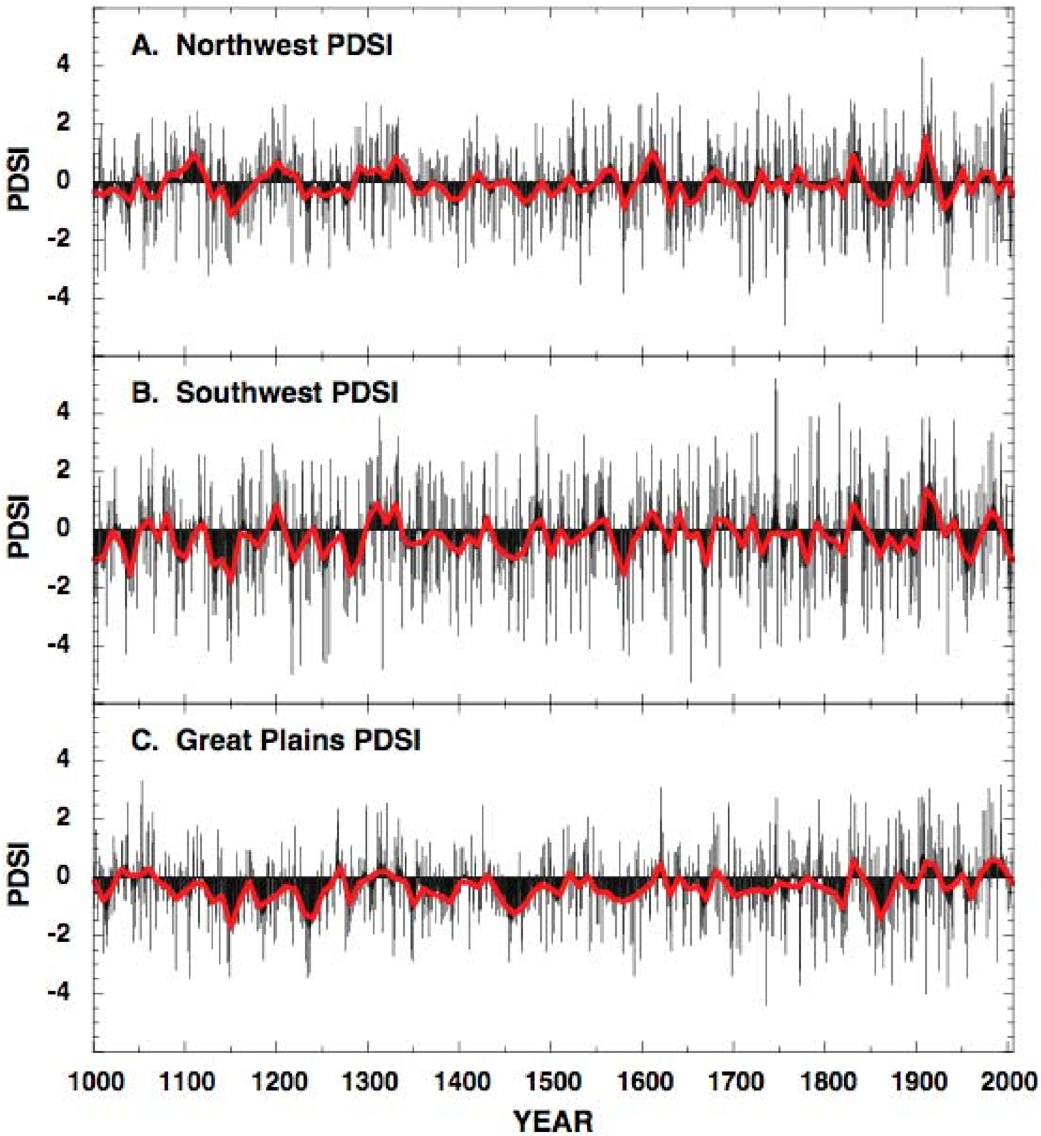

Figure 3.9. Reconstructed Palmer Drought Severity Index (PDSI) averaged over three regions of the Western and Central United States: the Northwest (A), the Southweset (B), and the Great Plains (C). Time is in calendar years A.D. See Figure 3.7 for the geographic locations of these regions. These series are all based on the updated version of North American Drought Atlas (NADA). Unlike the drought area index in Figure 3.8, where more-positive numbers mean larger areas affected by drought, more-negative numbers mean drier conditions here in accordance with the original PDSI scale devised by Palmer (1965).

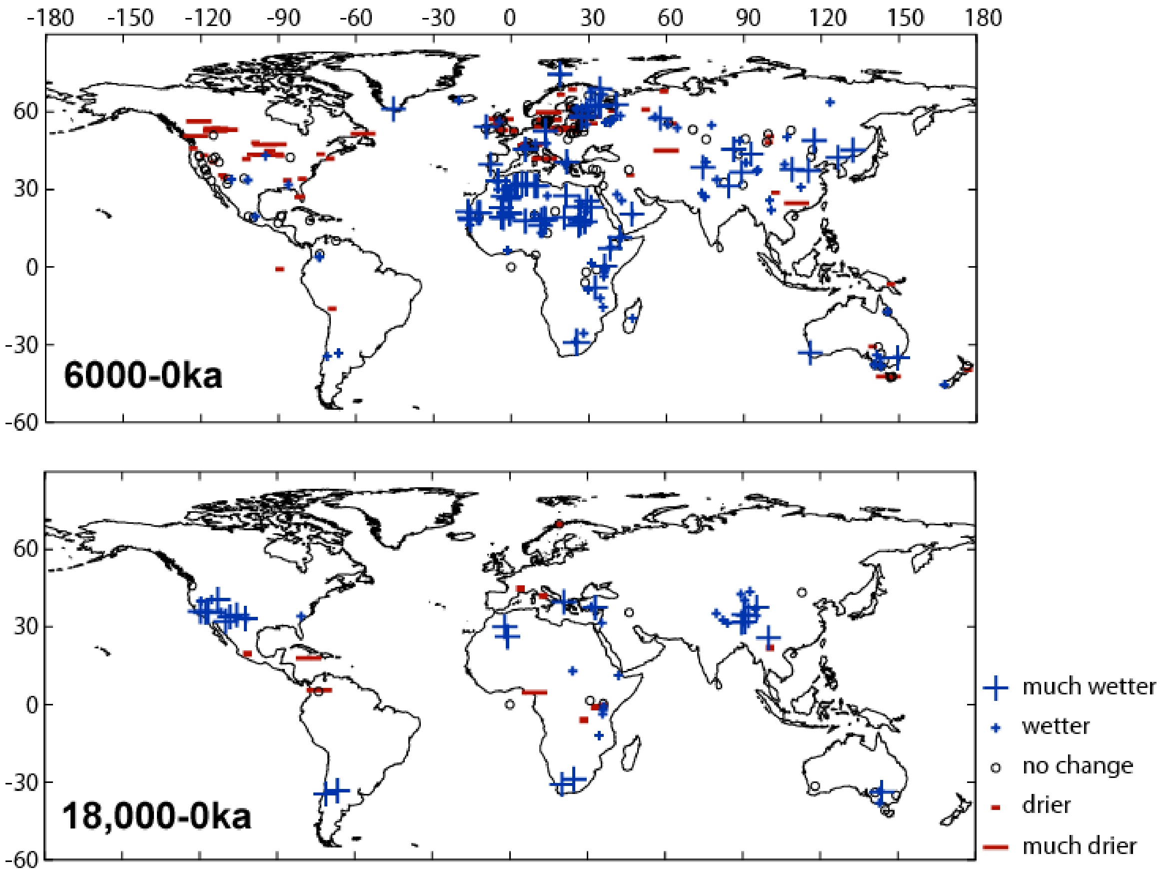

Figure 3.10. Global lake status at 6 ka (6,000 years ago) showing the large region that extends from Africa across Asia, where lake levels were higher than those of the present day as related to the expansion of the African-Asian monsoon. Note also the occurrence of much drier than present conditions over North America. (The most recent version of the Global Lake Surface Database is available on the PMIP 2 website http://pmip2.lsce.ipsl.fr/share/synth/glsdb/lakes.png.

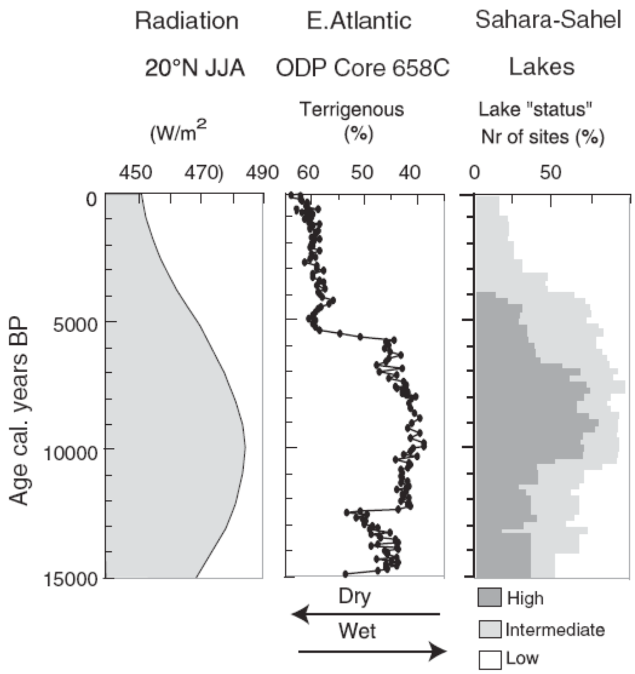

Figure 3.11. African Humid Period records (*Liu et al., 2007*).

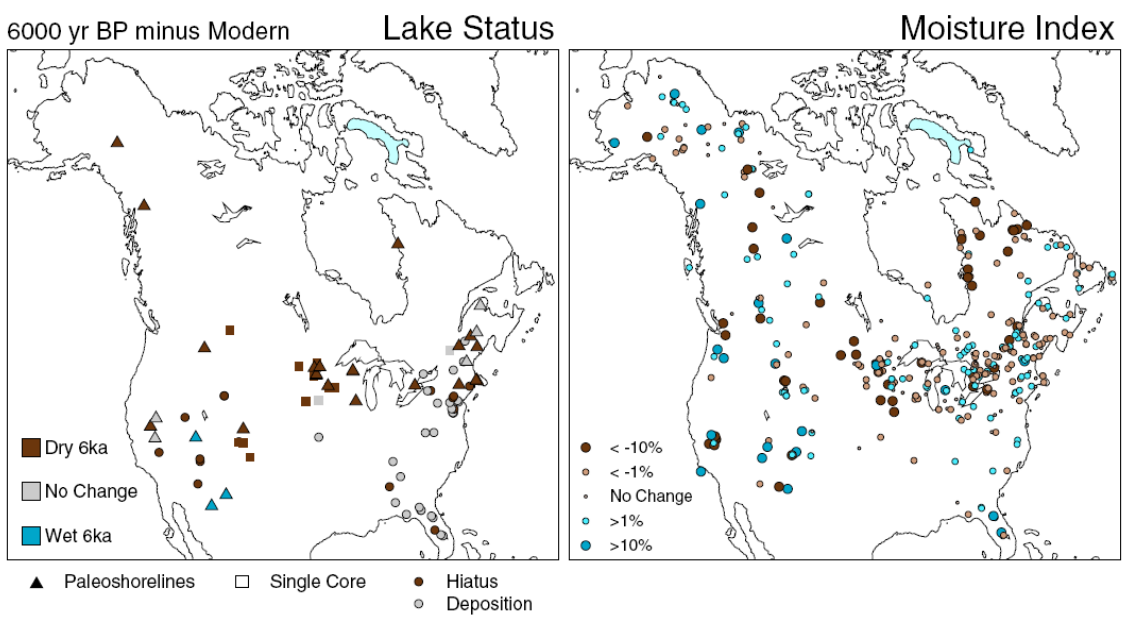

Figure 3.12. North American lake status (left) and moisture-index (AE/PE) anomalies (right) for 6 ka. Lake (level) status can be inferred from a variety of sedimentological and limnological indicators (triangles and squares), and from the absence of deposition (hiatuses, circles) (*Shuman and Finney, 2006*). The inferred moisture-index values are based on modern analog technique applied to a network of fossil-pollen data. Figure adapted from *Shuman et al. (in review)*.

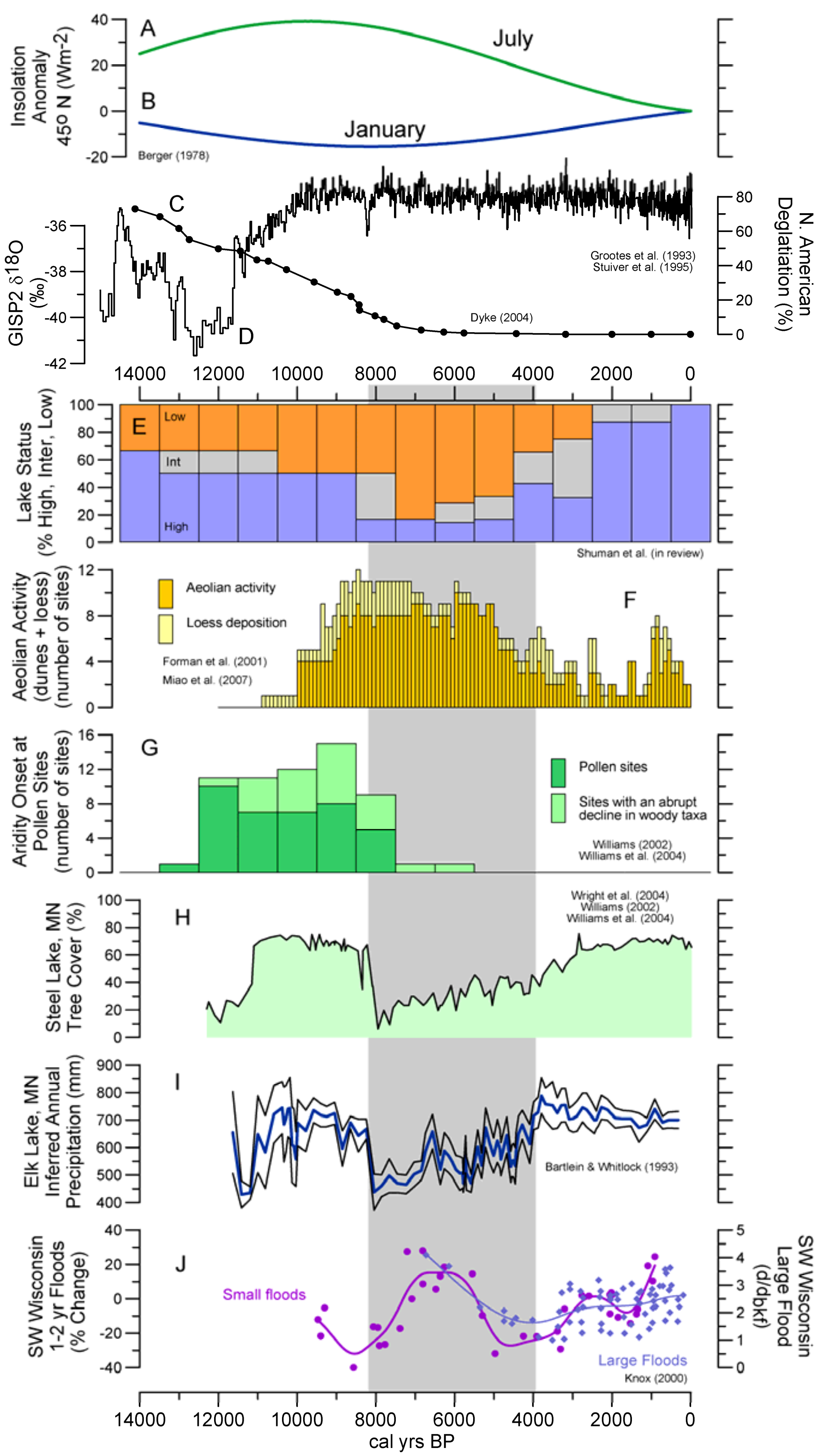

Figure 3.13. Time series of large-scale climate controls (A-D) and paleoenvironmental indicators of North American midcontinental aridity (E-I). A, B, July and January insolation anomalies (differences relative to present) (*Berger, 1978*). C, right-hand scale:

Deglaciation of North America, expressed as ice-sheet area relative to that at the Last Glacial Maximum (21 ka) (*Dyke, 2004*). D, left-hand scale: Oxygen-isotope data from the GISP 2 Greenland ice core (*Grootes et al, 1993; Stuiver et al., 1995*). Increasingly negative values indicate colder conditions. The abrupt warming at the end of the Younger Dryas chronozone (GS1/Holocene transition, 11.6 ka) is clearly visible, as is the "8.2 ka event" that marks the collapse of the Laurentide Ice Sheet. E, Lake status in central North America (*Shuman et al., in review*). Colors indicate the relative proportions of lake-status records that show lake levels that are at relatively high, intermediate, or low levels. F, Eolian activity indicators (orange, digitized from Fig. 13 in *Forman et al., 2001*) and episodes of loess deposition (yellow, digitized from Fig. 3 of *Miao et al., 2007*). G, Pollen indicators of the onset of aridity. Light-green bars indicate the number of sites with abrupt decreases in the abundance of woody taxa (data from *Williams, 2002; Williams et al., 2004*). H, Inferred tree-cover percentage at one of the sites (Steel Lake, MN) summarized in panel G (*Williams, 2002; Williams et al., 2004*; based on pollen data from *Wright et al., 2004*). I, Inferred annual precipitation values for Elk Lake, MN, a site close to Steel Lake (*Bartlein and Whitlock, 1993*). The inferred annual precipitation values here (as well as inferences made using other paleoenvironmental indicators) suggest that the precipitation anomaly that characterized the middle Holocene aridity is on the order of 350 mm y^{-1}, or about 1 mm d^{-1}. J, Frequency and magnitude of floods across a range of watershed sizes tracks climate variation during the Holocene. The gray shading indicates the interval of maximum aridity.

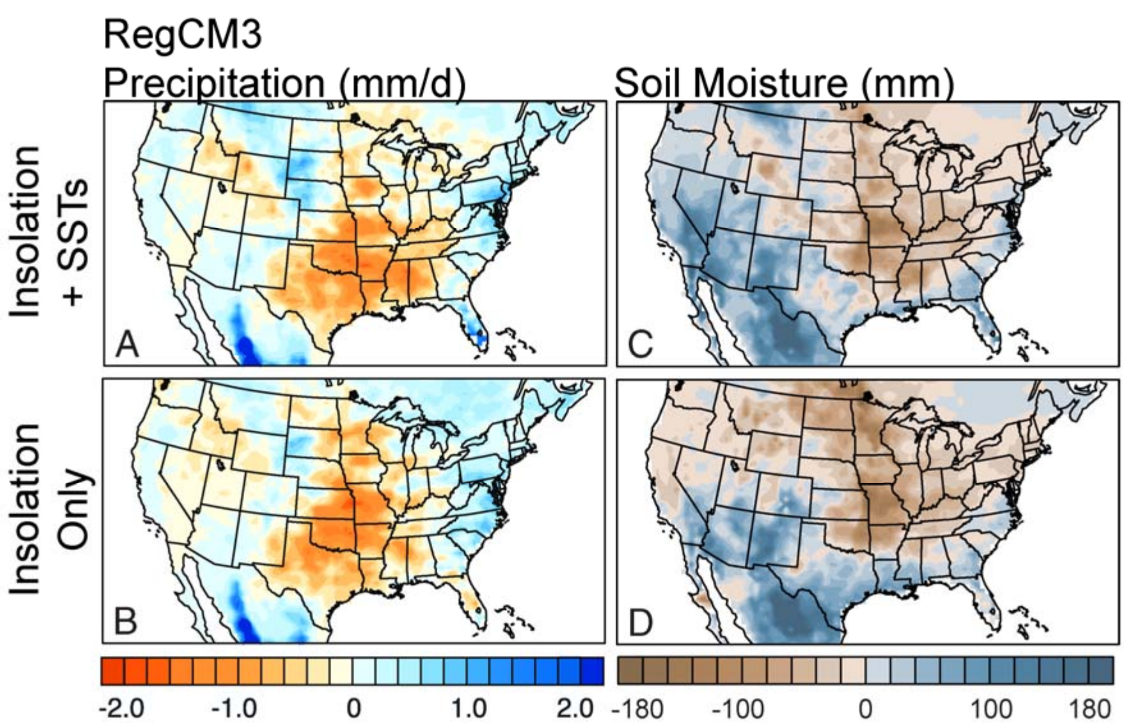

Figure 3.14. Regional climate model (RegCM3) simulations of precipitation rate (A, B) and soil moisture (C, D) for 6,000 years before present (6 ka) (*Diffenbaugh et al., 2006*,

land grid points only). RegCM is run using lateral boundary conditions supplied by CAM3, the atmospheric component of CCSM3. In panels A and C, the CAM3 boundary conditions included 6 ka-insolation, and time-varying sea surface temperatures (SSTs) provided by a fully coupled Atmosphere-Ocean General Circulation Model (AOGCM) simulation for 6 ka using CCSM3 (*Otto-Bliesner et al., 2006*). In panels B and D, the CAM3 boundary conditions included 6-ka insolation, and time-varying SSTs provided by a fully coupled CCSM simulation for the present. The differences between simulations reveal the impact of the insolation-forced differences in SST variability between 6 ka and present. mm, millimeters; mm/d, millimeters per day.

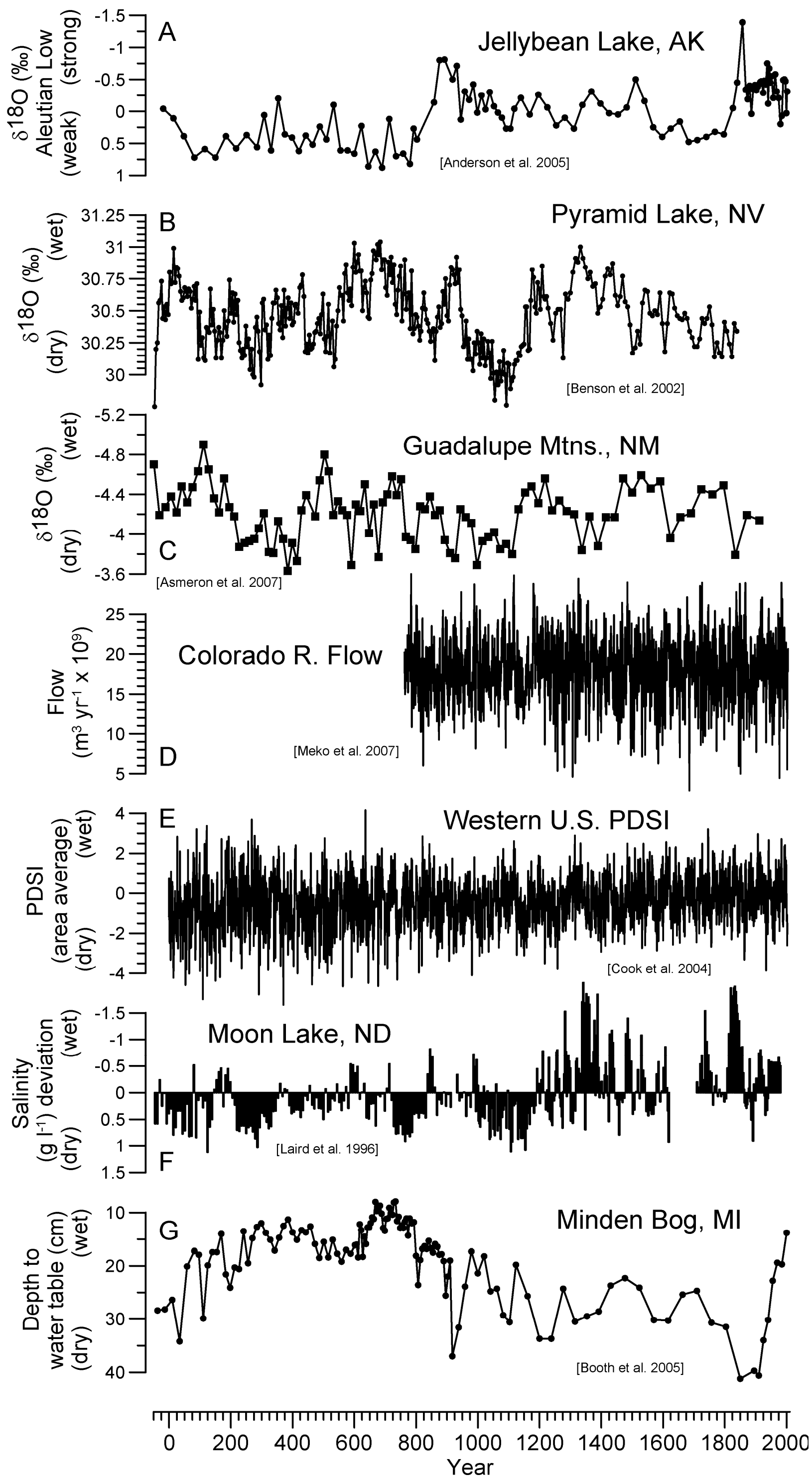

Figure 3.15. Representative hydrological time series for the past 2,000 years. A, oxygen-isotope composition of lake-sediment calcite from Jellybean Lake, AK, an indirect measure of the strength of the Aleutian Low, and hence moisture (*Anderson et al., 2005*). B, oxygen-isotope values from core PLC97-1, Pyramid Lake, NV, which reflect lake-level status (*Benson et al., 2002*); C, oxygen-isotope values from a speleothem from the

Guadalupe Mountains., NM, which reflect North American monsoon-related precipitation (*Asmerom et al., 2007*); D, dendroclimatological reconstructions of Colorado River flow (*Meko et al., 2007*); E, area averages for the Western United States of dendroclimatological reconstructions of PDSI (Palmer Drought Severity Index, *Cook et al., 2004*); F, diatom-inferred salinity estimates for Moon Lake, ND, expressed as deviations from a long-term average (*Laird et al., 1996*); G, depth-to-water-table values inferred from testate amoeba samples from a peat core from Minden Bog, MI (*Booth et al., 2006*). Abbreviations: ‰, per mil; $m^3\ y^{-1}$, cubic meters per year; $g\ l^{-1}$, grams per liter; cm, centimeter.

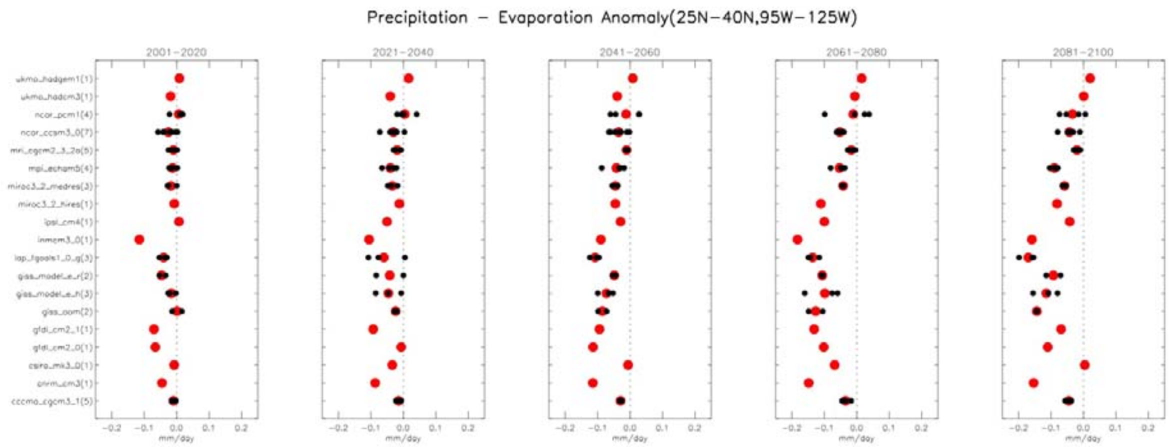

Figure 3.16. The change in annual mean precipitation minus evapotranspiration (P-E) over the American Southwest (125°W. – 95°W., 25°N. – 40°N., land areas only) for 19 models relative to model climatologies for 1950-2000. Results are averaged over 20-year segments of the current century. The number of ensemble members for the projections is listed by the model name at left. Black dots represent ensemble members, where available, and red dots represent the ensemble mean for each model. Units are in millimeters per day.

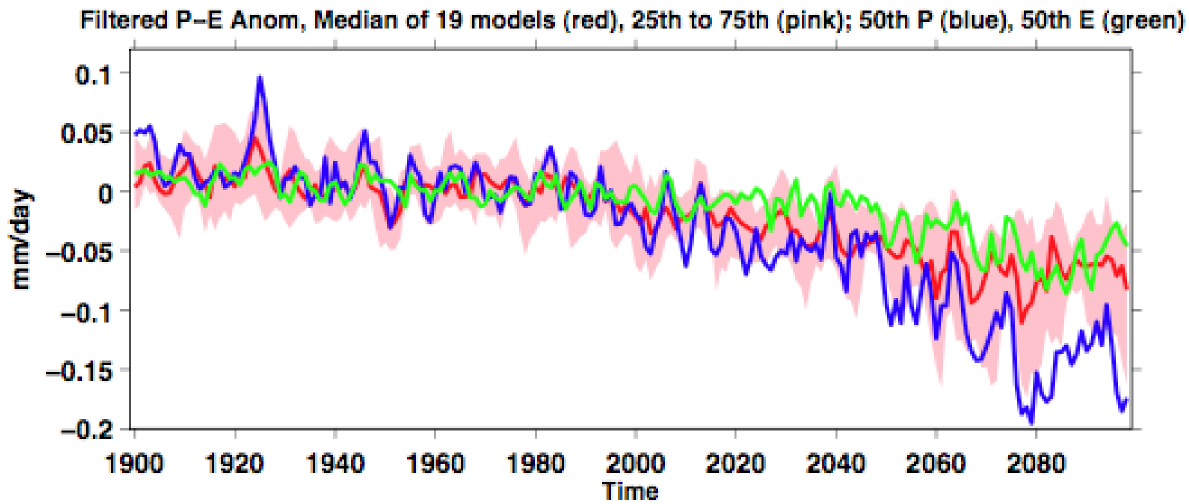

Figure 3.17. Modeled changes in annual mean precipitation minus evaporation (P-E) over southwestern North America (125° – 95° W., 25° – 40°N., land areas only) averaged over ensemble members for 19 models participating in IPCC AR4. The historical period used known and estimated climate forcings and the projections used the SResA1B emissions scenario (*IPCC, 2007*). Shown are the median (red line) and 25th and 75th percentiles (pink shading) of the P-E distribution amongst the 19 models, and the ensemble medians of P (blue line) and E (green line) for the period common to all models (1900-2098). Anomalies for each model are relative to that model's climatology for 1950-2000. Results have been 6-year low-pass filtered to emphasize low frequency variations. Units are millimeters per day (mm/day). The model ensemble mean P-E in this region is around 0.3 mm/day. From *Seager et al. (2007d).*