SAP 3.4 Abrupt Climate Change Report Chapter

Hydrological Variability and Change

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KEY FINDINGS

- Variations in water supply in general, and protracted droughts in particular, are arguably the greatest natural hazards facing the United States and the globe today and in the foreseeable future.

- In contrast to floods, which reflect both antecedent conditions and current meteorological events, and which are consequently more localized in time and space, droughts occur on subcontinental to continental scales, 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.

- On century-to-millennial time scales, droughts begin and end over intervals shorter than the time scales of variation that control global and regional climates, and again should be regarded as a class of abrupt climate changes.

- Empirical studies and climate model experiments conclusively show that droughts over North America and around the world are significantly influenced by the state of tropical sea surface temperatures (SSTs), with cool La Niña-like SSTs in the eastern equatorial Pacific being especially responsible for the development of droughts over the American West and northern Mexico. Unusually warm Indo-Pacific SSTs have also been strongly implicated in the development of global patterns of drought observed in recent years.

- 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. 1000 years ago) up to about A.D. 1600. Modeling experiments indicate that these megadroughts were also largely forced by cool SSTs in the eastern equatorial Pacific as well, but their exceptional duration has not been adequately explained.

- These megadroughts are significant, because they occurred in a climate system that was not being perturbed by major changes in its boundary conditions (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. The climate boundary conditions 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 truly abrupt hydroclimatic change that can be expressed by the climate system.

- Climate model scenarios of future hydroclimatic change over North America and the global subtropics indicate that subtropical aridity will intensify and persist due to future greenhouse warming. This drying is expected to extend poleward into the American West, thus increasing the likelihood of severe and persistent drought there in the future. The model results also indicate that this drying may have already begun.

RECOMMENDATIONS

- Predictive models of drought on the time scale of years to decades are needed to make future droughts less societally abrupt by increasing the lead time for preparedness, adaptation, and mitigation.

- 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 it is likely, strategies for response to this pending aridity, on both regional and global scales, are urgently needed.

- The development of predictive models with high levels of forecast skill requires a greatly improved understanding of the dynamics and fundamental causes of drought development. This will require taking into account both the evolution of natural climate variability and the internal and external forcing of that variability, including that from anthropogenic sources.

- 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 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.

- 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.

- 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.

I. 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. Yet, rising human populations everywhere are increasing the stress on currently available water supplies even without the anticipated impacts of climatic change. In many areas, the impacts of changing climate are going to make securing a reliable and adequate clean fresh water supply for all even more daunting. These concerns follow naturally from the general definition of drought used 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).

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). Already wet areas are likely to get wetter and already dry areas are likely to get drier, while some intermediate regions at slightly higher latitudes than the current subtropical dry zones will become increasingly arid. These anticipated changes will increase problems at both extremes of the water cycle, stressing water supplies in arid and semi-arid regions while worsening flood hazards and erosion in 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.

North America faces all of these problems. The semi-arid regions of the Southwest are projected to dry, with the model results suggesting that the transition may already be underway (Seager et al., 2007d). Intensity of precipitation is also expected to increase across most of the country (and world). 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. 3.6). 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 global hydroclimatic change into an international and cross-border issue with potential impacts on migration and social stability. The U.S. Great Plains could also experience changes in water supply that affect agricultural practices, grain exports, and biofuel production. 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. Other parts of North America, such as the agricultural interior and the heavily populated eastern half of the U.S., are also likely to see major hydrologic variations in the future.

In addition, the United States will also be affected by hydroclimatic changes in other regions of the world. The projected drying in the southern Europe-Mediterranean-North Africa-Middle East region will also affect a region already short on water resources and has the potential for increasing migration into western Europe and stoking regional conflicts that have global effects. Increased flooding risk, destruction of crops, and decline of water quality, combined with rising sea level, have the potential to force mass migration and cause social, economic, and political instability throughout Asia, again, with global implications.

The paleoclimatic record reveals dramatic changes in North American hydroclimate over the last millennium that had nothing to do with changes in greenhouse gases and human-induced global warming. In particular, tree ring reconstructions of the Palmer Drought Severity Index (PDSI) 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 (AD 800 to 1300) than in the last century. These megadroughts have been linked to changes in tropical Pacific SSTs with, perhaps, changes in solar irradiance and explosive volcanism being the drivers. If so, the megadroughts are dynamically distinct from projected future drying, but the paleoclimatic record raises the question as to whether the models lack the means for tropical climate to respond to external forcing and then impact climate globally. The processes involved in the development and persistence of 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 those expected during the next several decades and centuries. For example, the mid-continent of North America experienced conditions 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.

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 timescales 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, but 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, stood 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.

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 impacted 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).

2.1.1 Coupled ocean-atmosphere forcing of North American hydrological variability

The standard means to use models to demonstrate a link between SSTs and observed climate variability is to force an atmospheric general circulation model with observed SSTs as a lower boundary condition. 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 Fig. 3.1 and time series of modeled and observed precipitation over the Great Plains are shown in Fig. 3.2. 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 timescales, 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 but 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^{o}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 timescales 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) are of unknown origin. Kushnir (1994), Sutton and Hodson (2005), and Knight et al. (2005) have linked them to changes in ocean circulation even though it is unclear if they represent anything more than a local expression of external forcing.

The dynamics that link tropical Pacific SST anomalies to North American hydroclimate are better understood and, on long timescales, appear as analogues 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 (well described by 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., 2003b; Seager et al., 2005a; Seager et al., 2005b; Lau et al., 2006). During La Niñas both mechanisms force air to descend over western North America, which suppresses precipitation. Although models support the idea that warm subtropical North Atlantic SSTs can cause drying over western North America, the dynamics that underlay this are currently unknown. To date no model experiments suggest that extratropical SST anomalies influence North American hydroclimate variability.

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 (Namias, 1991; Oglesby, 1991; Oglesby and Erickson, 1989). Koster et al. (2004) used observations to show that on the timescale of weeks, precipitation in the Great Plains is significantly correlated with antecedent precipitation. Schubert et al. (2004b) compared models run with climatological SSTs and with and without variations in evaporation efficiency and showed that low frequency, multiyear, North American hydroclimate variability was significantly reduced in the latter case. 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 atmospheric moisture transport variability 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.

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), over the period from 1980 to 2006 droughts and heat waves were the second most expensive natural disaster in the U.S. behind tropical storms (a figure greatly inflated by the 2005 hurricane season, which included Katrina). The annual cost of drought to the U.S. is estimated to be in the billions of dollars. Thus, it is clear that drought leads to 1) crop losses resulting in a loss of farm income and an increase in Federal disaster relief funds and food prices, 2) disruption of recreation and tourism, 3) increased fire risk and loss of life and property, 4) reduced hydroelectric energy generation, and 5) enforced water conservation to preserve essential municipal water supplies and aquatic ecosystems.

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 mid 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 Indians 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. This conflict inevitably led to the catastrophic decline and near extinction of the American bison population (West, 1995; Isenberg, 2000). Drought did not nearly destroy the bison, but it did create conditions in which a multitude of human pressures came close to entirely eliminating one of America's few remaining species of megafauna.

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? 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. Once more the course of history of the West will be redirected by unanticipated environmental factors, but unlike prior times, this environmental change will be a consequence of human economic activity itself.

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, 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) it 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 rarely clean since regions such as the Mediterranean are also strongly impacted by other climate phenomena such as the North Atlantic Oscillation (NAO) (also sometimes called the Northern Annular Mode) (Hurrell, 1995; Fye et al., 2006).

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.

During 1998-2002, prolonged below normal precipitation and above normal temperatures caused the U.S. 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. 3.3). The Hoerling and Kumar (2003) study used climate model simulations to assess how the ocean conditions over the 4-year period influenced climate. 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. 3.4). 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 west equatorial Pacific.

Over the 1998-2002 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 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.

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 anthropogenic influence on North American precipitation amounts. The AR4 of the IPCC (IPCC, 2007) presents maps of the trend in precipitation over 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 Century 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 Ninos may have been caused by natural Pacific decadal variability (Huang et al., 2005). In contrast to the twentieth century record the southern parts of North America are projected to dry as a consequence of anthropogenic climate change. After the 1997/98 El Nino drought has indeed settled into the West but since it has gone along with a more La Nina-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. 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).

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, this conclusion is almost certainly not valid since, while these same models predict a weakening of the Walker Circulation in the current century, they also predict that western North America will dry as a consequence of rising greenhouse gases and global warming (see below).

Consequently, in the absence of compelling evidence to the contrary, it is reasonable to assume that the precipitation trends seen over the last century are of natural origin and related to naturally occurring decadal to multidecadal variations of tropical SSTs. The case is a little different if we restrict attention to the post-1976-77 climate shift period. The AR4 report shows that the 1979 to 2005 trend has clear drying over the Western U.S. This is because the post-1998 period has been one of drought and comes on the heels of the 1977 to 1998 pluvial. But this change may also have been of natural origin - after the 1997-98 El Niño the tropical Pacific entered a La Niña phase that lasted to 2002 and, as expected, this brought drought to the West and subtropical regions in much of the world (Hoerling and Kumar, 2003; Seager, 2007). What is more odd is that the drought persisted after the 1998-2002 La Niña and, while it was interrupted, at least regionally during the 2004-05 El Niño, has persisted across the region through the weak 2006-07 El Niño. However, 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. 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.

Box on impacts of all aspects of change in the atmospheric branch of the hydrological cycle for ground water and river flow

Introduction

Abrupt changes or shifts in climate regimes have had, and will continue to have, major impacts on society. Gradual shifts in the climate background state may modulate, and either constructively or destructively influence, the “typical” hydrologic impacts of seasonal to interannual climate variability. An example is the wetter or drier conditions that have been historically associated with the El Niño and the La Niña patterns of anomalously warmer and colder tropical SSTs in the Pacific and Indian Oceans. Southern States in the U.S. tend to receive higher than average winter-time precipitation during an El Niño and the Southwestern and Southeastern States tend to receive lower than average winter-time precipitation during a La Niña (Fig. 3.5). El Niños and La Niña also influence the hydrologic conditions in semiarid regions across Australia, South America, Africa, and Asia. In the semi-arid Southwestern U.S., the hydrologic impacts of past El Niños have been critical to refilling water supply reservoirs that were built to mitigate the impacts of drought.

The Department of Interior analysis of Western U.S. water supply issues (DOI Water 2025 Status Report, 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. This determination assumes a statistically stationary climate in the Western U.S. with no changes in moisture supply or demand in response to future changes in climate (Fig. 3.6). Any transient change in climate conditions that leads to an abrupt regime shift to more persistent and/or more severe drought will only compound these water supply conflicts and impact society.

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 flooding-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). Drought provides many examples of the impacts that may result from abrupt shifts in hydroclimate and will be the focus of this section.

Abrupt Change: Drought

Abrupt changes or shifts in climate, in particular those that lead to drought, have had major impacts on societies in the past. Paleoclimatic data document rapid shifts to dry conditions that coincided with downfall of advanced and complex societies. The history of the rise and fall of several empires and societies in the Middle East between 7000 and 2000 B.C. have been linked to abrupt shifts to persistent drought conditions (Weiss and Bradley, 2001, and others). Severe drought leading to crop failure and famine, followed by rebellion, have been suggested as causes for the decline and collapse of the Tang Dynasty and the Classic Maya, both in the mid-8th century (Yancheva et al., 2007). A more recent example of the impact of severe and persistent drought on society is the 1930s Dust Bowl in the Central United States, which led to a large-scale migration of farmers from the Great Plains to the Western United States. Societies in many parts of the world today may now be more insulated to the impacts of abrupt climate shifts in the form of drought through managed water resources and reservoir systems. However, population growth and over-allocation of scarce water supplies in a number of regions have made societies even more vulnerable to the impacts of abrupt climate change and consequent drought.

Abrupt climate change leading to persistent and/or severe drought can impact the water sector directly through deficits in surface and ground-water supplies. A reduction in surface-water supplies affects reservoir storage and operations, and delivery of water to users. Direct impacts of drought on surface water also lead to an impairment of in-stream flows and water quality. Impacts on ground water include drawdown of aquifers, increased pumping costs, subsidence, and reductions of adjacent or connected surface-water flows. Rapid climate changes leading to drought also challenge the management and maintenance of infrastructures for water storage and delivery, and wastewater treatment.

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 in the form of drought. In forests, which support the timber and recreation sectors, drought can lead to mortality due to insect infestation, and wildfire. Reductions in water supplies that impact any of these sectors 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). In addition, impacts on water supplies, both quantity and quality, can affect quality of life and human health, and well as ecosystem health.

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 U.S. 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. Many of these systems are extremely vulnerability to relatively small changes in runoff (e.g., Nemec and Schaake, 1982; Christensen and Lettenmaier, 2006). For example, in modeling experiments, Christensen and Lettenmaier (2006) report a 10% inflow change results in a 20% storage impact in the Colorado River system. In many parts of the Western U.S., surface water is administered through the prior appropriations doctrine, where severe drought conditions can lead to the curtailment of all but the most senior water rights, leaving junior water rights holders, who are often municipalities, to find alternative water supplies.

An example from the Colorado River

As an example of the potential impacts of a rapid change to more drought-prone conditions can be illustrated by the recent drought and its impacts on the Colorado River system. The Colorado River basin, as well as much of the Western U.S., 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 drought was the worst in the 20th century, paleoclimatic records indicate droughts as or more severe occurred as recently as the mid-19th century (Woodhouse et al., 2005). Impacts of this 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 over 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 could the system have handled drought conditions is uncertain, and at some point, a longer drought is certain to have much greater impacts. The Colorado River Compact and subsequent legal agreements currently require the Upper Basin to pass 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.

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 U.S. (see the 19th century instrumental data maps in Herweijer et al., 2006), which does not provides 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). The past 2,000 years is especially 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 U.S., based on tree rings and other paleo-proxy sources of hydroclimatic variability, can be found in Woodhouse and Overpeck (1998). In that paper, the authors introduced the concept of the ”megadrought,” one 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 multi-decadal 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 U.S., 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 U.S. 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 record.

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 Fig. 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 were also exceeded 1,000 years in length, which facilitated the creation of new PDSI reconstructions extending back into the megadrought period in the Western U.S. 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 U.S. 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 U.S. was extremely important.

Using the same basic methods as those in Cook et al. (1999) to reconstruct drought over the conterminous U.S., 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 Fig. 3.7, the irregular polygon delineates the boundaries of the area we refer to as the American West. 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 U.S.

Cook et al. (2004) examined the NADA contents back to A.D. 800 for the West to place the current drought there 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 severe 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 less data. These findings were rather sobering for the West because they 1) verified the occurrence of several past multi-decadal 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 even if the modeling results in Seager et al. (2007d) are wrong, the West still 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 U.S. (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 ”droughty” 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 does differentiate 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 U.S. 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 (henceforth NADAv1). Since the creation of NADAv1 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 NADAv2 (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 Herweijer et al. (2007).

Figure 3.9a-b shows the NADAv1 results for the West in a way very comparable to that in Herweijer et al. (2007). It shows a persistently dry MCA and the four megadroughts within it noted above. Figure 3.9c-d shows the NADAv2 results in the identical manner. While the relative patterns of variability are extremely similar throughout, the amplitude of overall aridity and the megadroughts in the MCA are considerably reduced in NADAv2. This difference reflects the improved spatial distribution of tree-ring chronologies used in NADAv2, which provides a more uniform geographic weighting in the average over the West. The intensity of drought during the MCA has not gone away, however. Rather, it is now focused more clearly toward the Southwest. This is shown in Fig. 3.10, which compares the Southwest and the Northwest as defined on the map in Fig. 3.7. This comparison indicates that the MCA aridity period is more strongly expressed in the Southwestern U.S. 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, NADAv2 still indicates the occurrence of multidecadal megadroughts that mostly agree with those of Herweijer et al. (2007) and an overall period of elevated aridity as described by Cook et al. (2004). From Fig. 3.10a, 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) based on NADAv1 (see also Fig. 3.9a).

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. Going a step farther Seager et al. (2007a) used the Palmyra modern and fossil coral records to reconstruct 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 suggest errors in the tropical Pacific SST reconstruction and/or a role for SST anomalies from other oceans.

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; Herwejer et al. 2007). 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). 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 the positive radiative forcing-La Niña-dry Southwest connection comes from analyses of the entire Holocene recorded in a New Mexico speleothem (a secondary mineral deposit formed in a cave) which shows a clear association between increased solar irradiance (as deduced from the atmospheric 14C content recorded in ice cores) and dry condition (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; Cane et al., 1997; Clement et al., 2000) while 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 to the response to rising greenhouse gases or it could be that the coupled GCMs respond to the latter incorrectly due to the many errors in their simulations of the tropical Pacific mean climate, not the least 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 U.S. because that is where the late-20th century drought began and has largely persisted up to the present time. This has been fortunate for U.S. agriculture because the present drought has largely skipped 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 NADAv2 to produce an average PDSI series for the Great Plains rectangle indicated in Fig. 3.7. That series is shown in Fig. 3.11 and it is far more provocative than even the Southwest series. The MCA period shows even more persistent drought, now on the centennial timescale, 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 timescale drought around that time. More ominously, in comparison the 20th century has actually been a period of relatively benign hydroclimatic variability, with the 1930s Dust Bowl and 1950s southern Great Plains droughts being rather unexceptional when viewed at the spatial scale being considered here. The closest historical analog to the extreme past megadroughts is the Civil War drought (Herweijer et al., 2006) from 1855 to 1865 (11 years) in NADAv2, followed closely by a multi-year 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 Fig. 3.11 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 with variations in the NAO (see also Section 2.2 of this chapter).

4. ABRUPT HYDROLOGIC CHANGES DURING THE HOLOCENE

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. These 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-timescale variations were interannual to millennial timescale 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 like those reviewed in the previous section. (On longer, or "orbital" timescales, the ice sheets and biogeochemically determined greenhouse gas concentrations 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; Saltzman, 2002.)

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 (Hegerl et al., 2003, 2007; Jansen et al., 2007). In other words, all of the variations in climate over the instrumental period and over the past millennia reviewed elsewhere have occurred in a climate system whose controls have not differed much from those of the 20th century. 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 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 then 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 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 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 20thor 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 mid-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 (12,000 calendar years ago) 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. In North America, drier conditions than present commenced in the mid-continent between 10 and 8 ka (Webb et al., 1993a; Thompson et al., 1993; 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).

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;
- paleoclimatic simulations suggest that ocean-atmosphere coupling played a role in determining the moisture status of these regions, as it has during the 20thcentury 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). 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).

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, or 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. 3.12), 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. 3.12) and 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 Projects (PMIP and PMIP 2; Joussaume et al., 1999; Cruxifix et al., 2005; Braconnot et al., 2007ab). 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) general circulation models with interactive oceans (AOGCMs), terrestrial vegetation (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, 2007b; 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.

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. Informed by a statistical analysis of vegetation-climate feedback in the AOVGCM, supplemented by the analysis of a simple conceptual model that relates a simple two-state depiction of vegetation to annual precipitation (Liu et al., 2006), they 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., in review), 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 can 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)).

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 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 (e.g., see Fig. 3.13 from Liu et al., 2006). 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.

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 an 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).

A variety of paleoenvironmental indicators reflect the spatial extent and timing of these moisture variations (Figs. 3.14 and 3.15), 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 most 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 U.S. Although the region of drier-than-present conditions extends into the Northeastern U.S. and eastern Canada, most of the multiproxy evidence for middle Holocene dryness is focused on the mid-continent, 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 paleoenvironmental indicators of the moisture changes are shown in Fig. 3.15. In addition to insolation forcing (Figs. 3.15a,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. 3.15c), 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 mid-continent 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 the Atlantic meriodional overturning circulation 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. 3.16d) 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 (Shuman et al., 2002; Dean et al., 2002), as well as indirectly, through its influence on Atlantic meridional overturning circulation (MOC), and related hemispheric atmospheric circulation changes.

The record of aridity indicators for the mid-continent 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. 3.15e, Shuman et al., in review) show a 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 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. 15g, Williams et al., in prep.) show a somewhat earlier onset of dryness than do the aeolian or lake indicators, reaching maximum frequency around 9 ka.

The pollen record from Steel Lake, MN, expressed in terms of tree-cover percentages (see Williams, 2002, for methods) illustrates 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 200years, Fig. 3.15h). 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. 15i, Bartlein and Whitlock, 1993; see also Webb et al., 1998). Expressed in terms of precipitation, the moisture anomaly in the mid-continent necessary to evoke the vegetation changes is about 350 mm y^{-1}, or about 1 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 mid-continent 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 mid-continent, 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 absence of such a mechanism in these early simulations, 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 a) 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, b) 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 c) 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 U.S. 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 U.S. 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 Fig. 3.16, 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 U.S. 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. 3.16). In contrast to the GCM simulations, the inclusion of 6 ka SST variability 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 mid-continental 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 (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.

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 already wet areas of the planet get wetter and already dry areas get drier, meaning that there is widespread drying in the subtropics (Held and Soden, 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).

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. 3.17). 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. 3.18). 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 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 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 timescale 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 timescale 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.

The Medieval megadroughts are probably also not analogs of future drying. As mentioned above, the leading theory for their origin is that the tropical Pacific SSTs were La Niña-like for up to decades at a time during the Medieval period, possibly as a response to increased surface solar radiation. Also, there is no evidence to date that the planet was overall warmer during the Medieval period than in subsequent centuries or as warm as the planet is expected to be in coming decades. If this is so, then future subtropical drying probably has no past analogs. But it is also 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, as suggested by the paleoclimate proxy data described before. Were this to be the case, presumably drying in the Southwest would be even 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 to 25 percent or more. 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 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. It seems unavoidable that there will have to be a renegotiation of water allocations between States, users within the States and between the U.S. and Mexico. 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. 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 Kelvin of warming. In contrast the global mean precipitation increases by about 1-2 % per Kelvin 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. 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.

7. CONCLUSIONS

Drought is arguably the greatest recurring natural hazard facing 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, 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.

Droughts can develop faster than the timescale 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 development of wiser ways to use the limited available water resources.

For examples of truly abrupt and long-lasting changes in hydroclimatic variability over mid-continental 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 might not to be as great. However, the rising level of greenhouse gas forcing happening now and in the foreseeable future is truly unprecedented over the Holocene. Therefore, the abrupt hydrologic changes in the Holocene ought to be viewed as useful examples of the amount of change that could conceivably occur in the future.

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 increasingly clear that 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. This has been shown to be the case for modern droughts, historical droughts, and even paleodroughts as far back as Medieval times over North America. 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 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, but the uncertainties remain large.

However true the importance of enhanced solar forcing has been in producing past megadroughts, the level of current and future radiative forcing due to greenhouse gases is much greater. 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 may in fact 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.

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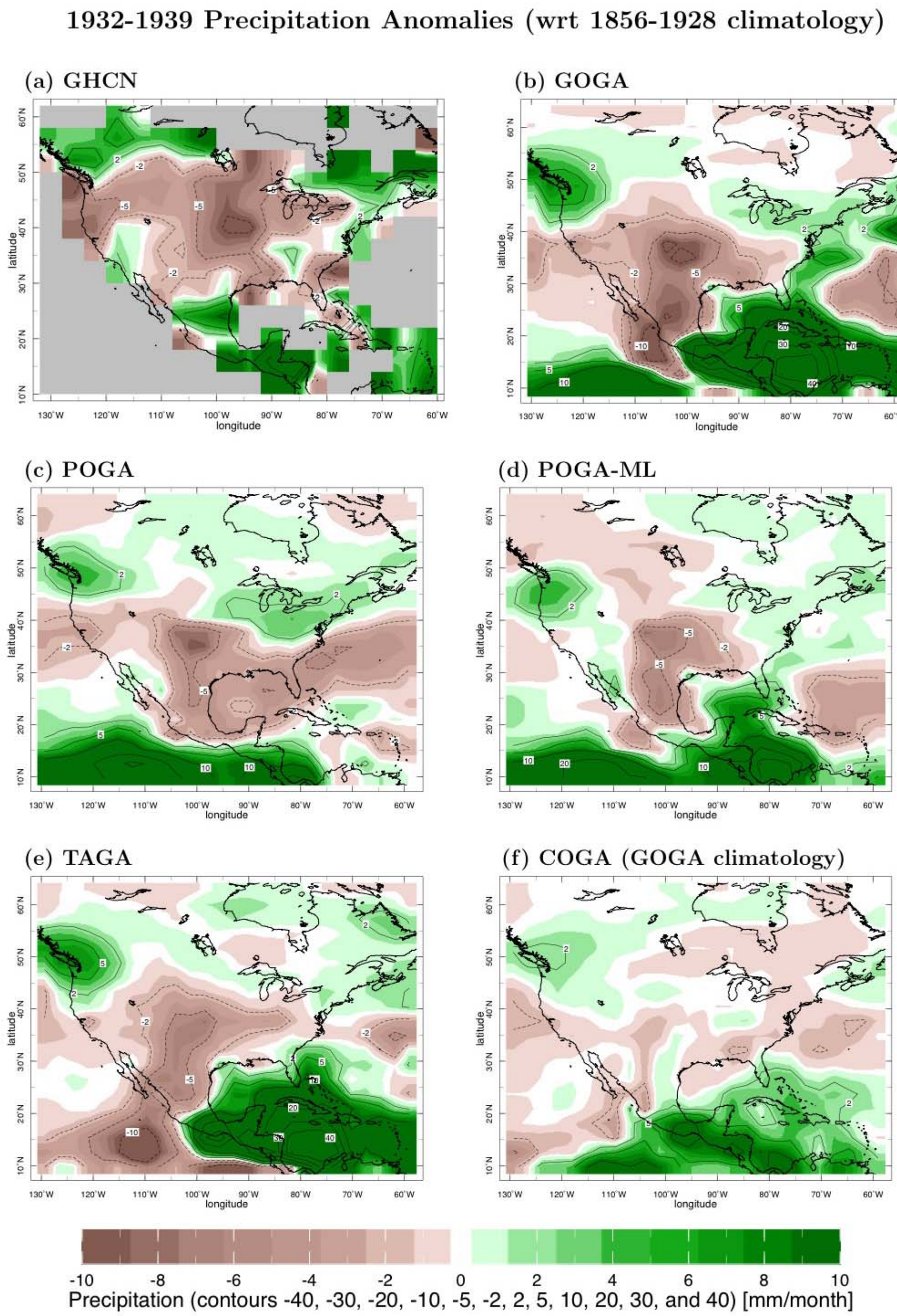

Figure 3.1. The observed (top left) and modeled precipitation anomalies during the Dust Bowl (1932 to 1939) relative to an1856 to 1928 climatology. Observations are from Global Historical Climatology Network (GHCN). The modeled values are model ensemble means from the ensembles with global sea surface 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).

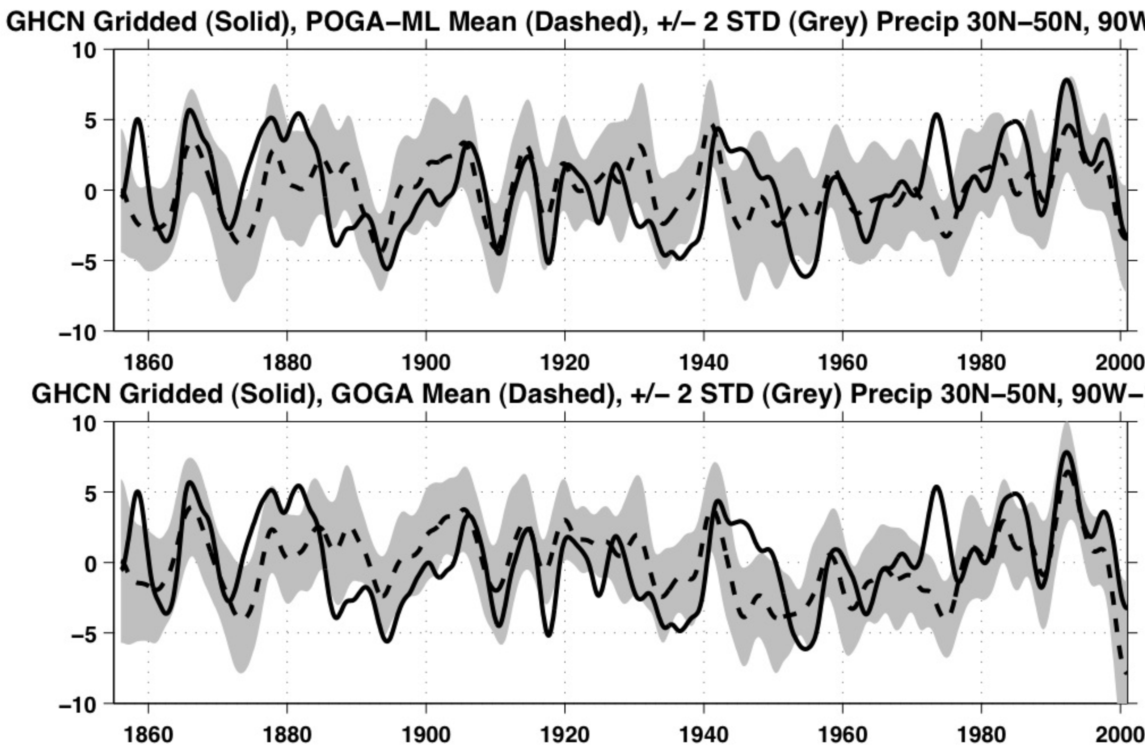

Figure 3.2. (top) The precipitation anomaly (in millimeters per month) over the Great Plains (30°N.-50°N., 90°W.-110°W.) for the period 1856 to 2000 from the POGA-ML ensemble mean with only tropical Pacific sea surface temperature (SST) forcing and from gridded station data. (bottom) Same as above but with GOGA ensemble mean with global SST forcing. All data has been 6-year low-pass filtered. The shading encloses the ensemble members within plus or minus of two standard deviations of the ensemble spread at any time. From Seager et al. (2005b). GHCN, Global Historical Climatology Network.

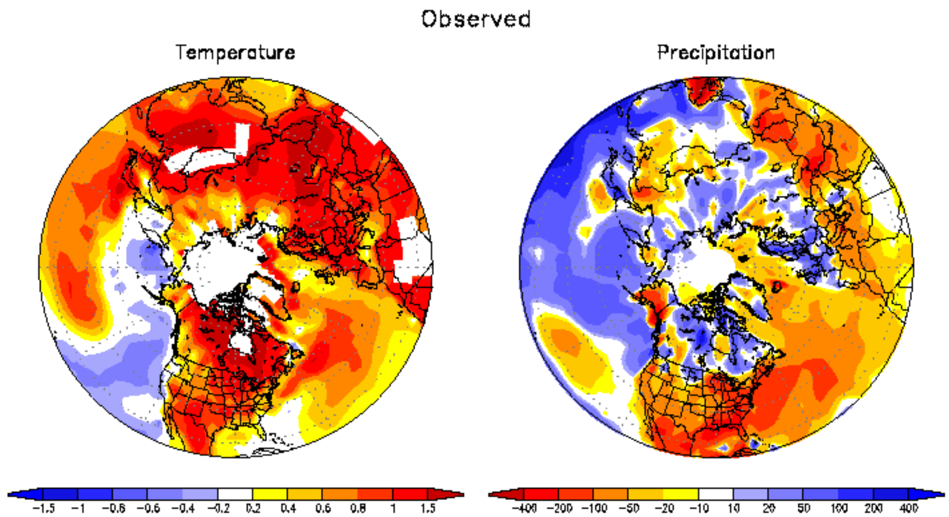

Figure 3.3. Observed temperature (°C) and precipitation (mm) anomalies (June 1998-May 2002).

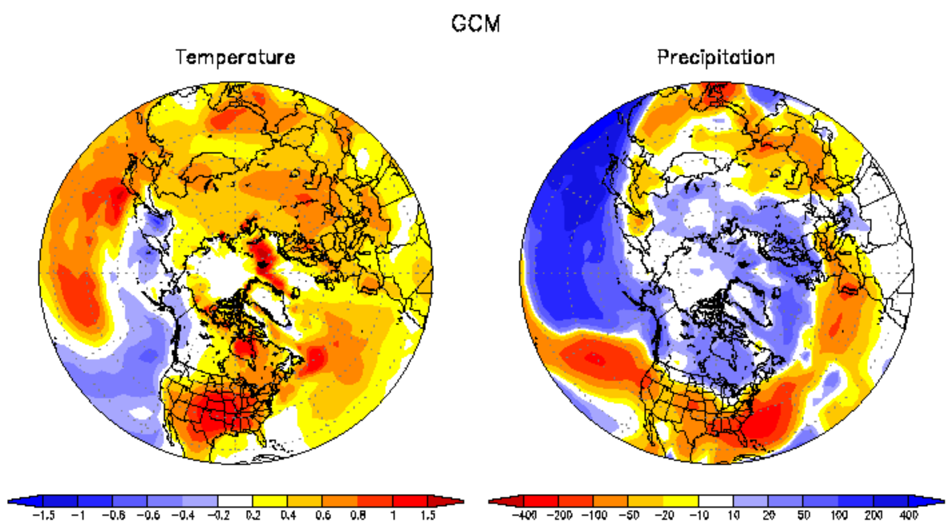

Figure 3.4. Model-simulated temperature (°C) and precipitation (mm) anomalies given observed SSTs over the June 1998 – May 2002 period. GCM, General Circulation Model.

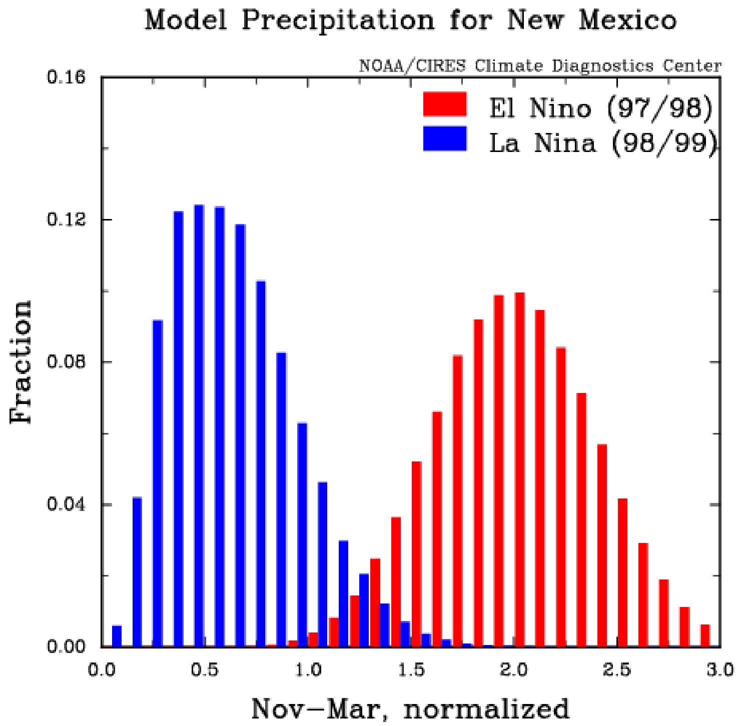

Figure 3.5. Differences in model precipitation for New Mexico for the two phases of El Niño/Southern Oscillation (ENSO): warm-wet El Niño and cool-dry La Niña conditions. There is very little overlap in the two distributions. These distributions illustrate the importance of El Niños to water supplies in New Mexico.

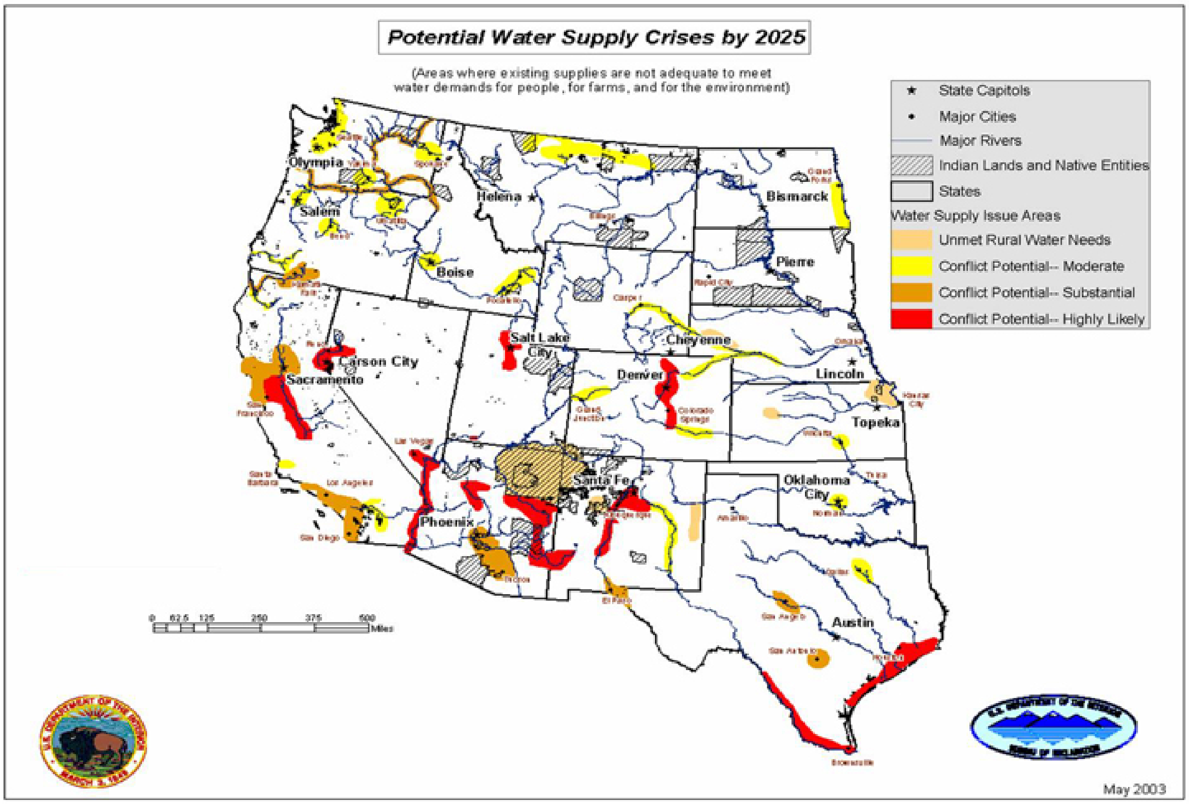

Figure 3.6. Interior Department analysis of regions in the West where water supply conflicts are likely occur by 2025 based on a combination of technical and other factors, including population trends and potential endangered species' needs for water. The red zones are where the conflicts are most likely to happen. See DOI Water 2025 Status Report (2005) for details. Note: There is an underlying assumption of a statistically stationary climate.

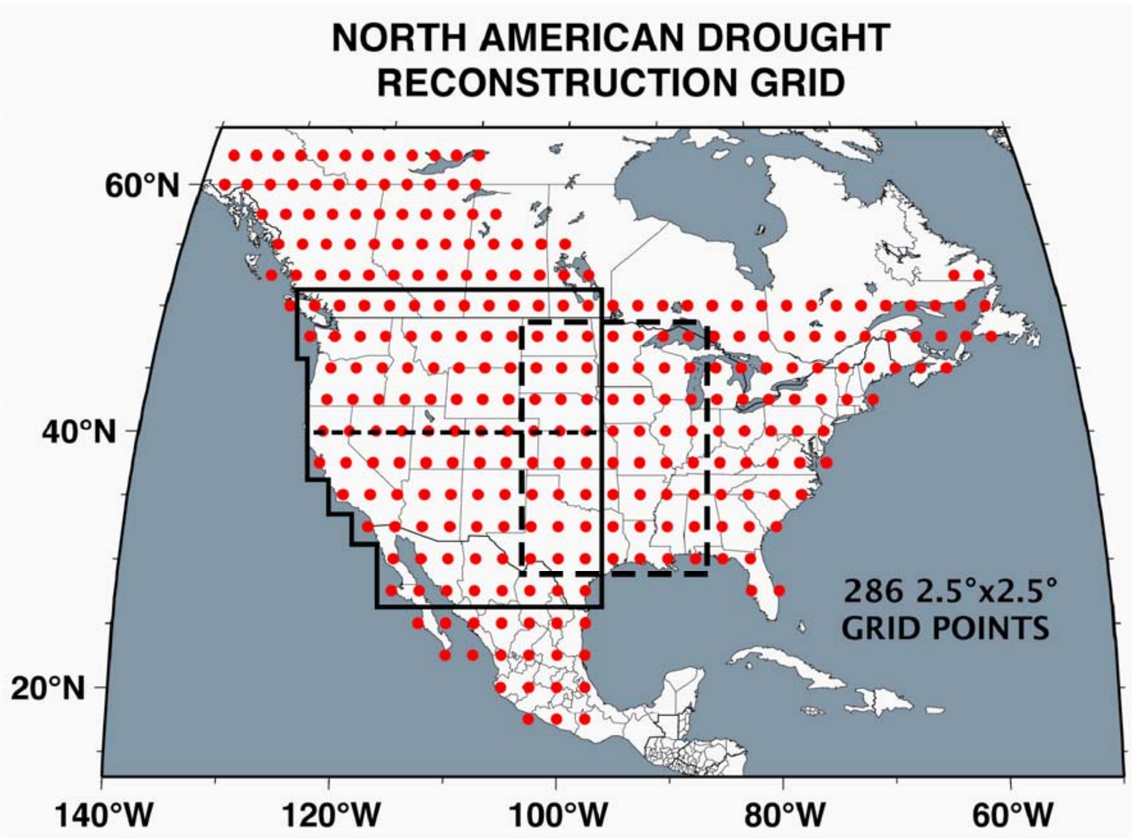

Figure 3.7. Map showing the distribution of 286 grid points of drought reconstructed for much of North America from long-term tree-ring records. The large, irregular polygon over the West is the area analyzed by Cook et al. (2004) in their study of long-term aridity changes. The dashed line at 40°N. 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.

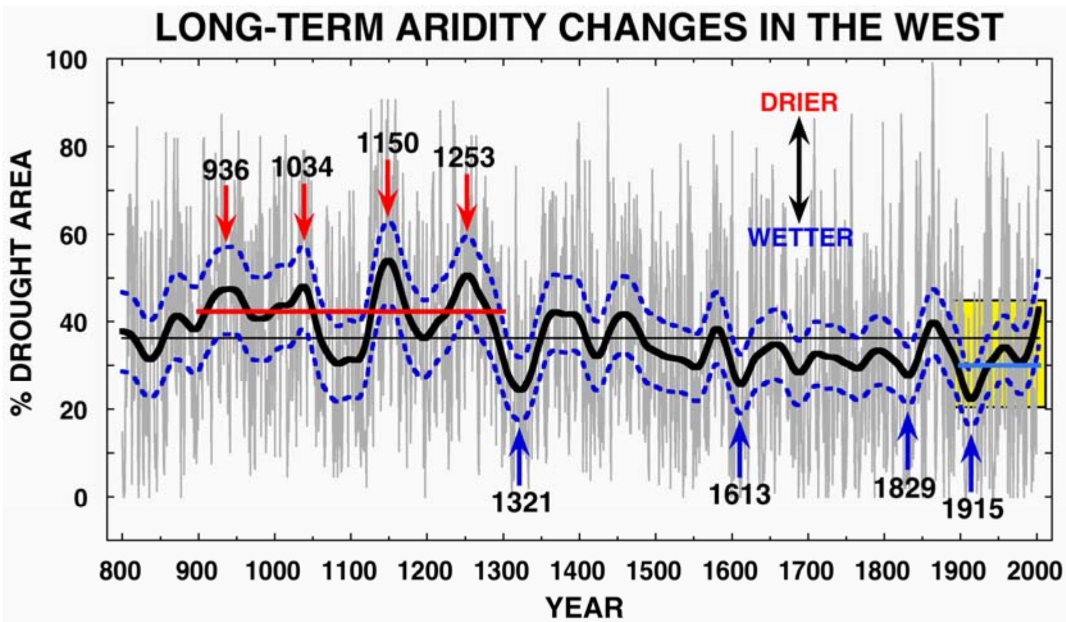

Figure 3.8. Percent area affected by drought (Palmer Drought Severity Index (PDSI) <-1) in the area defined as the West in Fig. 3.7 (redrawn from Cook et al., 2004). Annual data are in gray and a 60-year low-pass filtered version is indicated by the thick smooth curve. Dashed blue lines are 2-tailed 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.

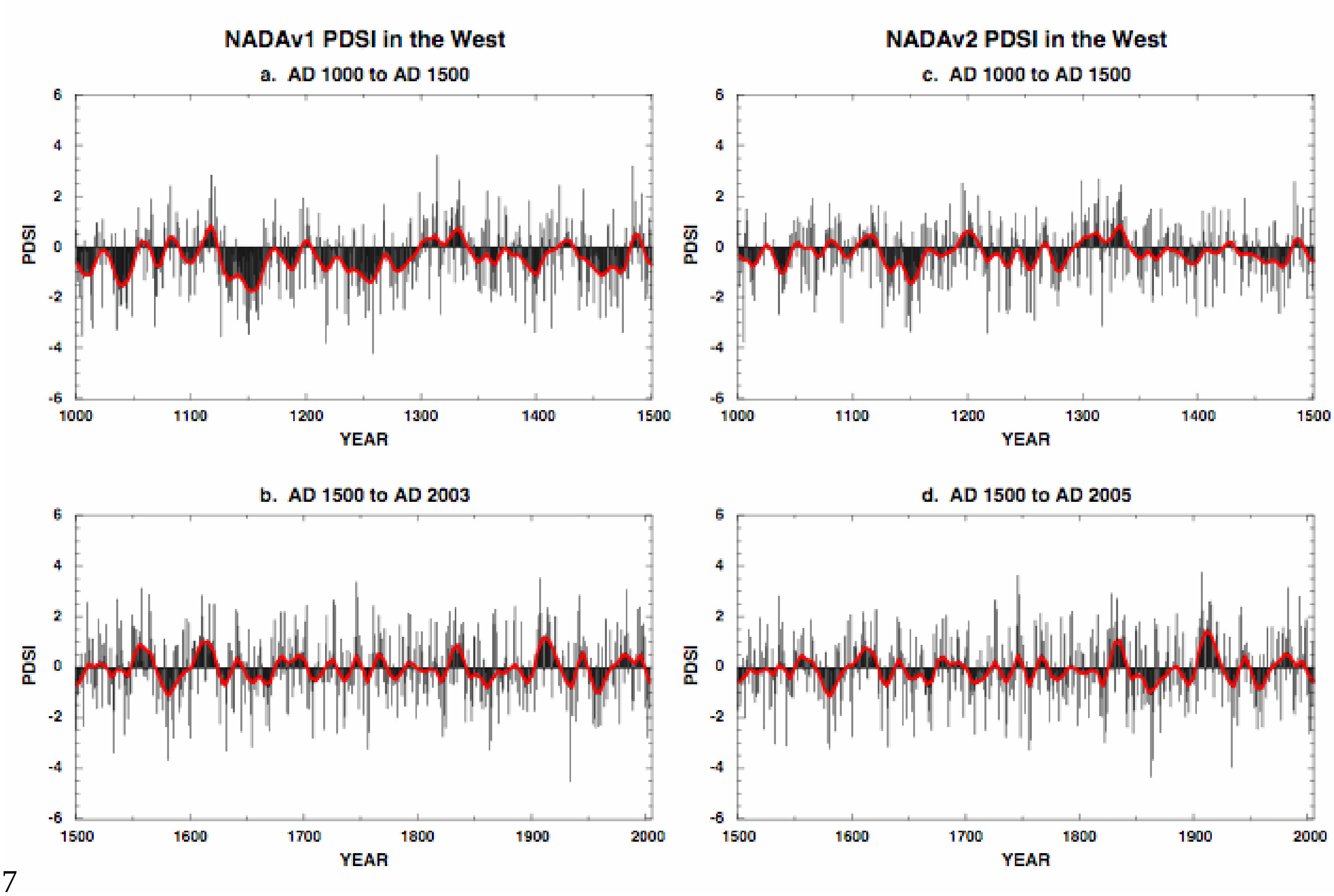

Figure 3.9. A comparison of average reconstructed Palmer Drought Severity Index (PDSI) for the West based on version 1 of the North American Drought Atlas (NADAv1) used by Cook et al. (2004) and a greatly improved version 2 (NADAv2) that has just been completed. Prior to A.D. 1300, the two series differ somewhat in the level of drought, with NADAv2 showing less “droughty” conditions. The reason for this is explained in the text.

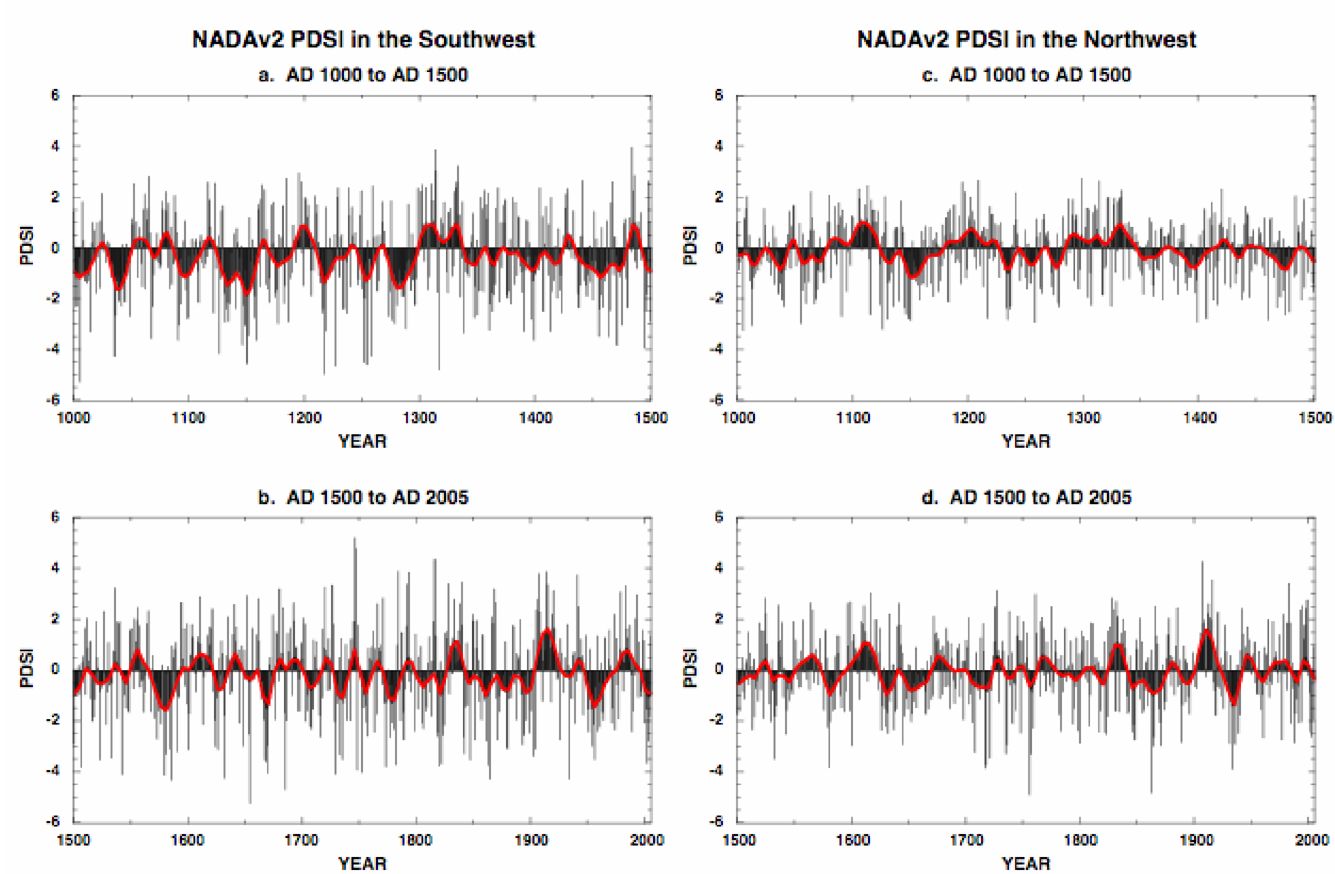

Figure 3.10. Average reconstructed Palmer Drought Severity Index (PDSI) for the West based on NADAv2 and now split into Southwest and Northwest regions (see Fig. 3.7). The difference in aridity between NADAv1 and NADAv2 prior to A.D. 1300 is due to the fact that the latter provides more sharply defined regional expressions of PDSI variability in Medieval times, with the increase in aridity reported by Cook et al. (2004) being primarily located in the Southwest.

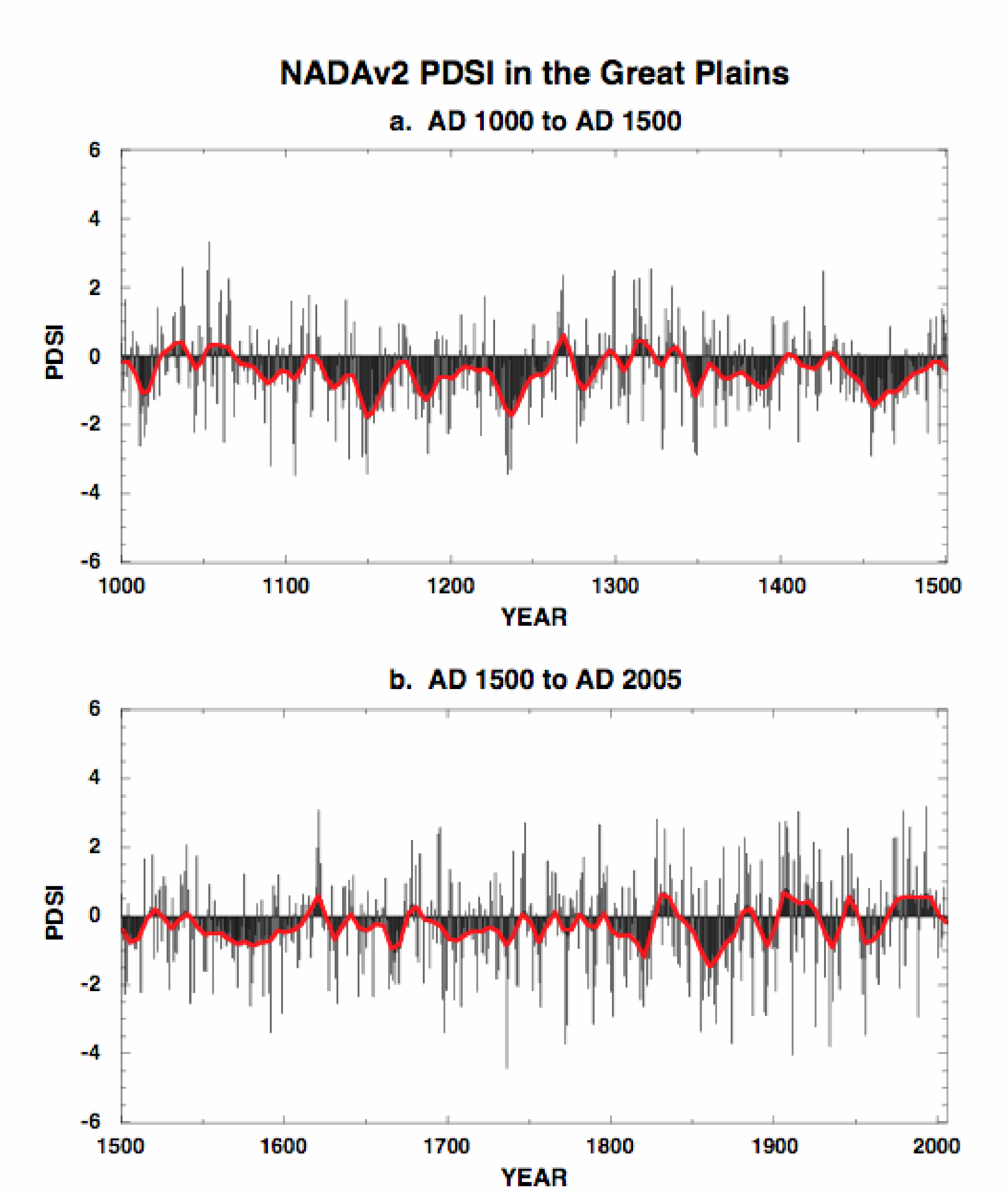

Figure 3.11. Average reconstructed Palmer Drought Severity Index (PDSI) for the Great Plains and Mississippi River valley (see Fig. 3.7). Drought in this region, which includes the "breadbasket" of America, is remarkably more common and persistent prior to A.D. 1500. A return to those conditions would be disastrous for agriculture and food supplies.

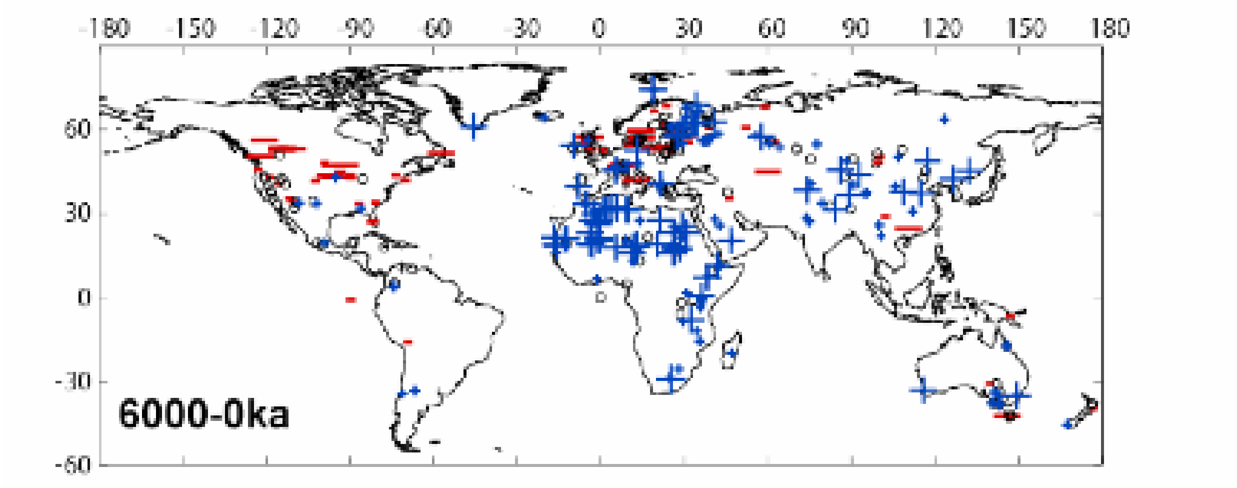

Figure 3.12. Global lake status at 6 ka (6,000 years ago) showing the large region that extends from Africa across Asia where lake levels were higher than those of the present day related to the expansion of the African-Asian monsoon. Note also the occurrence of much drier than present conditions over North America. (The most recent version of the Global Lake Status Database (GLSDB) is available on the PMIP 2 website http://pmip2.lsce.ipsl.fr/share/synth/glsdb/lakes.png.

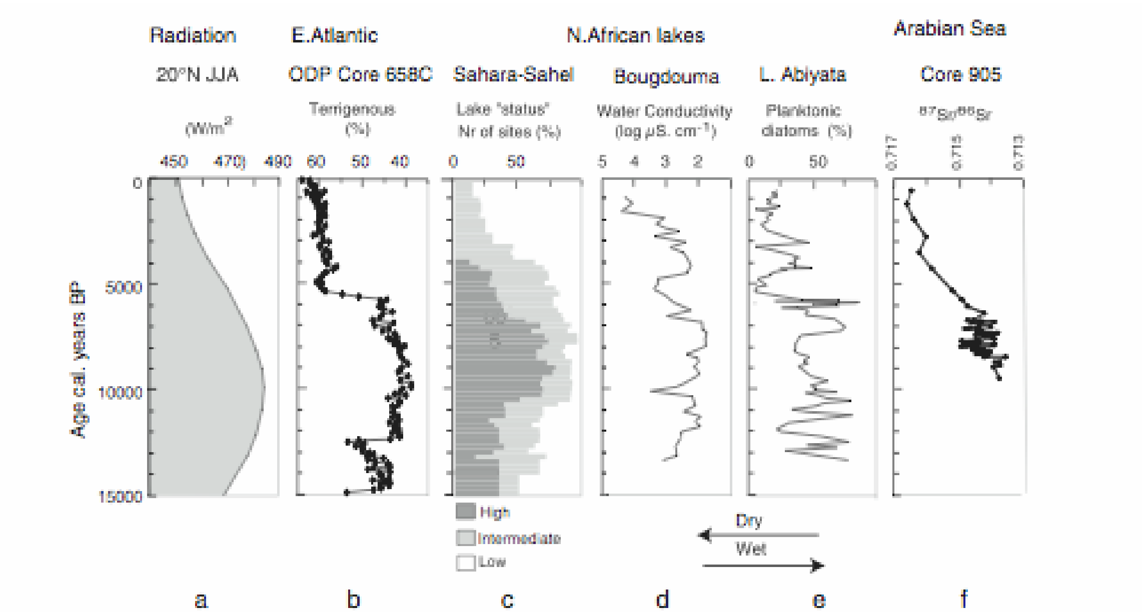

Figure 3.13. African Humid Period records. Reproduced or copied from Liu et al. (2007).

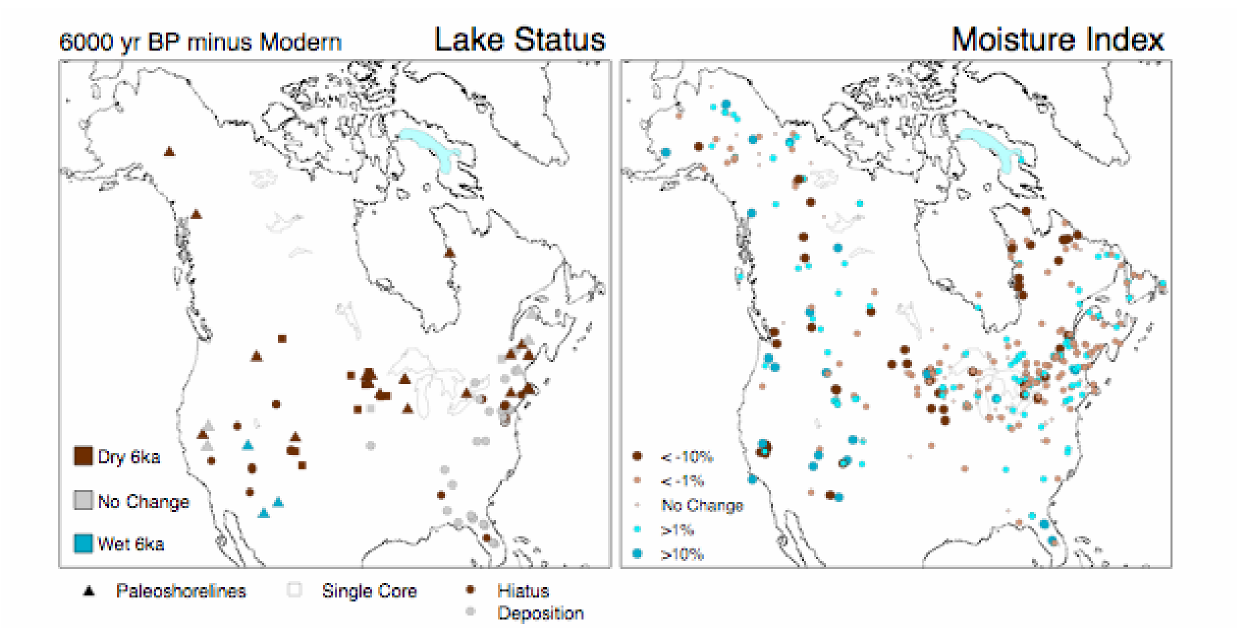

Figure 3.14. North American lake status (left) and moisture-index (AE/PE – actual evapotranspiration divided by potential evapotranspiration) anomalies (right) for 6ka. Lake (level) status can be inferred from a variety of sedimentological and limnological indicators (triangles and squares), and from the absence of deposition (hiatuses, circles) (Shuman and Finney, 2006). The inferred moisture-index values are based on modern analogue technique applied to a network of fossil-pollen data (Williams and Shuman, in prep.) Figure adapted from Shuman et al. (in review, *Quaternary Science Reviews*).

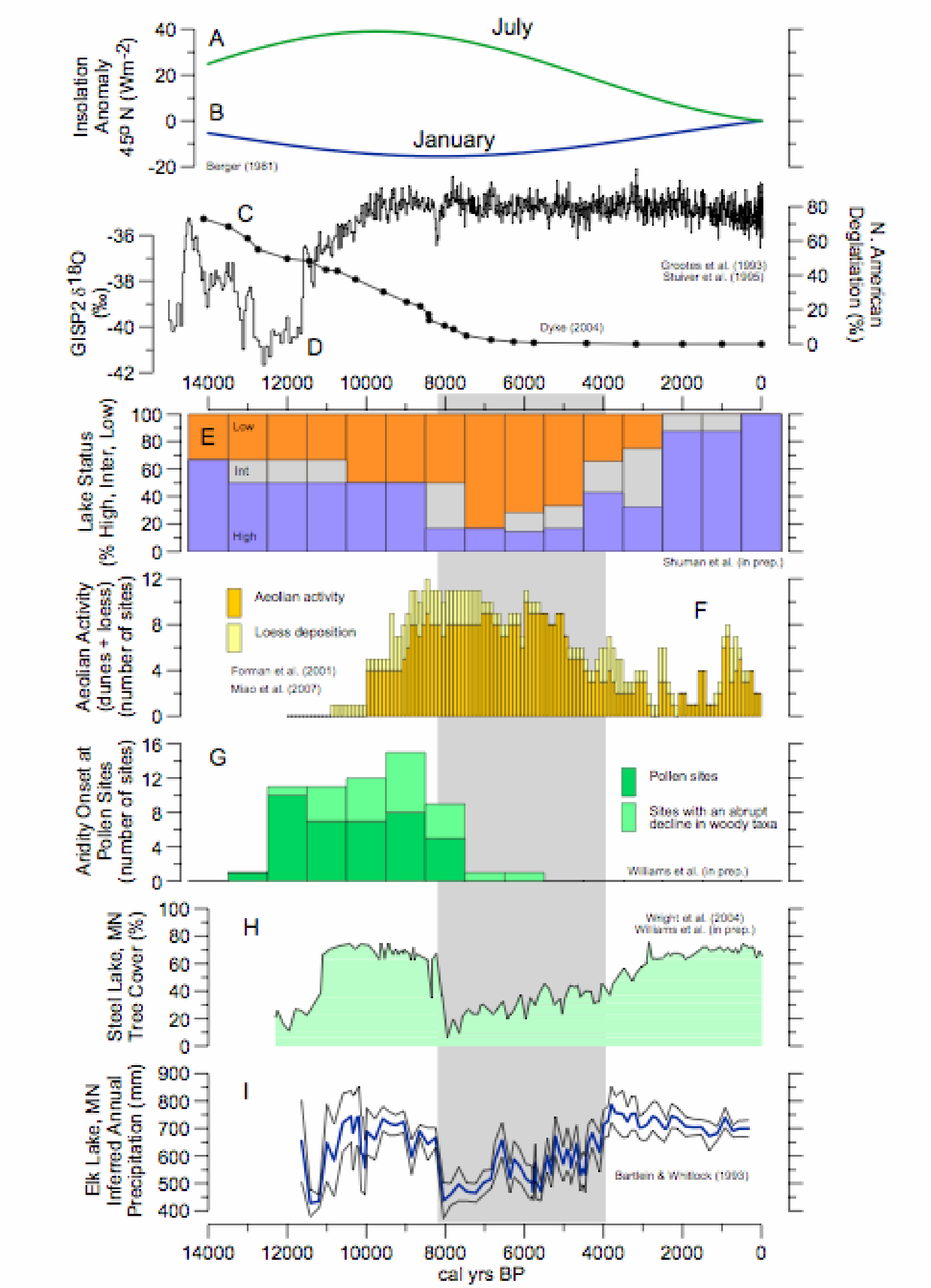

Figure 3.15. Time series of large-scale climate controls (A-D) and paleoenvironmental indicators of North American mid-continental aridity (E-I). A, B, July and January insolation anomalies (differences relative to present, Wm^{-2}) (Berger, 1978). C, right-hand scale: Deglaciation of North America, 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, $^{0}/_{00}$ (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 prep). 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 (number of sites, 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 (Williams et al., in prep.) H, Inferred tree-cover percentage at one of the sites (Steel L., MN) summarized in panel G (Williams et al., in prep.; based on pollen data from Wright et al., 2004). I: Inferred annual precipitation values (mm) 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}. The gray shading indicates the interval of maximum aridity.

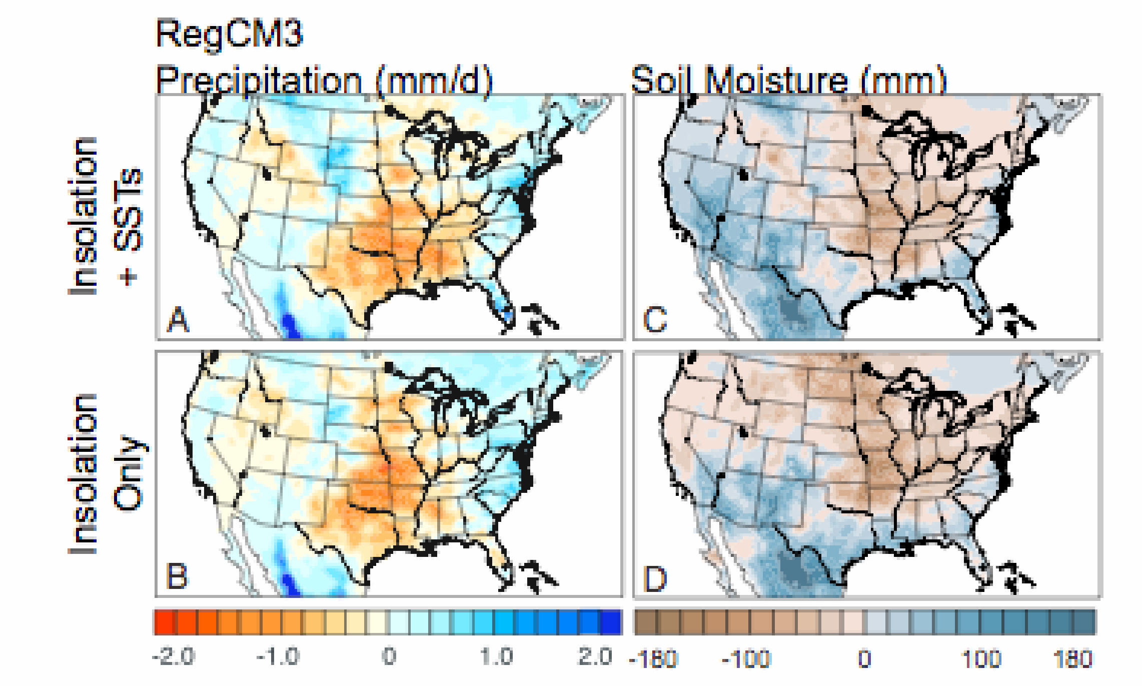

Figure 3.16. Regional climate model (RegCM3) simulations of precipitation rate (A, B) and soil moisture (C, D) for 6,000 years before present (6 ka) (Diffenbaugh et al., 2006, land grid points only). RegCM is run using lateral boundary conditions supplied by CAM3, the atmospheric component of CCSM3. 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.

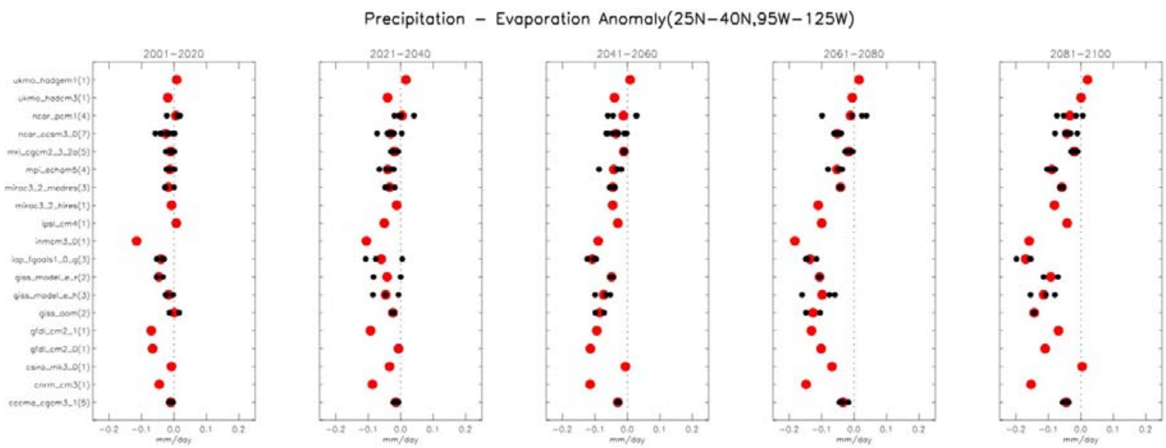

Figure 3.17. The change in annual mean precipitation minus evapotranspiration (P-E) over the American Southwest (125°W. – 95°W., 25°N. – 40°N., land areas only) for 19 models relative to model climatologies for 1950-2000. Results are averaged over t20-year segments of the current century. The number of ensemble members for the projections are 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.

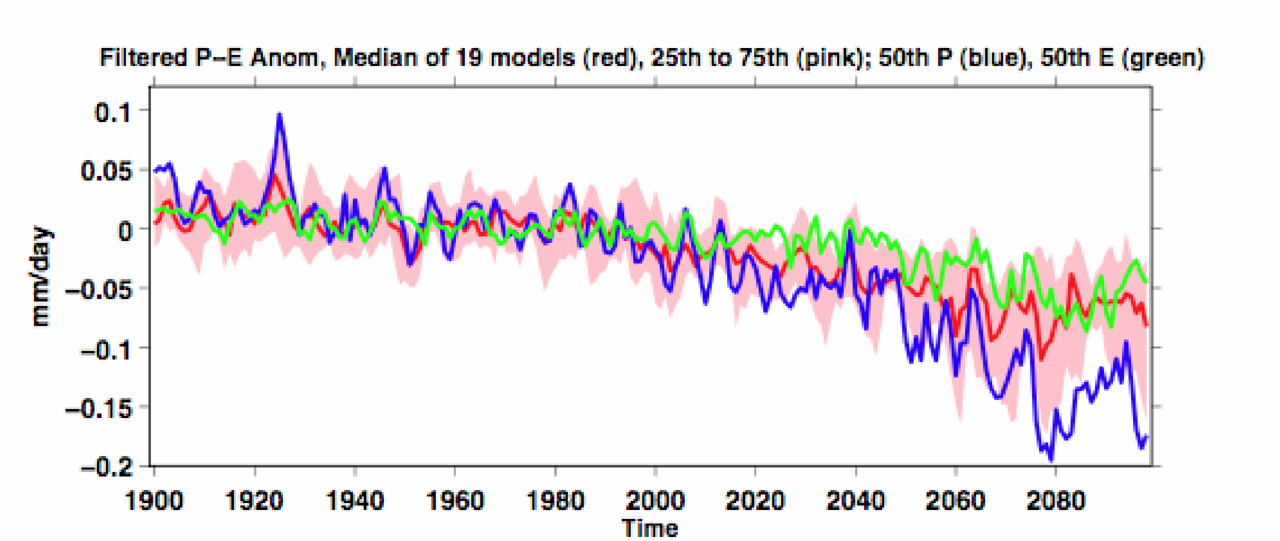

Figure 3.18. Modeled changes in annual mean precipitation minus evaporation (P-E) over southwestern North America (125W-95W, 25N-40N, land areas only) averaged over ensemble members for 19 models participating in IPCC AR4. The historical period used known and estimated climate forcings and the projections 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 six-year low-pass filtered to emphasize low frequency variations. Units are mm/day. The model ensemble mean P-E in this region is around 0.3 mm/day. From Seager et al. 2007d.